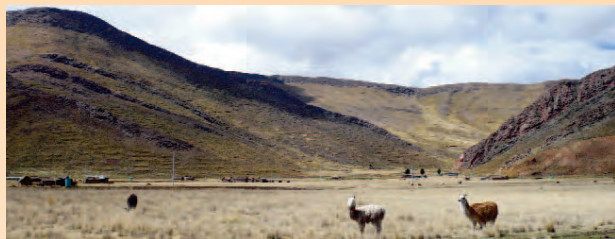


ORDOVICIAN OF THE WORLD



Editors: Juan Carlos Gutiérrez-Marco
Isabel Rábano
Diego García-Bellido



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ORDOVICIAN OF THE WORLD

Edited by

Juan Carlos Gutiérrez-Marco, Isabel Rábano and Diego García-Bellido

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Upper left: outcrops of the Late Ordovician glaciomarine Melaz Shuqran Fm, overlying Cambrian sandstones (Tihemboka Arch, Sahara desert, SW Libya).

Upper right: giant traces (> 11 m long) of marine worms in Early Ordovician quartzites from the Cabañeros National Park (central Spain), which serve as logo for the symposium.

Middle left: outcrops of the Late Ordovician Calapuja Fm (foreground mountains) in the Peruvian Altiplano, more than 4,500 m high.

Middle right: Global Stratotype Section at Point for the base of the Middle Ordovician series and of Dapingian stage, Huanghuachang section, Hubei province (South China).

Lower left: Early Ordovician shales (San José Formation) at the Inambari river, Amazonian basin (Eastern Peru).

Lower middle: A view of the Mount Everest (Tibet), whose summit (8,848 m) is formed by the Early-Middle Ordovician limestones of the Qomolangma Fm.

Lower right: Middle Ordovician dolomitic marls and mudstones of the Middle Guragir Fm at the key Kulyumbe river section (north-western part of the Siberian Platform, Russia).

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C/ Ríos Rosas, 23. 28003 Madrid

Tel.: +34 91 349 5700, Fax: +34 91 442 6216

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This book is dedicated to our mentors **Wolfgang Hammann** (Germany, 1942-2002) and **Michel Robardet** (France, 1939), who dedicated an important part of their lives to the Geology and Paleontology of the Ordovician of Spain

Both bestowed upon us their passion for the rocks and fossils of this period, and showed us how to study them with a modern vision and an open mind

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PREFACE

Among all the geological periods of the Earth's history, the Ordovician displays some of the most striking peculiarities, starting with an almost unique paleogeography, warm climates, high sea levels, the largest tropical shelf area of the Phanerozoic, kilometer-sized asteroid impacts, one of the two most significant biodiversification events on the planet, and the first of the "Big Five" mass extinctions, this one linked to a dramatic sea-level fall caused by the end-Ordovician glaciation.

The Iberian Peninsula comprises the most extensive outcrops of Ordovician rocks in Europe. They are mainly situated within the Iberian Massif and in its eastern extension in the Iberian Cordillera, as part of the Variscan Belt, and also in the Palaeozoic massifs of the Catalanian Coastal Ranges, the Pyrenees and the Betic Cordilleras, which have later been involved in the Alpine tectonic evolution. The celebration of an Ordovician meeting in Spain brings the opportunity to experience first hand the particular rocks and fossils representative of a special high-paleolatitudinal domain related to the southern polar margin of Gondwana, mainly represented by siliciclastic facies and with an interesting tectono-magmatic activity mostly linked with the opening of the Rheic ocean.

The present book, *Ordovician of the World*, is the proceedings volume for the 11th Symposium on the Ordovician System, sponsored by the Subcommittee on Ordovician Stratigraphy of the International Union of Geological Sciences. It contains 100 contributions, most of which in the form of short papers, which were delivered as oral presentations or posters in the symposium program. This volume represents a wealth of cutting-edge research on Ordovician rocks from around the world, and accommodate contributions from 228 authors and coauthors from 23 countries of four continents.

The book follows the trend of previous volumes devoted exclusively to the Ordovician. Most of them came after symposia arranged by the Ordovician Subcommittee, such as *The Ordovician System* (Birmingham, 1976), *Aspects of the Ordovician System* (Oslo, 1984), *Advances in Ordovician Geology* (St. John's, Newfoundland, 1988), *Global Perspectives on Ordovician Geology* (Sydney, 1992), *Ordovician Odyssey* (Las Vegas, 1995), *Quo vadis Ordovician?* (Prague, 1999) and *Ordovician of the Andes* (San Juan, Argentina, 2003). Other recent books like *The Great Ordovician Biodiversification Event* (Columbia University Press, 2004) and *The Ordovician Earth System* (The Geological Society of America, 2010) bear witness to a renewed interest for the Ordovician geology.

Starting from the 7th ISOS in Nevada, the proceedings volumes for the last five Ordovician symposia were distributed at the time of their respective meetings, and this book is not an exception. But *Ordovician of the World* could not have been ready for the Spanish symposium without the combined efforts of the authors of these high-quality works, the referees of the papers, and of the three editors that are research scientists at the Spanish Geological Survey (IGME, a veteran institution founded back in 1849) and from the Spanish Research Council (CSIC). Acknowledgement is also due to all institutions and private sponsors of the meeting, especially to the Spanish Ministry of Science and Innovation and to the members of the Portuguese Geological Survey, and the Spanish and Portuguese universities that made possible its organization in due time.

Rosa de Vidania
Director
Spanish Geological Survey (IGME)

CONTENTS

11th International Symposium on the Ordovician System.....	v
Preface	vii
 PRESIDENTIAL ADDRESS	
A SIXTH DECADE OF THE ORDOVICIAN PERIOD: STATUS OF THE RESEARCH INFRASTRUCTURE OF A GEOLOGICAL SYSTEM	3
D. A.T. Harper	
 KEYNOTE LECTURES	
THE LATE ORDOVICIAN GLACIAL RECORD: STATE OF THE ART.....	13
J.-F. Ghienne	
NEW INSIGHTS FROM EXCEPTIONALLY PRESERVED ORDOVICIAN BIOTAS FROM MOROCCO	21
P. Van Roy	
 PAPERS AND ABSTRACTS	
WHOLE-ROCK AND ISOTOPE GEOCHEMISTRY OF ORDOVICIAN TO SILURIAN UNITS OF THE CUYANIA TERRANE, ARGENTINA: INSIGHTS FOR THE EVOLUTION OF SW GONDWANA MARGIN.....	29
P. Abre, C. Cingolani, B. Cairncross and F. Chemale Jr.	
OCEAN CURRENTS AND STRIKE-SLIP DISPLACEMENTS IN WESTERN GONDWANA: THE CUYANIA HYPOTHESIS IN CAMBRIAN-ORDOVICIAN TIMES.....	35
F.G. Aceñolaza	
A NEW TRILOBITE BIOSTRATIGRAPHY FOR THE LOWER ORDOVICIAN OF WESTERN LAURENTIA AND PROSPECTS FOR INTERNATIONAL CORRELATION USING PELAGIC TRILOBITES	41
J.M. Adrain	
A PERI-GONDWANAN ARC ACTIVE IN CAMBRIAN-ORDOVICIAN TIMES: THE EVIDENCE OF THE UPPERMOST TERRANE OF NW IBERIA.....	43
R. Arenas, J. Abati, S. Sánchez Martínez, P. Andonaegui, J.M. Fuenlabrada, J. Fernández-Suárez and P. González Cuadra	
GEOCHEMICAL FEATURES OF THE ORDOVICIAN SUCCESSION IN THE CENTRAL IBERIAN ZONE (SPAIN).....	49
P. Barba, J.M. Ugidos, E. González-Clavijo and M.I. Valladares	
FAUNAL SHIFTS AND CLIMATIC CHANGES IN THE UPPER ORDOVICIAN OF SOUTH AMERICA (W GONDWANA).....	55
J.L. Benedetto, T.M. Sánchez, M.G. Carrera, K. Halpern and V. Bertero	

A SUMMARY OF THE ORDOVICIAN OF THE OSLO REGION, NORWAY – FUTURE CHALLENGES	61
D.L. Bruton	
PRELIMINARY REPORT ON <i>ARTHROHACHIS</i> HAWLE AND CORDA, 1847 (AGNOSTIDA) IN THE PRAGUE BASIN (BARRANDIAN AREA, CZECH REPUBLIC)	65
P. Budil, O. Fatka, P. Kolář and M. David	
GRAPTOLITE ZONATION FOR THE LOWER AND MIDDLE ORDOVICIAN OF THE GORNY ALTAI (SW SIBERIA, RUSSIA)	69
E.V. Bukolova	
ORDOVICIAN MAGMATISM IN THE EXTERNAL FRENCH ALPS: WITNESS OF A PERI-GONDWANAN ACTIVE CONTINENTAL MARGIN	75
F. Bussy, V. Péronnet, A. Ulianov, J.L. Epard and J. von Raumer	
REWORKED CONODONTS IN THE UPPER ORDOVICIAN SANTA GERTRUDIS FORMATION (SALTA, ARGENTINA)	83
J. Carlorosi, S. Heredia, G.N. Sarmiento and M.C. Moya	
A POST-GLACIAL BRYOZOAN FAUNA FROM THE UPPER ORDOVICIAN (HIRNANTIAN) OF THE ARGENTINE PRECORDILLERA	89
M. Carrera and K. Halpern	
ORDOVICIAN MAGMATISM IN NE IBERIA	95
J.M. Casas, P. Castiñeiras, M. Navidad, M. Liesa, J.F. Martínez, J. Carreras, J. Reche, A. Iriondo, J. Aleinikoff, J. Cirés and C. Dietsch	
CARBON ISOTOPE DEVELOPMENT IN THE ORDOVICIAN OF THE YANGTZE GORGES REGION (SOUTH CHINA) AND ITS IMPLICATION FOR STRATIGRAPHIC CORRELATION AND PALEOENVIRONMENTAL CHANGE	101
J. Cheng, Y.D. Zhang, A. Munneke and C. Zhou	
THE HIRNANTIAN-EARLY LLANDOVERY TRANSITION SEQUENCE IN THE PARANÁ BASIN, EASTERN PARAGUAY	103
C.A. Cingolani, N.J. Uriz, M.B. Alfaro, F. Tortello, A.R. Bidone and J.C. Galeano Inchausti	
DISTAL EFFECTS OF GLACIALLY-FORCED LATE ORDOVICIAN MASS EXTINCTIONS ON THE TROPICAL CARBONATE PLATFORM OF LAURENTIA: STROMATOPOROID LOSSES AND RECOVERY AT A TIME OF STRESS, ANTICOSTI ISLAND, EASTERN CANADA	109
P. Copper, H. Nestor and C. Stock	
LATE ORDOVICIAN GLACIAL DEPOSITS IN VALONGO ANTICLINE (NORTHERN PORTUGAL): A REVISION OF THE SOBRIDO FORMATION AND A CONTRIBUTION TO THE KNOWLEDGE OF ICE-MARGINAL LOCATIONS	113
H. Couto and A. Lourenço	
ABNORMAL ACRITARCHS IN THE RUN-UP OF EARLY PALAEOZOIC $\delta^{13}\text{C}$ ISOTOPE EXCURSIONS: INDICATION OF ENVIRONMENTAL POLLUTION, GLACIATION, OR MARINE ANOXIA?	119
A. Delabroye, A. Munneke, T. Servais, T. Vandenbroucke and M. Vecoli	

GEOCHEMISTRY OF LOWER PALAEOZOIC SHALES. A CASE STUDY IN A SECTOR OF THE IBERIAN VARISCIDES	121
I. Dias da Silva, E. González-Clavijo, P. Barba, M.I. Valladares and J.M. Ugidos	
EARLY SILURIAN VS. LATE ORDOVICIAN GLACIATION IN SOUTH AMERICA	127
E. Díaz-Martínez, M. Vavrdová, P.E. Isaacson and C.Y. Grah	
K-BENTONITES IN THE UPPER ORDOVICIAN OF THE SIBERIAN PLATFORM	135
A.V. Dronov, W.D. Huff, A.V. Kanygin and T.V. Gonta	
ORDOVICIAN OF BALTOSCANDIA: FACIES, SEQUENCES AND SEA-LEVEL CHANGES	143
A.V. Dronov, L. Ainsaar, D. Kaljo, T. Meidla, T. Saadre and R. Einasto	
POSSIBLE REMAINS OF THE DIGESTIVE SYSTEM IN ORDOVICIAN TRILOBITES OF THE PRAGUE BASIN (BARRANDIAN AREA, CZECH REPUBLIC)	151
O. Fatka, P. Budil and Š. Rak	
LATE ORDOVICIAN-EARLY SILURIAN SELECTIVE EXTINCTION PATTERNS IN LAURENTIA AND THEIR RELATIONSHIP TO CLIMATE CHANGE	155
S. Finnegan, S. Peters and W.W. Fischer	
GSSP BOUNDARY INTERVALS ARE CRITICAL FOR CHARACTERIZATION AND CORRELATION OF CHRONOHORIZONS THAT DEFINE GLOBAL STAGES, SERIES, AND SYSTEMS	161
S.C. Finney	
THE LATE TREMADOCIAN–EARLY ARENIG 2ND ORDER SEQUENCE OF THE CADENAS IBÉRICAS (NE SPAIN) AND ITS COMPARISON WITH BALTICA	163
J.A. Gámez Vintaned and U. Schmitz	
STRATIGRAPHIC EVIDENCE FOR THE HIRNANTIAN (LATEST ORDOVICIAN) GLACIATION IN THE ZAGROS MOUNTAINS, IRAN	169
M. Ghavidel-syooki, J.J. Álvaro, L. Popov, M. Ghobadi Pour, M.H. Ehsani and A. Suyarkova	
NEW DATA ON THE LATE ORDOVICIAN TRILOBITE FAUNAS OF KAZAKHSTAN: IMPLICATIONS FOR BIOGEOGRAPHY OF TROPICAL PERI-GONDWANA	171
M. Ghobadi Pour, L.E. Popov, L. McCobb and I.G. Percival	
A CONOP9 COMPOSITE-TAXON RANGE-CHART FOR ORDOVICIAN CONODONTS FROM BALTOSCANDIA: A FRAMEWORK FOR BIODIVERSITY ANALYSES	179
D. Goldman, S.M. Bergström, H.D. Sheets and C. Pantle	
BIOSTRATIGRAPHY OF THE GENUS <i>CALIX</i> (ECHINODERMATA, DIPLOPORITA) IN THE MIDDLE ORDOVICIAN OF THE SOUTHERN CENTRAL IBERIAN ZONE (SPAIN)	189
J.C. Gutiérrez-Marco and J. Colmenar	
A PRELIMINARY STUDY OF SOME SANDBIAN (UPPER ORDOVICIAN) GRAPTOLITES FROM VENEZUELA	199
J.C. Gutiérrez-Marco, D. Goldman, J. Reyes-Abril and J. Gómez	

ORDOVICIAN BRACHIOPOD DIVERSITY REVISITED: PATTERNS AND TRENDS IN THE OSLO REGION	207
J.W. Hansen, D.A.T. Harper and A.T. Nielsen	
ORDOVICIAN ON THE ROOF OF THE WORLD: MACRO- AND MICROFAUNAS FROM TROPICAL CARBONATES IN TIBET	215
D. A.T. Harper, R. Zhan, L. Stemmerik, J. Liu, S.K. Donovan and S. Stouge	
STRATIGRAPHIC RELATIONS OF QUARTZ ARENITES AND K-BENTONITES IN THE ORDOVICIAN BLOUNT MOLASSE, ALABAMA TO VIRGINIA, SOUTHERN APPALACHIANS, USA	221
J.T. Haynes and K.E. Goggin	
MAJOR ORDOVICIAN TEPHRAS GENERATED BY CALDERA-FORMING EXPLOSIVE VOLCANISM ON CONTINENTAL CRUST: EVIDENCE FROM BIOTITE COMPOSITIONS	229
J.T. Haynes, W.D. Huff and W.G. Melson	
MIDDLE DARRIWILIAN CONODONT BIOSTRATIGRAPHY IN THE ARGENTINE PRECORDILLERA	237
S. Heredia and A. Mestre	
CONVENTIONAL AND CONOP9 APPROACHES TO BIODIVERSITY OF BALTIC ORDOVICIAN CHITINOZOANS	243
O. Hints, J. Nölvak, L. Paluveer and M. Tammekänd	
ORDOVICIAN ROCKS IN JAPAN	251
Y. Isozaki	
DARRIWILIAN BIOSTRATIGRAPHY AND PALAEOECOLOGY DURING THE GREAT ORDOVICIAN BIODIVERSIFICATION EVENT – A NORTHERN GONDWANAN PERSPECTIVE	253
K.G. Jakobsen, D.A.T. Harper, A.T. Nielsen and G.A. Brock	
THE UPPER KATIAN (UPPER ORDOVICIAN) BRYOZOANS FROM THE IBERIAN CHAINS (NE SPAIN): A REVIEW	259
A. Jiménez-Sánchez	
CARBON ISOTOPE TREND IN THE MIRNY CREEK AREA, NE RUSSIA, ITS SPECIFIC FEATURES AND POSSIBLE IMPLICATIONS OF THE UPPERMOST ORDOVICIAN STRATIGRAPHY	267
D. Kaljo and T. Martma	
FOSSIL ASSEMBLAGES REFLECTING PROCESSES OF THE EARLY DEVELOPMENT OF THE PRAGUE BASIN (BOHEMIAN MASSIF, CZECH REPUBLIC)	275
P. Kraft, T. Hroch and M. Rajchl	
LATE KATIAN STRATIGRAPHY IN THE PRAGUE BASIN (CZECH REPUBLIC)	277
P. Kraft, J. Bartošová, T. Hroch, L. Koptíková and J. Frýda	
A GIANT <i>RUSOPHYCUS</i> FROM THE MIDDLE ORDOVICIAN OF SIBERIA	279
V.B. Kushlina and A.V. Dronov	
ON THE KATIAN/HIRNANTIAN BOUNDARY. APPLICATION ON THE NORTH-AFRICAN BORDER OF GONDWANA	287
Ph. Legrand	

CONODONT BIOSTRATIGRAPHY FROM SHALLOW WATER UPPER ORDOVICIAN PLATFORM ROCKS IN THE SUBSURFACE OF SOUTH TEXAS	295
S.A. Leslie, J.E. Barrick, J. Mosley and S.M. Bergström	
CONODONT BIOSTRATIGRAPHY AND STABLE ISOTOPE STRATIGRAPHY ACROSS THE ORDOVICIAN KNOX/BEEKMANTOWN UNCONFORMITY IN THE CENTRAL APPALACHIANS	301
S.A. Leslie, M.R. Saltzman, S.M. Bergström, J.E. Repetski, A. Howard and A.M. Seward	
DEMISE OF EARLY ORDOVICIAN OOLITES IN SOUTH CHINA: EVIDENCE FOR PALEOCEANOGRAPHIC CHANGES BEFORE THE GOBE	309
J. Liu, R. Zhan, X. Dai, H. Liao, Y. Ezaki and N. Adachi	
NEW INSIGHTS ON THE HIRNANTIAN PALYNOSTRATIGRAPHY OF THE RIO CEIRA SECTION, BUÇACO, PORTUGAL	313
G. Lopes, N. Vaz, A.J.D. Sequeira, J.M. Piçarra, P. Fernandes and Z. Pereira	
DARRIWILIAN (ORDOVICIAN) GRAPTOLITE FAUNAS AND PROVINCIALISM IN THE TØYEN SHALE OF THE KRAPPERUP DRILL CORE (SCANIA, SOUTHERN SWEDEN)	327
J. Maletz and P. Ahlberg	
GRAPTOLITE BIOSTRATIGRAPHY AND BIOGEOGRAPHY OF THE TABLE HEAD AND GOOSE TICKLE GROUPS (DARRIWILIAN, ORDOVICIAN) OF WESTERN NEWFOUNDLAND	333
J. Maletz and S. Egenhoff	
CORRELATION OF LOWER ORDOVICIAN (IBEXIAN) FAUNAS IN NORTH-EASTERN GREENLAND AND WESTERN NEWFOUNDLAND – NEW TRILOBITE AND LITHOSTRATIGRAPHIC DATA	339
L.M.E. McCobb, W.D. Boyce and I. Knight	
ICE IN THE SAHARA: THE UPPER ORDOVICIAN GLACIATION IN SW LIBYA – A SUBSURFACE PERSPECTIVE	347
N.D. McDougall and R. Gruenwald	
OSTRACODS IN BALTOSCANDIA THROUGH THE HIRNANTIAN CRISES	353
T. Meidla, L. Ainsaar and K. Truuver	
FAUNAL TURNOVER NEAR THE KATIAN/HIRNANTIAN BOUNDARY IN THE PRAGUE BASIN (CZECH REPUBLIC)	359
M. Mergl	
EARLY ORDOVICIAN ARTHROPOD TRACE FOSSILS IN THE PRAGUE BASIN (CZECH REPUBLIC)	367
M. Mergl	
NEW STABLE ISOTOPE DATA AND FOSSILS FROM THE HIRNANTIAN STAGE IN BOHEMIA AND SPAIN: IMPLICATIONS FOR CORRELATION AND PALEOCLIMATE	371
C.E. Mitchell, P. Štorch, C. Holmden, M.J. Melchin and J.C. Gutiérrez-Marco	
THE TREMADOCIAN DEPOSITS OF THE ARGENTINIAN EASTERN CORDILLERA: A SCANDINAVIAN SIGNAL IN THE CENTRAL ANDES	379
M.C. Moya and J.A. Monteros	

EARLY ORDOVICIAN MAGMATISM IN THE NORTHERN CENTRAL IBERIAN ZONE (IBERIAN MASSIF): NEW U-Pb (SHRIMP) AGES AND ISOTOPIC Sr-Nd DATA	391
M. Navidad and P. Castiñeiras	
A RE-CALIBRATED REVISED SEA-LEVEL CURVE FOR THE ORDOVICIAN OF BALTOSCANDIA	399
A.T. Nielsen	
NEW DATA ON UPPER ORDOVICIAN RADIOLARIANS FROM THE GORNY ALTAI (SW SIBERIA, RUSSIA)	403
O.T. Obut and A.M. Semenova	
DARRIWILIAN GRAPTOLITES FROM THE LINA RANGE, NORTHWESTERN PUNA OF JUJUY, ARGENTINA	409
G. Ortega, G.L. Albanesi and C.R. Monaldi	
PATTERNS OF ORIGINATION AND DISPERSAL OF MIDDLE TO LATE ORDOVICIAN BRACHIOPODS: EXAMPLES FROM SOUTH CHINA, EAST GONDWANA, AND KAZAKH TERRANES	413
I.G. Percival, L.E. Popov, R.B. Zhan and M. Ghobadi Pour	
RECENT DISCOVERIES AND A REVIEW OF THE ORDOVICIAN FAUNAS OF NEW ZEALAND	421
I.G. Percival, R.A. Cooper, Y.Y. Zhen, J.E. Simes and A.J. Wright	
ORDOVICIAN GRAPTOLITES AND ACROTARCHS FROM THE BARRANCOS REGION (OSSA-MORENA ZONE, SOUTH PORTUGAL)	429
J. Piçarra, Z. Pereira and J.C. Gutiérrez-Marco	
NEW INSIGHTS INTO THE STRATIGRAPHY AND STRUCTURE OF THE UPPER ORDOVICIAN ROCKS OF THE LA Cerdanya AREA (PYRENEES)	441
C. Puddu and J.M. Casas	
FINAL DESTINATION, FIRST DISCOVERED: THE TALE OF <i>OANDUPORELLA</i> HINTS, 1975	447
C.M.Ø. Rasmussen	
AN UNUSUAL MID-ORDOVICIAN ISLAND ENVIRONMENT ON THE WESTERN EDGE OF BALTICA: NEW PALAEOECOLOGICAL AND PALAEOBIOGEOGRAPHICAL DATA FROM HARDANGERVIDDA, SOUTHERN NORWAY	455
J.A. Rasmussen, A.T. Nielsen and D.A.T. Harper	
BIOSTRATIGRAPHY OF THE MIDDLE ORDOVICIAN BRACHIOPODS FROM CENTRAL SPAIN	463
J. Reyes-Abril, J.C. Gutiérrez-Marco and E. Villas	
STRATIGRAPHY AND STRUCTURE OF THE UPPERMOST PART OF THE LUARCA FORMATION IN ALTO BIERZO, LEÓN (ORDOVICIAN, NW SPAIN)	473
M.A. Rodríguez Sastre and L. González Menéndez	
ORDOVICIAN VS. "CAMBRIAN" ICHNOFOSSILS IN THE ARMORICAN QUARTZITE OF CENTRAL PORTUGAL	483
A.A. Sá, J.C. Gutiérrez-Marco, J.M. Piçarra, D.C. García-Bellido, N. Vaz and G.F. Aceñolaza	

ORDOVICIAN GEOSITES AS THE BASIS OF THE CREATION OF THE EUROPEAN AND GLOBAL AROUCA GEOPARK (PORTUGAL)	493
A.A. Sá, D. Rocha and A. Paz	
GRAPTOLOID EVOLUTIONARY RATES: SHARP CONTRAST BETWEEN ORDOVICIAN AND SILURIAN	499
P.M. Sadler and R.A. Cooper	
A BRIEF SUMMARY OF ORDOVICIAN CONODONT FAUNAS FROM THE IBERIAN PENINSULA	505
G.N. Sarmiento, J.C. Gutiérrez-Marco, R. Rodríguez-Cañero, A. Martín Algarra and P. Navas-Parejo	
THE LATE ORDOVICIAN GLACIAL EVENT IN THE CARNIC ALPS (AUSTRIA)	515
H.P. Schönlaub, A. Ferretti, L. Gaggero, E. Hammarlund, D.A.T. Harper, K. Histon, H. Prielwalder, C. Spötl and P. Štorch	
INTENSE VOLCANISM AND ORDOVICIAN ICEHOUSE CLIMATE	527
B.K. Sell	
NEW U-Pb ZIRCON DATA FOR THE GSSP FOR THE BASE OF THE KATIAN IN ATOKA, OKLAHOMA, USA AND THE DARRIWILIAN IN NEWFOUNDLAND, CANADA	537
B.K. Sell, S.A. Leslie and J. Maletz	
ORDOVICIAN REGIONAL CHRONOSTRATIGRAPHIC SCHEME OF THE GORNY ALTAI	547
N.V. Sennikov, O.T. Obut and E.V. Bukolova	
TRACES OF THE GLOBAL AND REGIONAL SEDIMENTARY EVENTS IN EARLY ORDOVICIAN SECTIONS OF THE GORNY ALTAI (SIBERIA)	553
N.V. Sennikov, O.T. Obut, E.V. Bukolova and T.Yu. Tolmacheva	
CONODONT BIODIVERSITY DYNAMICS FROM THE ORDOVICIAN OF BALTOSCANDIA	559
H.D. Sheets, D. Goldman, S.M. Bergström and C. Pantle	
THE DISTRIBUTION OF GONDWANA-DERIVED TERRANES IN THE EARLY PALEOZOIC	567
G.M. Stampfli, J. von Raumer and C. Wilhem	
MIDDLE ORDOVICIAN BIVALVES FROM BOHEMIA, SPAIN AND FRANCE	575
M. Steinová	
MIDDLE ORDOVICIAN (DARRIWILIAN) GLOBAL CONODONT ZONATION BASED ON THE DAWANGOU AND SAERGAN FORMATIONS OF THE WESTERN TARIM REGION, XINJIANG PROVINCE, CHINA	581
S. Stouge, P. Du and Z. Zhao	
THE BASE OF THE ORDOVICIAN SYSTEM – A HORIZON IN LIMBO	587
F. Terfelt, G. Bagnoli and S. Stouge	
THE LOWER TO MIDDLE ORDOVICIAN CONODONT BIOSTRATIGRAPHY OF NORTHERN TIAN SHAN (WESTERN PART OF THE KIRGYZ RANGE), KYRGYZSTAN	589
T.Yu. Tolmacheva, K.E. Degtyarev, L.E. Popov, A.V. Ryazantsev, A.B. Kotov and P.A. Aleksandrov	

COMPARATIVE ANALYSIS OF THE EARLY ORDOVICIAN BALTOGRAPTID SPECIES OF NORTHWESTERN ARGENTINA, BALTOSCANDIA AND SOUTH CHINA	597
B. A. Toro, J. Maletz, Y.D. Zhang and J. Zhang	
THE AGE OF THE <i>P. LINEARIS</i> GRAPTOLITE BIOZONE: A PROGRESS REPORT ON A POTENTIAL SOLUTION	605
T.R.A. Vandenbroucke, A.T. Nielsen and J.K. Ingham	
POLAR FRONT SHIFT AND ATMOSPHERIC CO₂ DURING THE GLACIAL MAXIMUM OF THE EARLY PALEOZOIC ICEHOUSE	607
T.R.A. Vandenbroucke, H.A. Armstrong, M. Williams, F. Paris, J.A. Zalasiewicz, K. Sabbe, J. Nölvak, T.J. Challands, J. Verniers and T. Servais	
CHITINOZOANS OF RIBEIRA DA LAJE FORMATION, AMÊNDOA-MAÇÃO SYNCLINE (UPPER ORDOVICIAN, PORTUGAL)	609
N. Vaz, F. Paris and J.T. Oliveira	
ORDOVICIAN COSMIC SPHERULES FROM THE CORDILLERA ORIENTAL OF NW ARGENTINA: PRELIMINARY SEM AND EDX INVESTIGATION	611
G.G. Voldman, G.L. Albanesi, C.R. Barnes, G. Ortega and M.J. Genge	
BIODIVERSITY PATTERNS AND THEIR IMPLICATIONS OF EARLY-MIDDLE ORDOVICIAN MARINE MICROPHYTOPLANKTON IN SOUTH CHINA	617
K. Yan, J. Li, and T. Servais	
EARLY-MIDDLE ORDOVICIAN ACRITARCH ASSEMBLAGE FROM CHENGKOU, CHONGQING CITY, SOUTH CHINA	619
K. Yan, J. Li and T. Servais	
BIOSTRATIGRAPHY AND PALEOENVIRONMENTS OF THE SANTA ROSITA FORMATION (LATE FURONGIAN–TREMADOCIAN), CORDILLERA ORIENTAL OF JUJUY, ARGENTINA	625
F.J. Zeballo, G.L. Albanesi and G. Ortega	
ON THE MACROEVOLUTION OF <i>EOSPIRIFER</i> SCHUCHERT, 1913 (SPIRIFERIDA, BRACHIOPODA)	633
R. Zhan, Y. Liang and L. Meng	
LATE DARRIWILIAN TO EARLY SANDBIAN GRAPTOLITE BIOSTRATIGRAPHY IN WESTERN ZHEJIANG AND EASTERN JIANGXI PROVINCES, SE CHINA	649
Y.D. Zhang, Y.Y. Song and J. Zhang	
DETRITAL SOURCE ANALYSES OF LATE ORDOVICIAN (HIRNANTIAN?) TO SILURIAN DEPOSITS OF NORTHWESTERN AND EASTERN ARGENTINA AND CONSTRAINTS FOR PALAEOTECTONIC EVOLUTION	659
U. Zimmermann	
FROM FORE-ARC TO FORELAND: A CROSS-SECTION OF THE ORDOVICIAN IN THE CENTRAL ANDES	667
U. Zimmermann	
Announcement of the new IGCP Project 591: The Early to Middle Paleozoic Revolution	675
Authors' index	677

PRESIDENTIAL ADDRESS

A SIXTH DECADE OF THE ORDOVICIAN PERIOD: STATUS OF THE RESEARCH INFRASTRUCTURE OF A GEOLOGICAL SYSTEM

D. A.T. Harper

Natural History Museum of Denmark (Geological Museum), University of Copenhagen, Øster Voldgade 5-7,
DK-1350 Copenhagen K, Denmark. Dharper@snm.ku.dk

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INTRODUCTION

Charles Lapworth's elegant solution to the great Cambrian-Silurian dispute (Secord 1986) involved the assignation of the overlapping middle ground of the Sedgwick's Cambrian and Murchison's Silurian (Sedgwick and Murchison, 1835) to a new system, the Ordovician (Lapworth, 1879). This bold action set the agenda for the next 130 years of research on this remarkable system, its rocks and its fossils. Most authorities agree that the Ordovician Period was special if not unique (e.g. Jaanusson, 1984; Ross, 1984). Thalassocratic (a word rarely used outside an Ordovician context) conditions were promoted by extensive, epicontinental seas, with virtually flat seabeds, and restricted land areas. Magmatic and tectonic activity was associated with rapid plate movements and widespread volcanic activity. Emergent island arcs and mountain belts provided sources for clastic sediment, in competition with the carbonate belts associated with most of the epicontinental seas. Most significant was the radiation of shelly organisms (Harper, 2006), including the suspension-feeding brachiopods, bryozoans, cephalopods, corals, crinoids, and stromatoporoids, predatory cephalopods and mainly deposit-feeding trilobites together with the nekto-benthic conodonts and the pelagic graptolites. Biogeographical differentiation was marked affecting plankton, nekton and benthos and climatic zonation, particularly in the southern hemisphere, where the continents were focussed. Ironically it was these very exciting aspects of Ordovician geology that would provide difficulties for both intra and intercontinental correlation within the system. The almost bewildering range of environments and facies, some without modern analogues, and the intense provincialism of Ordovician biotas appeared to be a formidable barrier to any acceptable global chronostratigraphy for the system. Nevertheless within the last thirty years a massive international effort has involved intense research, lively debate and a degree of compromise. Three global series and seven stages are in place (Bergström et al., 2009) providing a well-grounded infrastructure and a framework to address the problems of the origins of modern climate and modern ecosystems deep in the Palaeozoic.

STARTING WITH A COMPROMISE

'Now comes a curious sequel to our story. A proposal has been made to take all Sedgwick's Arenig and Bala beds, and Murchison's Llandeilo and Caradoc, and constitute not Upper Cambrian, not Lower Silurian, but Ordovician with a view to putting an end to controversy! One shell is given to Sedgwick, the other to Murchison, but who gets the oyster?' (Clark and Hughes, 1890, p. 555: The Life and Letters of the Reverend Adam Sedgwick, Cambridge University Press).

As most of us know, our system was born out of controversy, being the centre of a bitter territorial fight between Adam Sedgwick and Roderick Murchison during the mid 19th Century (Secord, 1986). Charles Lapworth provided a compromise. Writing in 1879, he explained his position thus:

'On this arrangement the Lower Palaeozoic Rocks of Britain stand as follows:

- (c) SILURIAN SYSTEM: Strata comprehended between the base of the Old Red Sandstone and that of the Lower Llandovery.
- (b) ORDOVICIAN SYSTEM: Strata included between the base of the Lower Llandovery formation and that of the Lower Arenig.
- (a) CAMBRIAN SYSTEM: Strata included between the base of the Lower Arenig formation and that of the Harlech Grits.

Every geologists will at last be driven to the same conclusion that Nature has distributed our Lower Palaeozoic Rocks in three subequal systems, and that history, circumstance, and geologic convenience, have so arranged matters that the title here for the central system is the only one possible.'

In many respects this tripartite division was already anticipated in Bohemia by Joachim Barrande in his three faunas (Bassett, 1979) and more specifically his Stage D, including Tremadocian-Hirnantian strata (Kriz and Pojeta, 1974), conforms more precisely to the modern concept of the Ordovician than Lapworth's original definition. But although the Ordovician was accepted by colleagues elsewhere in Europe and North America, relatively quickly, in Britain this compromise was not without criticism, particularly from the Cambridge school in deference to their late professor Adam Sedgwick; it was only in 1906 that the British Geological Survey accepted the term, three years after their American colleagues (Bassett, 1979). Whilst discussion commenced on the status of the system during the 1880s at the International Geological Congress, the Ordovician was only finally ratified during the 21st International Geological Congress held in Copenhagen in 1960, where it was suggested that the lower of the two systems between the Cambrian and Devonian should be named Ordovician, some eighty years after Lapworth's bold compromise. British series (e.g. Williams et al., 1972) have been widely used on the grounds of availability, historical priority and the 'colonial' geologists who left Europe to map the then remoter parts of the world. Later definitions have been formalized and modified to aid modern international correlation with the British series (Fortey et al., 1995, 2000) and a case can be made for their wider use in the greater Avalonian and Gondwanan regions (Cocks et al., 2010). But how would the original series and stage divisions, established mainly in shelly facies in England and Wales, stand scrutiny against a set of new international criteria for the establishment of a truly global chronostratigraphy?

FROM THE REGIONS TO THE GLOBE

During the first international symposium devoted entirely to the Ordovician, Alwyn Williams delivered a keynote address on the rocks and fossils of the period supporting the modification and refinement of the

English and Welsh type sections for the British series as global standards for the system. Yet, ironically, some of the key point of his address (Williams, 1976) indicated precisely why the adherence to a single region for a complete Ordovician stratotype would prove difficult if not impossible. He noted (op. cit.) that any attempt to establish a global correlation chart for the system must take into consideration firstly, the complex and perplexing distribution of the continents, microcontinents and their sedimentary facies and secondly the contrasting biofacies and provincial distribution of its faunas. Williams went further to suggest some criteria for the Ordovician type section: 1. Areas in close proximity; 2. Homogeneity of attendant faunas and their use in correlation; 3. Flexible approach to classification of regional successions. But some of the shortcomings of the type sections in the UK were already apparent to some congress participants, not least gaps in the sections and limited correlation value of at least some of the shelly faunas.

The International Subcommittee on Ordovician was established in 1974, initially with its main focus on the lower and upper boundaries of the system; work on the global series and stage divisions would follow later. The top of the system, defined by the base of the Silurian at Dob's Linn, southern Scotland was ratified in 1985; the graptolite zone effecting boundary correlation has since been modified, but the position of the boundary remains unchanged. The base of the Ordovician took a little longer to define; it was ratified in 2000, defined at Green Point in western Newfoundland and correlated on the basis of a conodont species. During the early stages of this process, Ross (1984) in a comprehensive and thoughtful review of the system outlined the many distinctive and exciting elements of Ordovician geology; for example, its length (80 myr), its significant biotas (fishes and land plants), fluctuating climates and sea level, widespread volcanism (with the opportunity to develop isotopic dating) together with the cosmopolitan distribution of Ordovician strata across every continent, even reaching the summit of Everest (see also Harper et al., this volume). Fundamental, however, to the development of a global chronostratigraphy was the detailed cataloguing of the many as possible of commonly distinctive regional Ordovician sections. The regional correlation charts published by the IUGS formed a formidable basis for the future discussions of a global stratigraphy.

Both Webby (1998) and Finney (2006) have documented in some detail the long and arduous progress towards the functional and pragmatic global chronostratigraphy (Bergström et al., 2009), a monument to the work of the previous chairs of the Subcommittee and the many dedicated working groups. Each of the bases of the seven stages are defined as a point in time (Table 1), by a hypothetical golden spike, and the point correlated by the first appearance of a key fossil, a conodont or a graptolite (Fig. 1). All the stratotypes are readily accessible, in deeper water facies in fossiliferous sections.

SYSTEM	GLOBAL SERIES	GLOBAL STAGES	KEY GRAPTOLITE/ CONODONT(C) BIOHORIZONS
ORDOVICIAN	UPPER	HIRNANTIAN	← <i>A. ascensus</i> (GSSP-Dob's Linn)
		KATIAN	← <i>N. extraordinarius</i> (GSSP-Wangjiawan North)
		SANDBIAN	← <i>D. caudatus</i> (GSSP-Black Knob Ridge)
	MIDDLE	DARRIWILIAN	← <i>N. gracilis</i> (GSSP-Fågelsång)
		DAPINGIAN	← <i>U. austrodentatus</i> (GSSP-Huangnitang)
	LOWER	FLOIAN	← <i>B. triangularis</i> (C) (GSSP-Huanghuachang)
		TREMADOCIAN	← <i>T. approximatus</i> (GSSP-Diabasbrottet)
			← <i>I. fluctivagus</i> (C) (GSSP-Green Point)

Figure 1. Ordovician chronostratigraphy indicating the three global series and seven stages (modified from various sources).

Stage	GSSP locality	Placement of spike	Correlation of spike
Hirnantian	Wangjiawan North section, North of Yichang City, Western Hubei Province, China	0.39 m below the base of the Kuanyinchiao Bed	FAD graptolite <i>Normalograptus extraordinarius</i>
Katian	Black Knob Ridge section, Atoka, Oklahoma, USA	4.0 m above the base of the Bigfort Chert	FAD graptolite <i>Diplacathograptus caudatus</i>
Sandbian	Sularp Stream section, Fågelsång, Scania, Sweden	1.4 m below phosphorite marker bed	FAD graptolite <i>Nemagraptus gracilis</i>
Darriwilian	Huangnitang section, Changshan, Zhejiang Province, China	Base of Bed AEP 184	FAD graptolite <i>Undulograptus austrodentatus</i>
Dapingian	Huanghuachang section, Northeast of Yichang City, Hubei Province, China	10.57 m above base of the Dawan Formation	FAD conodont <i>Baltoniodus triangularis</i>
Floian	Diabasbrottet, Hunneberg, Västergötland, Sweden	Lower Tøyen Shale, 2.1 m above top of Cambrian	FAD graptolite <i>Tetragraptus approximatus</i>
Tremadocian	Green Point section, western Newfoundland, Canada	101.8 m level within Bed 23	FAD conodont <i>Iapetognathus fluctivagus</i>

Table 1. Details of the Ordovician stratotype sections.

USING THE INFRASTRUCTURE

The establishment of a workable set of international series and stages, together with a range of more precise chronostratigraphic divisions, had immediate consequences. More accurate and reproducible global studies were now feasible, with much of the impetus rising from two IGCP projects (410 and 503) closely tied to the work of the Subcommittee. Estimates of changing global diversity were now possible (Webby et al., 2004) and a more holistic approach was possible for the two main events during the period, the Great Ordovician Biodiversification and the End Ordovician Extinction. In addition to the many recent publications on the Ordovician (listed in issues of Ordovician News) a range of thematic issues and volumes has been recently published: Early Palaeozoic palaeogeography and palaeoclimate (Munnecke and Servais, 2007); Ordovician biogeography and biodiversity change (Owen, 2008) Ordovician palaeoecology (Servais and Owen, 2010); Ordovician and Silurian sea-water chemistry, sea level, and climate (Munnecke et al., 2010), Ordovician Earth System (Finney and Berry, 2010) and Early Palaeozoic palaeobiogeography (Harper and Servais, in press), the last making use of BugPlates software (www.geodynamics.no/bugs/SoftwareManual.pdf). These many studies have sharpened our focus on the importance of climatic and environmental change for the evolution of Early Palaeozoic biotas while also identifying the need for the

continued careful sampling of regional sections against a well-constrained biostratigraphy across a wide range of palaeolatitudes. Many new geochemical proxies are now available to match our more refined biotic data. Essential to our understanding of Early Palaeozoic earth systems is the continued search for links and relationships between environments, ecosystems and evolution (Fig. 2) within a precise global chronostratigraphical framework; this, we have now.

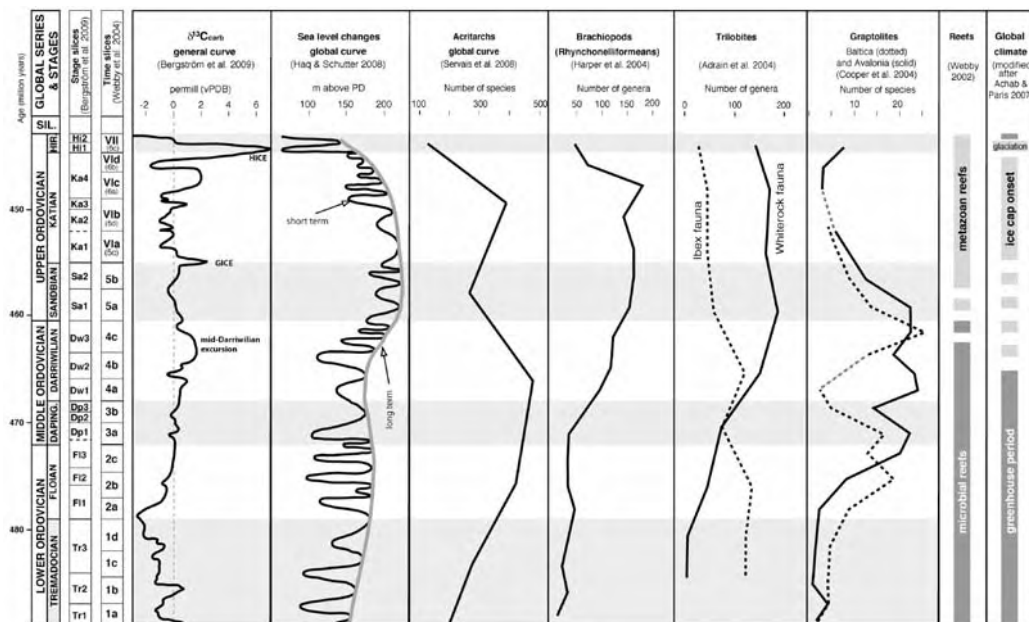


Figure 2. Compilation of biodiversity, seawater chemistry, sea-level and climate data [courtesy of Axel Munnecke; modified from data in Servais and Owen (2010) and Munnecke et al. (2010)].

CHALLENGES AHEAD

The work of the Ordovician Subcommittee, particularly a global stratigraphy, has driven fundamental and significant advances in our research infrastructure but also in the ways in which we tackle global problems. This new platform exposes a range of other challenges that the Subcommittee together with its colleagues in current and past IGCP projects and anyone interested in the Ordovician System must address. (i) An open debate on the formal definition of chronozones within the Ordovician System. This possibility arises from the time-slice concept of Webby *in* Webby et al. (2004) and the finer subdivision of the system presented by Bergström et al. (2009). This is also relevant to criticisms that the available global time divisions of the system are currently too crude for accurate correlation (Cope, 2007). (ii) Our existing boundaries may require review and thus a forum to assess the efficacy and utility of the newly-established international stages will be necessary. (iii) The new Ordovician chronostratigraphy will require a revision of regional correlation charts (and some geological maps) on the basis of new regional stratigraphic data and their relationship to the newly-established international stages. (iv) The work of IGCP 503 has brought sharply into focus the use of non-biologic methods of correlation of Ordovician strata, including

all manner of isotopic and other geochemical proxies together with sea-level change. (v) Interactive palaeogeographic maps are now available for the period using BugPlates software; biotic distributional can now be accurately plotted and analysed.

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KEYNOTE LECTURES

THE LATE ORDOVICIAN GLACIAL RECORD: STATE OF THE ART

J.-F. Ghienne

Institut de Physique du Globe, UMR7516 CNRS/ Université de Strasbourg, 1 rue Blessig,
67084 Strasbourg Cedex, France. ghienne@unistra.fr

Keywords: Glaciation, ice sheet, sea level, Gondwana, North Africa.

INTRODUCTION

The Late Ordovician glaciogenic strata were deposited over an up to 2000 m thick Cambrian – Ordovician wedge corresponding to a first-order retrogradational sequence set in an intracratonic platform setting (the so-called North Gondwana Platform, extending from Mauritania to Saudi Arabia, and from Spain to Oman). To the north (present-day coordinates), thick Cambrian – Ordovician successions are overlain by relatively ice-distal glaciogenics including one to three glacial sequences, Hirnantian in age. To the south (inner parts of the platform, e.g. towards the continental interior), thinner “pre-glacial” successions are severely truncated by glacial erosion surfaces related to a greater (at least 5) number of glacial sequences. Glacial erosion only rarely affects Panafrican or older basement rocks. Ice-flow orientations are to the WNW to NNW in Mauritania, essentially to the NNW to N in Algeria and Libya, and to the NE in Saudi Arabia and ice-stream pathways and correlative inter-stream areas have been described at the 100’s km scale. The late Ordovician glacier essentially corresponded to a large (giant?) ice sheet covering the main part of Africa. A platform setting (no mountains), and a substrate of Cambrian – Ordovician, essentially un lithified sediments characterise the Late Ordovician palaeoglacial setting.

GLACIAL FEATURES

Glacial features are distributed from the most internal part of the platform (where they are ubiquitous: Mauritanie, Algeria, Libya, Niger, Saudi Arabia) to the ice-distal zone (where they related to the maximum glacial advance: Morocco, Spain, Turkey). A wide spectrum of depositional/ deformational glaciogenic features is preserved in the Late Ordovician glacial record. Glacial erosion structures are associated with soft-sediment shear zones including intraformational striated surfaces (V-shaped striations), true ice-sediment striated interfaces, mega-scale glacial lineation (attenuated drumlins), rare roches moutonnées-like structures, liquefaction to fluidisation structures. Larger-scale features include giga-scale glacial

lineations in the form of kilometre-scale ridges and overdeepened incisions related to tunnel drainage systems. Large valleys may extend more than 50 km while channels are of limited extent (< 5 km). Incision depths range from 200 m (deepest valleys) to less than 30 m (channels) and widths from 5 km to 10 m. Palaeovalleys orientations are most often parallel to the main ice-flow orientations. Overpressures typify the subglacial environments, significantly contributing to ice flows and erosion.

Glacial deposits are poorly represented in the Late Ordovician glacial record though deformation tills, esker plugs or outwash fans (morainal banks) are often documented. More widespread are distal glaciomarine successions that are preserved both in formerly glaciated area and over the outer shelf setting beyond maximum ice fronts in present-day Europe. However, relative contributions of glacial ice (icebergs) or of sea-ice in the supply of ice-rafted debris is still to be elucidated.

NON-GLACIAL FEATURES

Glacially-controlled successions frequently include original stratigraphic architectures and facies that do not occur either in older or younger Lower Palaeozoic strata. Forced regression system tracts, involving major downward shift of facies belts are ubiquitous in the Katian to lower Silurian strata. Sharp-based shoreface deposits superimposed above lower offshore, or the abrupt occurrence of fluvial facies within a shallow-marine dominated succession are frequent.

As well, facies representing high-energy flow conditions are recurrent, in association with up to 100 m thick, prograding fluvio-deltaic systems that are demonstratively connected to the subglacial meltwater drainage. Large-scale antidune facies, or climbing-dune cross-stratification (CDCS) are ubiquitous in outburst-related, proglacial outwash successions. They occur generally as channel sandy plugs over the fluvio-glacial outwash plain or in mouth bar environments.

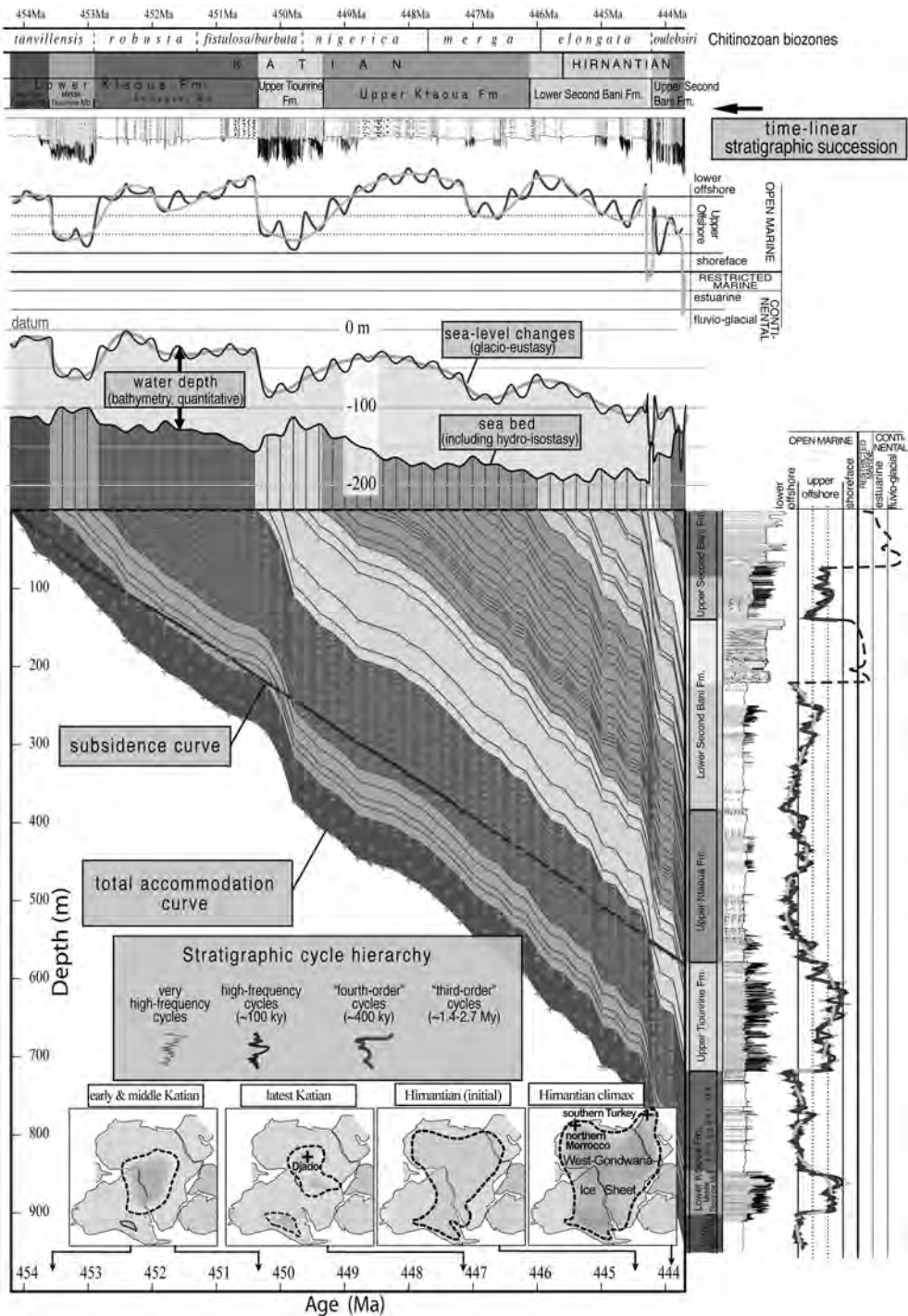
Other structures documenting a cold climate are ice-crystal imprints, grooves produced by floating ice masses in shallow-water environments (fluvial, mouth, tidal) and the occurrence of mini-basins (< 100 m in radius), the subsidence of which is controlled by the melting of a discontinuous permafrost. Ice-wedge features are at time unknown in the Late Ordovician but pingos have been described.

However, the main (up to 75%) part of the glacial record is made up of ordinary depositional facies, such as fluvial (braided streams, meanders), tidal (open-coast tidal flats, restricted marine), storm-dominated (shoreface to outer shelf) and turbiditic lobes. The latter occur preferentially in the far distal platform. Aeolian deposits have not been described even if aeolian processes were active as suggested by wind-blown sand grains, carried most probably by drifting sea ice and falling-out in offshore succession.

TIMING

The Late Ordovician glaciation was initially (1960-1990) thought to represent a long-lasting icehouse period. In the nineties, the idea was imposed of a much shorter glacial event, possibly catastrophic at the

Figure 1. Sequence analysis of the Bou Ingarf section (Late Ordovician, central Anti-Atlas, southern Morocco) and illustration of the backstripping procedure used to convert bathymetric changes in an eustatic curve. Sharp based sandstone-dominated units relate to glacio-eustatically controlled lowstand and subsequent transgressive events reflecting glaciations. Three Katian and two Hirnantian events are documented. Corresponding ice-sheet extents are figured. Only the latest Katian and the Hirnantian events are currently associated with a glacial record in North Africa. The Late Ordovician ice sheet reaches the Bou Ingarf area only during the late Hirnantian (modified from Loi et al., 2010).



geological timescale. The corresponding glaciation was limited to the Hirnantian, or even to a part of this 1-2 My long stage, as suggested by short-lived isotopic excursions and correlative biological turnovers. The sequence analysis of Late Ordovician successions in a palaeo-high latitude setting reveals that at least 3 glacial event similar to the Hirnantian event occurred in the Katian, as indicated by the stacking-pattern of shelf successions including forced regressive system tracts (Fig. 1). The latter seem to have their counterpart in low-latitude settings, either in the form of fully developed third-order regressive-transgressive cycles, or in karst horizons. The synchronous character of these inferred worldwide glacioeustatically controlled event is however difficult to ascertain as biostratigraphy is essentially based on an endemic spectrum of faunas.

Finally, the view of a long-lasting glaciation (> 25 My) including discrete short glacial events (< 1 Ma, intra-Katian events, Hirnantian, base Wenlock) prevails. Two essential questions arise. When (lower Late Ordovician, earlier?) and why the Late Ordovician glaciation or cooling phase began? Did a perennial ice sheet occupied central Africa or other relatively elevated areas in the time interval between two successive glacial events? In particular, carbonates of the Boda event (e.g. bryozoan mud mounds in Libya) may represent a strong warming event, with significant (complete?) retreat of the ice sheet, subsequent shoreline retreat and reduced clastic sedimentation favouring the development of carbonates mound (an alternative interpretation however consider they relate to "syn-glacial" cool-water carbonates).

Until recently, glacial features as listed above were ascribed to the Hirnantian glacial event(s). Only during the corresponding time interval ice fronts reach sedimentary basins around the Gondwana supercontinent, and hence were related to a glacial record. However, a pristine, Late Katian glacial record is possibly preserved as far as northern Niger, and pre-Hirnantian glaciomarine deposits are as well documented in Libya. Therefore, some of the lower glacial sequences in non-dated, Late Ordovician, glaciogenic successions may actually correspond to a pre-Hirnantian glacial record.

Dealing with the latest Ordovician, and succeeding to the late Katian Boda event, three major ice-sheet advances and intervening interglacials are recognized. A latest Katian glacial event ended just before the Katian/ Hirnantian boundary. An informal "lower Hirnantian" time interval was essentially free ice when considering sedimentary basins. An informal "upper Hirnantian" time interval is subdivided in two others glacial events, possibly as short as 100 or 400 ky, separated by a major interglacial (Fig. 1). Ice-sheet retreat from the North Gondwana resulted in the latest Hirnantian – earliest Silurian "postglacial" transgression while a further retreat (ice-free Africa?) later forced an early Llandovery flooding.

GLACIER EXTENTS

Ideas about former extents of the Late Ordovician ice sheets should ideally take into account two sets of data related to each of the glacial events and associated higher-frequency advance-retreat cycles. First, the reconstitution of ice fronts can use mappable distribution of glacial features in high palaeolatitude settings. Second, amplitudes of sea-level fall need to be estimated based on the sedimentary record of the low (carbonate platforms) to high palaeolatitude (siliciclastic, Fig. 1) settings. Sea-level fall estimates can only be appreciated in areas where no or minimal emersion occurred, discarding a number of shallow carbonate shelves. A difficulty arises regarding synchronicity of ice-sheet advances over the huge domain under consideration, as any faunal group has at time a sufficient temporal resolution. In addition, palaeogeographical reconstructions of the Gondwana margin limit ice-front reconstruction. For instance, tunnel valleys documented in NW Spain may be interpreted as the result of a satellite ice cap. As they are thought to be related to the main Gondwana ice sheet, a revision of the north Gondwana puzzle is favoured.

In North Africa, the ice-sheet sizes (latest Katian, Hirnantian) successively increased through time (Fig. 1). In Morocco and Turkey, the first Hirnantian glacial event was recorded by drastic sea-level fall and/or glaciomarine facies but no glacial surface has been documented. Only in the latest Hirnantian the ice sheet attained these areas far from the ice centres. This time interval corresponds to the glacial climax, during which continuous ice fronts are inferred from Mauritania, Morocco, northern Algeria and Libya, Turkey to SE Saudi Arabia, and Ethiopia. The southern extent of this vast ice sheet is largely debated. Did the north Gondwana ice sheet override Central Africa, joining southern ice fronts of South Africa and southern South America? Such an ice sheet would be approximately equivalent to an up to 200–250 m sea-level fall, i.e. 135–170 m of apparent sea-level fall at continental margins after corrections for hydro-isostasy of the oceanic crust. Such an estimate is apparently beyond all the proposed latest Ordovician sea-level fall estimates, which are currently in the 50–100 m range. An alternative view may consider fully diachronous development phases of more spatially restricted ice sheet comparable in size to the Late Pleistocene Laurentides.

CONCLUSIONS

Taking into account pre-Hirnantian ice sheets and related cumulated sea-level falls, the maximum reconstruction for the Late Ordovician ice-sheet extent is however achievable. To consider that a significant ice sheet recurrently or permanently occupied the centre of the Gondwana landmass throughout the Late Ordovician – and most likely the Early Silurian – may reconcile relatively moderate but time-restricted Hirnantian eustatic sea-level fall amplitudes, associated coeval ice-equivalent sea-level change in the 60–120 m range and palaeoglacial reconstructions that show at time of the glacial climax a giant ice sheet. The latter that should have resulted in an up to 250 m ice-equivalent sea-level change relative to Late Katian free-of-ice(?) conditions, may have cover up the entire West Gondwana from Arabia to North and West Africa and part of South America and South Africa. In this case, the Late Ordovician, glacial-maximum ice-sheet may have been the biggest ice masse of the Phanerozoic.

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NEW INSIGHTS FROM EXCEPTIONALLY PRESERVED ORDOVICIAN BIOTAS FROM MOROCCO

P. Van Roy

Research Unit Palaeontology, Department of Geology and Soil Science, Ghent University, Krijgslaan 281/S8, B-9000 Ghent, Belgium. peter.vanroy@ugent.be

Keywords: Burgess Shale, Cambrian, Ediacaran, Ordovician, Fezouata Biota, Tafilalt Biota, Great Ordovician Biodiversification Event, Konservat-Lagerstätten.

INTRODUCTION

The Great Ordovician Biodiversification Event was a pivotal episode in the history of life, replacing the Cambrian Evolutionary Fauna by the Palaeozoic Evolutionary Fauna which dominated the marine realm until the end-Permian mass extinction. During this Ordovician radiation, most phyla diversified more rapidly than at any other time in the Phanerozoic: diversity increased twofold at the ordinal level, three times at the family level, and nearly four times at the genus level (Droser and Finnegan, 2003; Harper, 2006; Webby et al., 2004).

Our knowledge of Ordovician communities is based almost entirely on the mineralised fossil record; although exceptionally preserved biotas have an important role to play in unravelling the evolution of Ordovician organisms and ecosystems, their contribution so far has been limited because of their scarcity and the fact that the few known Middle and Late Ordovician assemblages represent restricted marine environments (Farrell et al., 2009; Liu et al., 2006; Young et al., 2007). This situation recently improved markedly with the discovery of two exceptionally preserved biotas in the Ordovician of south-eastern Morocco.

THE FEZOUATA BIOTA

The first of these is the Fezouata Biota, which is encountered over an extensive area in the Draa Valley, north of Zagora (Fig. 1); it reveals the first Ordovician exceptionally preserved biotic complex from a normal, open marine setting. The assemblages range in age from latest Tremadocian to late Floian and represent the only exceptionally preserved fauna documenting the prelude to and early stages of the Ordovician radiation (Van Roy et al., 2010). The Fezouata biota shows considerable diversity and contains

a high number of taxa typical of Cambrian Burgess Shale-type faunas in association with more modern forms (Van Roy, 2006a; Van Roy and Briggs, 2011; Van Roy and Tetlie, 2006; Van Roy et al., 2010; Vinther et al., 2008; Fig. 2 A-D). The preservation of the fossils, entombment within and below fine-grained event beds, followed by early diagenetic pyritisation, is similar to that of the Early Cambrian Chengjiang fauna of China (Gabbott et al., 2004; Van Roy et al., 2010; Vinther et al., 2008).

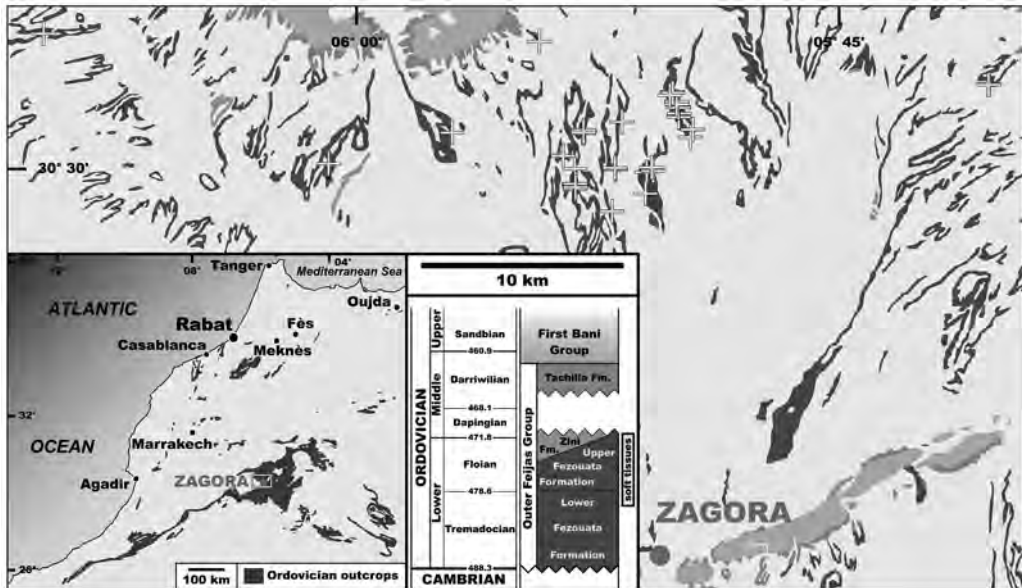


Figure 1. Ordovician outcrop map of the area north of Zagora, SE Morocco, showing localities (crosses) in the Lower and Upper Fezouata formations that yield exceptionally preserved fossils. Inset shows the position of the study area within Morocco and the stratigraphic context (from Van Roy et al., 2010).

THE TAFILALT BIOTA

The second exceptionally preserved biota occurs over a wide area in the Tafilalt region, in a triangle roughly demarcated by the towns of Erfoud, Rissani and Mesissi. Contrary to the deeper-water Fezouata Biota, the Tafilalt assemblages represent a shallow marine environment, and stretch temporally from the middle Sandbian to the middle Katian (Van Roy, 2006a). Only a relatively small number of soft-bodied taxa is preserved, with paropsonemid eldonioids (Fig. 2 E, F) and various holdfasts dominating (Alessandrello and Bracchi, 2003; Samuelsson et al., 2001; Van Roy, 2006a). These are invariably complemented by rich skeletal faunas and trace fossils (Alessandrello and Bracchi, 2006; Hunter et al., 2010; Lefebvre et al., 2010; Samuelsson et al., 2001; Van Roy, 2006a, 2006b), indicating that none of the sites represents a restricted environment. The soft-bodied organisms are preserved as moulds and casts in coarse sandstones, a mode of preservation which is strikingly similar to that of the terminal Neoproterozoic Ediacara biota.

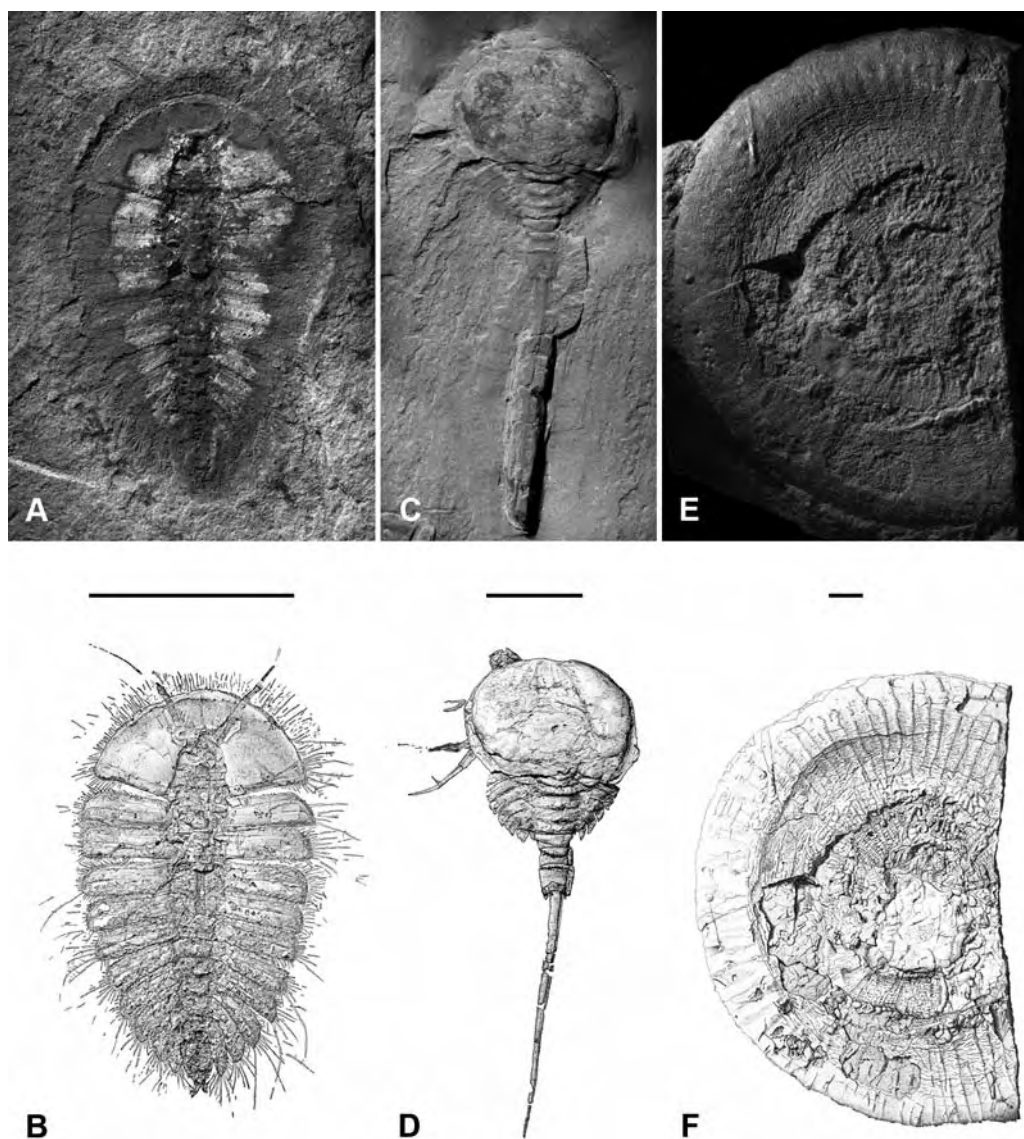


Figure 2. Exceptionally preserved fossils from the Ordovician of Morocco. A, B, Cheloniellid arthropod, top of Upper Fezouata Formation, Floian. C, D, Basal xiphosurid arthropod (horseshoe crab), showing fused preabdomen, base of Upper Fezouata Formation, Floian. E, F, Paropsonemid eldonioid, top of First Bani Group, Sandbian. All scale bars equal 10 mm.

CONCLUSIONS

The Fezouata Biota marks the first occurrence of Burgess Shale-type faunas after the Middle Cambrian, and shows that their perceived absence from younger deposits is a taphonomic artefact rather than the result of extinction and replacement of these biotas (Allison and Briggs, 1993; Aronson, 1992, 1993; Conway Morris, 1989; Orr et al., 2003). Burgess Shale-type taxa continued to impact the diversity and ecological structure of deeper marine communities well after the Middle Cambrian. This questions the concept of a sudden dramatic turnover between the Cambrian and Palaeozoic Evolutionary Faunas; concurrently, the presence of a number of advanced, typically post-Cambrian elements among the non-mineralised taxa indicates that significant diversification had already occurred prior to the Ordovician (Van Roy et al., 2010).

The Tafilalt Konservat-Lagerstätten show that the Ediacaran taphonomic window did not close with the onset of the Phanerozoic (Samuelsson et al., 2001; Van Roy, 2006a), and may call into question the absolute importance of microbial mats in Ediacara-style preservation (Gehling, 1999); while mats were no doubt of importance, the Moroccan fossils nevertheless suggest that preservation for a large part depended on the resistant properties of tough, leathery integuments (B. MacGabhann, pers. comm.). At the same time, the presence of abundant eldonioids unequivocally shows that Ediacara-style preservation of metazoan soft tissues is possible (Samuelsson et al., 2001; Van Roy, 2006a), undermining one of the main arguments for the Vendobionta hypothesis (Seilacher, 1989, 1992).

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PAPERS AND ABSTRACTS

WHOLE-ROCK AND ISOTOPE GEOCHEMISTRY OF ORDOVICIAN TO SILURIAN UNITS OF THE CUYANIA TERRANE, ARGENTINA: INSIGHTS FOR THE EVOLUTION OF SW GONDWANA MARGIN

P. Abre¹, C. Cingolani², B. Cairncross¹ and F. Chemale Jr.³

¹ Department of Geology, University of Johannesburg, PO Box 524, Auckland Park 2006, Johannesburg, South Africa.
paulinabre@yahoo.com.ar

² Centro de Investigaciones Geológicas, CONICET-Universidad Nacional de La Plata, Calle 1 n° 644, B1900TAC La Plata, Argentina.
ccingola@cig.museo.unlp.edu.ar

³ Núcleo de Geociencias, Universidade Federale do Sergipe, Brazil.

Keywords: Cuyania terrane, Ordovician to Silurian, provenance, geochemistry, U-Pb detrital zircon dating.

INTRODUCTION

The Cuyania terrane in central Argentina (Fig. 1) is characterized by a Mesoproterozoic (Grenvillian-age) basement with depleted Pb isotopic signatures and Mesoproterozoic Nd model ages resembling basement rocks of the same age from Laurentia (Ramos, 2004; Sato et al., 2004 and references therein). Several authors have proposed para-autochthonous (Aceñolaza et al., 2002; Finney et al., 2005) versus allochthonous (e. g. Ramos et al., 1986; Dalziel et al., 1994; Astini et al., 1995; Thomas and Astini, 1996) geotectonic models for the early Palaeozoic evolution of the Cuyania terrane. The tectonic evolution of the Cuyania terrane is a substantial part of the understanding of the evolution of the western border of southwest Gondwana. Several morphostructural units form the Cuyania composite terrane (Fig. 1; Ramos et al., 1996): The Precordillera *s.s.*, the Western Pampeanas Ranges and the San Rafael and Las Matras blocks. However, the boundaries of the terrane are still not well-constrained (Astini and Dávila, 2004; Porcher et al., 2004; Casquet et al., 2006).

A combination of several methodologies including geochemistry, Sm-Nd, Pb-Pb and U-Pb detrital zircon dating was applied to several clastic Ordovician (Los Sombreros, Gualcamayo, Los Azules, La Cantera, Yerba Loca, Empozada, Trapiche, Sierra de la Invernada, Portezuelo del Tontal, Las Vacas, Las Plantas and Alcaparrosa Formations) and Ordovician to Silurian (Don Braulio and La Chilca Formations) units of the Cuyania terrane (Fig. 2). The combination of these different approaches can give accurate information in order to constrain the probable sources that provided detritus to the Cuyania terrane and ultimately to constrain the existing models about its origin.

GEOLOGICAL SETTING

The fourteen units here studied crop out within the Precordillera *s.s.* Based on stratigraphy and structural features, the Precordillera *s.s.* has classically been divided into Eastern, Central, Western and Mendoza domains (Fig. 2). A carbonate platform overlaid by predominately clastic deposition within a shallow basin characterized the Eastern and Central domains (comprises the Gualcamayo, Los Azules, Las Vacas, Las Plantas, Trapiche, La Cantera, Don Braulio and La Chilca Formations).

Western Precordillera is characterized by turbidite deposition within a deep-sea basin with interlayered and intruded mafic to ultramafic igneous rocks and comprises the slope-type (olistostromic) deposits adjacent to the continental rise (Los Sombreros, Sierra de la Invernada, Portezuelo del Tontal Yerba Loca and Alcaparrosa Formations).

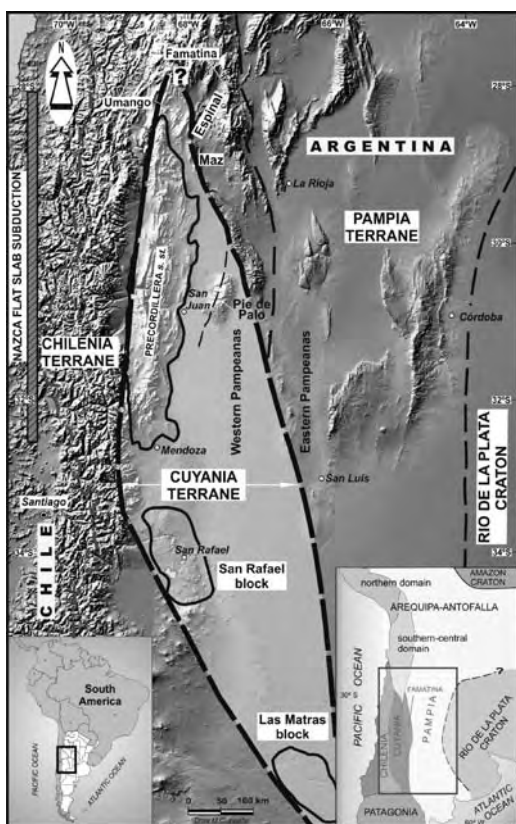


Figure 1. Satellite image based map showing Cuyania terrane boundaries as dashed lines and blocks boundaries as continuum lines. All the entities forming the Cuyania terrane develop a Grenvillian-age basement characterized by Nd, Sr and Pb depleted isotopic signatures (Ramos, 2004; Sato et al., 2004).
Righter inlet: location of neighboring terranes.

GEOCHEMISTRY

The Chemical Index of Alteration (CIA) quantitatively assesses the weathering effects on sedimentary rocks. The CIA values of the Mendoza, Western, Central and Eastern Precordillera range from 52 to 77, indicating intermediate to strong chemical alteration. In general, samples of all the units studied have Th/U ratios between 3.5 and 4, which is typical for derivation from upper crustal rocks. Samples with high Th/U ratios probably caused by loss of U due to weathering are also observed, particularly within units of the Western Precordillera.

Th/Sc ratios of the units studied show a general tendency indicating an unrecycled upper continental crust source composition, particularly for samples of the Western Precordillera. Effects of sedimentary recycling are evident for the Central and Eastern Precordillera where they are also accompanied by concomitant high SiO_2 concentration. Sandstones of the Precordillera of Mendoza show Zr/Sc values indicating the influence of recycling. The chondrite normalized REE patterns for all the sequence studied tend to be depleted in LREE and enriched in HREE compared with the PAAS (which reflects the average composition of the upper continental crust). A negative Eu-anomaly ($\text{Eu}_N/\text{Eu}^* = \text{Eu}_N / (0.67\text{Sm}_N + 0.33\text{Tb}_N)$) can be observed and samples also display a flat heavy REE distribution. Disturbed

pattern in few samples of the Los Sombreros Formation (Western Precordillera) indicate remobilization of REE.

ISOTOPE GEOCHEMISTRY

Nd isotopes indicate ϵ_{Nd} of -5.4, $f\text{Sm}/\text{Nd}$ -0.34 and T_{DM} 1.57 Ga in average for all the units studied. T_{DM} ages of two samples of the Los Sombreros Formation are aberrant due to REE fractionation, as indicated by geochemistry. The dataset is similar to those from other Ordovician to Silurian units of the Cuyania terrane (Cingolani et al., 2003; Gleason et al., 2007; Abre et al., 2011).

ϵ_{Nd} values for Ordovician to Silurian clastic rocks of the Cuyania terrane are within the ranges of variation of data from Famatina, the Laurentian Grenville crust, Central and Southern domains of the Arequipa-Antofalla Basement, basement of the Cuyania terrane and from the Western Pampeanas Ranges. The T_{DM} ages are comparable to T_{DM} ages for Mesoproterozoic and Palaeozoic rocks of the Cuyania terrane. Similar data are known from Mesoproterozoic rocks from Antarctica, Malvinas plateau and Natal-Namaqua Metamorphic belt and the Western Pampeanas Ranges.

$^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ range from 18.82 to 21.20, 15.67 to 17.27, and 38.7 to 42.93 respectively, for the Ordovician to Silurian units of the Cuyania terrane. Pb ratios are similar to values obtained for the Ponón Trehué Formation of the San Rafael block (Abre et al., 2011). Comparing the lead isotopic system with probable source areas it is evident that the datasets from the basement of the Cuyania terrane and from Proterozoic rocks of Eastern North America. Mesoproterozoic rocks from the Natal-Namaqua Metamorphic belt, Malvinas Microplate, West Antarctica and East Antarctica have different Pb isotopic signatures from those of the Cuyania terrane.

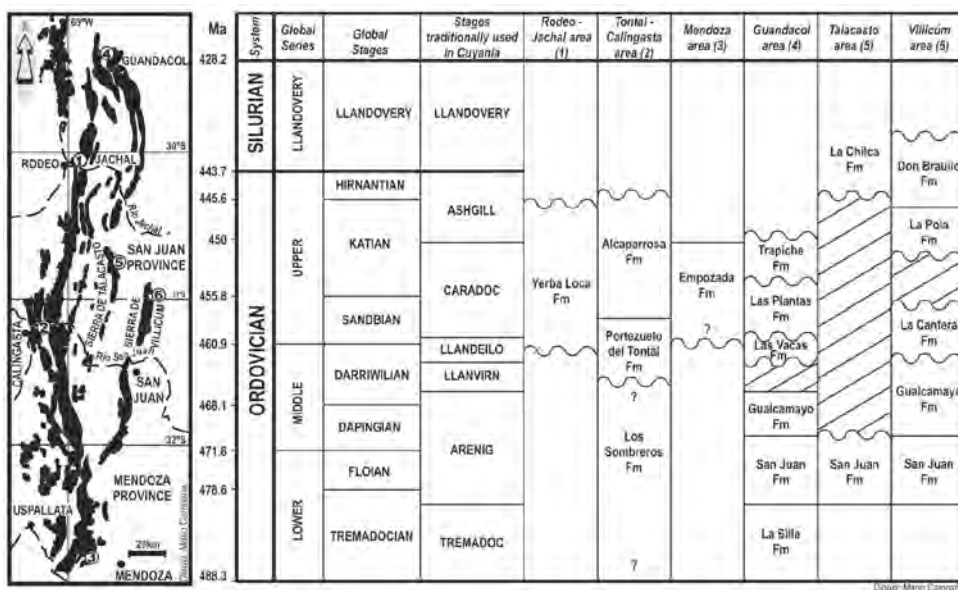


Figure 2. Ordovician to Silurian units correlation chart and location of the studied regions within the Precordillera of Western Argentina.

U–Pb dating of single detrital zircons was carried out in six samples of the Ordovician to Silurian record of the Cuyania terrane. Three units from the Eastern and Central Precordillera were analyzed: The La Cantera Formation ($n=38$) show peaks at 1140.6 Ma, 1351 Ma, and 1553 Ma in order of abundance. The Trapiche Formation ($n=60$) shows main peaks at 1024 Ma, 1089 Ma, 1162 Ma, 1360 Ma, 1456 Ma and 1265 Ma, with minor peaks at 661 Ma, 802 Ma and 1789 Ma. Detrital zircon ages main peaks for the Don Braulio Formation ($n=42$) cluster at 989 Ma, 1151 Ma, 1392 Ma, 658 Ma and 1553 Ma in order of abundance. Only three grains are Palaeoproterozoic in age (peak at 1941 Ma).

Two formations were studied from the Western Precordillera: Detrital zircon ages main peaks for the Yerba Loca Formation ($n=63$) are at 1023 Ma, 1099 Ma, 617 Ma, 1420 Ma, 1210 Ma, 526 Ma and 1361 Ma. Minor peaks are displayed at 760 Ma, 1567 Ma, 2224 Ma and 2499 Ma. Detrital zircon dates of the Alcaparrosa Formation ($n=49$) display main peaks at 1083 Ma, 544 Ma, 1275 Ma and 940 Ma in order of abundance. Minor peaks are found at 1825 Ma and at 1597 Ma. Three grains have an age in between 460 and 495 Ma.

The Empozada Formation ($n=38$), cropping out at Precordillera of Mendoza, shows main peaks at 1040 Ma, 1341 Ma, 1153 Ma and at 982 Ma. Minor peaks are displayed at 603 Ma and at 1106 Ma.

The detrital zircon dating here presented constrain the sources as being dominantly of Mesoproterozoic age, with a main peak in the range 1.0 to 1.3 Ga and a subordinate peak between 1.3 and 1.6 Ga, but inputs from both older (1.6 to 2.5 Ga) and younger (Neoproterozoic, Cambrian and Ordovician) sources are also recorded.

DISCUSSION

Several areas should be evaluated as sources with regards to the palaeogeography of the Ordovician to Silurian Cuyanian basin. Sedimentological characteristics such as palaeocurrents and the lack of important recycling tend to indicate that areas located far away of Cuyania and/or those located to the west can be ruled out as sources (e.g. the Amazon craton and the Chilenia terrane). Comparison of isotope data, including detrital zircon dating allow concluding that the Famatinian magmatic arc, the Grenville Province of Laurentia, the Natal- Namaqua Metamorphic belt, Malvinas Microplate, West Antarctica and East Antarctica were not sources for the Ordovician to Silurian basin of the Cuyania terrane.

On the other hand, Mesoproterozoic rocks that could have contributed to the bulk of detritus are: 1) the southern extensions of the basement of the Cuyania named the Cerro la Ventana Formation (Cingolani et al., 2005) of the San Rafael block; 2) the Western Pampeanas Ranges; 3) the Arequipa-Antofalla Basement, however, the northern termination of the Cuyania terrane (Jagüé area) is still under investigation and therefore the tectonic relationship with the Arequipa-Antofalla rocks is currently not well determined (Astini and Dávila, 2004). A provenance from the basement of the Cuyania terrane (Cerro La Ventana Formation) and from the Western Pampeanas Ranges was also deduced for clastic sedimentary Ordovician units of the San Rafael block (Cingolani et al., 2003; Abre, 2007; Abre et al., 2011).

CONCLUSIONS

The uniformity shown by the provenance proxies indicate that there were no important changes in the provenance within Eastern, Central, Western Precordillera and the Precordillera of Mendoza. Geochemical analyses indicate a dominant unrecycled upper crustal component. Sm-Nd and Pb-Pb data allow discarding

certain areas as probable sources. Detrital zircon dating further constrains the sources as being dominantly of Mesoproterozoic age, but with contributions from Ordovician, Cambrian, Neoproterozoic and Palaeoproterozoic sources. The combination of the different provenance approaches applied indicates that the Cuyanian basement and the Western Pampeanas Ranges (and less probably the Arequipa-Antofalla Basement) were the main sources.

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OCEAN CURRENTS AND STRIKE-SLIP DISPLACEMENTS IN WESTERN GONDWANA: THE CUYANIA HYPOTHESIS IN CAMBRIAN-ORDOVICIAN TIMES

F.G. Aceñolaza

INSUGEO, Miguel Lillo 205, 4000 Tucumán, Argentina.
facenola@infovia.com.ar

Keywords: Gondwana, Cuyania hypothesis, Cambrian, Ordovician, paleogeography.

INTRODUCTION

The existence of Laurentian faunas in the Cambrian-Ordovician of the Cuyania region (Precordillera) of west Argentina is known from the mid 20th century (among others: Harrington and Leanza, 1943; Borrello, 1971, with references). Even though these authors have recognized links of local trilobite faunas with similar taxa in Laurentia, they did not clearly state how they reached the outcrops in the provinces of San Juan and Mendoza.

After the full development of plate tectonics to Lower Paleozoic reconstructions, since the 1990's several authors have considered that these faunas arrived on a migrating microplate from Laurentia, which possibly drifted to the western side of Gondwana. Some of the main contributors to this hypothesis were the papers by Astini et al. (1995), Keller (1996), Astini and Thomas (1999) and Benedetto et al. (2009, with references), who supported their bases in the Cambrian-Ordovician faunal context.

The aforementioned hypothesis was widely accepted by those authors supporting that the formation of the gondwanic border was due to the accumulation of microplates derived from Laurentia (Cawood, 2005, with references). The paleontological context was used as the most convincing argument on which the physical and faunal relationship between both paleocontinents was founded.

An alternative hypothesis was proposed by which the presence of these fossils in the South American margin was due to the actions of ocean currents dispersing the Laurentian fauna and allowing it to reach Gondwana at appropriate paleolatitudes (Aceñolaza and Toselli, 2000; Aceñolaza et al., 2002, with references; Finney, 2007, with references). These authors have not only posed doubts about the paleontological arguments favoring the Cuyania allochthony, but are also skeptic about the geodynamic mechanisms behind the transport of such "microcontinent" in its migration from Laurentia.

OCEAN CURRENTS AS FAUNAL DISPERSING AGENTS

Nowadays, it is known that ocean currents are a relevant dispersing factor for marine organisms. This concept may also be applied to the geological past. Remarkably, currents are originated by the combined

action of wind, variations in atmospheric pressure and some other factors such as temperature, salinity, the Coriolis effect generated by the rotation of the planet, etc., affecting the motion of oceanic water masses (Hopkins, 1991).

The mechanics of oceanic and marine currents favoured dispersion of the organisms living therein in the way of eggs, larvae or cysts to huge distances without altering their vital possibilities. Besides, when environmental conditions are adequate, they may even hatch and develop (among others: Valentine, 1971; Skeltema, 1971; Sumich, 1999; Cecca, 2002).

When dispersion takes place in planktic larvae, some animals have proved to endure over six months before settling (Cecca, 2002); whereas if the same occurs as egg or cyst, this planktic phase may surpass a year and, consequently, may reach a greater dispersion.

In this context, we should consider the existence of present-day ocean currents, such as the "Gulf Stream", with an average speed of 7 km/h, the "Malvinas Current" with 2 km/h or the "Canarias Current" with 0.6 km/h, all of them in the Atlantic. In the Pacific Ocean, the "Cromwell Equatorial Current" reaches 1.5 m/sec; however, the "Antarctic Circumpolar Current" has a speed ranging from 4 to 6 km/h. Particularly in those currents of first magnitude, water masses has variable volume extending through several thousands of kilometers, either at a surface level or in deeper waters. One case is the "Greenland current", once it passes through equatorial latitudes, it sinks and eventually connects with the "Antarctic Circumpolar Current" (Hopkins, 1991).

The aforementioned references should be useful to show the relevance of sea currents as an efficient means of transport for water masses as well as for the fauna inhabiting them. In this sense, we must discuss some concepts referred to cosmopolitan and endemic faunas. The former includes taxa lacking geographical limitations and thus, reaching a global distribution; whereas in the latter, there can be several factors causing environmental, geographical and even genetic restrictions (Cecca, 2002).

Therefore, platforms, islands or microcontinents are appropriate regions for endemism; however, deep water regions are more prone to produce pandemic faunas. This implies that epibenthic organisms may easily colonize environments having more restrictions and being relatively shallower. Instead, pelagic forms display a greater dispersion without any limitations other than those granted by water turbidity, salinity or temperature.

These concepts were applied to Cambrian-Ordovician times by Ross (1975), where attention was paid to ocean currents as active agents for dispersion of fauna, especially trilobites. Based on it, he proposed models of ocean circulation during the Cambrian and Ordovician periods, pointing out that the existence of a Laurentian fauna in the west of Argentina was due to an "Andean Stream" which, as a means of transport, would have caused the dispersion. This criterion has also been adopted by Finney (2007, with references) to remark the possibility that ocean currents were the process by which Laurentian fauna was carried to Cuyania.

Evidently, if there was an "Andean Stream", water masses, due to the *Coriolis* effect, would have derived the larvae, eggs or cysts of most of the existing taxa in Laurentia in the direction of the Gondwanan margin. Once the platforms were colonized, this could have caused the appearance of endemic and pandemic species. In this case, the endemism took place on the Gondwanan sea platform, where different elements existing in Laurentia were found.

When authors such as Benedetto et al. (2009, with references) support that the displacement of fauna was on a migrating microcontinent, they ground this scheme on a limited number of taxa. This hypothesis does not mention anything about ocean current action –which undoubtedly existed–, which could make invalidate the model. Even if these authors remark that the faunal similarity between Laurentia and

Cuyania was particularly high in the Cambrian, the relation changes in the Ordovician, with the incorporation of Baltic elements. Thus, it is interpreted that the conditions of the ocean varied and subsequently, the ocean currents underwent alterations due to the fact that Gondwana moved to higher latitudes.

Remarkably, hardly ever have authors dealing with Cambrian-Ordovician Laurentian faunas referred to the ones in Antarctica (Trans-Antarctic and Ellsworth Mountains), in spite of its scarce, though recognized, existence. In that area, pandemic faunal elements are common with Cuyania. However in Cuyania there are either planktic, benthic or endemic faunas, which cannot be confronted to current findings related to Antarctica. Obviously, the increase of knowledge about the fauna in Ellsworth Mts. may cause a change in this context and the concept thereof.

"STRIKE-SLIP" AND CONTINENTAL DRIFT MECHANICS

Instances of a continent or a microcontinent model colliding against another, provide a set of fundamental geological data as a basis to analyze this type of tectonics:

First, the remarkable existence of a "collisional orogen" folded between the crushing blocks, and

The record of "high pressure" rock and mineral types (blue schists).

These factors, are both very relevant, and are not recorded in the contact zone between Cuyania and the neighboring Famatinian belt. Supposedly, in the first case either in the "microcontinent model" or the "Para-autochthonous model" of Gondwana, an inexistent important sedimentary prism was developed. When both prisms collided, they probably formed a significant orogenic structure which, after the impact, stood between them. There is no evidences along the Cuyania-Gondwana contact zone of such a structure. In addition, this suture did not leave any trace of rock exhibiting high pressure values and/or minerals featuring these metamorphic facies.

Neither of these processes took place: there is no "collisional orogen" standing between Cuyania and Gondwana, and there is no record of high pressure in metamorphic rocks of the area. Most of them are below 10 Kb, with very few local values slightly exceeding it. Likewise, research by Galindo et al. (2004) show the affinity of the Cuyanian western Sierras Pampeanas with Gondwana. Isotopic values by means of zircons, display identical data to those obtained in the eastern Sierras Pampeanas and Cordillera Oriental (Aceñolaza et al., 2010, with references).

This context led us to propose the existence of a strike-slip type of tectonic mechanism to explain the detachment and position of Cuyania within the margin of Gondwana as a para-autochthonous block (Aceñolaza and Toselli, 2000). The development of this tectonic system is widely known from cases where there is a notorious oblique collision causing strike-slip faults with lateral displacement. There are many examples of this type of tectonics, particularly on the West American border (Moore and Twiss, 1995; Storti et al., 2003; Cunningham and Mann, 2007). We must admit that even today on the Pacific Ocean border there are good examples of this sort of tectonism, which has already been noted in the San Andreas Fault System by Finney (2007).

In the light of the Laurentian origin for Cuyania supported by Astini et al. (1995), Astini and Thomas (1999), Keller (1996), Benedetto (1993) and Benedetto et al. (2009), an important incongruence can be pointed out: the inability to explain the allochthony following an adequate tectonic mechanism.

CONCLUSION

Following an actualistic criterion, it is considered that Laurentian faunas recorded in Cuyania have arrived due to an ocean current system carrying them as larvae, eggs and/or cysts. As a consequence, we may reject the possibility of a displacement on a hypothetical and migrating "microcontinent model of Laurentian origin". Its displacement mechanics is unlikely on the basis of supporting sedimentary and tectonic data, such as the inexistence of high pressure rocks, the absence of a true orogen, and the lack of a necessary sedimentary prism in the eastern border, where the hypothetical collision would have taken place.

Therefore, this contribution reasserts that Cuyania is a para-autochthonous gondwanic block displaced by strike-slip faults, derived from the oblique collision of the oceanic Paleo-Pacific Plate with the border of Gondwana.

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A NEW TRILOBITE BIOSTRATIGRAPHY FOR THE LOWER ORDOVICIAN OF WESTERN LAURENTIA AND PROSPECTS FOR INTERNATIONAL CORRELATION USING PELAGIC TRILOBITES

J.M. Adrain

Department of Geoscience, University of Iowa, 121 Trowbridge Hall, Iowa City, Iowa 52242, USA.
jonathan-adrain@uiowa.edu

Field based revision of classic sections in the type Ibexian area of western Utah (Hintze, 1951, 1953) and the Bear River Range of southeastern Idaho (Ross, 1949, 1951), along with sections in east-central Nevada, has revealed an order of magnitude more faunal information than was previously known. There are nearly continuous sequences of rich, closely spaced, and beautifully preserved secondarily silicified assemblages spanning the entire Lower Ordovician. The existing trilobite biostratigraphic scheme (Ross et al., 1997) was based on the original fieldwork carried out in the late 1940s, with only minor additions or modifications over the next half century. An extensive field sampling program permits the development of a much more detailed scheme. Ross et al. (1997), for example, recognized a total of five trilobite zones for the Tulean and Blackhillsian stages (the upper two stages of the Ibexian Series). Adrain et al. (2009), in contrast, recognized at least 15 distinct zones in this interval, and this number is now increased to 17. Revision in progress of the Stairsian Stage replaces the four zones of Ross et al. (1997) with 13 new or restricted zones. Revision of the upper part of the Skullrockian Stage replaces the single *Bellefontia-Xenostegium* Zone of Ross et al. (1997) with seven distinct zones.

The new biostratigraphic scheme is not based on more finely parsing stratigraphic distribution, nor is it a function of differing species concepts. It derives largely from extensive new discoveries in large swathes of the sections previously given only cursory treatment via undocumented faunal lists (or not sampled at all). In addition, existing horizon diversities reported by Hintze (1953) have in many cases been more than doubled, almost certainly as a result of greatly increased sample size.

Among the rich new faunas encountered are pelagic (mesonektic) telephinid trilobites belonging to the genera *Goniophrys* Ross, 1951, *Carolinites* Kobayashi, 1940, and *Opipeuterella* Fortey, 2005. These taxa include some of the few trilobite species with convincingly established intercontinental distributions during the Ordovician (e.g., Fortey, 1975; McCormick and Fortey, 1999). In addition to the very widely distributed *Carolinites genacinaca* Ross, 1951, which has its type horizon in the study area, the faunas include *Opipeuterella inconniva* (Fortey, 1974), described from Spitsbergen, and *O. insignis* (Henderson, 1983), described from Australia. All of these species are key to an emerging global framework for upper Floian trilobite biostratigraphy based on pelagics. This framework may be extended to the lower Floian via the discovery of a sequence of new, stratigraphically early species of *Carolinites* with which subsequent discoveries elsewhere may be matched, and particularly by a sequence of early, well preserved species of

Opipeuterella. These can be compared directly with a similar sequence described by Laurie and Shergold (1996) from the Emanuel Formation of Western Australia.

This latter comparison suggests, via tie points with the Australian graptolite scheme, that the Tremadocian/Floian boundary occurs much lower down in the western Laurentian succession than has previously been assumed. It may approximately correspond with the Stairsian-Tulean boundary and cannot be very far above it. The distinction between the Laurentian Tulean and Blackhillsian stages is trilobite-based, but there are few compelling faunal reasons to make a stadial distinction and the base of the Blackhillsian is unlikely to be easily recognized outside the western Laurentian region. The Tulean and Blackhillsian stages together approximately represent the Floian.

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A PERI-GONDWANAN ARC ACTIVE IN CAMBRIAN-ORDOVICIAN TIMES: THE EVIDENCE OF THE UPPERMOST TERRANE OF NW IBERIA

R. Arenas¹, J. Abati¹, S. Sánchez Martínez¹, P. Andonaegui¹, J.M. Fuenlabrada¹, J. Fernández-Suárez¹
and P. González Cuadra²

¹ Departamento de Petrología y Geoquímica e Instituto de Geología Económica (CSIC), Universidad Complutense de Madrid,
José Antonio Novais 2, 28040 Madrid, Spain. arenas@geo.ucm.es, abati@geo.ucm.es, s.sanchez@geo.ucm.es,
andonaeg@geo.ucm.es, jmfuenla@pas.ucm.es, jfsuarez@geo.ucm.es

² Instituto Geológico y Minero de España, Cardenal Payá 18, 15703 Santiago de Compostela, A Coruña, Spain.
santiago@igme.es

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The Variscan suture preserved in the NW of the Iberian Massif shows some stacked terranes generated in different tectonic settings (Figs. 1 and 2). These terranes are key elements to reconstruct the Paleozoic paleogeography in the peri-Gondwanan realm. The upper units of the allochthonous complexes of Galicia-Trás-os-Montes Zone were thrust over ophiolitic units, and they consist of a thick pile of terrigenous metasediments intruded by large massifs of gabbros and granitoids. The lower part of this terrane was affected by a high-P and high-T event dated at c. 410-390 Ma (Fernández-Suárez et al., 2007) related to the final assembly of Pangea, but most of the sections above did not record pervasive tectonothermal events after the Early Paleozoic. In this context, this uppermost terrane shows an evolution unique in the Variscan section of NW Iberia, with distinctive characteristics not presented neither in the autochthonous domains nor in the other terranes included in the allochthonous complexes. The uppermost terrane can be interpreted as a well preserved section of a peri-Gondwanan magmatic arc, active at least between Middle Cambrian and Early Ordovician times. The following characteristics of the metasedimentary series, magmatism, deformation and metamorphism are arguments favouring this interpretation.

The terrigenous metasedimentary series shows a low grade top turbiditic sequence with 3000 m of greywackes with average major and trace element compositions similar to PAAS (Post Archean Australian Shale), which is considered to reflect the composition of the upper continental crust. Their trace element composition is very consistent and records deposition within a convergent tectonic setting, probably in an intra-arc basin located in a volcanic arc built on thinned continental margin (Fuenlabrada et al., 2010). Detrital zircon populations suggest a Middle Cambrian maximum depositional age (530-500 Ma) for the turbiditic series.

The large igneous massifs intruded at c. 500 Ma (Abati et al., 1999). The Monte Castelo gabbro (~150 km²), is formed by three major compositional types, olivine gabbro-norites, amphibole gabbro-norites and biotite gabbro-norites. According to their general geochemical pattern, the gabbros show a close similarity to island-arc tholeiites. The most abundant lithology in the large felsic bodies, as the Corredoiras

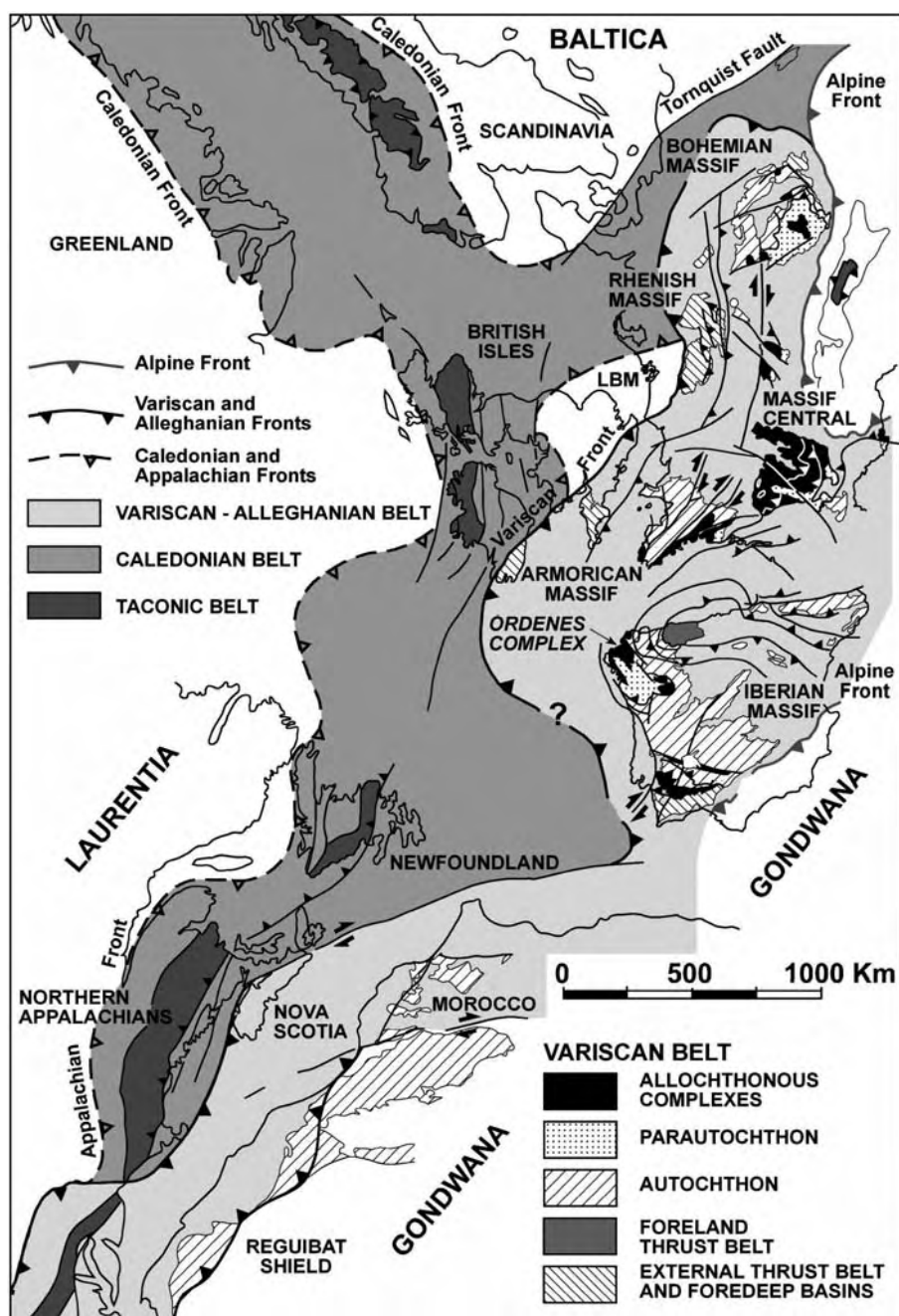


Figure 1. Sketch showing the distribution of the Paleozoic orogens in a reconstruction of the Baltica-Laurentia-Gondwana junction developed during the assembly of Pangea. The distribution of the most important domains described in the Variscan Belt is also shown, and also for reference the position of the Órdenes Complex in NW Iberia. LBM: London-Brabant Massif. From Martínez Catalán et al. (2002).

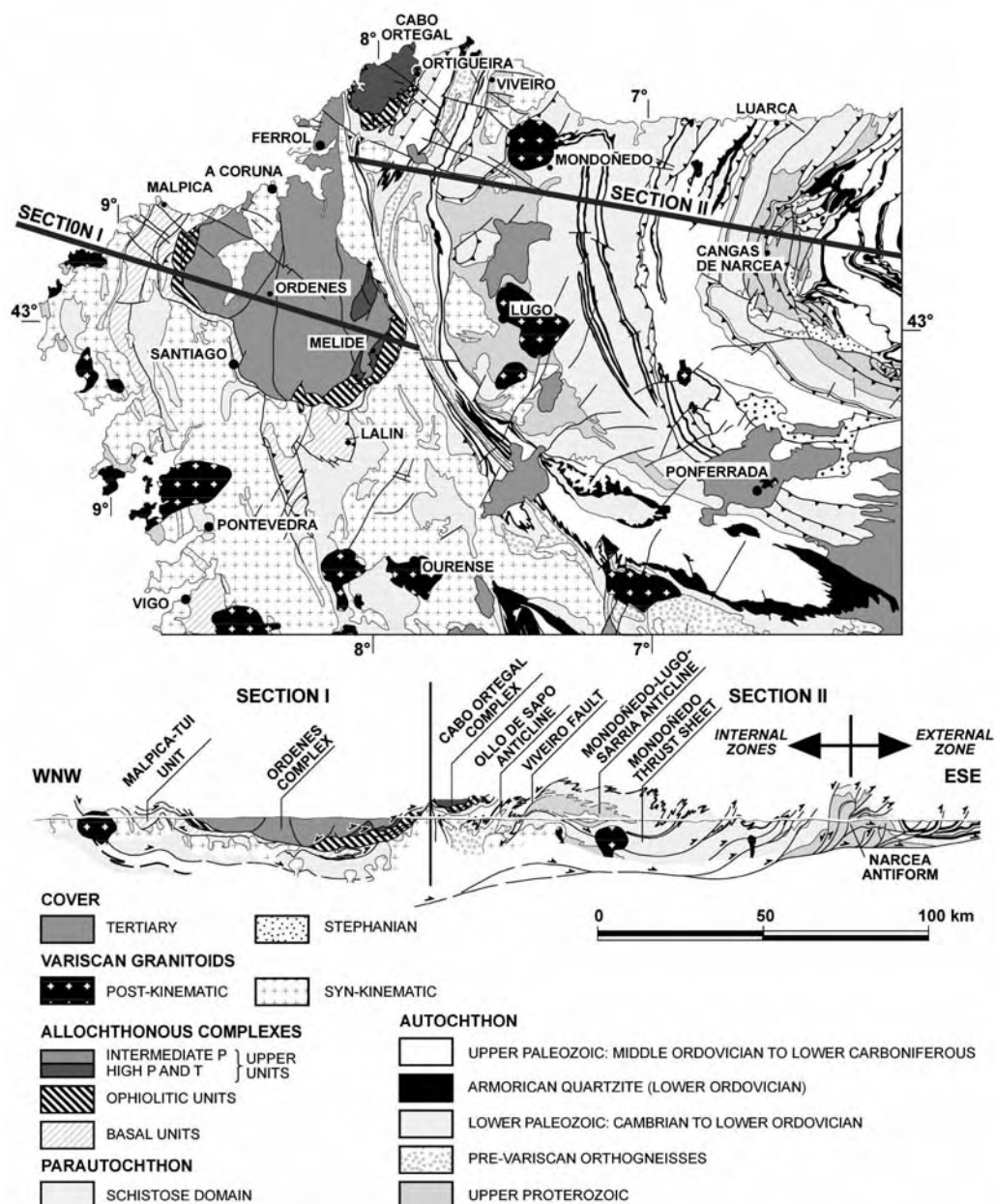


Figure 2. Geological map of NW Iberia. It shows the distribution of the Autochthon and Parautochthon domains and the main terranes involved in the allochthonous complexes located in the most internal part of the belt.

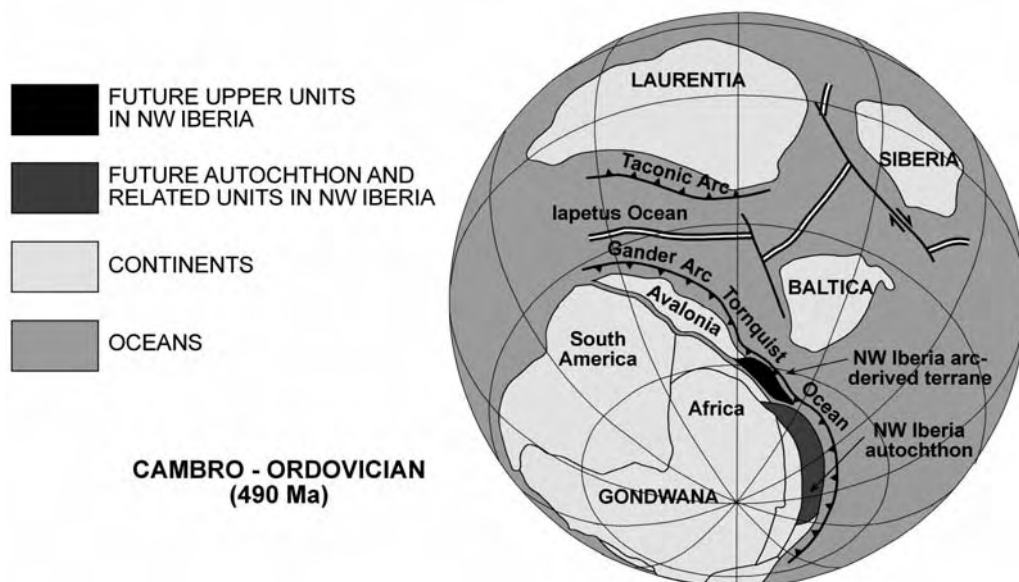


Figure 3. Paleogeographic reconstruction for the Cambrian-Ordovician limit showing the probable location of the peri-Gondwanan arc described in this contribution. The figure shows the moment immediately previous to the opening of the Rheic Ocean.

orthogneiss massif, is a hypidiomorphic granular coarse-grained granodiorite, with potassium feldspar and plagioclase phenocrysts. Minor bodies of tonalitic orthogneisses, amphibole-rich orthogneiss and gabbros also exist. The granodioritic orthogneisses are characterized by a highly fractionated trace element pattern, with a strong enrichment in Th and slightly enriched in Ce and Hf. They display significant negative anomalies in Ta, Nb and Zr, which together with their low contents in Y and Yb are characteristic of granitoids generated in volcanic arcs or subduction zones. The amphibole-rich orthogneisses also have negative anomalies in Ta and Nb and the same low content in Yb. Comparing our samples with the andesite-dacite-riolite association average from island and continental arcs from Drummond et al. (1996), the granodioritic and tonalitic orthogneiss patterns are similar to the continental arc association. The metagabbros pattern resembles continental arc basalts with high K (Pearce, 1996). All of the metagabbros have a negative anomaly in Nb, which is typical of igneous rocks generated in a subduction zone (Andonaegui et al., in prep.). Moreover, this huge magmatism allows to explain the anti-clockwise P-T paths described in the high-T lower sectors. These P-T paths are characterized by a first event with high-T and very low-P followed by a drastic compression, which can be only explain by a huge magmatic underplating taking place in the context of an active arc (Abati et al., 2003).

The regional tectonic fabrics dated in the uppermost terrane only recorded Early Paleozoic ages. In the lowest sectors with high grade metamorphism that can reach the intermediate-P granulite facies, different U-Pb data in monazite and zircon yielded ages in the range 496-482 Ma (Abati et al., 1999, 2007). Moreover, in the low grade top turbiditic series many diabasic dykes dated at c. 510 Ma intersect the regional schistosity (S_1+S_2) (Díaz García et al., 2010). These data, and the characteristics of the sedimentary series and igneous bodies, are only compatible with the dynamics of a peri-Gondwanan arc. This magmatic arc was raised during the activity of a subduction zone directed towards Gondwana

removing the pericontinental oceanic lithosphere. Some remnants of this ocean can be recognized accreted below the upper units, where they define one of the ophiolitic units described in the NW of the Iberian Massif (Bazar Ophiolite). New U-Pb data in this ophiolite yielded an age of c. 475 Ma for the main tectonothermal event with high-T and low-P characteristics (Sánchez Martínez et al., submitted). Moreover, the important contrast in the P-T path of the ophiolite in relation to the lowest sectors of the arc-derived terrane, clearly indicate that the accretion of the oceanic lithosphere occurred below a dissected arc, affected by important extension coeval to the accretionary activity or following its main development.

Finally, in relation to the location of the volcanic arc in the periphery of Gondwana, some scarce detrital zircon data seem to suggest that this arc was active in the periphery of the West Africa Craton (Fernández-Suárez et al., 2003). Additional information can be obtained using the Nd isotope data from the top metagreywackes which suggest mixed Ediacaran and Paleoproterozoic sources for the provenance of the greywackes, with T_{DM} ranging between 720 and 1215 Ma and an average of 995 Ma ($n=20$) - an age range unrepresented in the detrital zircon population. The Nd model ages are similar to those exhibited by West Avalonia, Florida or the Caroline terrane, but younger than those of Cambrian and Ordovician sandstones and shales from the autochthonous realm. These data suggest a westernmost location along the Gondwanan margin for the volcanic-arc represented in the upper terrane of NW Iberia (Fig. 3) relative to other terranes located in the footwall of the Variscan suture.

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GEOCHEMICAL FEATURES OF THE ORDOVICIAN SUCCESSION IN THE CENTRAL IBERIAN ZONE (SPAIN)

P. Barba¹, J.M. Ugidos¹, E. González-Clavijo² and M.I. Valladares¹

¹ Departamento de Geología, Facultad de Ciencias, 37008 Salamanca, Spain. barba@usal.es; jugidos@usal.es, valla@usal.es

² Instituto Geológico y Minero de España, Azafranal 48, 37001 Salamanca, Spain. egc@usal.es

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INTRODUCTION

It is generally accepted that fine-grained rocks are those that best reflect the chemical features of source areas (e.g., Condie, 1991; Cullers, 1995). However, the abundances of certain key elements or the values of certain element ratios may be changed in different sedimentary beds relative to the original ones as a consequence of possible mineral fractionation during sedimentary dynamics (McLennan and Taylor, 1991; Crichton and Condie, 1993, and references therein). Thus, geochemical data must be used carefully in studies dealing with tectonic settings or the provenance of detrital rocks before proposing a specific model for sedimentary environments or source areas of detrital material. In the present work major and trace elements are presented for 67 shales from different Ordovician synforms (Truchas, Alcañices, Tamames, Peña de Francia and Cañaveral) in the Central Iberian Zone (Fig. 1) in order to characterise the Lower, Middle, basal Upper and Upper Ordovician Series (samples 16, 23, 9 and 19, respectively) and to contribute to the assessment of the nature of this sedimentary record. In a previous work, restricted to the Truchas synform (Ugidos et al., 2004), the geochemical data revealed some differences among the different Ordovician formations. The aim of the present work is two-fold: to define the geochemical characteristics of the Ordovician succession in the Central Iberian Zone and to present an example of geochemical changes in the corresponding fine-grained rocks, probably related to mineral fractionation rather than to differences in the source.

GEOLOGICAL SETTING

According to the synthesis by Gutiérrez-Marco et al. (2002), Ordovician deposits overlie directly and unconformably Neoproterozoic and/or Lower Cambrian rocks in the Central Iberian Zone. Their sedimentation occurred on a siliciclastic marine shelf, where extensional tectonic processes controlled sedimentation and the presence of volcanic rocks. These deposits are generally of Arenig age, although in

some localities, such as in the Ollo de Sapo Antiform they are of Tremadoc age. During the Lower Ordovician, the sedimentation occurred on a storm-dominated shelf upwards evolving into shoreface sand bodies, and in some synforms sedimentation began as fluvial deposits. In the Middle Ordovician pelitic sedimentation dominates on an external shelf, in general under the action of storms, although the depth of the platform decreases from south to north in the Central Iberian Zone.

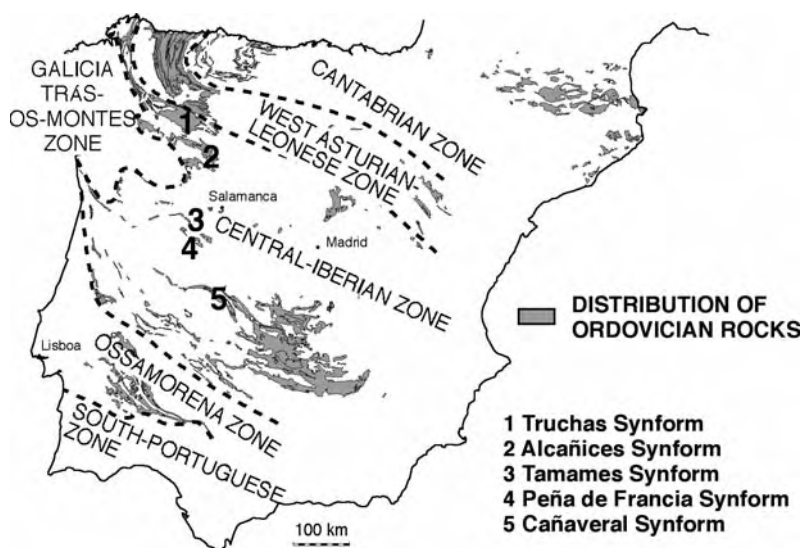


Figure 1. Simplified geological map of the distribution of Ordovician rocks in the Iberian Massif.

Upper Ordovician sediments in the Truchas synform were mainly deposited within shelf deep basins of semigraben type caused by listric normal faults. These basins filled with siliciclastic sediments, the lowermost ones being dominated by sandstones while top of the Upper Ordovician is dominated by shales with some intervals of diamictites, interpreted as corresponding to glaciomarine and/or mass flow deposits. In the roll-over antiform or in the upthrown fault block, the siliciclastic sedimentation was replaced by deposits of bioclastic shoals. In the synform of Cañaveral, the sandstones were deposited on a storm-dominated shelf. Extensional tectonic activity occurred to the south at the top of the Upper Ordovician, where the grain size is also fine, but with thick intervals of mass flow deposits, indicating the presence of sedimentary slopes of tectonic origin.

GEOCHEMICAL RESULTS

The results reveal similar geochemical features for the Lower Ordovician and basal Upper Ordovician shales (both showing the highest $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratios and Zr contents), than for the Middle and Upper Ordovician shales (those showing the lowest $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratios and Zr contents). Parameters such as the Zr/Y and Cr/Zr (Fig. 2), and Ti/Zr, Zr/Nb (see below) ratios clearly separate shales of each Series from the overlying or underlying one. However, the Ti, Nb, and Y contents of the shales from the different Series mostly overlap (Table 1). Moreover, there are rough positive covariations for SiO_2 -Zr and Al_2O_3 -Cr (Fig. 3).

Thus, trace element ratios involving Zr and Cr simply enhance the chemical differences. These features are common to all synforms sampled and strongly suggest that discriminating elements are dependent on the relative abundances of zircon and clays in the rocks studied as the most important minerals controlling the contents of Zr, Al_2O_3 and Cr (adsorbed on clays). In the Zr/Nb-Cr/Zr and Ti/Zr-Zr/Nb diagrams the 67 samples define hyperbolas (Fig. 4) where samples from the Upper Ordovician Series plot at an intermediate position between those from the other Series. This is an uncommon feature.

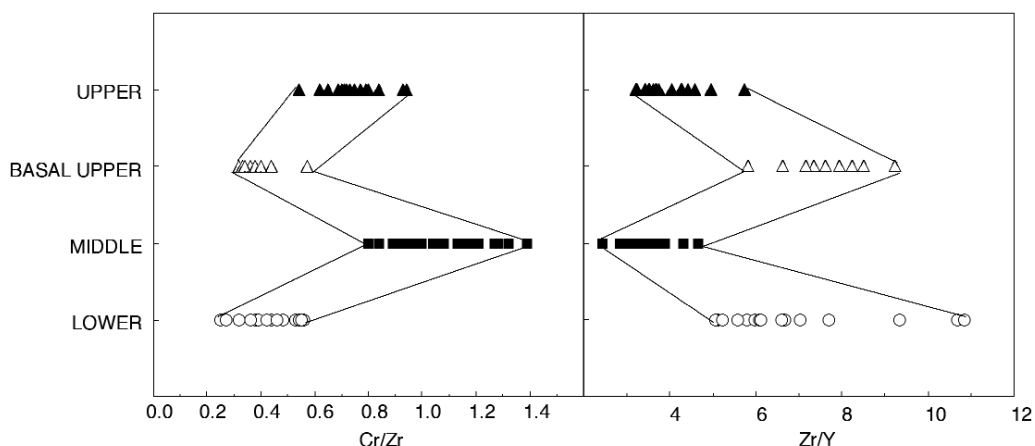


Figure 2. Examples of chemical parameters discriminating the Ordovician Series.

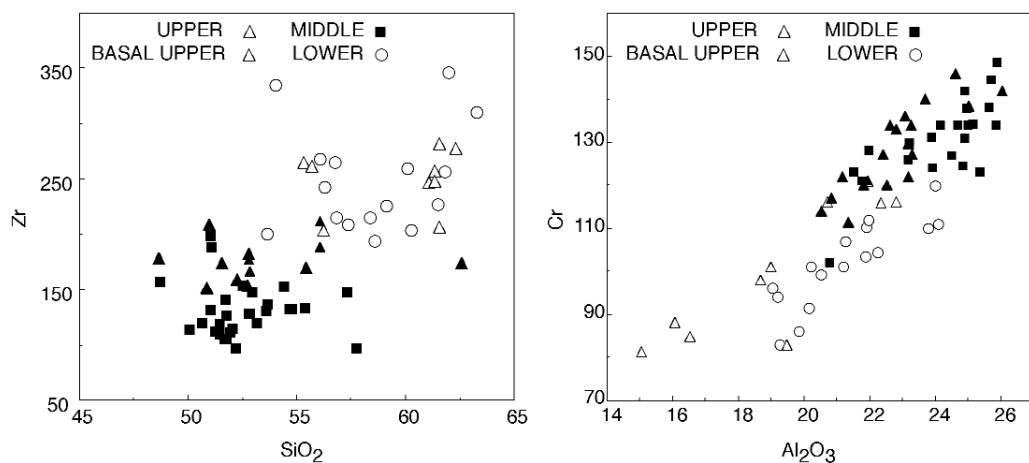


Figure 3. Diagrams showing rough SiO_2 -Zr and Al_2O_3 -Cr covariations.

	Lower Ordovician		Middle Ordovician		Basal Upper Ordovician		Upper Ordovician	
	Mean (n=16)	st. dev.	Mean (n=23)	st. dev.	Mean (n=9)	st. dev.	Mean (n = 19)	st. dev.
SiO ₂	58,51	2,86	52,82	2,04	59,61	2,91	52,48	2,09
TiO ₂	1,08	0,12	1,04	0,07	0,99	0,16	1,10	0,06
Al ₂ O ₃	21,30	1,68	24,17	1,50	18,96	2,73	22,82	1,42
Fe ₂ O ₃	7,20	1,20	9,03	1,45	7,43	1,17	9,29	0,99
MgO	1,62	0,39	2,13	0,36	2,02	0,30	2,70	0,42
MnO	0,04	0,01	0,06	0,03	0,05	0,03	0,07	0,03
CaO	0,20	0,16	0,22	0,21	0,88	1,14	0,41	0,26
Na ₂ O	0,50	0,43	0,96	0,27	1,33	0,28	0,87	0,43
K ₂ O	4,69	0,93	3,50	0,76	3,73	0,71	4,29	0,57
P ₂ O ₅	0,17	0,09	0,17	0,04	0,17	0,02	0,25	0,06
LOI	4,48	0,44	5,71	0,82	4,79	1,11	5,62	1,00
Total	99,78	0,56	99,81	0,35	99,95	0,55	99,86	0,45
Rb	182,00	26,30	165,00	24,00	141,00	24,60	174,00	21,80
Cs	8,23	1,54	7,86	1,87	5,69	0,95	7,36	1,22
Be	3,45	0,80	3,76	0,49	3,35	0,44	4,06	0,66
Sr	116,00	62,70	159,00	26,20	121,00	27,40	140,00	27,80
Ba	964,00	446,00	666,00	155,00	781,00	188,00	894,00	230,29
La	52,50	11,80	58,40	6,95	51,20	6,88	66,939	6,71
Ce	104,00	20,60	116,00	13,20	103,00	13,40	135,00	14,30
Pr	12,40	2,47	13,50	1,54	11,90	1,77	15,871	1,76
Nd	46,30	9,31	50,50	5,60	44,10	6,31	60,02	6,86
Sm	8,95	1,82	9,83	1,05	8,41	1,46	11,762	1,37
Eu	1,84	0,41	2,10	0,18	1,81	0,30	2,52	0,30
Gd	7,29	1,48	8,05	0,78	6,91	0,96	9,80	1,19
Tb	1,13	0,20	1,21	0,12	1,04	0,14	1,44	0,19
Dy	6,62	0,98	6,96	0,67	6,13	0,91	8,27	1,00
Ho	1,30	0,17	1,32	0,13	1,20	0,15	1,60	0,16
Er	3,71	0,42	3,62	0,36	3,34	0,47	4,26	0,50
Tm	0,56	0,06	0,54	0,05	0,50	0,07	0,63	0,08
Yb	3,77	0,37	3,58	0,35	3,33	0,39	4,22	0,44
Lu	0,59	0,06	0,54	0,05	0,50	0,07	0,62	0,07
Eu/Eu*	0,69	0,05	0,72	0,03	0,73	0,04	0,72	0,03
(La/Yb) _n	9,36	1,82	11,06	1,19	10,38	0,67	10,75	0,93
Y	37,10	4,71	36,20	3,06	33,20	4,69	43,60	5,01
Zr	247,00	48,10	123,00	14,00	250,00	28,10	172,00	21,10
Hf	6,76	1,28	3,45	0,33	6,76	0,73	4,81	0,50
Th	18,30	2,07	20,20	2,30	17,00	1,42	21,80	2,30
U	5,98	3,12	3,06	0,26	3,20	0,33	4,19	1,89
V	126,00	21,60	146,00	13,60	124,00	15,90	162,00	11,70
Nb	18,50	2,68	17,50	1,10	17,30	2,53	20,10	1,32
Cr	102,00	10,00	131,00	9,68	98,20	14,80	128,00	9,92
Co	16,60	4,11	17,70	6,69	23,30	8,13	23,90	6,43
Ni	35,60	9,50	43,90	9,70	46,40	13,60	55,50	7,47

Table 1. Mean analyses and standard deviation of Ordovician shales in the Central-Iberian Zone. n: chondrite normalized. Eu/Eu*: $Eu_p/(Sm_p \cdot Gd_p)^{1/2}$. Normalizing chondrite values after Taylor and McLennan (1985).

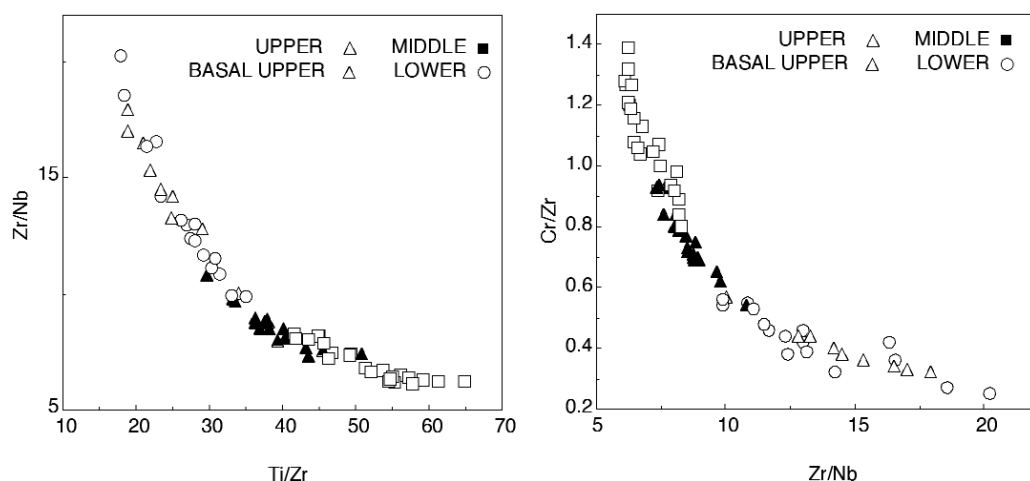


Figure 4. Geochemical parameters defining hyperbolas probably resulting from mineral fractionation (see text).

DISCUSSION AND CONCLUSIONS

The Lower Ordovician and lowermost Upper Ordovician Series are relatively enriched in SiO_2 and Zr while the other two are relatively depleted in these elements and in turn relatively enriched in Al_2O_3 and Cr. Given that these features seem to be associated with the different sedimentary settings, it seems logical to accept that sedimentary dynamics would have affected the mineral distribution, which in turn would have caused the chemical differences. It could be argued that mineral fractionation would also have affected other heavy minerals (e.g., Ti-minerals, monazite, xenotime), and consequently other major (e.g., TiO_2 , P_2O_5) or trace elements (e.g., rare earth elements, Y, Th). However, it must be taken into account that mineral fractionation in detrital rocks depends not only on the density of minerals but also on their size ranges, among other variables. Moreover, apart from Cr, other trace elements can be incorporated by clays and other minerals (e.g., rare earth elements, Honaty et al., 2008; Piasecky and Sverjensky, 2008; Galunin et al., 2010). In fact, in the present case zircon fractionation does not affect the $(La/Yb)_n$ ratio (all Series show similar ranges of values) even though it is a typical carrier of heavy rare earth elements, this suggesting that other minerals relatively rich in light rare earth elements would have fractionated together with zircon and compensated the Yb contribution of this mineral (the addition of minor quantities of monazite, for example, drastically increases the abundance of light rare earth elements in the corresponding rocks: McLennan, 1989). It is concluded that:

- (1) None of the geochemical features commented are related to changes in the source region but rather to differences in depositional environments. Apparently, during the Lower Ordovician and basal Upper Ordovician sedimentary dynamics was the same but different from that predominating during the Middle Ordovician sedimentation. This resulted in two extreme geochemical features, as shown by the hyperbolas in Figure 4. The intermediate position of the Upper Ordovician shales in the hyperbolas suggests that these rocks were deposited under less extreme dynamic conditions, intermediate between the other two. If this is really so, the other two groups would have resulted

from fractionation of the Upper Ordovician shales, which consequently should be the best representatives of the source composition.

- (2) Some geochemical parameters could be used for chemostratigraphic correlations on Ordovician successions, at least in the Central Iberian Zone.
- (3) The results strongly suggest that it is necessary to have a good knowledge of the stratigraphic and geochemical features of detrital sedimentary piles before proposing models about their provenance or geological settings.

Acknowledgements

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FAUNAL SHIFTS AND CLIMATIC CHANGES IN THE UPPER ORDOVICIAN OF SOUTH AMERICA (W GONDWANA)

J.L. Benedetto, T.M. Sánchez, M.G. Carrera, K. Halpern and V. Bertero

Centro de Investigaciones en Ciencias de la Tierra CICTERRA-CONICET, Facultad de Ciencias Exactas, Físicas y Naturales,
UNC, Av. Vélez Sarsfield 299, X5000JJC Córdoba, Argentina.
jbenedetto@efn.uncor.edu

Keywords: Precordillera, palaeoclimatology, Upper Ordovician, brachiopods, bivalves, gastropods.

INTRODUCTION

According to the microcontinental model the Precordillera terrane rifted off Laurentia in the late Early Cambrian and then drifted through a relatively narrow Southern Iapetus Ocean to finally collide with the Andean margin of Gondwana during the Late Ordovician (Benedetto, 2004 and references therein). Such a trajectory from low to intermediate/high southern latitudes has been well documented by a progressive decrease of Laurentian faunal affinities and a correlative input of Gondwanan taxa. By the end of the Ordovician the Precordillera basin was inhabited by the typical *Hirnantia* Fauna which occurs immediately above the glaciogenic deposits (Benedetto, 1986). Such paleogeographic changes took place under highly variable climatic conditions at global scale documented by lithologic, biological and stable isotopic data (mainly $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$). Most evidence used hitherto to establish global paleoclimatic models for the Late Ordovician comes from the continuous, essentially carbonate successions of Laurentia, Baltica and China, as well as from high-latitude basins of North Africa and perigondwanan terranes such as Iberia, Sardinia, Armorica and Perunica (e.g. Boucot et al., 2003; Fortey and Cocks, 2005; Ainsaar et al., 2010).

In this paper we analyze paleoclimatic evidence from the well-known Precordillera terrane compared to the autochthonous Central Andean basin of NW Argentina and Bolivia. Since carbon isotope data from the Precordillera are still limited (Marshall et al., 1997), we use lithofacial, stratigraphic and paleontologic evidence in order to infer (1) a relatively warm paleoclimate during the Late Sandbian, (2) a probable Katian (Ka2-Ka3) warming interval –partially equivalent to the Boda Event–, and (3) a cool-water postglacial transgression recording the first stages of development of the *Hirnantia* Fauna.

LATE SANDBIAN GREENHOUSE CONDITIONS

After the diachronic drowning of the carbonate platform during the upper Dapingian/lower Darriwilian, thick successions of graptolitic black shales were deposited in the deeper parts of the basin (Los Azules-

Gulcamayo-Las Plantas formations), although carbonate remnants locally persisted on structural highs. The upper part of the clastic successions is often punctuated by a few meters of calcareous silty shales, calcareous concretions and marls bearing graptolites of the *bicornis* Zone (Ottone et al., 1999). Of special interest is the debris flow succession named La Pola Formation exposed along the easternmost range (Sierra de Villicum) of the Precordillera. It consists of pebbly mudstones and bioclastic sandstones deposited in a proximal deep-marine through (Astini, 2001). Graptolites from the matrix of a debris flow bed about 10 m below the contact with the Hirnantian glaciogenic diamictite (Don Braulio Formation) indicate a late Sandbian age (Sa2 substage of Bergström et al., 2009) (Benedetto, 2003). Field data suggest that boulders come from a contiguous high-energy platform which supplied the fragments of ramoso bryozoan, thalli of solenoporacean red algae, and a few *Girvanella* remains present throughout the upper part of the formation (Carrera, 1997; Astini, 2001). Boulders from the debris flows as well as *in situ* bioclastic quartz-rich sandstone beds contain para-autochthonous assemblages of brachiopods, gastropods and bivalves (Sánchez, 1999; Benedetto, 2003; Bertero, in press). Similar fossiliferous boulders also occur sporadically within the overlying glaciogenic diamictite. Astini (2001) inferred that the bryozoan and solenoporacean-dominated pseudoreefal communities flourished in shallow subtidal settings. Since no biohermal structures are preserved *in situ* it is uncertain whether or not these organisms formed reef-like buildups. The absence of stromatoporoids along with the scarcity of corals and calcified green algae suggest that such bryozoan-rich beds developed in temperate rather than tropical waters.

Significant for paleoclimatic considerations is the influx of low-latitude brachiopods into the Precordillera terrane, which by the Late Ordovician was very close to the Gondwana margin by progressive closure of the interposed remnant seaway (Benedetto et al., 2009). According to the current paleogeographic models (e.g. Cocks and Torsvik 2002) the Precordillera basin was located by the late Sandbian at about 45°S. The La Pola Formation and coeval strata bearing calcareous nodules in the Las Plantas Formation have yielded a suite of genera (*Oanduporella*, *Dinorthis*, *Campylorthis*, *Hesperorthis*, *Ateleasma*, *Camerella*, *Glyptomena*) recorded elsewhere from Laurentia (Appalachians, Scotland) and/or from Australia-Tasmania, Baltica, China and Kazakhstani terranes marginal to Siberia (Benedetto, 2003). Also interesting is the presence in the Precordillera of *Anchoramena*, an endemic genus closely related to *Sowerbytes*. Both genera belong to a clade within the palaeostrophomenins that evolved in low-latitude areas (Candela, 2010). Gastropods include, among others, *Tetranota bidorsata* (Hall), *Sinuities* aff. *reticulatus* Perner, *Cyclonema* aff. *bilis* Hall, and *Clathrospira subconica* (Hall) all of them recorded elsewhere from Laurentia. Bivalves are unusually diverse and endemic in relation to other Gondwanan assemblages leading Sánchez (1999) to suggest that this radiation event was promoted by a warming of ocean water. Warm-water taxa coexisted in the Precordillera basin with a few brachiopods distinctive of the Mediterranean Province (*Tissintia*, *Aegiromena*, *Drabovia*). In contrast, the Central Andean basin of NW Argentina and Bolivia was inhabited almost exclusively by cold-water Mediterranean brachiopods and bivalves, including *Drabovia*, *Eorhipidomella*, *Aegiromena*, *Heterorthis alternata* (Sowerby), *Rafinesquina* aff. *pseudoloricata* (Barrande), *Drabovinella* cf. *erratica* (Davidson), *Cadomia typa* de Tromelin, and *Cardiolaria* (Benedetto et al., 2009), suggesting that water was too cold for tropical or subtropical organisms. There, the sole and non conclusive evidence of climate amelioration is the occurrence of thin calcareous horizons bearing conodonts of Sandbian age (Albanesi and Ortega, 2002), and sporadic calcareous boulders within the glaciogenic diamictite bearing the widespread brachiopod *Dinorthis*. It should be noted that during the Late Ordovician the vast Central Andean basin was located at higher latitudes (c. 52-55°S) than the Precordillera basin, which could explain such differences in water temperature and consequently in the provincialism.

In conclusion, evidence from different benthic groups suggests that by the late Sandbian waters in the Precordillera basin were warmer than it can be expected relative to its paleolatitude supporting greenhouse global conditions.

A MID KATIAN WARMING EVENT?

The youngest Ordovician carbonate succession of the Precordillera basin is the Sassito Formation, a c. 25 m thick unit composed of calcareous-rich shales, thin-bedded calcarenites and bioclastic grainstones with hummocky cross stratification and wave ripples, suggesting a shallow-ramp, high-energy environment (Astini and Cañas 1995; Keller and Lehnert 1998). The upper grainstones have yielded conodonts ranging from the *A. tvaerensis* to *A. ordovicicus* zones (encompassing essentially the *A. superbus* Biozone) which indicate the Ka2-Ka3 substages (Albanesi and Ortega, 2002). Megafossils (brachiopods, crinoids) are poorly preserved excepting bryozoans which are very abundant in the bioclastic grainstone facies. According to Ernst and Carrera (2008) the absence of typical tropical carbonate components (oncolites, oolites) and framework-building photozoans (e.g. green calcified algae) is suggestive of relatively cool waters. However, the presence of the ramose cryptostomids *Moyerella* and *Phylloporina* in the Sassito Formation is interesting as they have been previously recorded in paleocontinents placed consistently within the tropical belt in the Late Ordovician, such as Baltica, Siberia and Laurentia. Available evidence indicates that the Sassito Formation is comparable to other bryomol-type carbonate units of Katian age widely developed in North Gondwana from Morocco to the Indian Himalaya (Pin Formation), as well as in several peri-Gondwanan terranes (e.g. Iberia, Armorica, Sardinia, Carnic Alps) (Jiménez-Sánchez and Villas, 2010).

Development of carbonates for several hundred kilometers along the N Gondwana margin at mid to high latitudes was interpreted as evidence of a global warming named the Boda Event by Fortey and Cocks (2005), although Cherns and Wheeley (2007) postulated that such pre-Hirnantian bryozoan carbonates and mud mounds formed in response to an episode of glacioeustatic lowstand during a cooling event. Recent sequential analysis combined with facies interpretation in the Anti-Atlas of southern Morocco revealed a series of high-frequency stratigraphic sequences regarded as reflecting glacioeustatic oscillations (Loi et al., 2010). The Katian is also characterized by an alternation of positive and negative shifts in the $\delta^{13}\text{C}$ curve, which coincides with the onset of climate cooling that led to development of a vast ice sheet over Gondwana at the end of the Ordovician (Ainsaar et al., 2010). If correct, significant glacial episodes took place prior to the glaciation climax during the Hirnantian. According to the eustatic sea-level change curve calculated from the Moroccan sections three large-amplitude sea-level drops occurred during the Katian. Of them, the mid Katian sea-level rise may be tentatively correlated with deposition of the Precordilleran Sassito Formation, which unconformably overlies the Mid Ordovician San Juan Formation. Its basal erosive surface is thought to represent either the sequence boundary and the transgressive surface, whereas the lower calcipelites represent the transgressive systems tract which culminates with high stand regressive deposits (Astini and Cañas, 1995). Deposition of temperate-type bryozoan-rich carbonates on the opposing side of the South Pole relative to the N Gondwanan (Mediterranean) margin during the mid Katian was probably promoted by a global climate amelioration leading to the almost complete melting of the ice cap (Fig. 1) (Loi et al., 2010).

A very different succession of conglomerates, debris flows, turbidites, amalgamated sandstones and shales, named Trapiche Formation, crops out in the northern Precordillera. According to the conodonts of

the *A. superbus* Zone (Albanesi and Ortega, 2002) recovered from its lower part this unit is partially equivalent to the Sassito Formation. Shelly fauna includes the brachiopods *Reuschella* sp., *Rhynchotrema* sp. and *Destombesium argentinum* Benedetto, and a new ambonychiid bivalve. It should be noted that during the Mid-Upper Ordovician ambonychiids diversified mostly on low latitude carbonate platforms, being absent from definitely cold-water settings. These thick open-shelf deposits may also be correlated to the mid Katian post-glacial transgression, and its basal unconformity on the *gracilis/bicornis*-bearing shales (Gualcamayo Formation) may be tentatively correlated with the preceding sea-level drop. Such an abrupt sea-level fall has also been reported from Baltica and Laurentia (Calner et al., 2010).

THE HIRNANTIAN TRANSGRESSION

Sedimentological evidence indicates that the lower part of the Don Braulio Formation was deposited from wet-base grounded glaciers transitional to shallow marine settings peripheral to the main Gondwana ice sheet (Astini, 1999). Diamictites culminate with glacio-lacustrine and channelized glacio-fluvial deposits which consist of matrix-supported conglomerates and amalgamated coarse-grained sandstone beds, the latter probably representing proglacial sheetflows. Reworked marine fossils first appear within a thin but widespread quartz-conglomerate/sandstone bed at the top of the diamictite. This bed, rich in bioclasts and brachiopods, bivalves, trilobites and bryozoans marks the onset of the glacio-eustatically driven marine transgression during the Hi2 substage. The assemblage is dominated by large specimens of *Hirnantia sagittifera* (M'Coy) associated with *Dalmanella testudinaria* (Dalman), *Cliftonia oxoplectoides* Wright, numerous stick-like bryozoans of the genus *Helopora* (Carrera and Halpern, this issue), the gastropod *Holopea*, and two genera of ambonychiid bivalves. *Holopea* has been reported from the Hirnantian of Baltica, China, Laurentia and New Zealand. As stated above, in the Late Ordovician ambonychiids were almost completely confined to tropical-subtropical regions, which suggests that postglacial conditions were more temperate than had previously been supposed. A few meters above, an interval of near-shore calcareous siltstones contain the low diversity *Hirnantia-Modiolopsis* community described by Sánchez et al. (1991).

Acknowledgements

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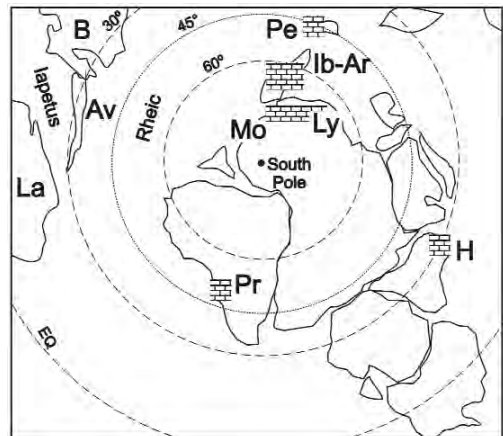


Figure 1. Late Ordovician palaeogeography based on Cocks and Torsvik (2002) and Fatka and Mergl (2009) showing mid-to high-latitude Katian carbonates. Av: Avalonia; B: Baltica; H: Indian Himalayas; Ib-Ar: Ibero-Armorica terrane; La: Laurentia; Ly: Lybia; Mo: Morocco; Pe: Perunica terrane; Pr: Precordillera terrane.

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A SUMMARY OF THE ORDOVICIAN OF THE OSLO REGION, NORWAY – FUTURE CHALLENGES

D.L. Bruton

The Natural History Museum (Geology), University of Oslo, Postboks 1172 Blindern, N-0318 Oslo, Norway.
d.l.barton@nhm.uio.no

Keywords: Ordovician, Oslo Region, Caledonide Orogen.

The Ordovician rocks of the Oslo Region crop out in a graben of Permian age some 40-70 km in width and with a total length of 115 km. The maximum thickness of strata is 1 km and the succession is autochthonous and parautochthonous in the south but becomes allochthonous northwards with movement from northwest towards the southeast on a sole thrust above the Cambrian alum shale (Bruton et al., 2010). In the south, the base of the Ordovician is biostratigraphically well defined in a continuous succession of black shales with carbonate nodules (Bruton et al., 1988). Bed by bed correlation of equivalent units of carbonates and mudstones in Sweden is possible up to the middle Darriwilian but above this level, correlation becomes less precise. This is because of changes in sedimentary rates and marked lateral facies changes caused by local faulting and the fact that the Oslo Region occupied an intermediate position between the stable platform to the east and the developing Caledonide orogen to the west (Jaanusson, 1973, p. 29-30). Thus, in the early Ordovician, thin units, some only a few metres thick can be traced from east to west over great distances and occur in some of the early thrust nappes with origins outboard of the present Norwegian west coast (Bruton and Harper, 1988). A marked break in sedimentation occurs at the top of the Ordovician and a relict fauna of deep-water brachiopods occurs in the overlying Lower Silurian (Baarli and Harper, 1986). A rapid fall in sea-level related to the end Ordovician glaciation has been used to explain this (Brenchley and Newall, 1980) and a rapid erosion of reef carbonates prior to this produced block-filled submarine gullies and channels (Braithwaite et al., 1995). The Katian carbonates in the Mjøsa area are associated with a warm water conodont fauna of North American midcontinent type (Bergström et al., 1998). These carbonates and equivalent lime-mud units in the south contain evidence for the global Guttenberg Carbon Isotope Excursion (GICE) previously recognised in Sweden, Estonia, North America, Thailand, and China (Bergström et al., 2010). One of the several rapidly deposited late Sandbian K-bentonites occupies the same stratigraphic position as the thick Millbrig K-bentonite in eastern North America (Huff et al., 1992, 2010).

Absolute Age (Ma)	System	Global Series	Global Stages	British Series	Series	Baltic		Lithostratigraphy of the central Oslo Region (stage for reference only)
						Stages	Graptolites	
443.7	ORDOVICIAN	UPPER	HIRNANTIAN	ASHGILL	HARJU	Porkuni	<i>persculp. extraordi.</i>	Langoyene Fm. (5b)
445.6						Pirgu	<i>(anceps)</i>	Husbergøya Fm. (5a)
			<i>complanatus</i>				Skogerholmen Fm. (4d)	
			Vormsi				<i>linearis</i>	Skjerholmen Fm. (4cγ) Grimsoya Fm. (4cβ)
			Nabala				Venstøp Fm. (4cα)	
			CARADOC	Rakvere		<i>clingani</i>	 Solvang Fm. (4bδ1-2)	
				Oandu			Nakkholmen Fm. (4bγ)	
				Keila			Frognerkilen Fm. (4bβ)	
				Jõhvi Idavere		<i>foliaceus</i>	Arnestad Fm. (4bα)	
			VIRU	Kukruse	<i>gracilis</i>	Vollen Fm. (4aβ)		
				Uhaku	<i>teretiusculus</i>	Elnes Fm. (4aα)		
460.9		MIDDLE	DARRIWILIAN	LLANVIRN	Lasnamagi		<i>distichus</i>	
					Aseri		<i>elegans</i>	
					Kunda	Aluoja	<i>fasciculatus</i>	
						Valaste	<i>lentus</i>	Svartodden Mbr (3cγ)
468.1			Hunderum	Volkhov		<i>hirundo</i>	Lysaker Mbr (3cβ)	
							Huk Fm.	
			OELAND	DAPINGIAN	ARENIG	Billingen	<i>elongatus</i>	Hukodden Mbr (3cα)
471.8							<i>densus</i>	
		<i>balticus</i>						
		Hunneberg				<i>phylograp- toides</i>	Hagastrand Mbr (3bα)	
				<i>copiosus</i>				
				<i>murrayi</i>				
478.6		LOWER		TREMADOCIAN	TREMADOC	Varangu	<i>supremus</i>	Bjorkåsholmen Fm. (3aγ)
						Pakerort	<i>hunneberg. Rhabdinop.</i>	Alum Shale Fm. (2e-3aβ)
488.3								

Figure 1. The correlation of the Ordovician succession of the central Oslo Region with the standard British and Baltic sequences. Note that the relative durations of the chronostratigraphical units are not equivalent to their absolute durations but are scaled to fit the detail of the Oslo Region succession. Based on Owen et al. (1990), Nielsen (2004), Gradstein et al. (2004) and Dronov and Rozhnov (2007 pars). From Bruton et al. (2010).

CONCLUSIONS

Reference to Bruton et al. (2010), highlights the need to solve future details around the following topics:

1. Further revision of key Ordovician groups such as the brachiopods, trilobites and graptolites in the light of the basin's unique position which allows the inter-fingering of deep and shallow water facies.
2. The combination of faunas and facies and study of palaeoenvironments in the basin (Hansen J. et al., 2009; Hansen, T. et al., 2010).
3. The importance of bentonites for correlation and their influence on the palaeoenvironment.
4. Extended work on Carbon isotopes and their use in interregional correlation.
5. The relationship of tectonism to basin development.
6. Study of shifts in sedimentary environments and the timing of orogenic events.

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PRELIMINARY REPORT ON *ARTHRRORHACHIS* HAWLE AND CORDA, 1847 (AGNOSTIDA) IN THE PRAGUE BASIN (BARRANDIAN AREA, CZECH REPUBLIC)

P. Budil¹, O. Fatka², P. Kolář³ and M. David⁴

¹ Czech Geological Survey, Klárov 3, Praha 1, CZ-118 21, Czech Republic. petr.budil@geology.cz

² Department of Geology and Palaeontology, Faculty of Science, Charles University, Albertov 6, Praha 2, CZ-128 43, Czech Republic. fatka@natur.cuni.cz

³ Charles University Botanical Garden, Na Slupi 16, Praha 2, CZ-128 43, Czech Republic.

⁴ Rožmberská 10, Praha 9, CZ-198 00, Czech Republic. ordovik@seznam.cz

Keywords: Agnostida, Upper Ordovician, Prague Basin, Barrandian area, Czech Republic.

INTRODUCTION

The agnostids are characteristic elements of many Ordovician shelly faunas of peri-Gondwana. In the Upper Ordovician, they became rare and usually of low taxonomic diversity. The last, but locally still abundant representatives, including *Arthrorhachis* Hawle and Corda, 1847 are known from the Upper Ordovician of the European peri-Gondwana, ATA (Armorican Terrane Assemblage): Italy and Bohemia; Avalonia: South Wales, North Wales, Northern England; Baltica: Norway, Sweden, Bornholm, Denmark, Poland; Kazakhstan: Kazakhstan, Uzbekistan and Northern China.

This short contribution summarizes the results of thorough study, which was submitted to the Bulletin of Geosciences.

REVIEW OF RESULTS

We follow Fortey (1980) in restricting *Trinodus* M'Coy, 1846 to the holotype of its type species, *T. agnostiformis* M'Coy, 1846. This decision was based on the observation, that the type specimen of the type species shows none of the most critical features used for the identification of metagnostid genus (see also Fortey 1997). The concept of Fortey (1980) has been accepted by Fortey and Owens (1987), Ahlberg (1989), Romano and Owen (1993), Hammann and Leone (1997), Whittington et al. (1997), Nielsen (1997), Vaněk and Vokáč (1997), Shaw (2000) and Vaněk and Valíček (2001) but not by Pek and Prokop (1984), Pek and Vaněk (1989) and Bruton and Nakrem (2005). Owen and Parkes (2000) consequently published sparse new material of *T. agnostiformis*, a poorly preserved pygidium, coming from the same horizon as the lectotype specimen. They also tentatively affiliated another pygidium, the lectotype of *Agnostus limbatus* Salter, 1848 and *Agnostus trinodus* Salter, 1848 to *T. agnostiformis*. Subsequently, they proposed to consider the name *Arthrorhachis* Hawle and Corda, 1847 as a subjective junior synonym of

Trinodus Mc'Coy, 1846. Because of morphological differences between *T. agnostiformis* and *A. tarda*, they also suggested to retain the name *Arthrorhachis* as a subgenus of *Trinodus* to encompass the *A. tarda* species group recognized by Nielsen (1997). This opinion followed by Turvey (2005), Owens and Fortey (2009) and Owen and Romano (2010), while Jell and Adrain (2002) consider *Arthrorhachis* as a valid genus. After careful evaluation of all arguments, we have to consider all the known material of *Trinodus agnostiformis* (including the newly published specimens) as still insufficient and especially as poorly preserved. New, better preserved material is necessary for a closer comparison of both species and thus also for the final decision of the *Arthrorhachis*/*Trinodus* relation. Therefore, we consider the Fortey's (1980) proposal as still relevant and the most reasonable solution of the question in the present-day stage of knowledge. The rare Sandbian species *Trinodus agnostiformis* (see Owen and Romano, 2010) even in evaluation of all the most recently gathered material, is still very poorly documented to be compared with the very abundant and often well- to excellently preserved specimens of *Arthrorhachis tarda* of Katian age.

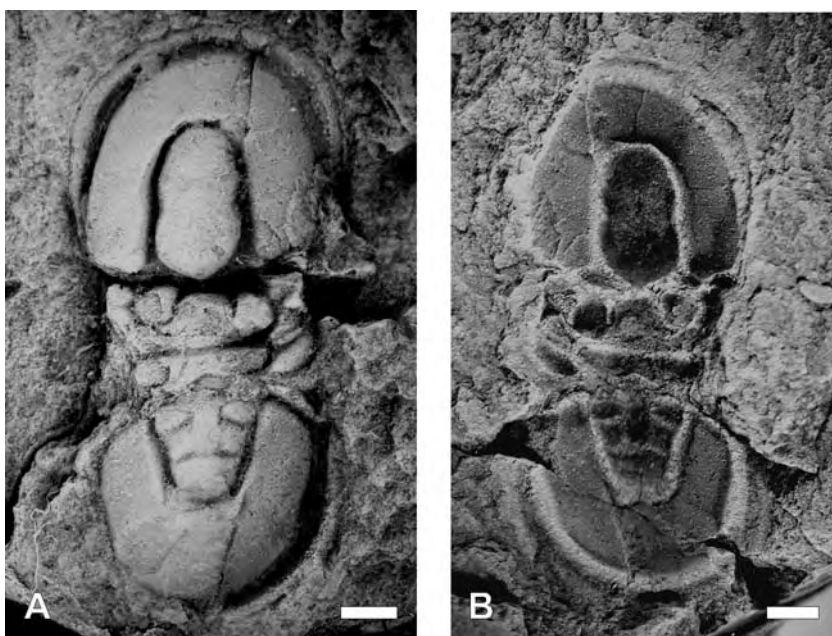


Figure 1. A-B: *Arthrorhachis tarda* (Barrande, 1846). Upper Katian, Králův Dvůr Formation. Libomyšl (very probably, Lejškov near Libomyšl). Neotype NM L 16534 (original No. ČD 1812), an internal mould (A) of a nearly complete specimen and its negative counterpart (B). The specimen was illustrated by Barrande (1852, pl. 49, figs. 1-2) and Pek (1977, pl. 8, fig. 2), and was incorrectly selected as a lectotype by Přibyl (in Horný and Bastl 1970, p. 308). The scale bar represents 1 mm.

So far, the limited material of *A. tarda* from the Prague Basin has been discussed by Pek (1977), Whittington (1950), Šnajdr (1983), and Fortey (1997). Its original lectotype specimen (NM L 16534, older No. ČD 1812), was incorrectly chosen by Přibyl in Horný and Bastl (1970) because this specimen does not belong to the authentic collection of Barrande (1846) but was collected slightly later and was published by Barrande (1852). Therefore, status of this specimen as neotype is preferred. The second questionable lectotype specimen was not originally selected by Šnajdr (1983) but already also by Přibyl in Horný and

Bastl (1970, p. 307) but as lectotype specimen of *Arthrorhachis tarda* (*sensu*) Hawle and Corda, 1847. A short diagnoses is given and the morphological variability is evaluated. All the mentioned results support the "wide concept" of the species as a highly variable form [see Ahlberg (1989), Hammann and Leone (1997) and Shaw (2000)]. Ahlberg (1989) also noted that the measured data may be highly affected by deformation. We agree with this opinion. It is very complicated to distinguish between the variability and diverse preservation in the material originating from claystone and siltstone.

Comparatively poorly known *Arthrorhachis pragensis* (Příbyl and Vaněk, 1968) was also subject of our study. Its diagnosis is supplemented and the relations to *A. tarda* discussed. The validity of *A. pragensis* is supported (see also Vaněk and Vokáč 1997), being partially supplemented also by biometrics (only limited material is available). *A. pragensis* differs from the *A. tarda* in having distinct and stout posterolateral spines with robust bases, shallower pygidial border furrow, narrower cephalic border, more (sag.) elongated cephalon of more rectangular outline, and possibly also by lower convexity of the pygidial pleural field. In *A. pragensis*, the cephalic border furrow in its anterior part is prominently wider than posterolaterally and the cephalic border is narrower. Basal glabellar lobes are tr. longest, and in contact with the glabellar axis.

Analyses of palaeogeographical distribution supports the almost cosmopolitical distribution of *A. tarda* (some of its occurrences are, however, questionable). On the other hand, *A. pragensis* could be considered as an endemic species, restricted to the Prague Basin only.

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GRAPTOLITE ZONATION FOR THE LOWER AND MIDDLE ORDOVICIAN OF THE GORNY ALTAI (SW SIBERIA, RUSSIA)

E.V. Bukolova

Trofimuk Institute of Petroleum Geology and Geophysics SB RAS, Acad. Koptug av. 3, 630090 Novosibirsk, Russia.
BukolovaEV@ipgg.nsc.ru

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INTRODUCTION

After ratification of new stages of the Ordovician System, chronostratigraphic position of boundaries of local and regional stratigraphic units are revised in all regions of Russia. GSSPs were officially selected for all boundaries of International Stratigraphic Chart, where their position was defined as the First Appearance Datum (FAD) of particular index-species of some graptolite and conodont zones. Chronostratigraphic position of substage boundaries for the new Ordovician standard scale, based on other graptolite and conodont zones, were also informally defined (*stage slices* of Bergström et al., 2008).

GRAPTOLITE ZONATION FOR THE ORDOVICIAN OF THE GORNY ALTAI

On the territory of Gorny Altai, the first Ordovician graptolites were found in the south-west by Z.E. Petrunina in 1956 (Petrunina, Severgina, 1960). Early Ordovician graptolite species were recorded in 1964 by M.V. Romanenko in south-eastern Gorny Altai (Romanenko, 1966). Among them *Tetragraptus bigsbyi* (Hall), *Expansograptus* aff. *suecicus* (Tullberg), *Expansograptus* sp., *Paratetragraptus approximatus* (Nicholson), *Paratetragraptus* aff. *acclinans* (Keble) and *Corymbograptus* sp.

Later, numerous Ordovician graptolites were collected by N.V. Sennikov and the first regional graptolite zonation for the Gorny Altai was proposed (Obut and Sennikov, 1986; Sennikov, 1996). A revised zonation was published by Sennikov et al. (2008). These subdivisions were defined as complex zones with lower boundaries marked by the first appearance of index-species. Zonation was made taking into account maximal potential for high-resolution correlation with regard to the standard British graptolite zones, series and stages: Tremadoc, Arenig, Llanvirn (including the Llandeilian), Caradoc and Ashgill. Thus, in Gorny Altai some of the British graptolite zones such as the *H. teretiusculus*, *N. gracilis*, *A. multidentis* etc. were recognized. A revision of British chronostratigraphy started in 1991 proved among others, that some

stages overlap each other, and that boundaries of British graptolite zones are not suitable to define the boundaries of newly proposed stages subdivision (Fortey et al., 1991, 2000).

The need to revise graptolite zonation for the Ordovician of the Gorny Altai arose after ratification of the new Ordovician stages by the International Commission on Stratigraphy (Ogg et al., 2008). New zones and new zonal index-species were defined as new studies on graptolite distribution in Ordovician sections of Gorny Altai. So the *D. protobifidus* Zone was based in the first identification of the index-species in the «Tuloi» section (north-eastern Gorny Altai). Beside it, the following taxa were identified in the assemblage: *Eotetragraptus harti* (T.S. Hall), *Expansograptus extensus* (Hall), *Expansograptus taimyrensis* Obut et Sobolevskaya, *Expansograptus suecicus suecicus* (Tullberg), *Acrograptus pusillus* (Tullberg), *Phyllograptus densus densus* Törnquist, *Phyllograptus ilicifolius glaber* Monsen, *Pseudophyllograptus angustifolius elongatus* (Bulman), *Pendeograptus* aff. *pendens* (Elles) and *Corymbograptus* sp.

The *I. gibberulus* Zone is subdivided into a lower *C. deflexus* Subzone and upper *I. maximo-divergens* Subzone, both of them established in the «Pridorozhnyi» section (north-eastern Gorny Altai). Besides the index-species, the *C. deflexus* assemblage comprises *Pseudisograptus manubriatus* (Hall), *Acrograptus nicholsoni* (Lapworth), *Isograptus paraboloides* Tzaj, *Isograptus* aff. *walcottorum* Ruedemann, *Corymbograptus* sp. and *Paradelograptus* sp. In the *I. maximo-divergens* Subzone along with index-species were identified *Pseudotrigranograptus ensiformis* (Hall), *Isograptus reduncus* Tzaj, *Isograptus primulus* Harris, *Isograptus* aff. *schrenki* Obut et Sobolevskaya, *Pseudisograptus manubriatus janus* Cooper and Ni and *Isograptus elegans* Tzaj.

The *E. hirundo* Zone was subdivided into a lower *I. caduceus imitatus* Subzone and an upper *U. sinodentatus* - + plus? *Cardiograptus* Subzone in «Maralicha» section (western Gorny Altai). The graptolite association of the *I. caduceus imitatus* Subzone, besides the index-species includes *Glossograptus* aff. *acanthus* Elles et Wood and *Pseudoclimacograptus* sp. In the *U. sinodentatus* - *Cardiograptus* Subzone, together with the index-species, we have recorded *Pseudotrigranograptus angustus* Mu et Lee, *Acrograptus cognatus* (Harris and Thomas), *Expansograptus extensus* (Hall), *Loganograptus logani* (Hall) and *Undulograptus sinodentatus* (Mu and Lee).

Chronostratigraphic position of the lower boundary for the Middle Ordovician Darriwilian Stage was defined internationally by the FAD of *U. austrodentatus* (Chen and Bergström, 1995). The *U. austrodentatus* Zone was defined in the «Maralikha» section of western Gorny Altai.

As a result of the revision of the graptolite zonation for the Ordovician of the Gorny Altai, the zonal species defining the base of most of the global series, stages and "stage slices" (= substages) have also been recognized. These are *Tetragraptus approximatus* for the lower boundary of the Floian stage, *Didymograptus protobifidus* for the base of the Floian 3 substage, *Undulograptus austrodentatus* for the basal boundary of the Darriwilian stage and Middle Ordovician series, *Nemagraptus gracilis* for the lower boundary of the Sandbian stage and Upper Ordovician series, *Climacograptus bicornis* for the base of the Sandbian 2 substage, *Diplacanthograptus caudatus* Zone for the lower boundary of the Katian, and *Pleurograptus linearis* for the lower boundary of the Katian 2 substage.

It should be noted that the lower part of the Gorny Altai zonal scale is more consistent with graptolite zonation of China (for the Tremadocian–lower Darriwilian interval). Among the same name zones are *C. tenellus*, *T. approximatus*, *C. deflexus*, *I. caduceus imitatus* and *U. austrodentatus*. The upper part of the zonal scale fits better with the graptolite zonation of Baltoscandia (upper part of Darriwilian–Katian interval). Among the same name zones are *H. teretiusculus*, *N. gracilis*, *D. clingani* and *P. linearis* (Fig. 1) (Webby et al., 2004).

System	Series	Stages of SS	Graptolite zonation of the Gorny Altai (Sennikov et al., 2008)	Graptolite zonation of China (Webby et al., 2004)	Graptolite zonation of Baltoscandia (Webby et al., 2004)
ORDOVICIAN	LATE	HIRNANTIAN	persculptus	persculptus	
			supernus	ornatus	
		KATIAN		extraordinarius / ojsuensis	
				pacificus	
				mirus typicus	
			supernus	complexus	complanatus
				complatatus	
		quadrimum-cronatus	linearis	quadrimumcronatus / johnstrupi	linearis
				pygmaeus	
				spiniferus	clingani
			clingani	clingani	
		SANDBIAN	multidens	wilsoni	foliaceus
			bicornis	bicornis	
			wilsoni	gracilis	gracilis
			peltifer / antiquus lineatus		
			gracilis / serratulus		
	MIDDLE	DARRIMILIAN	teretiusculus	murchisoni	teretiusculus
			jakovlevi / coelatus	jiangxiensis	elegans
				elegans	
			balchaschensis / kirgizicus	fasciculatus	distichus
			dentatus	ellessae	fasciculatus
			austro-dentatus	austro-dentatus	lentus
				sinicus	
				zhejiangensis	
		DAPINGIAN	hirundo	clavus	hirundo
			sinodentatus / Cardiograptus		
			caduceus imitatus	caduceus imitatus	
			gibberulus	suecicus	
			maximo-divergens	deflexus	
			deflexus		
	EARLY	FLOIAN	angustifolius elongatus / broggeri	fruticosus	elongatus
			proto-bifidus		
			densus		densus
			balticus		balticus
			approximatus	approximatus	phyllograptoides
			?	Adelogr. / Clonogr.	H. copiosus
			osloensis / hyperboreus		Ar. murrayi
			tenellus / kiaeri		K. supermus
	TREMADOCIAN		?	Psigraptus	Ad. hunnebergensis
				R. flabelliformis - dichotomus	R. socialis/flabelliformis / desmograptoides/parabola

Figure 1. Correlation of regional stratigraphic subdivisions for the Ordovician of the Gorny Altai, China and Baltoscandia.

Graptolites are common in the entire Ordovician succession of the Gorny Altai, but taxonomic diversity of the zonal associations varies greatly (Fig. 2). Such zones as *E. hirundo*, *U. austro-dentatus*, *U. dentatus* and in the interval of *N. gracilis* to *C. supernus* zones are characterized by more than 30 species. The taxonomic diversity is 12-19 species in *P. densus*, *P. angustifolius elongatus*, *E. broggeri* and *I. gibberulus* zones. In *E. jakovlevi-A. coelatus* and *N. persculptus* zones, 6 and 5 species are identified, respectively. Within *E. balchaschensis*, *E. kirgizicus* and *H. teretiusculus* zones no more than two species are identified.

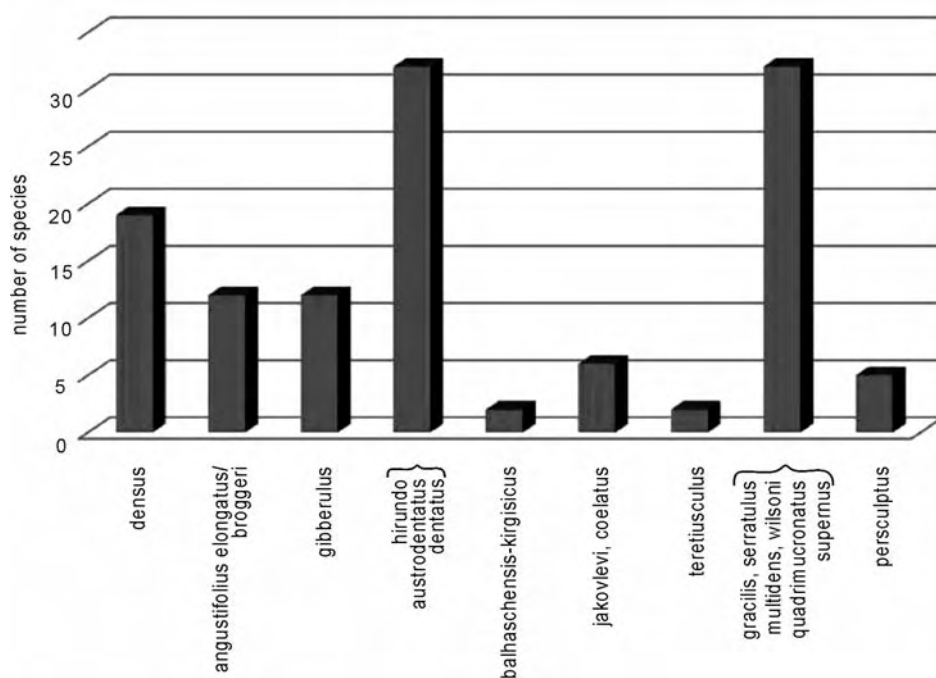


Figure 2. Number of species recorded in the Gorny Altai in the different biostratigraphic intervals according to graptolite zones.

CONCLUSIONS

Lower and Middle Ordovician deposits play an important role in the Paleozoic history of the Gorny Altai. Graptolite research has brought a substantial supplement to our knowledge of the Ordovician strata in this area. New graptolite zonation established for the Ordovician of the Gorny Altai allows direct correlation of regional zones with standard zones in the frame of the new global Ordovician chronostratigraphic scale.

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ORDOVICIAN MAGMATISM IN THE EXTERNAL FRENCH ALPS: WITNESS OF A PERI-GONDWANAN ACTIVE CONTINENTAL MARGIN

F. Bussy¹, V. Péronnet^{1,2}, A. Ulianov¹, J.L. Epard² and J. von Raumer³

¹ Institute of Mineralogy and Geochemistry, Anthropole, University of Lausanne, CH-1015 Lausanne, Switzerland.
francois.bussy@unil.ch, alexey.ulianov@unil.ch, jean-luc.epard@unil.ch

² Institute of Geology, Anthropole, University of Lausanne, CH-1015 Lausanne, Switzerland

³ Dept. of Geosciences, University of Fribourg, Switzerland. juergen.vonraumer@unifr.ch

Keywords: Ordovician magmatism, Western Alps, granite, zircon, geochronology.

INTRODUCTION

The pre-Mesozoic basement areas of the external Alpine domain (e.g. Aiguilles-Rouges Mont-Blanc, Aar Gotthard crystalline massifs) are underlain by former early Palaeozoic sedimentary and magmatic rock units, which underwent a high-grade metamorphic overprint during the Carboniferous Variscan orogenic events. On the other hand, they were fairly well preserved from the Tertiary Alpine metamorphism, which reached only lower greenschist facies conditions (von Raumer et al., 2009, with references) and moderate deformation. Pre-Variscan lithologies are particularly well documented in the Aiguilles Rouges massif, west of Chamonix (France). Here we present new age determinations on several magmatic bodies of this massif. Together with pre-existing geochronological and geochemical data, they document a major magmatic event of Ordovician age which can be related to an active margin geodynamic environment.

GEOLOGICAL FRAMEWORK

The Aiguilles Rouges massif (ARM) is one of the so-called external crystalline massifs of the Alpine belt and corresponds to a huge Alpine basement antiform structure of 20 by 45 km surrounded by Mesozoic sedimentary cover units (geological maps and lithologic descriptions in von Raumer and Bussy, 2004). The lithologies include various metasedimentary rocks like metagreywackes, banded paragneisses, quartzites, micaschists, as well as orthogneisses and metabasites like garnet-amphibolite and eclogite boudins. Pre-Mesozoic metamorphic assemblages record at least two distinct P-T events. In the Lake Cornu area, mafic eclogites preserve high-pressure garnet-omphacite metamorphic assemblages, recording P-T conditions of > 1.1 to 1.4 GPa and 700°C, respectively (Liégeois and Duchesne, 1981). The age of this high-P event is unknown, but predates the Variscan high-T event described below. In the Lake Emosson area, metagrawackes have partially melted yielding migmatites with up to 25 vol.-% leucosome. According to Genier et al. (2008), anatexis was triggered by water fluxing of metagreywackes in a transcurrent shear-

zone at 0.3-0.4 GPa and 640-670°C. Monazite from a leucosome vein yielded a U-Pb date of 321 Ma (Bussy et al., 2000), interpreted as the age of leucosome crystallization. More recently, Schulz and von Raumer (2011) obtained electron microprobe dates of ca. 440 Ma on monazite grains included in garnet in micaschists. These dates document a pre-Variscan high-T event distinct from that of the Eosson migmatites.

ORDOVICIAN MAGMATISM

Apart from late Carboniferous continental detrital deposits (Salvan-Dorénaz syncline, Capuzzo and Bussy 2000) and granite intrusions (Vallorcine granite, Fully granodiorite, Bussy et al. 2000), all other lithologies of the ARM predate the 320 Ma-old Variscan high-T metamorphic event, but no relative chronology can be established among them, as no primary contacts are preserved.

Mafic magmatism is of little volumetric importance. It is mainly documented as swarms of metre-long garnet-amphibolite boudins within micaschists; they have a geochemical signature of continental tholeiites (von Raumer et al., 1990) and their age is unknown. More continuous outcrops are found in the Lake Cornu area, with layers up to several tens of metres long. They consist of mafic eclogites variously retrogressed into garnet-amphibolites. The eclogites are either massive and isotropic or banded with alternating cm-thick dark layers of garnet-amphibole-quartz-plagioclase and light layers of coronitic garnet-clinopyroxene-plagioclase. This layering has been interpreted by Liégeois and Duchesne (1981) as evidence for a volcanic-sedimentary origin. Alternatively, Péronnet (2009) has shown that the only difference between light and dark layers is the relative proportion of amphibole and diopside originating in the retrogression of the original eclogitic assemblage. Thus the limited amount of available water at time of retrogression might have controlled the development of the banding in originally homogeneous eclogites. Nevertheless, some light layers of the banded eclogites are relatively enriched in Al and Sr and display strong positive Eu anomalies in chondrite-normalized REE patterns. This is indicative of plagioclase accumulation and might record in situ mineral fractionation during basalt crystallization. Liégeois and Duchesne (1981) interpret the Lake Cornu eclogites as various terms of the low pressure differentiation of a continental tholeiite and suggest emplacement in a thinned continental crust environment. This is in agreement with the data of Paquette et al. (1989), who concluded that the massive eclogites have N-MORB REE signatures and positive initial epsilon Nd values between 5.9 to 6.8. Paquette et al. (1989) also performed isotope-dilution U-Pb dating on large zircon fractions extracted from a Lake Cornu eclogite. They obtained an upper intercept date of 453 $\pm$ 3/-2 Ma interpreted as the magmatic age of the mafic protolith of the eclogite.

Interestingly, a few ultramafic boudins of serpentinite up to 100 m long are wrapped in paragneisses in the Lake Cornu area. Their chemistry points to former Iherzolites and pyroxenites (Pfeifer and von Raumer, 1996; von Raumer and Bussy, 2004).

Granitic magmatism is expressed as large volumes of orthogneisses in the ARM. The most common facies is a porphyritic biotite±muscovite augengneiss with K-feldspar megacrysts up to 5-10 cm long (e.g. Eosson, Bérard, Lac Noir). Other lithologies include amphibole-biotite orthogneisses (Bérard) and some leucogneisses. The augengneisses are peraluminous (A/CNK=1.4) granodiorites to monzogranites. Their zircons display morphologies typical of S-type granites (Pupin, 1980). The amphibole-biotite orthogneisses are metaluminous granodiorites to tonalites typical of I-type calc-alkaline series, as confirmed by zircon morphology and whole-rock geochemistry (von Raumer and Bussy, 2004).

GEOCHRONOLOGY

Five rock samples have been dated by U-Pb zircon geochronology. In situ isotopic measurements have been performed by LA-ICPMS using an Element XR sector-field spectrometer interfaced to an UP-193 excimer ablation system. The instrument was calibrated using a GJ-1 zircon as external standard. Accuracy was monitored by analysing the Harward 91500 standard as an unknown. The systematic error of the 91500 standard measurements is <1% for any of the sequences. The U-Pb ages reported here correspond to the weighted mean of individual $^{206}\text{Pb}/^{238}\text{U}$ age determinations.

Banded eclogite sample ViP44 (Swiss grid coordinates 89.758/554.105) yielded rounded zircons about 50 to 100 microns in diameter. Cathodoluminescence (CL) imaging (Fig. 1a,b) reveals the common association of an oscillatory zoned rounded core and an unzoned rim of homogeneous colour, which can be either darker (higher in U) or lighter (lower in U) than the core. The core is interpreted as magmatic, whereas the homogeneous rim is considered metamorphic. 43 out of 55 concordant measurements on zircon cores were statistically selected as a coherent group by the “zircon age extractor” subroutine of Isoplot (Ludwig, 2009); they yield a mean $^{206}\text{Pb}/^{238}\text{U}$ age of $463 \pm 3/-2$ Ma (Fig. 2a) interpreted as the magmatic age of the mafic protolith. Measurements in the rims yielded inconsistent dates older than the magmatic cores (Fig. 1a).

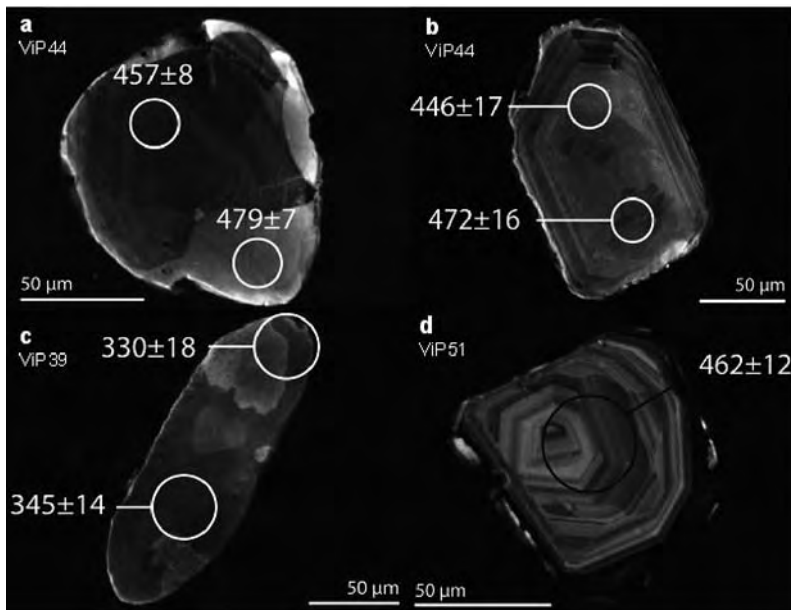


Figure 1. Cathodoluminescence images of zircon crystals with position of the ablation craters and corresponding individual $^{206}\text{Pb}/^{238}\text{U}$ ages (± 2 sigma).

Eclogitic amphibolite sample ViP39 (Swiss grid coordinates 89.757/554.083) yielded rounded zircons very similar to those of the banded eclogite. 54 out of 75 concordant measurements on zircon cores were statistically selected as a coherent group (Ludwig, 2009); they yield a mean $^{206}\text{Pb}/^{238}\text{U}$ age of 458 ± 5 Ma (Fig. 2b), identical within errors to the age of the massive eclogite ViP44. Among the remaining

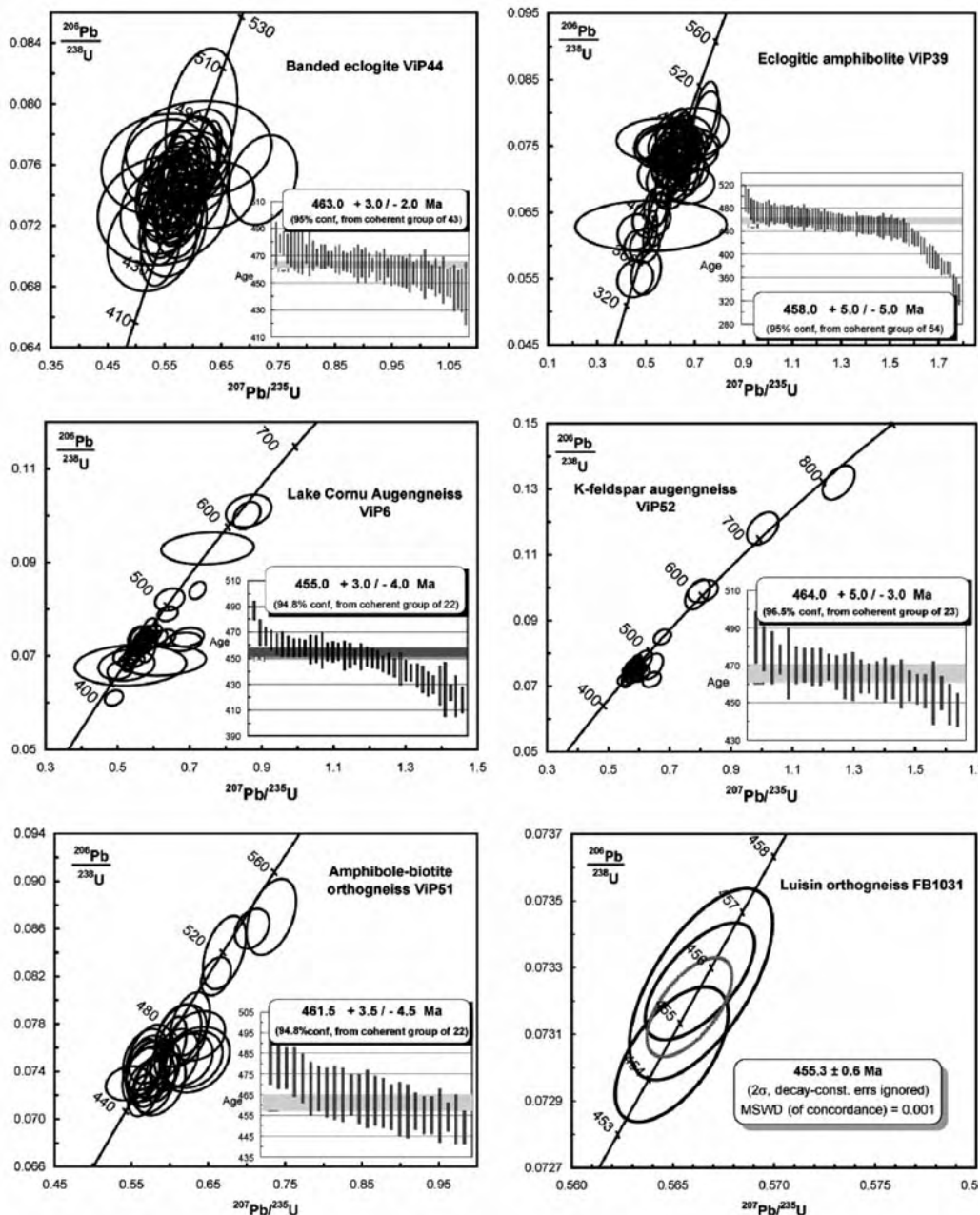


Figure 2. U-Pb concordia plots for the dated lithologies; reported ages are mean $^{206}\text{Pb}/^{238}\text{U}$ ages calculated with the “zircon age extractor” subroutine of Isoplot (Ludwig, 2009), see inserts. Sample FB1031 has been dated by the isotope dilution (IDTIMS) technique and the reported age is the Concordia age (Ludwig, 2009). All errors are given at the two-sigma level.

measurements, two are older and discordant ($^{206}\text{Pb}/^{238}\text{U}$ dates of 492 and 496 Ma, respectively); they probably contain an inherited component. 16 other data points spread down to a concordant point at 345 Ma ($^{206}\text{Pb}/^{238}\text{U}$ age = 345 ± 14 and $^{207}\text{Pb}/^{235}\text{U}$ age = 346 ± 43 Ma) (Fig. 1c and 2b); they are interpreted as mixed ages between a magmatic component at 458 Ma and a metamorphic event close to 345 Ma, which might correspond to the high-pressure metamorphic event.

Lake Cornu augengneiss ViP6 (Swiss coordinates 90.423/554895) is representative of the widely distributed biotite-muscovite peraluminous K-feldspar orthogneisses. It yielded relatively big zircons with a well-developed {211} pyramid typical of Al-rich melts (Pupin, 1980). The internal CL structure shows inherited cores wrapped by large oscillatory growth zones of magmatic origin. Metamorphic overgrowths are usually lacking. A statistically consistent group of 22 out of 45 analyses yield a mean $^{206}\text{Pb}/^{238}\text{U}$ date of $455 +3/-4$ Ma, interpreted as the crystallization age of the porphyric granite (Fig. 2c). Older concordant dates range from 470 to 1035 Ma and are interpreted either as mixed ages or as related to inherited cores. Younger dates spread down to 384 Ma and are interpreted as data points disturbed by metamorphic remobilization.

The Val Bérard ViP52 K-feldspar augengneiss is similar to sample ViP6 and yielded zircons of the same kind, except that the {211} pyramid is less developed and that no inherited cores have been observed on CL images. A statistically consistent set of 23 out of 32 analyses yield a mean $^{206}\text{Pb}/^{238}\text{U}$ date of $464 +5/-3$ Ma, interpreted as the magmatic age of the porphyric granite (Fig. 2d). It is slightly older than the Lake Cornu augengneiss.

The Val Bérard ViP51 amphibole-biotite metaluminous orthogneiss (French grid coordinates 950.125/121.855) yielded zircons up to 150 microns long with some inherited cores and a generally well-developed oscillatory zoning (Fig. 1d). Many crystals are metamictic and were partly dissolved during the leaching procedure. 22 out of 32 measurements define a statistically coherent group which yields a mean $^{206}\text{Pb}/^{238}\text{U}$ date of $461.5 +3.5/-4.5$ Ma, interpreted as the magmatic age of the granodiorite (Fig. 2e). Some older concordant dates ranging between 500 and 735 Ma point to inherited cores or crystals.

In addition, a sample from the calc-alkaline metaluminous orthogneiss of Mt Luisin north of Lake Emosson (FB1031) has been dated in the nineties at the Royal Ontario Museum by U-Pb zircon geochronology using the isotope dilution technique (ID-TIMS). The applied analytical procedure is described in Bussy and Cadoppi (1996). Three small zircon fractions yielded a concordia age of 455.3 ± 0.6 Ma (MSWD = 0.001) (Fig. 2f).

DISCUSSION

All pre-Variscan magmatic rocks of the ARM emplaced in a relatively short time span between 455 and 465 Ma. The same is true for a large augengneiss body from the neighbouring Mont Blanc massif dated at 453 ± 3 Ma (Bussy and Von Raumer, 1994). The amphibole-biotite granodiorite (ViP51) and the Mt Luisin granodiorite (FB1031) are I-type metaluminous granitoids with mafic microgranular enclaves typical of calc-alkaline magmatic series. On the other hand, the voluminous K-feldspar augengneiss found both in the Aiguilles Rouges (Bérard (ViP52), Lac Noir (ViP6), Mont-Blanc (Morard, 1998)) are relatively peraluminous as confirmed by their zircon morphology, but they do not display the characteristic features of classical S-type granites like restitic enclaves and schlieren or large amounts of primary muscovite. We interpret the corundum-normative character of these intrusions as acquired through high-pressure (>0.8 GPa) fractional crystallization processes of standard metaluminous calc-alkaline melts in lower crustal levels (see e.g. Alonzo-Perez et

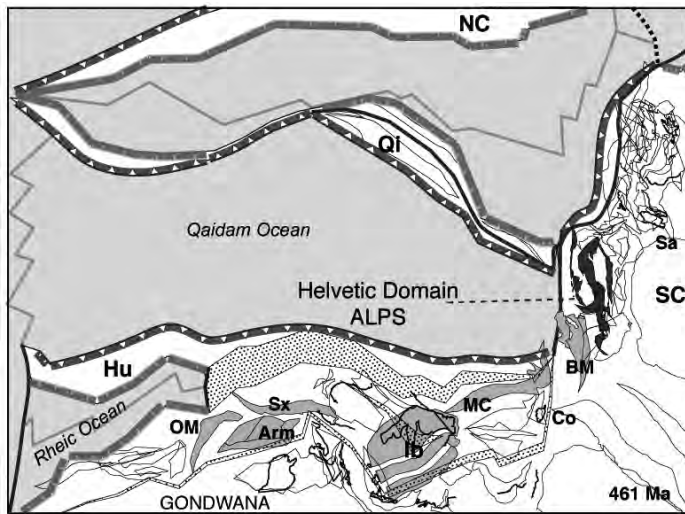


Figure 3. Ordovician (461 Ma) plate tectonic reconstruction (after Stampfli et al., this volume: Fig. 1D), showing the future Alpine Geodynamic units (dark grey), in the frame of the Ordovician basement areas, at the eastern limits of the Qaidam Ocean, spanning between the Qilian basement in the north and the Hunic terranes (Hu), still located at the Gondwana margin. The future eastern branch of the Rheic Ocean is not yet opened. Specific basement areas in light grey: Arm, Armorican terrane assemblage; BM, Bohemian massif and Barrandian areas; Co, Corsica; Ib, Iberian terrane assemblage; MC, Massif Central; NC, North China; Qi, Qidam; Sa, Sardinia; SC, South China; Sx, Saxothuringian domain; OM, Ossa Morena; dotted spaces, sedimentary troughs with detrital sediments (e.g. Armorican quartzite).

al. 2009). The close spatial association of the calc-alkaline granite plutons with minor volumes of mafic rocks of tholeiitic affinity is probably not original, but might result from Variscan tectonics. Indeed, the distribution of the (retro)eclogite mafic boudins in paragneisses is reminiscent of the tectonic accretion channel of a suture zone (Engi et al., 2001). If true, the basaltic sills(?) might have emplaced away from the granite plutons, like in an extensional basin in a back-arc geodynamic environment.

Ordovician magmatism is recorded all over the Alpine basement units. It extends from 470 to 440 Ma in the external crystalline massifs (see review in von Raumer et al., 2002) and from ca. 480 to 450 Ma in the Penninic and Austroalpine domains (e.g. Guillot et al., 2002; Schulz et al., 2008; Liati et al., 2009). The general picture is the following (see review in Schaltegger and Gebauer, 1999): an early mafic activity locally coeval with orthogneisses is documented by gabbros in various units (Silvretta nappe, Gotthard, Tavetsch and Aar massifs) between 471 and 467 Ma. Many large granite plutons emplaced between 470 and 450 Ma, whereas a regional high-T metamorphic event with partial melting is recorded in the Aar and Gotthard massifs at ca. 445-450 Ma and in the ARM at ca. 440 Ma by monazite in garnet micaschists (Schulz and von Raumer, 2011).

The Ordovician magmatism in the future Alpine realm is quite distinct from that developed during the subsequent Variscan orogeny. The latter is characterized by 340 Ma-old Mg-K-rich monzogranites emplaced along lithospheric-scale transcurrent faults tapping an enriched subcontinental mantle source (Debon and Lemmet, 1999), and by abundant migmatites and cordierite-bearing peraluminous granites like in the Velay dome (French Massif Central), which evidence a very high thermal flux unknown in Ordovician lithologies. The Ordovician context is more reminiscent of an active continental margin of western north American type.

Current geodynamic reconstructions generally agree on the existence of a Cambrian active continental margin all along northern Gondwana, consequence of the southward subduction of the Iapetus ocean to the west and the Prototethys ocean to the east (e.g. von Raumer and Stampfli, 2008; Schulz et al. 2008; Guillot and Ménot, 2009). On the other hand, models significantly diverge in detail for the post-Cambrian evolution of this margin (compare e.g. Stampfli et al., this volume; Guillot and Ménot, 2009). The widespread Ordovician magmatism in the future Alpine realm on the one hand and the new geochronological data in the ARM on the other hand bring two important constraints to geodynamic models. First, the future Alpine terranes must be located relatively close (up to a few hundreds km) to the oceanic trench of the north Gondwanan active margin from 480 to 450 Ma (see e.g. Stampfli et al. 2011), as large volumes of calc-alkaline granites emplaced during that time. Second, the contemporaneous mafic rocks of tholeiitic affinity most probably emplaced in a different setting, possibly in a back-arc extensional basin locally floored by exhumed subcontinental mantle, now preserved as mega serpentinite boudins in the ARM. The mafic rocks were subsequently subducted during the Variscan orogeny (possibly at ca. 345 Ma, sample ViP39) and involved in a tectonic accretion channel which brought them back to mid-crustal levels as eclogitic boudins.

The thermal event recorded by the 450 Ma-old migmatites in the Aar massif (Schaltegger and Gebauer, 1999) and 440 Ma-old monazites in high-grade micaschists from the ARM (Schulz and von Raumer, 2011) might either be linked to the voluminous magmatic activity in these terranes at that time or to a high heat flux related to crustal extension, possibly the back-arc extension which will ultimately open the Paleotethys ocean (see reconstructions by Stampfli et al., this volume).

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REWORKED CONODONTS IN THE UPPER ORDOVICIAN SANTA GERTRUDIS FORMATION (SALTA, ARGENTINA)

J. Carlorosi¹, S. Heredia², G.N. Sarmiento³ and M.C. Moya⁴

¹ Instituto Superior de Correlación Geológica (CONICET-UNT), Miguel Lillo 205, 4000 San Miguel de Tucumán, Argentina.
josefinacarlorosi@hotmail.com

² CONICET-IIM, Universidad Nacional de San Juan, Av. Libertador y Urquiza, 5400 San Juan, Argentina. sheredia@unsj.edu.ar

³ Departamento de Paleontología, Universidad Complutense de Madrid, José Antonio Novais 2, 28040 Madrid, Spain.
gsarmien@geo.ucm.es

⁴ CONICET-CIUNSA, Universidad Nacional de Salta, Buenos Aires 177, 4400 Salta, Argentina. moyacris@arnet.com.ar

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INTRODUCTION

The Sierra de Moajoroto in the Eastern Cordillera represents one of the best Lower Paleozoic successions where reworked Middle Ordovician fossils may occur in the Upper Ordovician Santa Gertrudis Formation. The geology of this region was described by Harrington (1938,1957), Ruiz Huidobro and González Bonorino (1953), Ruiz Huidobro (1955, 1968, 1975), Moya (1988), Hong and Moya (1993), Moya et al. (1994), Malanca (1996), Waisfeld (1996) and Moya (1998) among others.

The first mention of conodonts in the Santa Gertrudis Fm is due to Monaldi and Monaldi (1978), followed by Sarmiento and Rao (1987) and Albanesi and Rao (1996), who described part of its conodont fauna.

The main purpose of this contribution is to offer a review of the conodont species described for this unit, giving a precise age for the Santa Gertrudis Formation.

STRATIGRAPHY

The Santa Gertrudis Formation was defined by Harrington (1957). This unit crops out at the Gallinato and Santa Gertrudis Creeks, 14 km north of Salta City (Fig. 1), and is composed by quartz wackes intensively bioturbated (lower member) and by grey siltstones alternating with limestones (upper member). The entire unit reaches 80 m of thickness in the Gallinato Creek, where the Santa Gertrudis Fm paraconformably overlies the Lower Ordovician Mojotoro Formation (Fig. 2). The fossiliferous levels are in the upper member, present in thin carbonate-phosphatic beds. The fossil record comprises trilobites (Harrington, 1957; Monaldi and Monaldi, 1978; Monaldi, 1982), bivalves (Sánchez, 1986), brachiopods (Benedetto, 1999) and conodonts.

METHODOLOGY

Six conodont samples were collected from limestone beds (upper member) at random intervals (Fig. 2). All the elements have a color alteration index of 2 (60–140°C) (Epstein et al., 1977). The conodonts are housed in the collection of the INSUGEO at the Instituto Miguel Lillo (Tucumán), under the code MLC–C.

CONODONTS

Previous work on conodonts from the Santa Gertrudis Fm (Sarmiento and Rao, 1987; Albanesi and Rao, 1996; Moya et al., 2003) mentioned the conodont species *Erismodus quadridactylus* (Stauffer), *Bryantodina* aff. *typicalis* Stauffer, *Plectodina* n. sp. A, *Erraticodon* cf. *gratus* (Moskalenko), *Erismodus typus* Branson and Mehl, *Icriodella* n. sp. A, *Polyscaulodus* sp. and *Semiacontiodus* sp.

According with our own conodont collection, *Baltodina triangularis* (Lindström), *Erraticodon patu* (Cooper), *Trapezognathus quadrangulum* Lindström, *Bryantodina* nov. sp. A and *Erismodus quadridactylus* (Stauffer) were identified in this Ordovician unit.

This conodont association indicates a mixture of conodont faunas pointing out to erosion of former deposits of earliest Middle Ordovician age. In the other hand, the autochthonous conodont fauna includes only a few elements of *Erismodus quadridactylus* that are indicative of the *E. quadridactylus* Zone, which have a Late Ordovician age according to the Midcontinent chronobiostratigraphic chart (Sweet, 1984). Also, *Erismodus* suggests a high energy environment which is consistent with the tempestitic features of the intercalated coquinoid beds.

In our opinion the confirmed presence of *E. quadridactylus* demonstrates an Upper Ordovician age for the upper part of the Santa Gertrudis Fm, allowing its correlation with the Capillas Formation, which also bears *E. quadridactylus* and is widely represented on the eastern flank of the Eastern Cordillera and in the Subandean Sierras (Andean Basin).

Albanesi et al. (2007) recorded conodonts from the Capillas Fm, suggesting a late Darriwilian age for this unit. Based on the conodont record, these authors proposed a correlation with the Santa Gertrudis Fm. We agree with this last statement after comparing the conodonts illustrated by Albanesi et al. (2007) with our conodont collection (both with autochthonous and reworked species). We also propose a younger age for the

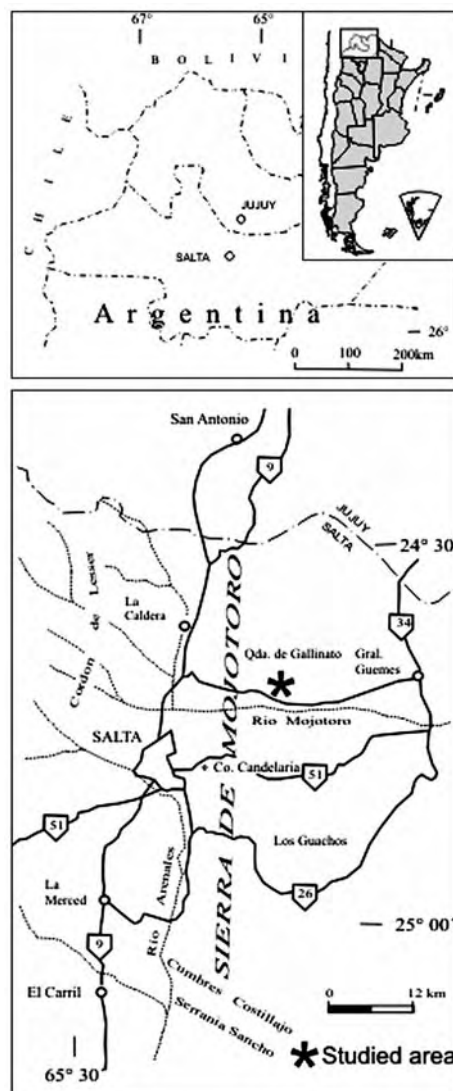


Figure 1. Location map of the studied outcrops.

conodont-bearing strata from the upper member of the Santa Gertrudis Fm, as well as from the Capillas Fm, based in the common record of *Erismodus quadridactylus* and various conodont taxa reworked from younger rocks.

Baltoniodus triangularis (Lindström) is here recorded for the first time in the Andean basin. *B. triangularis* was recently proposed as an index-conodont for the earliest Dapingian (Middle Ordovician Series) (Wang et al., 2003a, 2003b, 2009; Stouge et al., 2005). This species is a very unusual conodont in the Ordovician of South America. There is only one mention of a single P element from the Precordillera (Albanesi et al., 1998). The appearance of *B. triangularis* in the recovered residues of Santa Gertrudis Formation represents a hidden history of former deposits.

Trapezognathus quadrangulum Lindström is recognized here for the first time in Gondwana and the Andean basin, and it ranges from the *Baltoniodus triangularis* Zone to the *Baltoniodus norrlandicus* Zone of the Middle Ordovician. Stouge and Bagnoli (1990) restricted *T. quadrangulum* to the early Middle Ordovician.

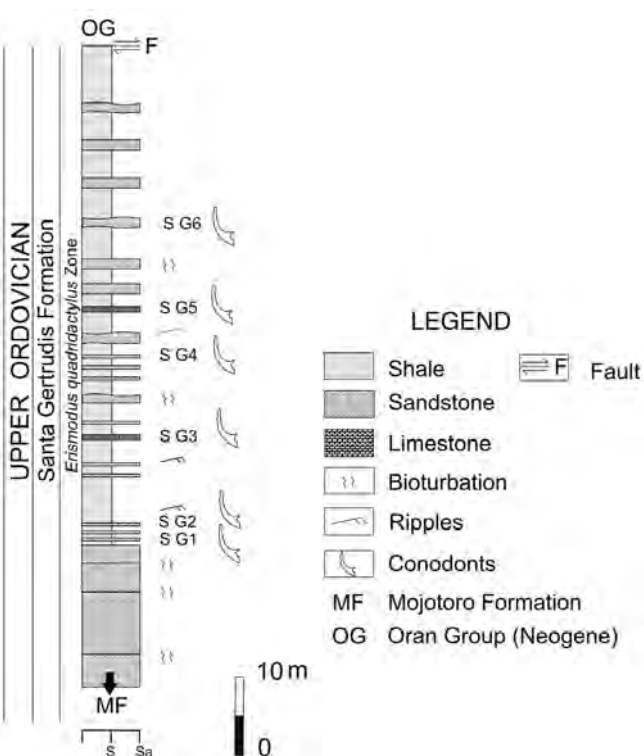


Figure 2. Gallinato Creek section showing sampled levels with Sandbian conodonts.

CONCLUSIONS

The conodont record of the studied section provides a significant increase in the knowledge of Ordovician conodont faunas of the Andean Basin, and turns out to be of great interest for the reconstruction of the sedimentary history of this area. Most of this conodont fauna is composed by reworked species, while the autochthonous conodonts are few, being specially represented by *Erismodus quadridactylus*. The age of the upper member of the Santa Gertrudis Formation is constrained to the *E. quadridactylus* Zone. In the other hand, the biostratigraphical meaning of allochthonous index species allow the dating of the depositional time for the reworked sediments, pointing out a younger depositional cycle over the Acoite Formation (uppermost Floian) developed at least during the *B. triangularis* Zone (basal Dapingian), followed by emersion and partial erosion. During the Sandbian (*Erismodus quadridactylus* Zone) open shelf marine sedimentation recycled these older deposits (Aceñolaza et al., 2010).

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A POST-GLACIAL BRYOZOAN FAUNA FROM THE UPPER ORDOVICIAN (HIRNANTIAN) OF THE ARGENTINE PRECORDILLERA

M. Carrera and K. Halpern

CICTERRA-CONICET, Facultad de Ciencias Exactas, Físicas y Naturales. Av. Vélez Sarsfield 299, 5000 Córdoba, Argentina.
mcarrera@efn.uncor.edu, karenhalpern@gmail.com

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INTRODUCTION

The Late Ordovician extinction was the second largest loss of diversity in the history of life. Almost 48% of marine genera disappeared during this event. Nevertheless, the communities rapidly recovered due to low structural changes in terms of ecological organization. Detailed taxonomic studies are frequent in the literature on this subject. On the contrary, studies on paleoecological dynamics of this event are scarce or treated broadly. However, these preliminary attempts are of enormous value and allow continuing research.

An accepted paleoclimatic interpretation of the Latest Ordovician, based on different sources of data, points out that the polar region of Gondwana was covered by an ice sheet for a short time, and it is consistent with the presence of glacial deposits in Africa and in South America (Sutcliffe et al., 2001 and references therein).

Since the pioneer studies of Benedetto (1986, 1990) and Sánchez et al. (1991) very few taxonomic or paleoecological studies on the *Hirnantia* fauna in the glacial deposits of the Argentine Precordillera have been carried out. Only some taxonomic studies of brachiopods or paleogeographic implications of their distribution were made.

Bryozoans have been listed as very scarce components of this fauna. New fossil collections in the post-glacial deposits in the Don Braulio section, Villicum range (Fig. 1), allow us to recognize numerous bryozoan specimens. The brachiopod *Hirnantia* and the bivalve *Modiolopsis* were considered as the dominant components of the fossil communities previously described (Sánchez et al., 1991). These dominance values should be reevaluated in the light of the amount of bryozoans found in the new collected material.

In this contribution, we report the occurrence of two bryozoan genera in the Upper Ordovician (Hirnantian) Don Braulio Formation, Argentinean Precordillera. They are associated with the typical *Hirnantia* Fauna (Benedetto, 1986), representing the first community that flourished after the Late Ordovician Glaciation.

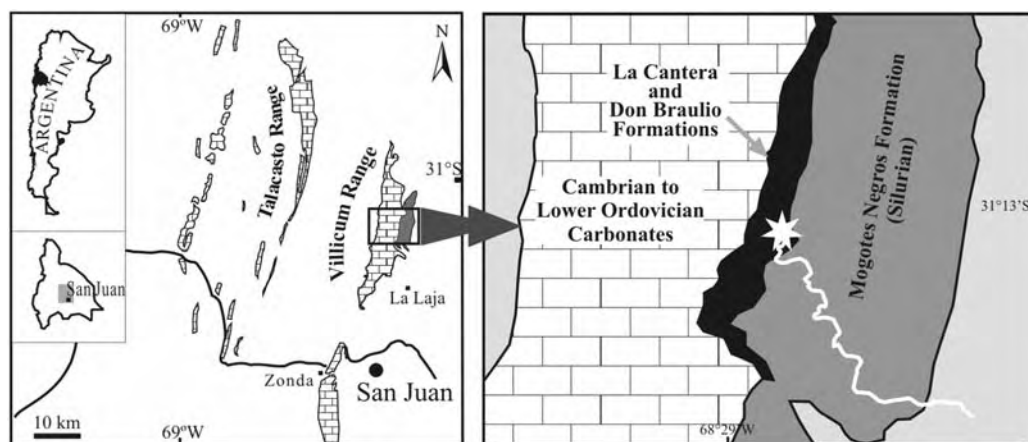


Figure 1. Geological map and location of the studied section in the Don Braulio Formation, Villicum range.

GEOLOGICAL SETTING AND STRATIGRAPHY

The area of study is located at the Eastern flank of the Villicum Range in the Argentine Precordillera, at Don Braulio section, where detailed sedimentological studies have been carried out by different authors (Fig. 1). The Don Braulio Formation rests on La Cantera Formation (Darrivillian to Katian) and underlies the Mogotes Negros/Rinconada Formation (Ludlovian to Pridolian). The Hirnantian age of this unit was established by the presence of brachiopods and the record of *P. persculptus* (Brussa et al., 2003).

The Don Braulio deposits were interpreted of glacial origin (Buggisch and Astini, 1993; Peralta and Carter, 1990). The unit is usually divided into two members. The Lower Member starts with mud-supported diamictites bearing clasts often polished and striated, that alternate with channel-like deposits filled with sandstones and grain-supported conglomerates. The upper member mainly consists of greenish bioturbated mudstones and sandstones, some containing carbonate cement and commonly macrofossils (bryozoans, brachiopods, trilobites, bivalves and graptolites). The specimens come from the upper member of the type-section at Don Braulio Creek (Figs. 1, 2).

BRYOZOAN TAXONOMY

In this preliminary contribution we mention the morphologic characteristics and make a brief discussion of the two bryozoan genera found, the complete taxonomic study will be part of a forthcoming contribution.

The most abundant specimen is a stick like cylindrical form, few centimeters long identified as *Helopora* sp. (Fig. 2). No branching forms have been found. The specimens have oval autozooeal apertures 0.22-0.25 mm in maximum diameter, arranged in irregular rows. Diaphragms in autozoecia are rare or absent. Acanthostyles are long, abundant, three to five surrounding autozoecia apertures. Mesopores are polygonal, small, with diaphragms and 0.05-0.06 mm in diameter. Axial region with thin slightly define linear axis.

The colonies look similar to those described as *Moyerella* by Ernst and Carrera (2008) in the late Katian

Sassito Formation. However, this genus has the autozoecia disposed in regular diagonal rows and it has conspicuous tectitozoecia which are absent in *Helopora* sp.

The genus *Helopora* has been previously mentioned by Rusconi (1956) as a form found in the Ordovician of Mendoza in the south of Precordillera. However, the colony figured by Rusconi is a ramose form without description that could be included in any kind of Ordovician genus.

The other bryozoan form is a small ramose colony represented by several fragments. The colony is a reticulate form with anastomosing branches in a similar way of the family Phylloporinidae. Phylloporinid specimens have been previously reported in Argentina in older units (Carrera and Ernst, 2010; Ernst and Carrera, 2008).

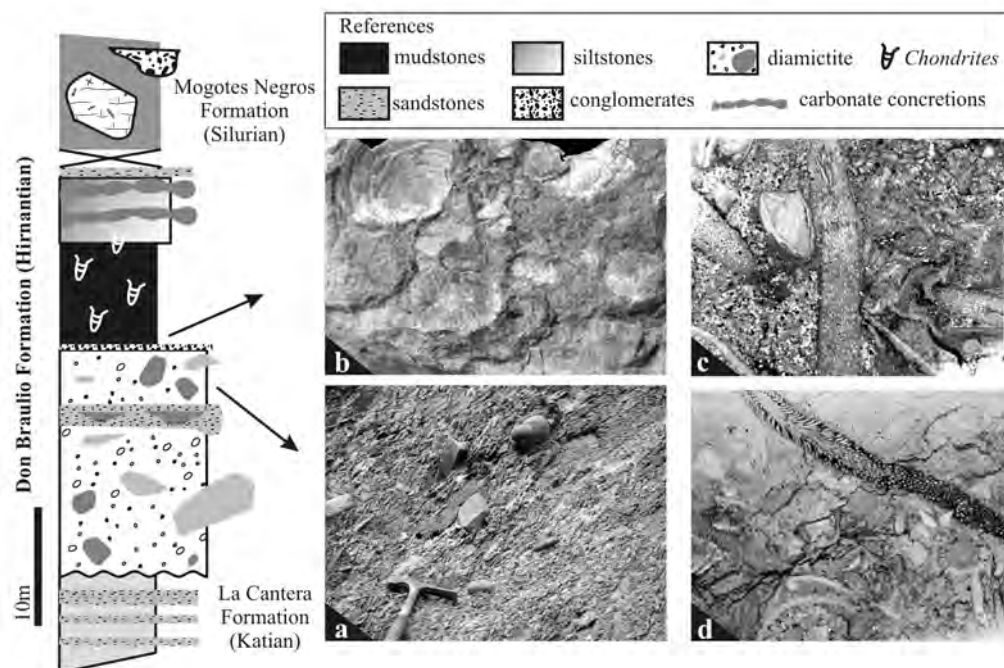


Figure 2. Stratigraphic column of the Don Braulio Formation (Hirnantian) in the studied section. (a) Glacial diamictites (b) Shell bed concentration with *Hirnantia* brachiopod valves. (c, d) Thin sections of b, including bryozoan specimens of *Helopora* sp. and Phylloporinidae indet.

BRYOZOAN DISTRIBUTION AND PALEOGEOGRAPHY

Paleogeographical approaches on Precordillera terrain have been made since 1990's, starting with the proposal made by Benedetto (2004, and references therein) and later several other studies that support the allocthonous history of the terrain. The latitudinal variation in faunal composition and the changing depositional sedimentary regime, show that the Precordillera traveled from equatorial to higher latitudes (Astini, 1998; Benedetto et al., 1999; Keller et al., 1998). Bryozoans are no the exception and match this pattern. After typical warm water carbonates in the Cambrian and Lower Ordovician, Middle Ordovician units developed at mid latitude (30-35°) locations, and the Katian Sassito limestones into the temperate belt, at higher latitudes (Ernst and Carrera, 2008). The Hirnantian glacialigenic rocks of the Don Braulio

Formation are the last step in the climatic wandering path of the Precordillera terrain (Astini, 1998; Benedetto et al., 1999).

As recently mentioned by Ernst and Carrera (2008) bryozoans have been documented in different locations and stratigraphic positions in the Argentine Precordillera during the Ordovician. They have been studied by Carrera (2003 and references therein), Ernst and Carrera (2008) and Carrera and Ernst (2010). Nevertheless, most of the paleontological record comes from Middle to Upper Ordovician sequences (Darrivillan tropical carbonates to Katian temperate water carbonates).

The record of uppermost Ordovician Bryozoans is scarce worldwide. There have been Ashgillian (Upper Katian) reports from Baltica, Laurentia, Siberia, Avalonia, North Western Africa and Mediterranean (Jiménez-Sánchez and Villas, 2010). However, there are few reports of definite Hirnantian bryozoans, such as those from the Anticosti Island (Canada) (Ernst and Munnecke, 2009) and the Late Katian to Silurian interval in Northern India (Suttner and Ernst, 2007).

The genus *Helopora* has been mentioned in the Middle Ordovician of North America (Ohio, Indiana, Michigan, Minnesota) and Estonia, and in the Upper Ordovician of India (Suttner and Ernst, 2007). During the Silurian and Devonian the genus is recorded in Canada, Russia, and China, showing a more widespread distribution.

Both previous Hirnantian bryozoan records in Anticosti (Ernst and Munnecke, 2009) and India (Suttner and Ernst, 2007) show an important diversity (13 and 29 species respectively) which contrast with the clearly low diversity found in the Don Braulio Formation with two species reported. The paleoenvironmental features of the Anticosti and the Indian localities may explain the high diversification of their bryozoans assemblages. Both are developed in shallow tropical to subtropical areas and the Anticosti bryozoans belong to a reef related community (Ernst and Munnecke, 2009).

The glacial related bryozoans reported here follow the pattern of low diversity values found in Paleozoic temperate to cold climates in contrast to post Paleozoic distribution (Ernst and Carrera, 2008; Taylor and Allison, 1998).

Although the Silurian of Argentina has been extensively studied and includes important fossil collections of sessile attached organisms, such as, crinoids and corals, no bryozoans have been found to date. *Helopora* and the undetermined phylloporinid were the last representatives of the phylum in Precordillera until few representatives reappeared in the Carboniferous. *Helopora* has several Silurian and Devonian species, and it was one of the successful genera that crossed the Ordovician-Silurian boundary.

ACCOMPANYING FAUNA AND PALEOECOLOGY

Just above the glacial deposits of Don Braulio Formation, a conspicuous shell bed occurs, followed by bioturbated mudstones (Fig. 2). This first fossil association consists primarily of brachiopods and some bivalves, resembling the *Hirnantia-Modiolopsis* community (Sánchez et al., 1991). Among these brachiopods and bivalves we recognize very common fragments and entire colonies of *Helopora* and phylloporinid specimens.

A preliminary study of the fossiliferous association indicates some physical influence during its development, i.e.: fragmentation and disarticulation both implying high energetic conditions. However, almost the same taxa are also found in the overlaying fossiliferous mudstones beds and thus the shell bed can be considered as an *in situ* reworked para-autochthonous concentration. Besides some fossil remains such as the bryozoans colonies are almost complete suggesting not so high degree of reworking.

The samples contain frequent orthid and strophomenid brachiopods: *Dalmanella testudinaria* (58%), *Paromalomena polonica* (20%), and *Hirnantia sagittifera* (15%), among few bivalves (*Modiolopsis* sp., 7%) and two specimens of Dalmanitoid trilobites. Bryozoan colonies represent 26% of the *Hirnantia-Modiolopsis* 'community'.

Sedimentary studies indicated a deepening-upward sequence after the diamictite deposits. Temperature increases causing the ice cap melting and a sea level rise.

Classically, the *Hirnantia* fauna has been interpreted of cold to temperate shallow waters. Some of the morphofunctional characteristics of the fauna support the shallow environmental setting. *Modiolopsis* sp. is a bisally attached semi-infaunal bivalve and *H. sagittifera* is a pedicle attached epifaunal brachiopod, typically found in shoreface facies. In this section, the dominance of *D. testudinaria* is not conclusive, for being one of the most eurytopic species in the *Hirnantia* fauna (Rong and Li, 1999). *P. polonica* is also a pedicle attached form, although is adapted for quiet and deeper water regimes (Rong and Li, 1999).

The low diversity values found in this association and the particularly poor bryozoan fauna points to temperate to cold waters. The presence and abundance of stick-like bryozoans allow us to narrow up the environmental conditions known to date: hard substrates, low turbidity and a nutrient-rich environment can be reassured.

More sampling and detailed analysis of the vertical distribution of the *Hirnantia* fauna in this and other localities will help to understand the evolution of environmental conditions after the latest Ordovician Glaciation on the Argentine Precordillera and the impact caused on the biota resulting in its extinction.

Acknowledgements

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ORDOVICIAN MAGMATISM IN NE IBERIA

J.M. Casas¹, P. Castiñeiras², M. Navidad², M. Liesa³, J.F. Martínez⁴, J. Carreras⁴, J. Reche⁴, A. Iriondo⁵,
J. Aleinikoff⁶, J. Cirés⁷ and C. Dietsch⁸

¹ Dept. de Geodinàmica i Geofísica, Universitat de Barcelona, Institut de recerca Geomodels, Martí Franquès s/n,
08028 Barcelona, Spain. casas@ub.edu

² Depto. de Petrología y Geoquímica-Instituto de Geología Económica (UCM-CSIC), Facultad de Ciencias Geológicas.
Universidad Complutense, 28040 Madrid, Spain. castigar@geo.ucm.es; navidad@geo.ucm.es

³ Dept. de Geoquímica, Petrologia i Prospecció Geològica, Universitat de Barcelona, Martí i Franquès s/n, 08028 Barcelona, Spain.
mliesa@ub.edu

⁴ Dept. de Geologia, Universitat Autònoma de Barcelona, 08193 Bellaterra (Cerdanyola del Vallès), Spain.
Francisco.Martinez@uab.cat; Jordi.Carreras@uab.cat; joan.reche@uab.es

⁵ Centro de Geociencias, Universidad Nacional Autónoma de México, Campus Juriquilla, 76230 Juriquilla, Querétaro, México.
iriondo@geociencias.unam.mx

⁶ US Geological Survey, MS 963, Denver, CO 80225, USA. jaleinikoff@usgs.gov

⁷ Institut Geològic de Catalunya, Balmes 209-211, 08006 Barcelona, Spain. jcires@igc.cat

⁸ Dept. of Geology, University of Cincinnati, Cincinnati, OH 45221-0013, USA. dietscc@uc.edu

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INTRODUCTION

An important Ordovician magmatic event has been documented in the pre-Variscan rocks cropping out in the NE of Iberia (Pyrenees and Catalan Coastal Ranges) as in the rest of the European Variscides (Fig. 1). In the Pyrenees, large granitic orthogneissic bodies were interpreted initially as forming the core of metamorphic massifs that represented a Cadomian basement overlain by a lower Paleozoic cover (Autran et al., 1966; Guitard, 1970). In contrast, pioneering geochronological works pointed to an Ordovician age for the orthogneisses (Jäger and Zwart, 1968). Improvement in U-Pb geochronology (U-Pb SIMS, ID-TIMS and Laser ablation) yielded new data that confirmed the Early Ordovician age of part of the orthogneisses. The granitic protoliths of the gneisses intruded in a Neoproterozoic-Early Paleozoic metasedimentary sequence and deformed during the Variscan orogeny, thus invalidating the basement-cover model (Deloule et al., 2002; Cocherie et al., 2005; Castiñeiras et al., 2008). In this contribution we present a summary of the recent U-Pb geochronological data obtained in the pre-Variscan igneous rocks cropping out in the NE of Iberia which confirm the existence of a well developed Ordovician magmatism in this zone.

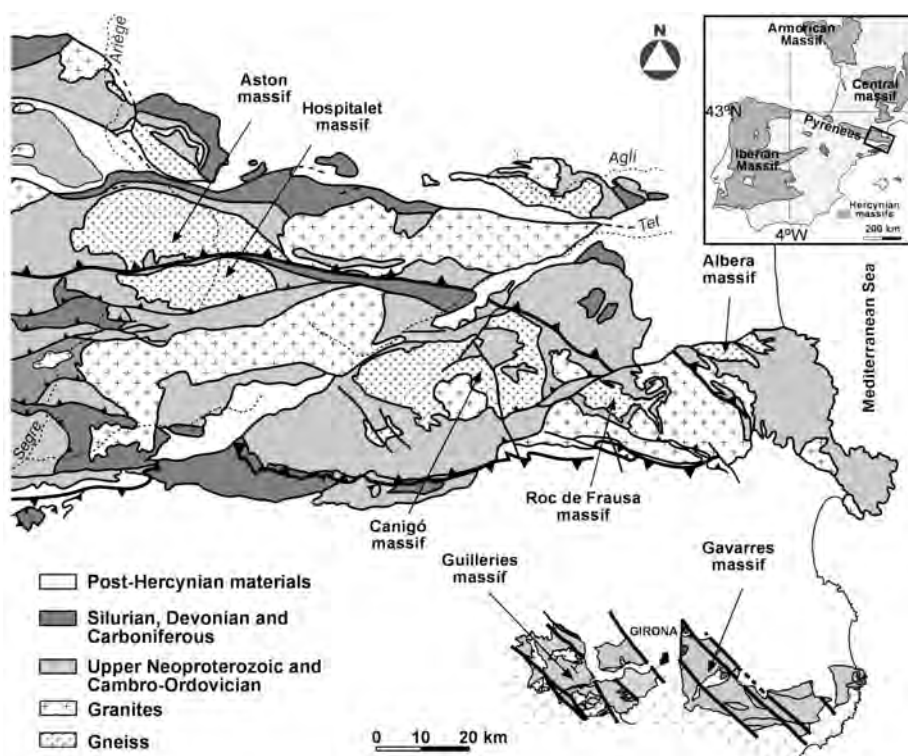


Figure 1. Geological sketch of the Variscan basement rocks cropping out in the NE of Iberia.

THE GEOCHRONOLOGICAL DATA

The large gneissic bodies

Large gneissic bodies with laccolithic morphology crop out at the core of dome-like massifs in the backbone of the Pyrenees (Fig. 1). They derive from granites ranging in thickness from 500 to 3000 m emplaced in the Neoproterozoic metasediments. From west to east, the protoliths of these augengneiss have yielded 472 ± 2 and 470 ± 6 Ma in the Aston-Hospitalet massifs (Denele et al., 2009); 472 ± 6 to 467 ± 7 in the Canigó massif (Cocherie et al., 2005); 477 ± 4 and 476 ± 5 Ma in the Roc de Frausa massif (Cocherie et al., 2005; Castiñeiras et al., 2008) and 470 ± 3 Ma in the Albera massif (Liesa et al., 2011). That is Floian (Late Early Ordovician) to Dapingian (Early Mid Ordovician) ages predominate. It should be noted that intermediate to basic coeval magmatic rocks have not been described and that no volcanic equivalents have been reported, except in the Albera massif, where subvolcanic rhyolitic porphyroid rocks yielded similar ages than those of the main gneissic bodies: 465 ± 4 , 472 ± 3 , 473 ± 2 and 474 ± 3 Ma (Liesa et al. 2011; Liesa et al., unpubl.). In the Les Guillerics massif (Catalan Coastal Ranges), Martínez et al. (2010) report ages ranging from 488 ± 3 to 459 ± 3 Ma, that is Early Ordovician (Tremadocian) to Late Ordovician (Sandbian), for fine grained gneisses forming tabular lenses interlayered in pre-Upper Ordovician metapelites.

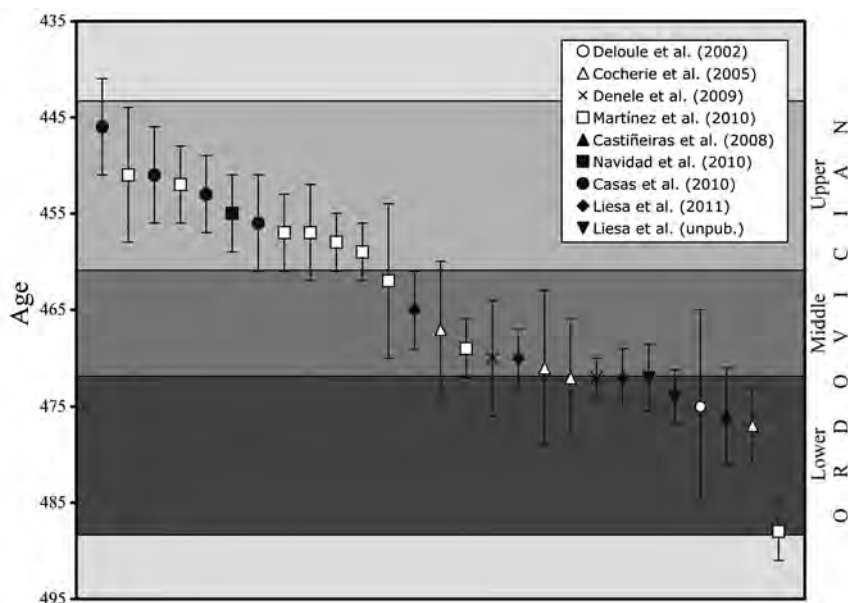


Figure 2. Summary of the U-Pb geochronology data of the Ordovician magmatism of the NE of Iberia.

The Upper Ordovician magmatic rocks

A well developed Late Ordovician volcanism has been widely described in the Pyrenees (Robert and Thiebaut 1974; Martí et al., 1986; Calvet et al., 1988) and the Catalan Coastal Ranges (Navidad and Barnolas, 1991; Barnolas and García-Sansegundo, 1992). This Late Ordovician magmatic event is mainly represented by calc-alkaline ignimbrites, andesites, volcanoclastic rocks, diorites and various types of granitic bodies. Recent dating allows us to confirm the Late Ordovician age for volcanic rocks interbedded in the Upper Ordovician sequence in the Les Gavarres massif, 455 ± 2 Ma (Navidad et al., 2010) and in the Les Guillerries massif, 452 ± 4 Ma (Martínez et al., 2010). We would also like to emphasize the Late Ordovician age obtained for the isotropic Ribes granophyre, 458 ± 3 Ma (Martínez et al., 2010), and for the Núria gneisses, in the southern slope of the Canigó massif (457 ± 4 and 457 ± 5 Ma, Martínez et al., 2010). In fact, the deepest rocks cropping out in the core of the Canigó massif, the Casemí and Cadí gneisses and some microdiorite bodies, have also yielded Late Ordovician ages:

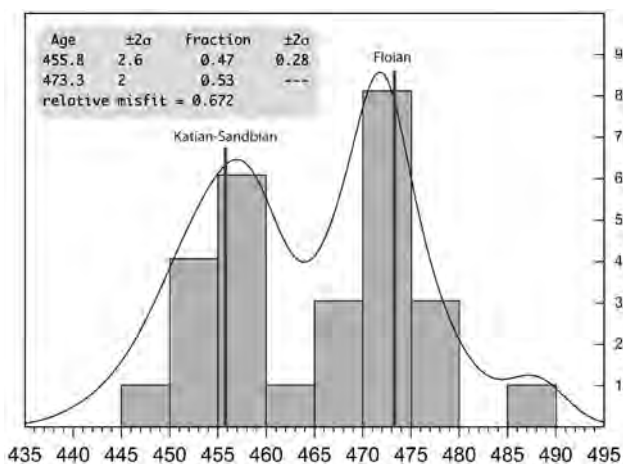


Figure 3. Distribution of the geochronological data.

446±5 and 451±5 Ma for the Casemí gneisses, invalidating a previous obtained Silurian age (Delaperrière and Soliva, 1992), 456±5 Ma for the Cadí gneisses and 453±4 Ma for a microdiorite body (Casas et al., 2010). Thus, the Cadí and Casemí gneisses, together with the Núria gneisses and the Ribes granophyre, can be regarded as the plutonic equivalent of coeval Late Ordovician volcanic rocks interbedded in the Upper Ordovician sequence of the Pyrenees.

DISCUSSION

From the aforementioned ages it follows that a continuous magmatic activity took place in the northeastern Iberian Peninsula during the Ordovician (Fig. 2). However, in spite of the limited number of available data some points should be emphasized. Magmatic activity seems to be more relevant during the late Early Ordovician to the Mid Ordovician (465–480 Ma) and during the early Late Ordovician (450–460 Ma). A period of minor activity between the 460 to 465 Ma can be envisaged (Fig. 3). The Mid to Early Ordovician magmatic activity has been widely related to the break-up of the northern Gondwana margin whereas the second magmatic event is coeval with normal fault development in Upper Ordovician rocks (Casas et al., 2010). A question arises whether there exists a period of lesser magmatic activity separating two differentiated events. The lack of magmatic rocks may be a sampling bias caused by the limited number of study areas or may be caused by the uncertainties inherited of the analytical results. If a period of lesser magmatic activity exists it would be compatible with the development of the Mid Ordovician deformation episode described in the Pyrenees (Casas, 2010). Whatever the case, both magmatic events display marked differences: The Early to Mid Ordovician magmatic activity gave rise to large bodies of aluminous granites with no coeval basic or intermediate rocks and subordinate subvolcanic rocks, whereas the Late Ordovician magmatism is responsible for calc-alkaline volcanic rocks and various types of metaluminous and aluminous granites and diorites.

The similarity of the isotopic signatures of some of the Early Ordovician aluminous magmatic protoliths of the Iberian Massif (such as the Ollo de Sapo volcanic Formation and the Guadarrama orthogneisses) suggest the repeated extraction of crustal melts from a common source during the Ordovician as previously proposed (e.g., Fernández Suárez et al., 2000). This crustal recycling would account for the volcanic arc signature of some of the samples. This signature was probably inherited by melting of pre-existing Neoproterozoic–Early Paleozoic calc-alkaline crust (Navidad et al., 2010; Martínez et al., 2010). Crustal recycling has been invoked to explain the volcanic arc affinity of the Early Ordovician Ollo de Sapo magmatic rocks (Díez Montes et al., 2010) in the Iberian Massif. It should be noted, however, that in the Pyrenees and the Catalan Coastal Ranges the described Early to Mid Ordovician magmatic event is younger than in the rest of the Iberian Massif, where Late Cambrian/Early Ordovician ages are common and coevally with this magmatism thick (up to 4,500 m) Early Ordovician detrital sediments were deposited (Pérez-Estaún et al., 1990; Valverde-Vaquero et al., 2005; Díez Montes et al., 2010). Moreover, in the Iberian Massif Late Ordovician magmatic activity is scarce (Valverde-Vaquero et al., 2007), and evidence of Ordovician deformation is limited (Martínez-Catalán et al., 1992). These features suggests that the Pyrenees and the Catalan Coastal Ranges were probably located in a different position on the northern Gondwana margin from that occupied by the rest of the Iberian Massif, and that both areas evolved differently following the Early Ordovician birth of the Rheic Ocean.

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CARBON ISOTOPE DEVELOPMENT IN THE ORDOVICIAN OF THE YANGTZE GORGES REGION (SOUTH CHINA) AND ITS IMPLICATION FOR STRATIGRAPHIC CORRELATION AND PALEOENVIRONMENTAL CHANGE

J. Cheng¹, Y.D. Zhang¹, A. Munnecke² and C. Zhou¹

¹ State Key Laboratory of Palaeobiology and Stratigraphy, Nanjing Institute of Geology and Palaeontology,
39 East Beijing Road, Nanjing 210008. ydzhang@nigpas.ac.cn

² GeoZentrum Nordbayern, University Erlangen-Nürnberg, Löwenichstrasse 28, D-91054 Erlangen, Germany.
axel.munnecke@pal.uni-erlangen.de

New carbon isotope ($\delta^{13}\text{C}_{\text{carb}}$) data of the Ordovician rocks from the Yangtze Gorges region, South China are presented. The Ordovician rocks are well exposed and dominated by carbonates intercalated with shales yielding abundant graptolites and shelly fossils. 534 samples were collected from five sections: Lianghekou–Chenjahe (158), Jingshan (85), Laomatou (89), Gaoluo (22), Houping (180). Some 263 samples in total have been processed for $\delta^{13}\text{C}$ values. The results suggest: (1) The $\delta^{13}\text{C}$ values are steadily negative with a slight decrease in Tremadocian to Dapingian, and increase progressively to positive values in the Darriwilian, and further positive in the Sandbian to early Katian. (2) The timing of the transition from negative to positive values falls basically within the middle Ordovician, but shows some variations among the five sections. This pronounced shift from negative to positive values may be an important indicator for significantly changing palaeoenvironments. (3) There is a prominent increase of $\delta^{13}\text{C}$ in mid-late Tremadocian in most of the five sections, and a negative $\delta^{13}\text{C}$ excursion near the Tremadocian/Floian boundary. (4) No significant Mid Darriwilian positive $\delta^{13}\text{C}$ excursions are recognized herein, except for one section (Jingshan) where a minor excursion is observed. (5) In the early Katian, a positive excursion of $\delta^{13}\text{C}$ is well recognized in all the five sections.

THE HIRNANTIAN-EARLY LLANDOVERY TRANSITION SEQUENCE IN THE PARANÁ BASIN, EASTERN PARAGUAY

C.A. Cingolani^{1,2}, N.J. Uriz¹, M.B. Alfaro¹, F. Tortello^{2,4}, A.R. Bidone¹ and J.C. Galeano Inchausti³

¹ División Geología, Museo de La Plata, Facultad de Ciencias Naturales y Museo de La Plata, Paseo del Bosque s/n, 1900, La Plata, Argentina. cingola@fcnym.unlp.edu.ar, nuriz@fcnym.unlp.edu.ar, malfaro@fcnym.unlp.edu.ar, andreabidone@fcnym.unlp.edu.ar

² CONICET.

³ Ministerio de Obras Públicas y Comunicaciones de Paraguay.

⁴ División Paleoinvertebrados, Museo de La Plata, Facultad de Ciencias Naturales y Museo, La Plata, Argentina.

Keywords: SW Gondwana, Paraná Basin, Eastern Paraguay, *Normalograptus persculptus* Biozone, Hirnantian-Llandovery transition.

INTRODUCTION

The Lower Paleozoic Itacurubí Group (Harrington, 1972) is exposed in the western border of the intracratonic Paraná Basin of eastern Paraguay (Fig. 1). This group (c. 350 m thick) includes from base to top the Eusebio Ayala, Vargas Peña and Cariy siliciclastic formations. It was traditionally assigned to the Llandovery (e.g. Harrington, 1950; Dyck, 1991; Benedetto et al., 1992; Benedetto, 2002; Galeano Inchausti and Poiré, 2006; Uriz et al., 2008a, 2008b and references therein) based on a marine fossil record mainly of graptolites, shelly fauna, and palynofacies assemblages. The new records of graptolites and some trilobites from the Eusebio Ayala Formation, exposed in clay quarries to the east of Asunción allow the comparison with other sequences bearing similar faunal associations known in west Gondwana. These records are discussed here, taken into account that the mentioned interval was a relevant paleobiogeographical time slice during the Lower Paleozoic (Cocks, 2001).

GEOLOGY

The great South American Paraná basin extends from the Asunción arch, as a western boundary near Paraguay River, to the south and southeast of Brazil, the central region of Uruguay, and northeastern Argentina (Milani et al., 2007). The geological evolution of this intracratonic basin was influenced by the geodynamics of southwestern Gondwana, with compressional stresses derived from an active convergent margin. During the Late Ordovician–Lower Devonian, the basin was filled by continuous and thick siliciclastic sequences named from base to top: Caacupé and Itacurubí groups. The latter represents a complete transgressive-regressive cycle, where the sandstones of the Eusebio Ayala Formation mark the base of this cycle, which is composed by yellowish, brownish, and reddish to purple micaceous sandstones with intercalated mudstone-siltstone beds with iron rich levels. The sandstones are laminated and wave-

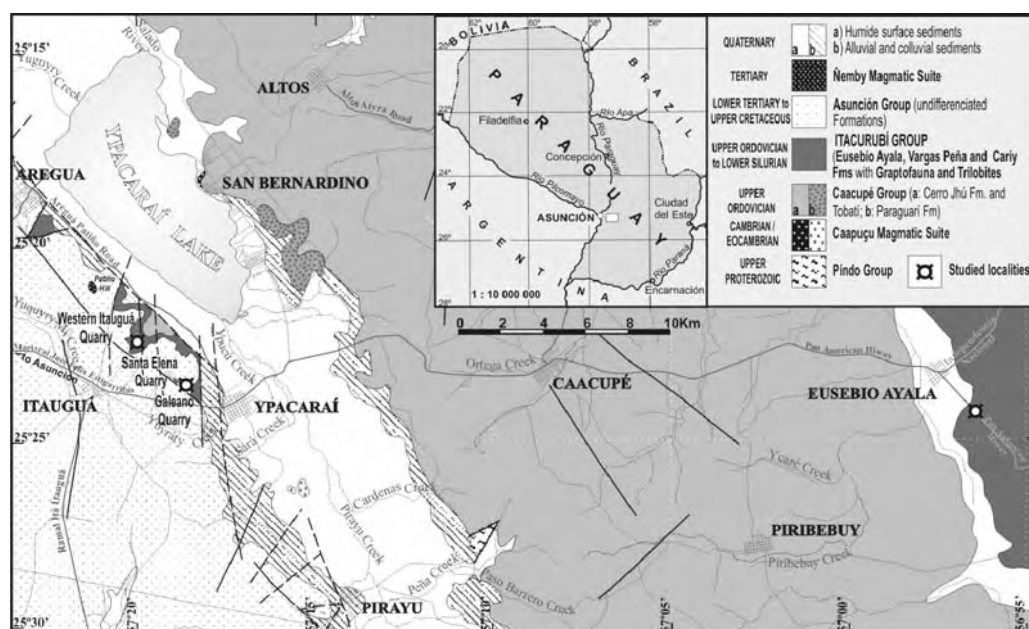


Figure 1. Geological sketch map of the Itauguá-Eusebio Ayala region (east of Asunción; based on Dionisi, 1999) with location of the studied outcrops.

cross stratifications are frequent. Reddish fine sandstone levels are fossiliferous and have bedding planes covered by detrital micas. Mudstones and siltstones show some bioturbation and wavy-linsen structures. The sandstone unit that bears the invertebrate fossils was deposited at the beginning of the transgression in a shallow marine environment during a flooding event described in the Paraná Basin (Milani et al., 2007).

GRAPTOLITE-TRILOBITE RECORD FROM THE HIRNANTIAN-LLANDOVERY INTERVAL

A low diversity graptolite fauna composed of *Normalograptus persculptus*, *Normalograptus normalis* and *Normalograptus medius*, in association with the trilobite *Mucronaspis sp.* was recently described from the Eusebio Ayala Formation, in beds also yielding brachiopods, bivalves and cephalopods (Alfaro et al., 2011). The identification of the *N. persculptus* Biozone allows us to assign the studied stratigraphical levels close to the Hirnantian-Llandovery interval (Fig. 2). As we can see in the composite graptolite-trilobite range chart, the Hirnantian-Rhuddanian transition shows a low-diversity graptolite assemblage in the upper part of the Eusebio Ayala Formation. During the Early Llandovery, the record of the first Silurian graptolites and trilobites (Tortello et al., 2008a, 2008b) was accompanied by the arrival of a diverse fauna, during a flooding event that improved the environmental conditions on the Paraná Basin under a warm-water influx, favoring a biological colonization. *Climacograptus innotatus brasiliensis*, an apparently endemic South American graptolite (Underwood et al., 1998) was also recorded in the succession. The sequence of graptolite taxa would reflect distinct faunal events, also known in other Gondwanan outcrops, related to a dramatic change of environmental conditions during the Hirnantian-Llandovery transition.

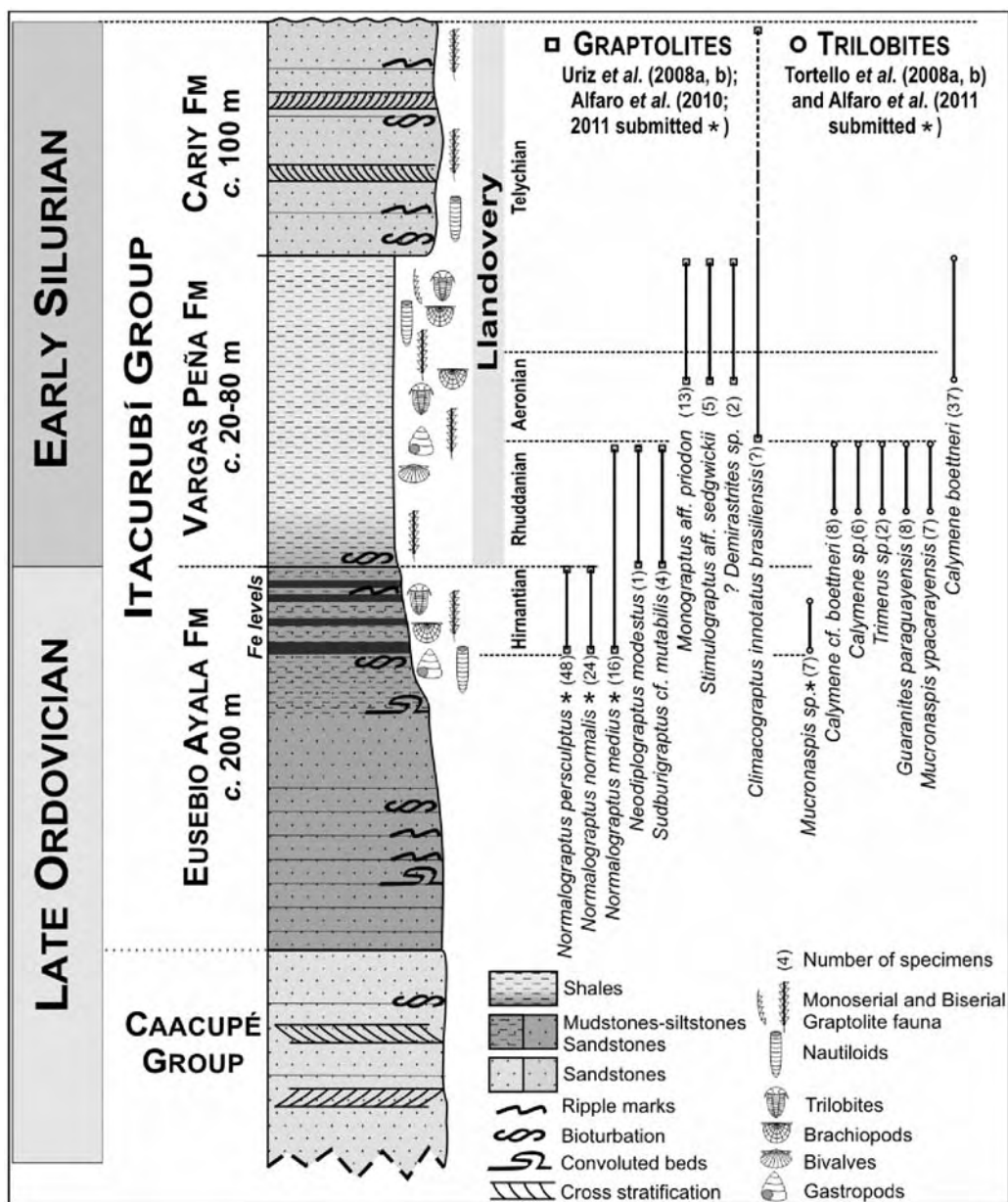


Figure 2. Composite graptolite-trilobite range chart for the Ordovician-Silurian boundary interval, in the Itacurubí Group of eastern Paraguay.

CORRELATION WITHIN SW GONDWANA

In Western Gondwana, a large ice sheet is assumed to have covered most of Africa and South America. The South Pole would have been located in west-central Africa at the time (Underwood et al., 1998; Cocks, 2001; Ghienne, 2003; Legrand, 2009). Late Ordovician glacial deposits are found in the Pakhuis Formation in the Western Cape Fold Belt, South Africa (Young, 2004). The black shales of the Soom unit have been assigned to the Late Ordovician *N. persculptus* graptolite Biozone.

In South America, there is evidence of the record of this glacial event in the 'Central Andean Basin', in the Precordillera region (Cuyania terrane), and in the Amazonas, Parnaíba and Paraná intracratonic basins. In Perú, Bolivia and northwestern of Argentina, as part of the 'Central Andean Basin', identification of diamictites and an erosional surface near the Ordovician-Silurian boundary characterizes the setting of a glaciogenic environment that would have extended to Silurian times (Benedetto et al., 1992; Díaz-Martínez, 1997; Díaz-Martínez and Grahn, 2007). These glaciogenic conditions in northwestern Argentina are recognized in the Late Ordovician levels of the Zapla Formation and in the lowermost levels of the Lipeón Formation (Monteros et al., 1993), as well as in their equivalent units from southern Bolivia (Schönian et al., 1999). In the Precordillera region (part of the Cuyania terrane) of San Juan, Argentina, tillite levels were recorded at the base of the Don Braulio Formation (Peralta and Carter, 1999). Benedetto (1986) recognized a brachiopod association at the base of this unit and referred it to the Hirnantia fauna, while Peralta and Baldi (1990) described *N. persculptus* towards the top of the same sequence. Also in the regions of Talacasto and Cerro del Fuerte-Cerro La Chilca (San Juan) it was possible to define the boundary between both systems by the record of the *N. persculptus* and *P. acuminatus* biozones (Cuerda et al., 1988; Astini and Benedetto, 1992; Rickards et al., 1996). On the Río de la Plata craton (Tandilia System, Argentina) the presence of a diamictite level was mentioned in the Balcarce Formation; Zimmermann and Spalletti (2009) based on mineralogical provenance studies suggested a possible Hirnantian age for this glacial event. In the Amazonas and Parnaíba intracratonic basins, northeastern Brazil, there are potential tillite deposits referred to the Late Ordovician, suggesting a glaciogenic influx.

For the western border of the Paraná Basin (Paraguay), in the quarries bearing *N. persculptus*, *N. medius* and *N. normalis* within the Eusebio Ayala Formation, typical glacial sediments were not found, although tillites were described from drill cores (Figueredo, 1995). Fifty meters of tillites were described at the base of these cores, followed by 150 m of sandstones with conglomeratic levels, and 200 m of fine sandstones with interbedded shales and claystones. Preliminary palynostratigraphic studies for the interval between -198 to -385 m reveal Upper Ordovician-Lower Silurian ages for the section (González Nuñez et al., 1999), while Steemans and Pereira (2002) described an interesting Llandovery palynomorph assemblage coming from three boreholes from central Paraguay. A graptolite association collected in the upper levels of the Eusebio Ayala Formation (Uriz et al., 2008a) indicated a Rhuddanian age. Another record that proves the continental glaciation in other sectors of the Paraná Basin is revealed in the Ponta Grossa structural arch in the Apucarana sub-basin (Brazil) where the less than 20m-thick Iapó Formation is essentially composed of diamictites covering large areas, and included in the 'Río Ivai Supersequence' (Milani et al., 2007).

The Paraguayan lower Itacurubí Group that documented the Ordovician-Silurian transition (Fig. 2) and recorded graptolites and other invertebrate groups, could be a suitable sequence for high-resolution studies on stable isotope chemostratigraphy. These may constraint shallow-water environmental changes associated with a mass extinction in Western Gondwana, and may be used to correlate the organic-inorganic carbon isotope excursion models known from other paleocontinents such as Baltica, Laurentia, South China and North Gondwana (Finney et al., 2007).

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DISTAL EFFECTS OF GLACIALLY-FORCED LATE ORDOVICIAN MASS EXTINCTIONS ON THE TROPICAL CARBONATE PLATFORM OF LAURENTIA: STROMATOPOROID LOSSES AND RECOVERY AT A TIME OF STRESS, ANTICOSTI ISLAND, EASTERN CANADA

P. Copper¹, H. Nestor² and C. Stock³

¹ Loupicoibas, 46220 Prayssac, France. pcopper@laurentian.ca

² Institute of Geology, Tallinn University of Technology, Tallinn, Estonia. hnestor@gi.ee

³ Department of Geology, University of Alabama, Tuscaloosa, USA. cstock@geo.ua.edu

The Anticosti carbonate platform-ramp that developed from Katian and Hirnantian through Llandovery (Early Silurian) time was situated in a tropical setting at ca. 25° latitude south of the equator, on the southern side and eastern flanks of Laurentia. This was separated from glaciated Gondwana to the south and subtropical Baltica to the east by relatively wide, 2500 km+ oceans. Sediments were deposited in a mid to outer shelf monsoonal environment at, or periodically below, typhoon wave base.

During the late Katian (Richmondian), the columnar, 'tree trunk-like' aulacratids (order Labechiida) were the volumetrically dominant stromatoporoids on Anticosti, some 3-4 m long and up to ca. 40 cm in diameter, anchored or rooted in the carbonate sediment by early cemented 'fence-posting' strategies, but usually found as storm-redeposited, broken fragments (Cameron and Copper, 1994). These are now assigned to the genus *Aulacera* Plummer 1843, who identified them as nautiloid conchs, but were a decade later described by Billings (1857) as a new genus, *Beatricea*, naming two species from the Vaureal Fm, and pigeon-holed taxonomically as 'Plantae'. These *Aulacera*, with large central, sometimes off-set megacysts, and layers of laterally concentric microcysts penetrated by pillars, are typical of the upper Vaureal Fm. They vanished in a first extinction wave of Katian benthic faunas, below the base of the Ellis Bay Fm, named by Schuchert and Twenhofel (1910) for strata of post-Richmondian (post-Katian) and pre-Silurian age sediments of Anticosti, and identified as their new 'Gamachian' Stage. The only tabular to domal stromatoporoids of the Vaureal Fm belong to the rare labechiid genus *Pseudostylodictyon*, found in reefs of the upper Katian Mill Bay Mbr, ca. 50 m below the top of the formation: it also disappeared in the first end-Richmondian extinction wave.

During the Hirnantian, the columnar aulacratids were initially still the most abundant volumetrically, but a new genus, with large axial cysts, surrounded by layers of micro-cysts, in turn enveloped by an outer layer of laminae penetrated by numerous dense pillars, provisionally named '*Quasiaulacera*', arrived in the Ellis Bay strata, replacing *Aulacera*. The final Ordovician aulacratids, found on the tops and flanks of Laframboise Mbr coral patch reefs, include a genus with only axial macro- and outer microcysts (no pillars), and '*Cryptophragmus*'. In the same Hirnantian patch reefs are common, but small-sized domical clathrodictyids *Clathrodictyon* and massive *Labyrinthodictyon* (mostly in shallow subtidal oncolitid facies), and the actinodictyid *Ecclimadictyon*, marking the advance and take-over of 'Silurian-type' stromatoporoids. Except for the aulacratids, no other families of stromatoporoids went extinct at the O/S

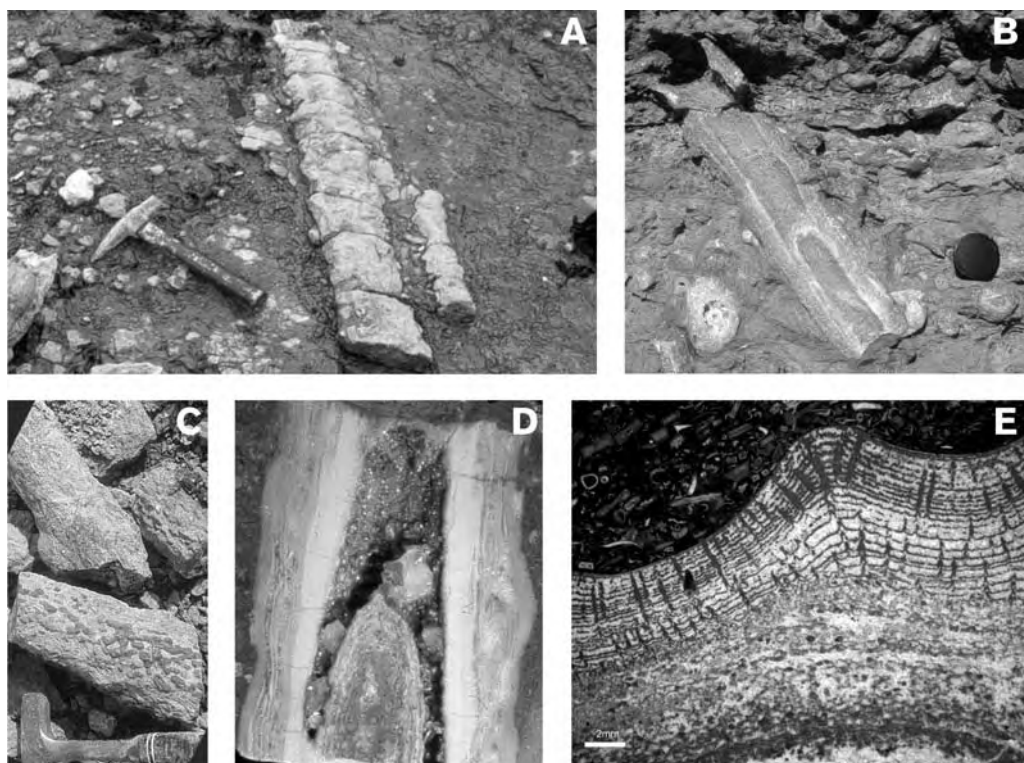


Figure 1. A, Bedding plane view of recumbent *Quasiaulacera* sp. B on the tidal flat of the west side of Ellis Bay, Anticosti Island, eastern Canada. Photo taken at low tide of a specimen ca. 1.2 m long, 20 cm diameter, broken away from its base during a storm event; Lousy Cove Mbr equivalent, Ellis Bay Fm, Hirnantian, locality A1175, NTS 22H/16, 00280:17530. B, Coastal bluff on west side of Lousy Cove, eastern Anticosti Island, showing a jumbled, storm-disturbed aulaceratid 'forest' bed at the base of the Prinista Mbr, Ellis Bay Fm; locality A315, 12F/5, 80920:64780 [this site is periodically buried by beach gravels after severe winter storms]. Most of the specimens in this ca.1 m thick bed are broken fragments, with rare specimens in life position, or tilted, and with numerous 'in situ' attachment bases. Specimen at oblique angle shows the central cavity with megacysts. The specimens shown are mostly representative of an earlier Hirnantian species of *Quasiaulacera* sp. A. C, Broken samples, largely of a single specimen of a ca. 1 m long *Aulacera nodulosa* (Billings, 1857), arranged for photography on the outcrop bedding plane surface. Beacon road outcrop, 3.4 km W of Naticotek road junction, locality A1286, Mill Bay Mbr, upper Vaureal Fm, late Katian, 12E/10, 35250:86480. D, Polished slab of longitudinal section of the base of *Quasiaulacera* sp. B from the type locality on the tidal flats at the mouth of Laframboise Creek: such 'bases' remained in situ, cemented on the former seafloor. Lousy Cove Mbr, Ellis Bay Fm ca. 10 m below the reefal Laframboise Mbr, Ellis Bay Fm, Hirnantian; locality A972, 22H/15, 97550:17920. E, Thin section photomicrograph of *Quasiaulacera* sp. B (holotype specimen GSC129346), showing the unique nature of this aulaceratid skeleton, consisting of a 1-2 cm thick outer layer of laminae and pillars, covering a layer of microcysts, and an axial column of megacysts, unknown from other aulaceratids (from locality A972, as above).

boundary on Anticosti, defined at the top of the patch reef and inter-reef crinoidal/calcarene/oncolitic facies, with sharp δO_{18} and δC_{13} excursions indicating oceanic cooling. There is no evidence for anoxia/hypoxia at the boundary, and none for emergence or even intertidal conditions (no karst, no paleosols, no stromatolites, no isotopic signatures for fresh water cements, etc.). O/S stromatoporoid biodiversity losses nowhere match those seen at the stepped Frasnian/Famennian mass extinctions, and are paralleled in other benthic groups from Anticosti such as brachiopods, crinoids, nautiloids, and corals.

The slow recovery of stromatoporoids in the overlying Rhuddanian (Early Silurian) Becscie Fm limestones was impoverished in the lower Fox Point Mbr (ca. 35 m thick): only *Clathrodictyon* is present, and skeletons were very rare, small, spheroidal forms barely reaching 5 cm diameter, a phase estimated to have lasted ca. 0.5 myr. A modest stromatoporoid recovery began in the upper Becscie Fm (Chabot Mbr) with large skeletons of *Ecclimadictyon* and *Clathrodictyon* of up to 50-70 cm diameter, and the return of a labechiid, *Pachystylostroma*, at the top of the Becscie Fm. Diversity expanded modestly to 5 genera in the late Rhuddanian Merrimack Fm (some 1.2 myr after the O/S events) and the arrival of *Forolinia*, and two new genera, branching *Desmidodictyon* and domical *Camptodictyon*. Reefs did not return to the Anticosti platform until deposition of the late Aeronian East Point Mbr of the Jupiter Fm (ca. 3-4 myr post O/S), and with it a fully diverse stromatoporoid expansion to 6 genera (10 spp.). Stromatoporoids did not play a volumetrically, nor biodiversity dominant role in reefs of the Late Ordovician and Early Silurian of Anticosti, compared to tabulate and rugose corals.

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LATE ORDOVICIAN GLACIAL DEPOSITS IN VALONGO ANTICLINE (NORTHERN PORTUGAL): A REVISION OF THE SOBRIDO FORMATION AND A CONTRIBUTION TO THE KNOWLEDGE OF ICE-MARGINAL LOCATIONS

H. Couto¹ and A. Lourenço²

¹ Universidade do Porto, Faculdade de Ciências, Departamento de Geociências, Ambiente e Ordenamento do Território, Centro de Geologia, Rua do Campo Alegre 687, 4169-007 Porto, Portugal. hcouto@fc.up.pt

² Universidade do Porto, Reitoria, Centro de Geologia, Praça Gomes Teixeira, 4099-002 Porto, Portugal. aafonso@reit.up.pt

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INTRODUCTION

In this work we made a revision of Sobrido Formation of Romano and Diggens (1974) tacking in account a more complete succession with lateral facies variation not completely exposed in the type section described by these authors. Romano and Diggens (1974) interpreted this formation to be, at least in part, Caradoc (i.e. Katian) in age based on the similarities with the Upper Ordovician succession of Cáceres (Spain). By contrast, Robardet (1981) suggested a possible early Silurian age for the “tillites” of Northern Portugal. Later, Robardet and Doré (1988) noted strong similarities with the Upper Ordovician glacial deposits of North Africa. Oliveira et al. (1992), meanwhile, considered the upper member of Sobrido Formation to extend from the Upper Ashgill or Lower Silurian. On the basis of an unconformity that can be mapped throughout the area, and noting that the Caradoc and part of Ashgill were missing from the succession, these latter workers interpreted a discordance that was probably caused by the late Ordovician glaciation. By 1997, a more specific Hirnantian age was suggested for the Sobrido Formation by Couto et al. (1997). Other studies concerning the Upper Ordovician of Valongo were published by Ribeiro et al. (1997), Sá et al. (2006, 2009) and Couto and Lourenço (2008).

REGIONAL GEOLOGICAL SETTING

The Valongo Anticline placed east of Porto extends approximately 90 km from Esposende (north Valongo) to Castro Daire (south Valongo), and is located near the southwestern boundary of the axial part of the Hercynian fold belt in the Central Iberian Zone.

The basement comprises metasediments of marine origin ranging from probable Neoproterozoic and Cambrian through Devonian (Couto, 1993). Younger, Upper Palaeozoic metasediments of continental

origin (Carboniferous) occurs to west of Valongo Anticline (Wagner and Sousa, 1983). The metasediments are intruded by Hercynian granites (Ribeiro et al., 1987).

Romano and Diggins (1974) differentiated three formations in the Ordovician succession of the Valongo Anticline. In ascending stratigraphic order, these are the Santa Justa Formation (Lower Ordovician), the Valongo Formation (Middle Ordovician) and the Sobrido Formation (Upper Ordovician). The Sobrido Formation has its type section about 750m WNW of Santa Justa which we illustrate in the present manuscript. The lower member consists of massively bedded, whitish-grey quartzites. The poorly bedded deposits at its base are overlain by more flaggy beds of quartzite alternating with thin grey siltstones and mudstones. Above, laminated cross-bedded siltstones occur, and in the top of the lower member, two quartzite beds with large load structures at their bases are present. The upper member begins with thinly laminated and cross-bedded dark grey mudstones up to 1.5m thick that are overlain by pebbly greywackes. Romano and Diggins (1974) considered three types of greywacke, namely (1) pebbly greywacke with angular to sub-rounded pebbles of sandstone, siltstone and granite, (2) those with large, ellipsoidal, grey calcareous concretions and (3) those with conspicuous banded weathering.

LATE ORDOVICIAN DEPOSITS IN VALONGO ANTICLINE

In the present study some areas along Valongo Anticline, between Esposende (north) and Arouca (south), including the type section of Sobrido Formation studied by Romano and Diggins (1974) in Valongo, have been selected.

A detailed cartographic and petrographic study allowed a better knowledge of the Sobrido Formation described by these authors. In this work two members were considered as well.

The lower member of Sobrido Formation begins with massive immature whitish-grey quartzite unconformably overlying the nodule-bearing schists of Valongo Formation suggesting a hiatus (paraconformity) with the lack of upper Darriwilian, Sandbian and Katian strata. Massive quartzites alternate with thin-bedded grey siltstones and mudstones exhibiting climbing-ripples, being again present to the top of this sequence.

Occasionally the transition between massive immature whitish-grey quartzites of the lower succession and the overlying deposits of the upper succession seems gradual but most often a ferruginous level accounts for an erosive contact. In this case a thin laminated mudstone horizon with thinly laminated black and white layers, interpreted as varves (Fig. 1A) is present at the base of the upper member. To the top of this deposit a thin millimetric shelly bed appears (Couto and Lourenço, 2008). Overlying these mudstones, sandstones clasts bearing can be observed (Fig. 1B). This upper member is dominated by diamictites, also occurring quartzites, conglomerates and schists.

The diamictites (sandstones clast-bearing) can show different facies. They can be massive or laminated, clast poor or clast-rich, with an argillaceous, siltitic or arenaceous matrix. Laminated diamictites usually overly massive diamictites and are in general clast-poor.

Massive diamictites, exhibit often thin ferruginous horizons with iron and occasionally manganese, phosphate and chamosite oolitic. These horizons sometimes correspond to sedimentary layers, but often correspond to remobilized iron controlled by joints or outlining banded weathering. Soft sedimentary deformation affects these clast bearing sandstones.

The quartzite layers observed to northeast of Valongo in Sobrado area, are very variable in width. Quartzite can be more or less laminated, poorly- sorted, with sharp contacts to the confining diamictites.

Groove-casts are present in the bedding planes of some quartzitic beds. Quartzites occasionally show hummocky cross-stratification. Conglomerates occur intercalated in this sequence, evidencing graded bedding. Sometimes conglomerate have dominantly shaly clasts.

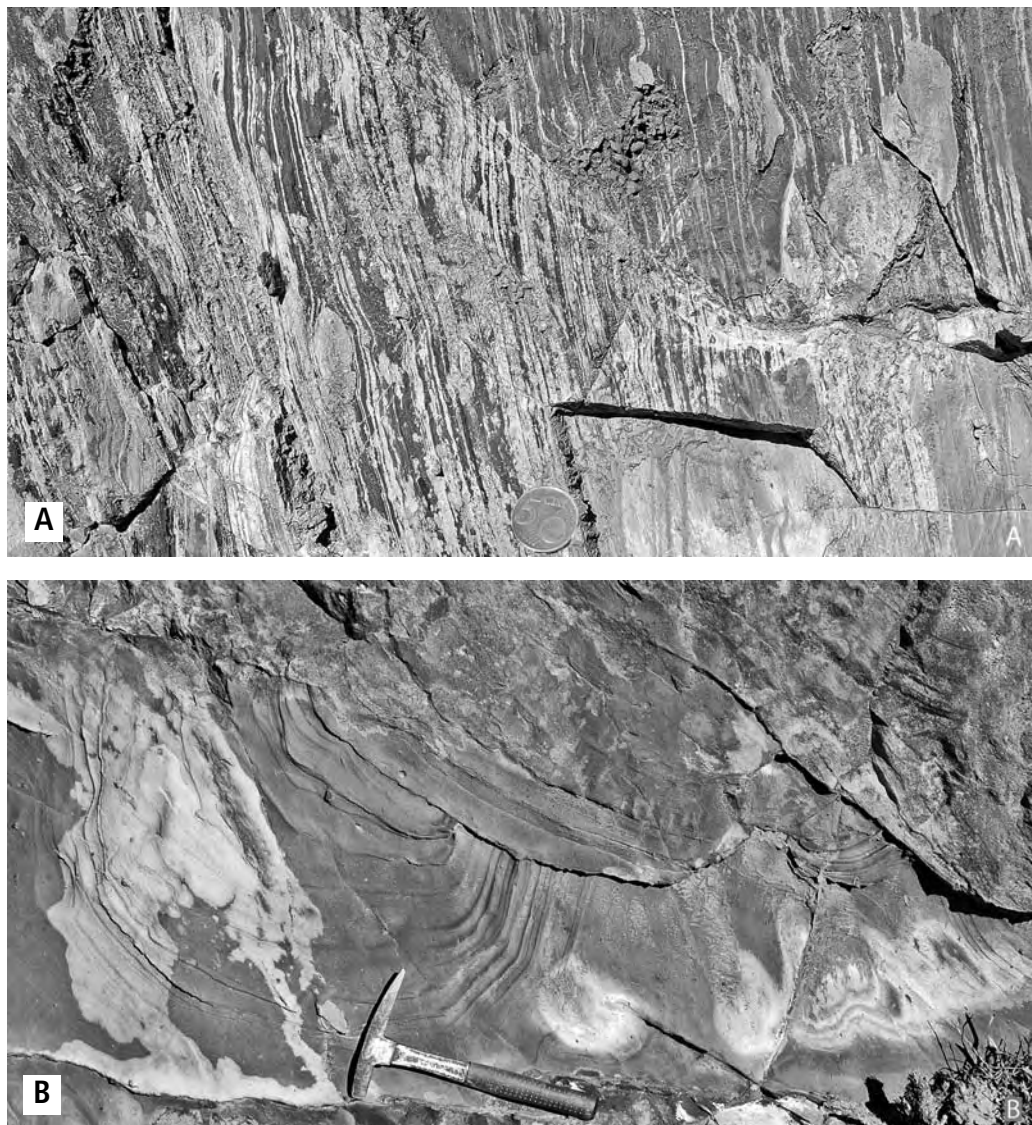


Figure 1. Late Ordovician in Valongo (type section of Sobrido Formation described by Romano and Diggins, 1974). A, Laminated mudstones, interpreted as varves, of upper member unconformably overlying the massive quartzites of lower member (visible elements are clasts). B, Sandstones clast-bearing with ferruginous horizons outlining banded weathering in upper member.

To the west another type of conglomerate was observed in the middle of the diamictitic sequence, including rounded quartz and quartzite clasts, supported by a ferruginous shaly matrix rich in muscovite and with dark oxidized nodules.

The contact between Ordovician and Silurian strata is variable. Black quartzites are sometimes present in the Silurian base. More frequently the upper Ordovician is in contact with black-shales or dark grey schist bearing Middle to Upper Llandovery graptolites (Romariz, 1962).

DISCUSSION

According to Romano and Diggins (1974), the basal quartzites of Sobrido Formation were deposited in shallow water, high-energy conditions. These quartzites overlying erosionally schists of Valongo Formation deposited during the maximum of glaciation. Thin laminated mudstone (varves) with alternating black and white layers developed at the base of the upper succession of Sobrido Formation represent probably the beginning of deglaciation.

According the same authors the origin of impure pebbly sandstones overlying basal quartzites, was more difficult to understand but was considered the result of turbidity current or submarine slumping. Nevertheless Romano and Diggins (1974) citing Schermerhorn and Stanton (1963) and Destombes (1968), discussed the hypothesis of a glacial origin for these rocks by comparison with the rocks of Upper Ordovician of Africa.

Massive diamictite exhibiting ferruginous horizons with manganese, phosphate and oolites of chamosite, are interpreted to have formed near surface conditions with oxidation during oscillatory sea-level low-stands, which allowed the drainage of fresh water from continent creating oxidizing conditions with the formation of Fe and Mn with terrigenous contribution. According Ghienne et al. (2000) the Mn-rich crusts interbedded within glacially-related Hirnantian deposits of Sardinia result from starved sedimentation in isolated sub-basins that resulted from low glacio-eustatic sea-levels on the North-Gondwan platform. According to Young (1989) the widespread Ashgill ooidal chamositic ironstones of the Ibero-Armorican region are related to cool if not glacial climate.

From east to west of the area under study, a lateral facies variation seems to occur. A more ice-distal facies dominated by diamictites occurring in west changes to a more ice-proximal facies with diamictites, conglomerates, quartzites and schists to east around Sobrado (NE Valongo). In the proximal facies of the upper member, quartzites with hummocky stratification indicate a deposition in a shallow shelf setting. Conglomerates within the upper member, erosively based and occurring interbedded with schists in rhythmic sequences may be interpreted as ice-contact front fans (terminoglacial fans canalized deposits). This ice-proximal facies correspond to a great part of Sobrado Formation of Pereira and Ribeiro (1992) (excluding upper member), until now considered of Silurian or Devonian age.

Conglomerates with shally ferruginous matrix and oxidized nodules that occur to the west of the area are probably associated with subaquatic debris-flow.

Resemblances can be noted between these facies and deposits described by Ghienne (2003), Ghienne et al. (2007), Le Heron (2007), Le Heron et al. (2007) in North Africa, namely ice contact deposits, tidal deposits, glaciomarine deposits, turbidites and debris-flow conglomerates.

According the interpretations of Fortuin (1984) about Late Ordovician deposits in the Sierra de Albarracín (Spain), grounded ice sheets were present in the Iberian Peninsula. Based on the discovery of tunnel valleys in Spain, and on the co-occurrence of soft sediment striated surfaces, Gutiérrez-Marco et al. (2010) have proposed that the Late Ordovician African ice sheet may hence have reached Europe.

CONCLUSION

The Sobrido Formation, actually considered as Hirnantian in age, contains the record of glacial sedimentary processes and can be split into two members. Two members were also considered by Romano and Diggens (1974), but the present work allows to define a more complete upper member with successions not exposed in the areas studied by these authors.

We emphasize the consideration of a genesis associated with glaciers at least for the pebbles of Sobrido Formation, according these new data. The glacial record involves depositional successions of both ice-proximal and ice-distal areas, reflecting diverse environments. Some of these deposits evidence continental origin related with the North Gondwana ice sheet.

The detailed mapping also allowed recognising that a part of "Sobrado Formation" of Pereira and Ribeiro (1992) is really part of Sobrido Formation.

It is hoped that the description and re-interpretation of Sobrido Formation will contribute to our understanding of the palaeogeography of the Late Ordovician, particularly with respect to the ice-marginal locations.

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ABNORMAL ACRITARCHS IN THE RUN-UP OF EARLY PALAEOZOIC $\delta^{13}\text{C}$ ISOTOPE EXCURSIONS: INDICATION OF ENVIRONMENTAL POLLUTION, GLACIATION, OR MARINE ANOXIA?

A. Delabroye¹, A. Munnecke², T. Servais³, T. Vandenbroucke³ and M. Vecoli³

¹ Université Paul Sabatier, UMR 5563 CNRS, L.M.T.G., 31400 Toulouse, France. delabroy@lmtg.obs.mip.fr

² GeoZentrum Nordbayern, Fachgruppe Paläoumwelt, Universität Erlangen-Nürnberg, Loewenichstrasse 28, 91054 Erlangen, Germany. axel.munnecke@pal.uni-erlangen.de

³ Université de Lille 1, FRE 3298 CNRS, Laboratoire Géosystèmes, SN5, Cité Scientifique, 59655 Villeneuve d'Ascq, France. Thomas.Servais@univ-lille1.fr, Thijs.Vandenbroucke@univ-lille1.fr, Marco.Vecoli@univ-lille1.fr

The Late Ordovician and Silurian are characterised by several pronounced, globally recognised, short-lived positive $\delta^{13}\text{C}_{\text{carb}}$ excursions. At least three of them (Hirnantian, early Wenlock and late Ludlow) exceed +5‰ and thus belong to the strongest excursions of the Phanerozoic. Although the excursions share many geochemical, sedimentological, and palaeontological characteristics there is to date no general agreement on the steering mechanisms. One of the most striking feature is the fact that the onsets of the excursions correlate with extinctions of several groups of organisms, especially conodonts, graptolites and trilobites, but also acritarchs, chitinozoans, ostracods, brachiopods, and corals. Sometimes, the first extinctions even precede the excursions indicating that some as yet totally unknown processes occurred prior to the $\delta^{13}\text{C}$ increase, and that both first extinctions and $\delta^{13}\text{C}_{\text{carb}}$ excursions are likely the result of these enigmatic processes.

Own investigations in the Hirnantian (latest Ordovician) have shown that the onset of the major $\delta^{13}\text{C}$ excursion (HICE) is characterised by very high abundances of acritarchs showing abnormal, teratological growth forms. High abundances of teratological growth form in modern marine protists are commonly observed in environments with high degree of, e.g., organic or metal pollution, ash pollution, eutrophication, or even thermal and radioactive waste of nuclear power plants. In the fossil record, however, it is much more difficult to attribute abnormal growth forms to specific environmental factors. Abnormal acritarchs, for example, have only been rarely described in the literature, but a critical literature survey implies that they are somehow related to the global carbon cycle, i.e. to carbon isotopic composition of the ambient sea water. In the present paper we present a review of published reports of abnormal Ordovician and Silurian acritarchs, and we correlate the occurrences to the global $\delta^{13}\text{C}$ curve. We will document that high abundances of teratological growth forms of acritarchs are often related to the run-up of $\delta^{13}\text{C}$ excursions, and finally we discuss possible environmental implications.

GEOCHEMISTRY OF LOWER PALAEOZOIC SHALES. A CASE STUDY IN A SECTOR OF THE IBERIAN VARISCIDES

I. Dias da Silva¹, E. González-Clavijo¹, P. Barba², M.I. Valladares² and J.M. Ugidos²

¹ Instituto Geológico y Minero de España, Azafranal 48, 37001 Salamanca, Spain. if.silva@igme.es, e.clavijo@igme.es

² Dept. Geología, University of Salamanca, Plaza de la Merced s/n, 37008 Salamanca, Spain.
barba@usal.es, valla@usal.es, jugidos@usal.es

Keywords: Geochemistry, shales, Cambrian, Ordovician, Central Iberian Zone, Spain, Portugal.

INTRODUCTION

Geochemical data may be relevant to study detrital rocks especially in areas where fossil record is absent or misleading. In these cases distinctive parameters can be proposed as possible criteria to discriminate units of different ages, used for stratigraphic correlation and to characterise sedimentary environments.

Fine-grained rocks are those that best reflect the chemical features of source areas (Condie, 1991; Cullers, 1995). The aim of this work is to apply geochemical data of shales to contribute to unravel some geological problems related to differences in age interpretations. Also as a practical test on the use of these geochemical characteristics in one area where strong deformation is present and fossil record is absent or ambiguous.

The present study was carried out in an area of the Central Iberian Zone (CIZ) along the Spanish-Portuguese border (Duero/Douro River Canyon Section) (Fig. 1). This zone yields out some interesting results with practical relevance on the recognition of lithostratigraphic units on the basis of a combination of geological and geochemical data.

Here presented data are part of a bigger sampling survey realized through the last decade by this research team. Previous results in the Spanish Central Iberian Zone – in areas with a good stratigraphic, paleontological and structural control – led to the identification of different geochemical clusters that successfully separate the Upper Neoproterozoic from the Cambrian (Ugidos et al., 2010, Ugidos et al., 1997) and this one from the Lower Ordovician (Valladares et al., 2009).

GEOLOGICAL SETTING

The study area (Fig. 1) belongs to the Central Iberian Zone of the Iberian Massif. The Lower Palaeozoic mostly consists of siliciclastics with local carbonate and calcsilicate rocks with scarce diagnostic fossils. The

high shear deformation overprints almost all fossil record, making difficult identification and stratigraphic correlation of lithological units (in some cases several strong regional tectonic foliations were observed).

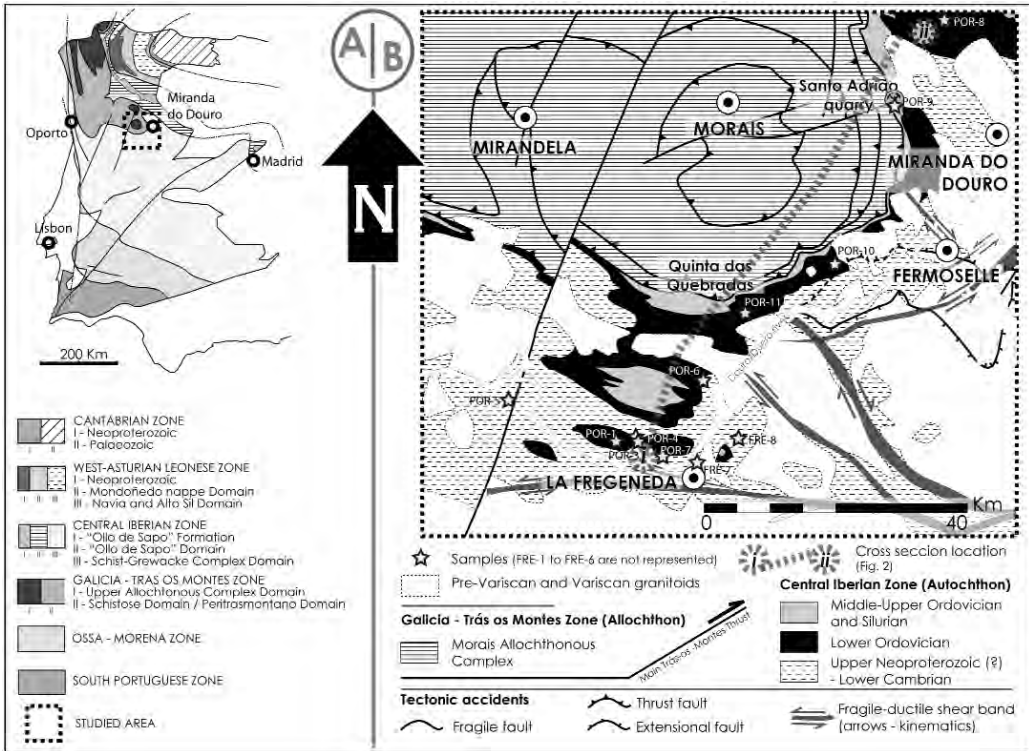


Figure 1. Situation map of the study area. A, Position on the Iberian Massif (after Vera, 2004); B, Simplified geologic map (modified from Mapa Geológico de España, escala 1:1.000.000) with main tectonic structures, sample location and main stratigraphic groups. See cross section i-ii on Fig. 2.

Two major Lower Palaeozoic unconformities were identified (Fig. 2): one located at the base of the Lower Ordovician siliciclastic sequence (Marão and Vale de Bojas Formations; Sá, 2005; Sá et al., 2003, 2005) and other at the base of the Upper Ordovician limestones (Santo Adrião Formation; Sá, 2005; Sá et al., 2003, 2005) and possibly also under the corresponding detrital rocks (Maceiras/Guadramil Formations; Sá, 2005; Sá et al., 2003, 2005). Both are witness of the partial erosion of the units that lie below. The former is an angular unconformity locally marked by a basal conglomerate (Quinta da Ventosa/Vale de Bojas Formations; Sá, 2005; Sá et al., 2003, 2005; Pereira et al., 2006) usually deposited on richly bioturbated Lower Cambrian sediments [Desejosa Formation; Pereira et al. (2006), and Mazouco formation (this work)]. The second is marked by a cartographic unconformity that affects all the Middle Ordovician record (Moncorvo Formation; Sá, 2005; Sá et al., 2003, 2005) and a part of the Lower Ordovician in the Santo Adrião quarry zone (Dias da Silva, 2010; Dias da Silva et al., 2010) and possibly also in Quinta das Quebradas area, in Portugal.

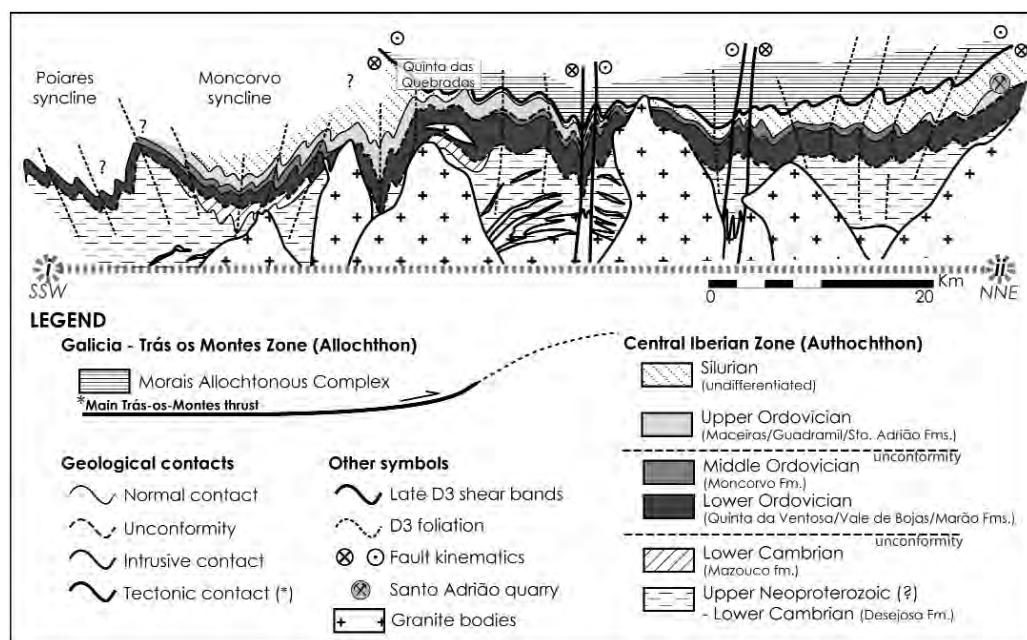


Figure 2. Schematic cross section i-ii (see Fig. 1) with main stratigraphic units, unconformities (and other geologic contacts) and structures. The newly recognized upper Lower Cambrian unit (Mazouco formation) was represented at top of Desejosa Formation. It is possible to see how lateral continuity of the Middle and Upper Ordovician units changes dramatically, possibly related with a cartographic scale Upper Ordovician unconformity (see text).

GEOCHEMISTRY OF LOWER PALAEOZOIC SHALES

Seventeen shale samples were collected in several points of the study area (Fig. 1B – samples FRE-1 to 6 were collected outside this map coverage). Chemical analysis were performed at Service d'Analyses de Roches et Minéraux of the CRPG-CNRS (Nancy, France) by AES (major elements) and ICP-MS (trace elements).

Chemical analyses of shales from well known Upper Ordovician (12 samples), Lower Ordovician (12 samples) and Lower Cambrian Series (24 samples) in Spain are used here as reference compositions in diagrams (Fig. 3) (Ugidos et al., 2010; Valladares et al., 2009).

Geochemical clusters are better expressed in TiO_2 -MgO, TiO_2 - Fe_2O_3 /MgO and TiO_2 -MgO/ TiO_2 diagrams (Fig. 3). To ease data interpretation of the newly collected samples we plotted them in these diagrams.

With the exception of three samples, all of them group clearly in two populations: on the Lower Cambrian and on the Lower Ordovician. This means that no sample is older than Lower Cambrian, even those collected well inside the Slate and Greywacke Complex of the CIZ. Some differences are observed between the Portuguese and the Spanish Lower Ordovician shales, which could be related with paleogeographic constraints. The three samples that fit less perfectly (FRE-8, POR-5 and POR-9) present an anomalous and sometimes ambiguous distribution and further work will be necessary to understand this behaviour. Nevertheless, one single sample of dark pelite from the top of the calcareous Kralodvorian unit in Santo Adrião quarry (sample POR-9) shows a better fit for the Upper Ordovician. In the case of sample

FRE-8 a basal Ordovician age fits better due to field criteria – it belongs to a detritic series identical to the local Lower Ordovician series. The age of these pelites will soon be tested with complementary data.

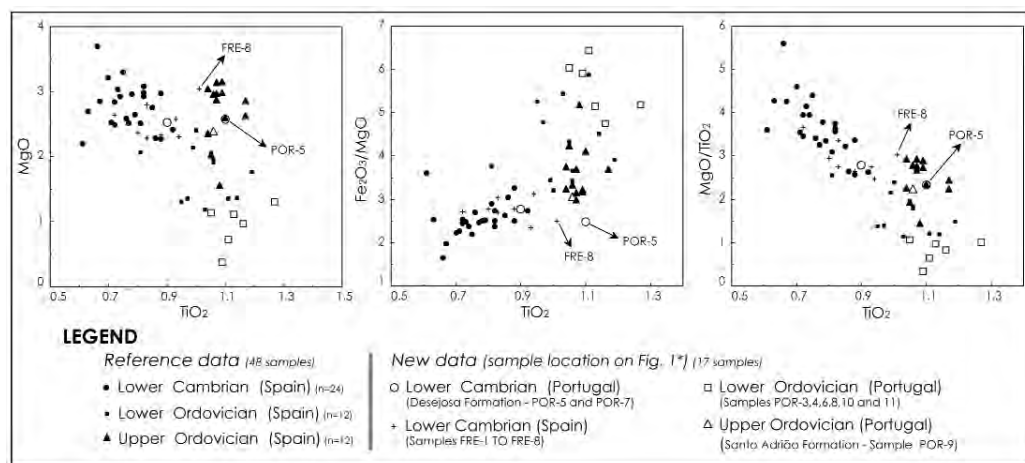


Figure 3. Data plots on TiO_2 -MgO, TiO_2 - $\text{Fe}_2\text{O}_3/\text{MgO}$ and TiO_2 -MgO/ TiO_2 diagrams. The apparently anomalous samples are marked with an arrow and the corresponding label.

CONCLUSIONS

The geochemical data of samples collected on Desejosa Formation, suggest that its age cannot be older than Lower Cambrian (Fig. 3). This assumption is supported by intense bioturbation frequently observed in the detrital rocks of this Formation.

All samples collected in the Lower Ordovician series confirmed this age, although the samples from Portugal show MgO contents lower than those of the Spanish equivalents. This could be related to differences in the relative paleogeographic positions of these two groups of samples along the continental margin but current data are not enough to suggest a solution. Also, the three samples that do not clearly fit in any diagram may be essential in future discussions when a better knowledge of the regional geology is reached.

The results in the present work can be useful for the actualization of the geological map of the region, especially to define formations, their correlations and continuity through the border of the two countries.. Complementarily, they allow a new working line the geological structure of the area thus getting better knowledge of the Variscan evolution in this important sector of the Iberian Variscides, close to the limit between the Central Iberian Zone and the exotic overthrust Galicia-Trás-os-Montes Zone.

Future fossil record studies leading to the possible confirmation of the data here presented are now in progress.

Acknowledgements

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EARLY SILURIAN VS. LATE ORDOVICIAN GLACIATION IN SOUTH AMERICA

E. Díaz-Martínez¹, M. Vavrdová², P.E. Isaacson³ and C.Y. Grahn⁴

¹ Geological Survey of Spain (IGME), Ríos Rosas 23, E-28003 Madrid, Spain. e.diaz@igme.es

² Institute of Geology, Academy of Sciences of the Czech Republic, Rozvojová 135, 16500 Praha 6, Czech Republic.
vavrdova@gli.cas.cz

³ University of Idaho, Moscow, Idaho, USA. isaacson@uidaho.edu

⁴ UERJ/Fac. de Geologia, Rua São Francisco Xavier 524, sala 4001, CEP 20550-013, Rio de Janeiro, RJ, Brasil.
yngvegrahn@hotmail.com

Keywords: Glaciation, Gondwana, South America, Llandovery, Silurian.

INTRODUCTION

Late Ordovician and Early Silurian glacial events have been extensively described and discussed in the literature. Late Ordovician glaciations are documented in Arabia, central and southern Europe, North Africa, the Paraná basin of Brazil, and the Precordillera of Argentina (Hambrey, 1985; Vaslet, 1990; Eyles, 1993; Grahn and Caputo, 1994; Assine et al., 1996; Caputo, 1998; Steemans and Pereira, 2002; Ghienne, 2003; Ghienne et al., 2007; Le Heron et al., 2009; Finnegan et al., 2011, and references therein) and Early Silurian glaciations have been reported from southern Libya, the Amazonas and Parnaíba Brazilian intracratonic basins, and the Peru-Bolivia basin (Grahn and Caputo, 1992; Semtner and Klitzsch, 1994; Caputo, 1998; Díaz-Martínez, 1998; Suárez-Soruco, 2000; Díaz-Martínez and Grahn, 2007; and references therein) (Fig. 1). The record of the Ordovician-Silurian transition within the western (South American) margin of Gondwana is characterized by a diamictite-bearing unit which overlies several different Ordovician units, and underlies mid Silurian shales. The recent review of the lithostratigraphy, sedimentology and biostratigraphy of this diamictite unit (Díaz-Martínez and Grahn, 2007) allowed to revise stratigraphic relationships and several ideas and interpretations previously proposed. We herein summarize its main conclusions and provide some new biostratigraphic data which further corroborates its Llandovery age for Peru. Turbidite and shale interbeds within these diamictites indicate a deep marine environment, with interbedded mud flows, debris flows, slumps and large displaced slabs providing evidence for sediment instability and resedimentation. Acritarch and chitinozoan biostratigraphy indicates a Llandovery age, in contrast with previous proposals of a Hirnantian age (see for example Schönian et al., 1999; Astini, 2003; and Schönian and Egenhoff, 2007). The revised Llandovery age suggests the need for more detailed sedimentologic and biostratigraphic studies, and a reassessment of this diamictite unit in western Gondwana. The resedimented character of the deposit explains some of the Ordovician fauna previously described, which must be considered as recycled from underlying units (see discussion in Díaz-Martínez and Grahn, 2007). *In situ* fauna needs to be reassessed, as it may be poorly calibrated endemic, diachronic and/or indicative of migrations within Gondwana. Glacially-faceted and striated clasts, as well as large granitoid boulders within the resedimented materials, provide evidence for glaciation

of the source area, and are interpreted as recycled from former glaciogenic deposits. The evidence found in the Central Andes indicates a glaciomarine origin for the diamictites and corroborates glaciation of a western source area prior to the late Llandovery (Telychian). The precise age of glaciation in this part of Gondwana cannot be confirmed due to a lack of true tillites (ice-contact deposits). Tectonic deformation and the resulting relief are respectively identified as the origin for sediment instability and for local glaciation along the active margin of western Gondwana during the Late Ordovician and Early Silurian. Paleogeographic reconstruction of the Early Silurian Peru-Bolivia basin (Fig. 1) depicts the extension of the diamictite unit southwards into northern Argentina and western Paraguay, connecting with the Paraná basin through the Asunción Arch. To the north it continued into northern Peru and Ecuador, connecting with the Early Silurian record in Colombia and Venezuela, and to the northeast into Brazil (Solimões and Amazonas basins). The sedimentary record of glaciation in the Peru-Bolivia basin was due to the development of local ice fields to the west of the basin, related with reliefs along the active margin of Gondwana, and most probably did not coincide with the main development of the North African late Ashgillian ice cap, but instead, took place afterwards.

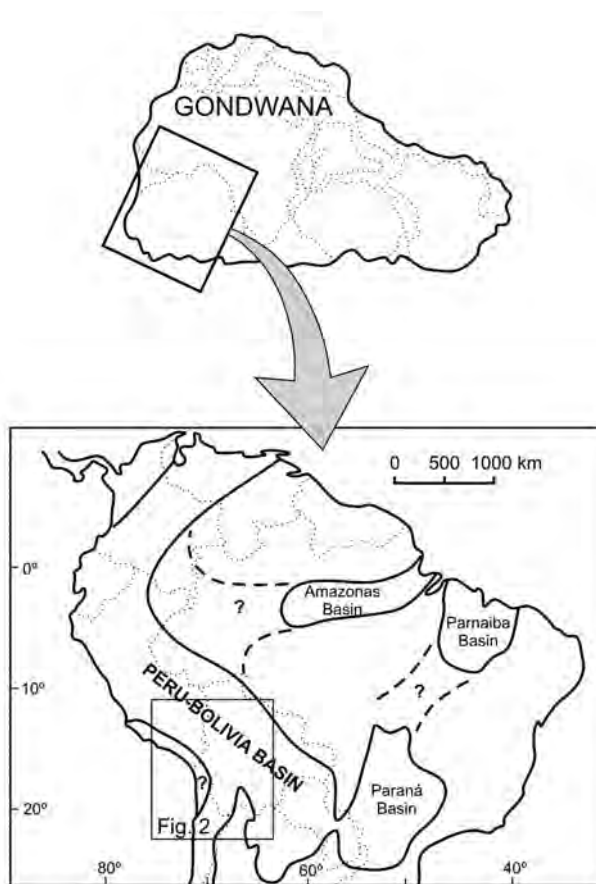


Figure 1. Location of the study area within Gondwana and the Peru-Bolivia basin, and extent of Late Ordovician and Silurian basins in South America. Inset indicates location of Fig. 2.

LITHOSTRATIGRAPHY

In the Central Andes, a diamictite unit is present near the Ordovician-Silurian boundary which has been traditionally used as a stratigraphic marker within the thick and otherwise monotonous Lower Palaeozoic siliciclastic sequences (Boucot, 1988). The diamictites are commonly interbedded with sandstones and shales, and frequently display slumps and contorted beds. The unit is currently known as San Gabán Formation in Peru, Cancañiri Formation in Bolivia, and Zapla Formation in northern Argentina, and extend over an area exceeding 400 km wide and 1600 km long (Fig. 2). We herein refer to this correlative diamictite unit as SGCZ (acronym for San Gabán, Cancañiri and Zapla formations). The variable character of the underlying unconformity, and the common recycled character of most of the fossils found within

these deposits, have led to strong discussions about its age, which have been recently reviewed and reassessed in the light of recently published and new palynological data (Díaz-Martínez and Grahn, 2007; Vavrdová et al., submitted).

The continental-scale diachronism and glacial character of the late Ordovician and early Silurian diamictite deposits in Gondwana have been used in paleogeographic reconstructions as evidence for the displacement of this megacontinent across the pole. However, there is an ongoing discussion regarding the precise age and paleoenvironmental interpretation of these deposits in South America. In an attempt to contribute towards an integrated model of paleogeographic evolution, we reviewed previous work and presented evidence suggesting (a) the Llandovery age of the SGCZ diamictites in the western Gondwana margin, and (b) the predominant glaciomarine and resedimented character of the SGCZ diamictites, in the form of sediment gravity flows which recycled previous glacial deposits. At the same time, we drew conclusions on the implications of this unit for regional tectonism and correlations, as well as suggestions for future research. Evidence for Early Silurian (Llandovery) glaciation seems to be rather systematically ignored by the current scientific paradigm, which tends to simplify early Palaeozoic glaciation to a single and brief glacial event of Hirnantian age. However, as proposed by Ghienne (2003) and Le Heron et al. (2009) for northern Gondwana, Hirnantian glaciation corresponds only to the glacial maximum of a longer-lived glaciation.

A wealth of evidence has been published on Early Silurian glaciations during the last 4 decades which is not taken into account by global palaeogeographic and palaeoclimatic reviews (e.g., Gibbs et al., 2000; Cocks and Torsvik, 2002; Fortey and Cocks, 2003; Raymond and Metz, 2004; Finnegan et al., 2011). In case this is because of the accessibility of the information (mostly in Spanish and spread out in many different local journals and conference proceedings), our intention with our review was to synthesize the information available for the Peru-Bolivia basin, in an attempt to contribute towards its better understanding.

The terminology we have used considers the term diamictite as it was originally proposed, i.e., strictly descriptive and referring to a siliciclastic sedimentary rock with variable grain size ranging from boulder, cobble or gravel size to silt and clay size. This use is irrespective of the origin of the unit or the interpretation of the processes involved in its formation.

The early Palaeozoic Peru-Bolivia Basin (Sempere, 1995) extended along the western margin of Gondwana (Fig. 1), covering most of today's Peru and Bolivia, as well as large parts of Venezuela, Colombia, Ecuador, northern Argentina and Paraguay, and westernmost Brazil. In Peru and Bolivia, this basin was separated from the Proto-Pacific (Iapetus, Rheic) Ocean to the west by an active magmatic arc and a deformational front. Therefore, the geodynamic setting of the basin was a retroarc foreland, and its evolution was strongly influenced by tectonism along the active margin (Sempere, 1995; Díaz-Martínez et al., 1996; Gagnier et al., 1996; Díaz-Martínez, 1998; Jaillard et al., 2000; Díaz-Martínez et al., 2001). Towards the interior of the paleocontinent, the early Palaeozoic Peru-Bolivia Basin was connected with the Paraná intracratonic basin, and probably also with the Amazonas Basin (Fig. 1).

Knowledge about the SGCZ in the Central Andes (Peru, Bolivia and northern Argentina) varies greatly from country to country depending on the different coverage by geological surveys, and on the economic interest of the unit at each region. The SGCZ spans across these administrative boundaries always within the same sedimentary basin, and with correlative facies and thicknesses (Fig. 2). Cenozoic Andean thrusting and telescoping of the basin must also be kept in mind in order to understand rapid changes across tectonic boundaries.

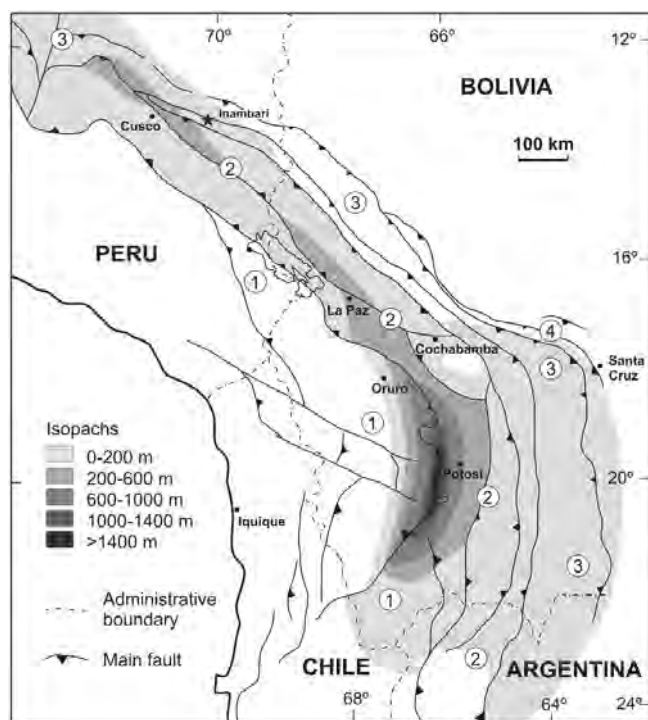


Figure 2. Distribution of the Early Silurian diamictites of the San Gabán Formation (Perú), Cancañiri Formation (Bolivia) and Zapla Formation (Argentina) between 12 and 24°S (Díaz-Martínez and Grahn, 2007). Numbers indicate main tectonostratigraphic domains: 1, Altiplano and Puna; 2, Eastern Cordillera; 3, Subandean; 4, Chapare and Boomerang Hills. The star indicates the location of the Inambari section with new biostratigraphic data mentioned in the text.

BIOSTRATIGRAPHY

The age of the SGCZ diamictites has been difficult to constrain due to an apparent lack of diagnostic fossils. The geological literature includes many references to macrofossils, but most have little chronostratigraphic value. Only very recently have palynological studies contributed to the solution of the problem. Díaz-Martínez and Grahn (2007) suggested that more work is still needed on the Ordovician and Silurian endemic invertebrate macrofauna before we consider its chronostratigraphic value. A Hirnantian age has been extended to different species of the same invertebrate genus at different locations, originating a strong confusion with the ages provided by macrofauna due to the poorly calibrated ages of the biozones corresponding to such endemic species.

Limachi et al. (1996) found a palynological association in the Cancañiri Formation at Pongo (half way along the road between Oruro and Cochabamba) consisting of *Neoverhachium carminae* and *Leiofusa* cf. *bernesgae*, with reworked Ordovician species (*Rhabdochitina* cf. *magna*), which they assigned a Silurian age. Contrary to most previous works in Argentina following the Hirnantian glaciation paradigm (see review in Díaz-Martínez and Grahn, 2007), Grahn and Gutiérrez (2001) concluded from their study of chitinozoans in the Zapla Formation at its type locality that this unit is most probably no older than Aeronian (middle Llandovery), and no younger than late Telychian (late Llandovery). The association found

in the Zapla Formation includes *Angochitina* sp. 1, *Cyathochitina* sp. B, *Cyathochitina* sp. cf. *C. campanulaeformis*, *Conochitina elongata*, and *Conochitina proboscifera*. These authors mention the striking similarities of the chitinozoan faunas found in the Zapla and Santa Barbara ranges with those found in contemporary rocks in the same Peru-Bolivia Basin to the north, and with the Paraná Basin in Paraguay and Brazil as described by Grahn et al. (2000). Grahn and Gutiérrez (2001) suggested that the Zapla Formation corresponds in time partly or completely to the Cancañiri Formation in Bolivia, the Vargas Peña Formation in eastern Paraguay, and the Vila Maria Formation in southern Brazil, and that all these formations represent one or two deglaciation events during the Aeronian and Telychian.

Our study of chitinozoans in the Cancañiri Formation at the section of La Cumbre, along the road from La Paz to Coroico, included the identification of *Belonechitina* cf. *postrobusta* (Rhuddanian), *Cyathochitina* sp. B Paris 1981 (Rhuddanian-Telychian), and *Conochitina* cf. *elongata* (Aeronian-Telychian) at different sampling locations within the section. These new results from chitinozoan biostratigraphy suggest that the most probable age for the Cancañiri Formation near La Paz (Bolivia) is late Rhuddanian-early Aeronian, and therefore strictly Llandovery and coinciding with the first Silurian deglaciation event identified in Gondwana (Grahn and Caputo, 1992; Grahn and Paris, 1992; Grahn et al., 2000).

In southern Peru, recent work on samples from the base of the San Gabán Formation yielded a limited assemblage of distinctly dwarfed, small-sized acritarchs. Representatives of the genus *Veryhachium*, and other polygonomorphid acritarchs such as *Unicisphaera* sp., dominate the assemblages. Several damaged specimens of *Neoverhachium carminae* and an occurrence of *Deunffia* sp. and *Domasia* sp. inhibit a precise age determination for this unit. Nevertheless, damaged specimens of *Hoegklintia visbyensis* (Eisenack) Dorning 1981 and *Polygonium polygonale* (Eisenack) Le Herissé 1981 suggest an Early Silurian (late Llandovery to early Wenlock) age, thus confirming the Early Silurian age also obtained for Bolivia and Argentina by Díaz-Martínez and Grahn (2007).

A last comment is offered regarding the current scientific paradigm on Hirnantian glaciation. The glaciogenic character and Silurian age of the SGCZ diamictites and their correlatives in Brazilian intracratonic basins have been mentioned in the literature for more than 30 years (see above, and references below). This Early Silurian glaciation agrees with the apparent wander path of the southern pole obtained from recent syntheses on paleomagnetism (McElhinny et al., 2003), which locate this pole on Brazil for the Early Silurian (Fig. 3). However, it is sad that all this evidence is commonly ignored in recent global palaeogeographic and palaeoclimatic reviews (e.g., MacNiocaill et al., 1997; Gibbs et al., 2000; Cocks and Torsvik, 2002; Fortey and Cocks, 2003; Raymond and Metz, 2004).

CONCLUSIONS

The San Gabán Diamictite Formation of southern Peru, the Cancañiri Diamictite Formation of Bolivia, and the Zapla Diamictite Formation of northern Argentina are all lateral equivalents and form part of the same lithostratigraphic unit deposited in the same foreland basin along the western (Proto-Andean) margin of Gondwana. According to recent and new palynomorph biostratigraphy, the age of this unit is Llandovery and therefore synchronous with the glaciomarine record of adjacent intracratonic basins of South America. Evidence for glaciation of the source area during or immediately before its deposition is in the form of glacially-faceted and striated clasts, as well as boulder-size granitoid clasts, included within sediment gravity flow deposits. Our overall assessment of the evidence for Early Silurian glaciation in South America supports the small ice-sheet hypothesis proposed by Le Heron and Dowdeswell (2009), and the

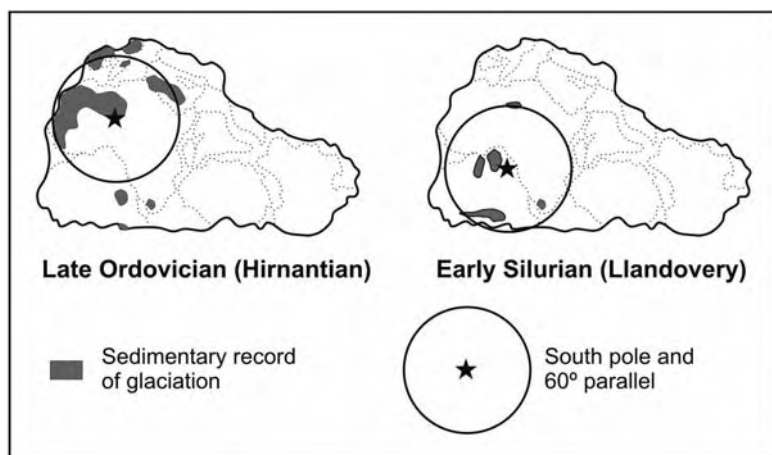


Figure 3. Distribution of Gondwana basins with sedimentary record of glaciation. Compiled with data from Vaslet (1990), Grahn and Caputo (1992, 1994), Eyles (1993), Semtner and Klizsch (1994), Assine et al. (1996), Caputo (1998), Díaz-Martínez (1998), Grahn et al. (2000), Grahn and Gutiérrez (2001), Fortey and Cocks (2003), Ghienne (2003), McElhinny et al. (2003), Ghienne et al. (2007), Le Heron and Dowdeswell (2009) and Le Heron et al. (2009).

isotopic evidence for glaciation-related short-term climatic changes in the early Silurian (Lehnert et al., 2010).

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K-BENTONITES IN THE UPPER ORDOVICIAN OF THE SIBERIAN PLATFORM

A.V. Dronov¹, W.D. Huff², A.V. Kanygin³ and T.V. Gonta³

¹ Geological Institute, Russian Academy of Sciences, Pyzhevsky per.7, 119017, Moscow, Russia.
dronov@ginras.ru

² Department of Geology, University of Cincinnati, OH 45221, USA. warren.huff@uc.edu

³ Trofimuk Institute of Petroleum Geology and Geophysics, Siberian Branch of Russian Academy of Sciences, Acad. Koptuyug 3, 630090, Novosibirsk, Russia. KanyginAV@ipgg.nsc.ru

Key words: K-bentonites, volcanism, paleogeography, Upper Ordovician, Siberia.

INTRODUCTION

Ordovician K-bentonite beds have a long history of investigation all around the world. They have been reported from Gondwana (Ramos et al., 2003), the Argentine Precordillera (Huff et al., 1998), the Yangtze Platform (Su et al., 2003), Laurentia, Baltica, and numerous terrains between Gondwana and Baltica, which now constitute a part of Europe (Huff et al., 2010 and reference therein). In recent years several K-bentonite beds have also been discovered in the Upper Ordovician of the Siberian Platform. This discovery is significant not only for their value in local and regional chronostratigraphic correlation but also for global geochronology, paleogeography, paleotectonic and paleoclimatic reconstructions. All in all, 8 individual K-bentonite beds have been identified in the Baksian, Dolborian and Burian regional stages, which correspond roughly to the Upper Sandbian – Katian Global Stages (Bergström et al., 2009). In this short paper we will present preliminary results of the study of the 4 lowermost beds from the Baksian and Dolborian Regional Stages.

GEOLOGICAL SETTING AND STRATIGRAPHY

During the Cambrian, Ordovician and Silurian the Siberian Platform, which constitutes the core of the Siberian paleocontinent, was located in the low latitude tropical area migrating slowly from the southern hemisphere in the Cambrian and Lower Ordovician to the northern hemisphere in the Upper Ordovician and Silurian (Cocks and Torsvik, 2007). The central part of this continental bloc was occupied by the extensive intracratonic Tungus basin (Markov, 1970; Kanygin et al., 2007). The Lower Ordovician and the lower part of the Middle Ordovician series (from Nyaian to Kimaian regional stages) of the basin are represented by succession of warm-water tropical-type carbonates. The Upper Ordovician series (from Chertovskian to Burian regional stages) by contrast is represented by succession of cool-water carbonates dominated by bioclastic wackestone and packstone beds intercalated with fine-grained terrigenous

sediments. The two carbonate successions of contrasting lithologies are separated by a unit of pure quartz sandstones up to 80 m thick (Baykit Sandstone) that is overlain by the fine-grained terrigenous deposits of the Volginian and Kirensko-Kudrinian regional stages (Dronov et al., 2009; Kanygin et al., 2010).

All K-bentonite beds have been found within the Upper Ordovician cool-water carbonate succession. The four lowermost K-bentonite beds, which were sampled, are located within the Mangazea and Dolbor Formations (Baksian and Dolborian regional stages respectively). Precise biostratigraphic correlation of the Siberian regional stages to the Global Ordovician Stages remains problematic due to the endemic character of the Siberian fauna, but these beds appear to be located near the Sandbian/Katian boundary (Fig. 1). The Mangazea Formation is interpreted as a highstand systems tract of the Mangazea depositional sequence (Kanygin et al., 2010). In the outcrops along the Podkamennaya Tunguska River valley and its tributaries it is represented by greenish-gray siltstones alternating with bioclastic limestone beds. The bioclasts are predominantly fragments of brachiopods and trilobites as well as echinoderms, ostracods, and bryozoans. The limestone interbeds sometime show ripple marks on its upper bedding plane. The intercalations of siltstones and bioclastic limestones of the Mangazea Formation are interpreted as cool-water carbonate tempestites deposited in the middle ramp settings.

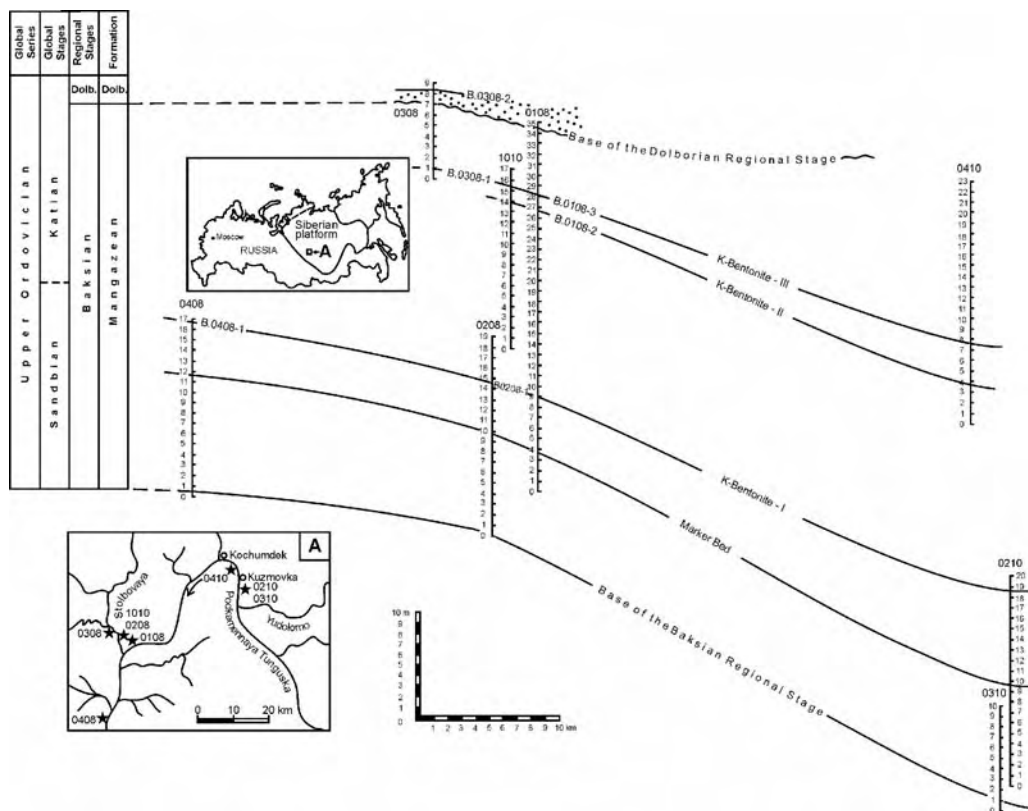


Figure .1. Stratigraphic distribution and correlation of Upper Ordovician K-bentonite beds in the Siberian Platform.

The K-bentonites from the Mangazea Formation are generally represented by thin beds (1-2 cm) of soapy light gray or yellowish plastic clays. They are different by consistence and color from the enclosing sediments and usually easily identifiable in the outcrops. The lowermost (K-bentonite-I) bed of the Mangazea formation was traced over a distance of more than 60 km along the Podkamennaya Tunguska River valley. The other two beds (K-bentonite-II and K-bentonite-III) were traced over at least 40 km (Fig. 1).

MINERALOGY AND GEOCHEMISTRY

A portion of each sample was suspended in distilled water after particle separation by ultrasonic disaggregation. The $<0.2 \mu\text{m}$ size fraction was recovered by ultracentrifugation and was used to make oriented slides by the smear technique for powder X-ray diffraction (XRD) analysis. After drying and vapor-saturation with ethylene glycol for 48 hours at 50°C , the slides were analyzed by powder X-ray diffraction using a Siemens D-500 automated powder diffractometer. Slides were scanned at $0.2^\circ 2\theta/\text{minute}$ using CuK α radiation and a graphite monochromator. Powder diffraction patterns of illite/smectite were modeled using the NEWMOD computer program of Reynolds (1985). The ethylene glycol-saturated diffraction patterns shown in Fig. 2 have essentially the same clay composition. There is an R3-ordered mixed-layer illite-smectite component with about 90% illite represented by peaks at $10.9\text{\AA}$, $9.7\text{\AA}$, $5.1\text{\AA}$ and $3.3\text{\AA}$. There is chlorite and probably some corrensite (mixed-layer chlorite-smectite) at $15\text{\AA}$, $14.5\text{\AA}$, $7.1\text{\AA}$, $4.7\text{\AA}$ and $3.52\text{\AA}$. A small amount of quartz is also present at $4.2\text{\AA}$ and $3.3\text{\AA}$. These are fairly typical patterns for K-bentonites that have undergone a very slight amount of low-grade metamorphism (Krekeler and Huff, 1993).

Modeling of the diffraction tracings using NEWMOD (Reynolds, 1985) showed the samples to contain 80% illite and 20% smectite. The same conclusion results from consideration of the $\Delta 2\theta$ value of $5.1\text{\AA}$, which measures the difference between the 001/002 and 002/003 reflections of I/S (Moore and Reynolds, 1997). Huff et al. (1991) described long-range or R3 ordered I/S in Llandoverly K-bentonites from northern Ireland and the Southern Uplands of Scotland. Batchelor & Weir (1988) interpreted powder diffraction analysis of K-bentonite clays from the Southern Uplands as R0 ordered I/S; however their XRD tracings clearly show R3 ordering. Silurian K-bentonites from Podolia, Ukraine, contain R0 ordered I/S in carbonate facies and R1-R3 ordered I/S in the shale facies (Huff et al., 2000). The preservation of randomly ordered I/S is frequently interpreted as indicative of a shallow burial history with a history

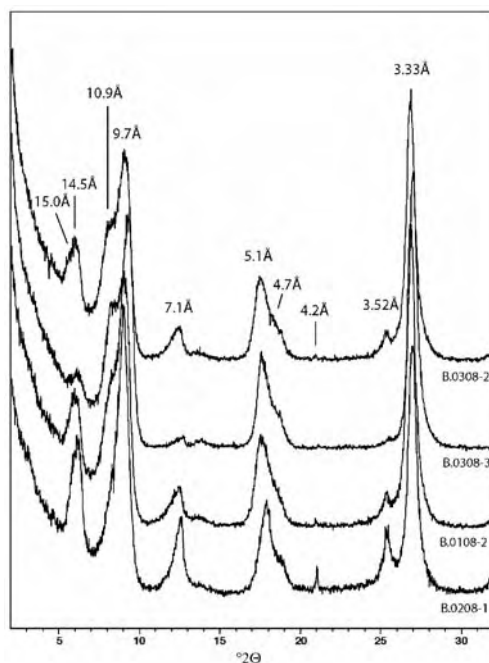


Figure 2. Ethylene glycol-saturated diffraction patterns of the clays from K-bentonite samples.

of relatively low temperatures. Such clays would be expected to show a transition to more highly ordered forms during increased burial metamorphism (Altaner and Bethke, 1989), particularly in shales and mudrocks undergoing basinal subsidence. However, the Siberian sequence of K-bentonites, which occurs on the edge of the Siberian Platform, contains illite/smectite ratios that seem to vary more with rock type than with depth showing no systematic depth-dependent variation in illite percent. This relationship suggests that facies composition and K-availability factors rather than thermal history may have played a leading role in determining clay mineral characteristics. Heavy minerals in the K-bentonite layers provide further evidence of a volcanogenic origin in the form of euhedral apatite phenocrysts (Pl.1, figs.1, 2, 3, 4.).

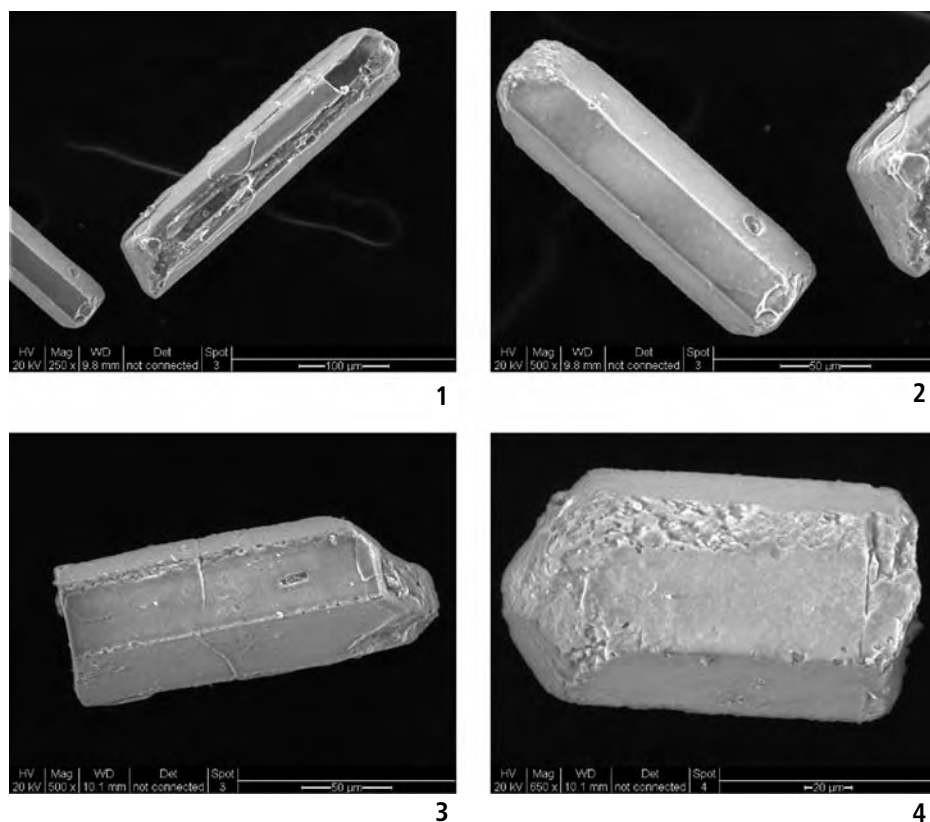


Figure 3. Euhedral apatite phenocrysts from the Siberian K-bentonite beds. 1, sample B.0108-3; 2., sample B.0108-3; 3, sample B.0208-1; 4, sample B.0108-2.

PALEOTECTONIC POSITION AND VOLCANIC SOURCE

The K-bentonite beds discussed in this paper are situated on the southwestern margin of the big intracratonic basin. Exact location of the volcanoes that produced these ash beds still remains unknown. It seems reasonable to suggest that the source of volcanic ash was at or near the border of the Siberian Platform. In our case it is a southwestern border in present day orientation. Volcanic rocks of Ordovician

age are known from Tuva, Eastern Kazakhstan (Chingiz Range) and supposedly existed within the basement of Western Siberia under the Mezo-Cenozoic cover of the West Siberian basin (Dergunov, 1989). Sengor and Natal'in (1996) introduced a term Kipchak Arc for a collage of terranes from Altai-Sayan area, Northern Tian Shan and Kazakhstan. But existence of this enormous island arc, which they believed to have stretched between Baltica and Siberia does not agree perfectly with the data from regional geology. The best preserved fragments of an undoubtedly Ordovician island arc which collided with the Siberian craton in the Late Ordovician – Early Silurian time are known from the Chingiz-Tarbagatay Range in eastern Kazakhstan. It is usually called the Chingiz-Tarbagatai Arc (Dobretsov, 2003).

DISCUSSION

Previous studies of Ordovician K-bentonites in eastern North America and northwestern Europe contain mixed layer illite/smectite clay with 75 to 90% illite. Besides their regularly interstratified illite/smectite clay ratio between 3:1 and 4:1 (e.g. Reynolds and Hower, 1970), Ordovician K-bentonites contain differing amounts of primary and secondary non-clay minerals, some of which provide additional stratigraphic and tectonomagmatic information. The main primary minerals, mostly in the form of isolated, euhedral phenocrysts are quartz, biotite, plagioclase and potassium feldspar, ilmenite, apatite, zircon and magnetite.

Considerable information has been published in recent years on the smectite to illite conversion during diagenesis, and its correlation with organic maturity (Velde and Espitalié, 1989). While many studies indicate that multiple factors influence the progress of clay diagenesis, including the initial composition of smectite, fluid composition, and the rock to water ratio (Freed and Peacor, 1989), most authors have generally considered time, temperature, and K⁺ availability to be the most important factors (Hoffman and Hower, 1979; Huang et al., 1993; Pollastro, 1993). Clay minerals derived from the alteration of felsic volcanic ash are sensitive to the thermal conditions and the geochemical environments, which have characterized their post-emplacement history. The transition of I/S to R3 ordering occurs during burial metamorphism at about 150-175°C, and under equilibrium conditions complete the transition to illite at about 250°C (Hoffman and Hower, 1979; Rateyev and Gradusov, 1970). Previous studies have shown that I/S in K-bentonites as well as in shales is a diagenetic product of smectite alteration (Altaner et al., 1984; Bethke et al., 1986; Brusewitz, 1988; Anwiller, 1993) and that further alteration to C/S occurs under low grade metamorphic conditions (Krekeler and Huff, 1993). However, more recent work (Essene and Peacor, 1995; Sachsenhofer et al., 1998) has cautioned against the unequivocal use of interstratified illite/smectite as a geothermometer, and has provided further evidence that factors such as pore fluid chemistry and rock to fluid ratios can have an important role in determining the reaction progress of clay mineral diagenesis.

CONCLUSIONS

- 1) The K-bentonite beds from the Upper Ordovician Mangazea and Dolbor formations of the southwestern part of the Tungus basin in Siberia seem to be derived from the alteration of volcanic ash falls. Their appearance points to the intensive explosive volcanism on or near the western (in present day orientation) margin of the Siberian craton in Late Ordovician time.

- 2) Timing of volcanism in the Ordovician of Siberia is surprisingly close to the period of volcanic activity of the Taconic arc near the eastern margin of Laurentia. It looks like both arcs were activated by the same plate tectonic reorganization.
- 3) Similar to the situation in North America the Upper Ordovician K-bentonite beds in Siberia are associated with cool-water carbonates.

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ORDOVICIAN OF BALTOSCANDIA: FACIES, SEQUENCES AND SEA-LEVEL CHANGES

A.V. Dronov¹, L. Ainsaar², D. Kaljo³, T. Meidla², T. Saadre⁴ and R. Einasto⁵

¹ Geological Institute, Russian Academy of Sciences, Pyzhevsky per.7, 119017, Moscow, Russia.
dronov@ginras.ru

² Institute of Ecology and Earth Sciences, University of Tartu, Ravila 14a, 50411 Tartu, Estonia. leho.ainsaar@ut.ee;
tonu.meidla@ut.ee

³ Institute of Geology, Tallinn University of Technology, Ehitajate tee 5, 19086 Tallinn, Estonia. kaljo@gi.ee

⁴ Geological Survey of Estonia, Kadaka tee 82, 12618 Tallinn, Estonia. t.saadre@egk.ee

⁵ University of Applied Sciences, Pärnu mnt 62, 10135 Tallinn, Estonia.

Key words: Sequence stratigraphy, sea-level changes, Ordovician, Baltoscandia.

INTRODUCTION

Sea level changes are the main mechanism controlling facies dynamics, stratal geometry and temporal variations in fossil assemblages in marine sedimentary basins. Besides having direct influence on the water depth sea-level changes could serve as a trigger for various oceanographic, climatic, sedimentary, chemical, and biotic events. Patterns of large-scale evolutionary radiation and mass extinction may be related in complex ways to large-scale fluctuations of the sea level. In recent years a number of new sea-level reconstructions for specific Ordovician basins around the world (Munnecke et al., 2010) as well as the refinement of the global chronology of the Ordovician depositional sequences (Haq and Schutter, 2008) have been suggested. Bathymetry of the Ordovician basin of Baltoscandia is considered in numerous papers and has been summarized in regional sea-level curves (Nestor and Einasto, 1997; Dronov and Holmer, 2002; Nielsen, 2004) which differ in several stratigraphic levels. The purpose of this article is to present a refined review of sequence stratigraphy and sea-level changes for the relatively shallow-water Estonian part of the basin.

ORDOVICIAN BASIN OF BALTOSCANDIA

The Baltoscandian Palaeobasin is one of the largest and best studied Lower Palaeozoic sedimentary basins in the world. It represents a typical epicontinental interior sag basin which covers a vast territory (over the 1000000 square kilometres) and is characterized by a relatively low rate of subsidence. Despite a large number of small breaks in shallow shelf areas the Ordovician succession is almost complete from the biostratigraphic point of view, with no essential parts missing. Latitudinal migration of the Baltic continent is reflected in the succession of shallow-water facies from siliciclastic sands in the Tremadoc

through cool-water carbonates in the Floian – Sandbian to warm-water tropical carbonates in the Katian – Hirnantian. Shift of depocentres from relatively deep-water to shallow-water settings, clearly visible on the profile across the basins margin (Fig. 1), reflects transition from cool-water to warm-water carbonate factories (Schlager, 2007). A general facies and bathymetric zonation of the basin was described by Männil (1966), Jaanusson (1982) and Harris et al. (2004). Deep-water off-shore facies are confined to the Scanian Confacies belts whose lithology is dominated by black shales. The Central Baltoscandian Confacies belt occupies an intermediate position between the Scanian and North Estonian ones. The dominant facies types are marine red beds with thin black shale units and grey limestone-marl intercalations at some levels. The sediments seem to be deposited in a hemipelagic environment below the storm wave base. The shallow-water deposits with traces of storm, wave and tidal activity occur in the North Estonian Confacies belt. In this environment the development of regional unconformities at sequence boundaries and shifts of facies at transgressive surfaces are better expressed. This makes a shallow-water setting more favorable for detecting major sea-level fluctuations which principally should affect the whole basin.

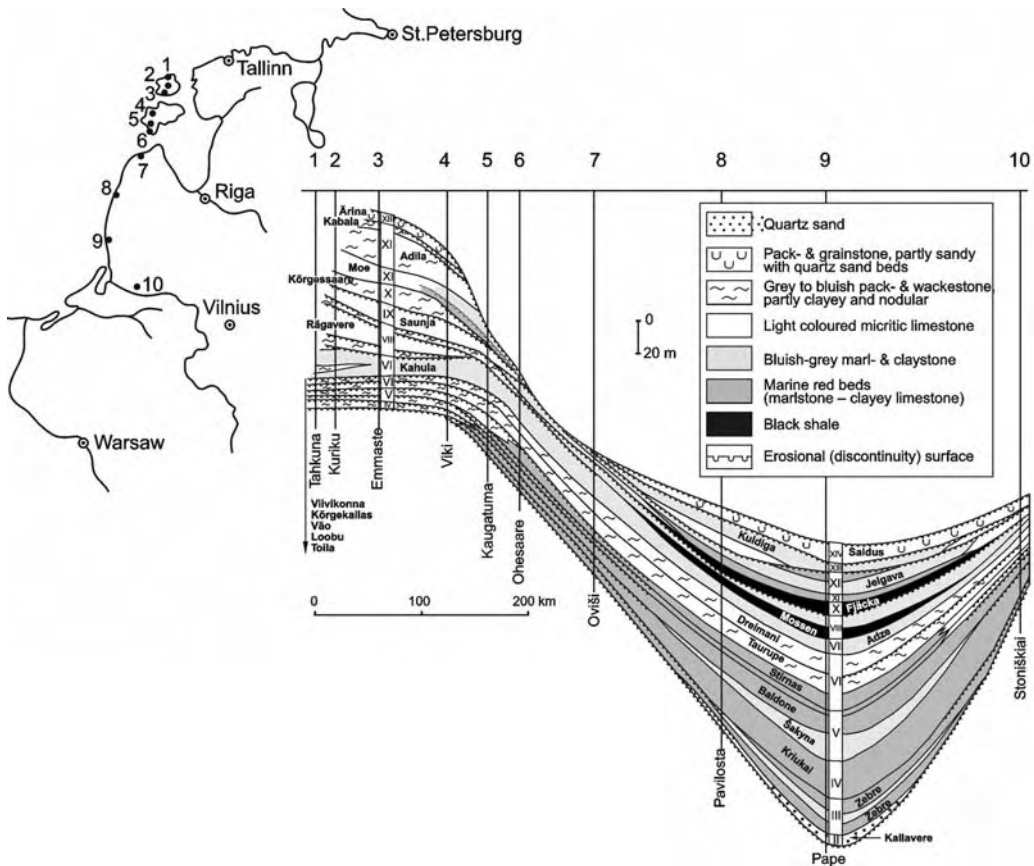


Figure 1. Profile across the Livonian Tongue displaying facies of shallow-water North Estonian and deep-water Central Baltoscandian Confacies belts. Vertical lines mark boreholes and Roman numbers mark the sequences.

DEPOSITIONAL SEQUENCES

On the basis of outcrop and drill core data the Ordovician succession of Baltoscandia is here subdivided into 14 depositional sequences, modifying the former models by Dronov and Holmer (1999) and by Harris et al. (2004). The sequences represent third-order cycles of relative sea-level changes and have duration from 0.9 to 12 my (Fig. 2). We concentrate primarily on the changes made in the sequence stratigraphic subdivision of the succession in comparison with the previous works. Sequence (I) agrees with the Pakerort sequence of Dronov and Holmer (1999). Sequence II coincides with the Varangu Regional Stage. In relatively deep-water settings this sequence is represented by the "*Ceratopyge* shale" (glauconite sandstone) overlain by the Björkåsholmen Formation (former "*Ceratopyge* limestone") and bound by unconformities at the base and top of the sequence. Previously the Varangu deposits have been interpreted as a lowstand systems tract of the Latorp sequence (Dronov and Holmer, 1999). Sequence III corresponds to the Hunneberg and Billingen regional stages. The bounding surfaces are an unconformity on top of the Björkåsholmen Formation and an erosional surface at the base of the Volkhov Regional Stage. In the eastern Baltic area sequence III comprises sandstones of the Leetse Formation grading into the limestones of the Toila Formation, and the upper Zebre Formation within the Livonian Tongue. The sequence has the longest duration (about 12 my) and a minimal average thickness rarely exceeding 1.5 m. It represents a highly condensed stratigraphic interval of diverse lithology, varying from quartz and glauconite sands to clays and bioclastic limestones. Gaps and erosional surfaces are common. Only fragments of previous sedimentary units are preserved and the small thickness makes a sequence stratigraphic analysis of this interval extremely difficult. It could not be excluded that in future two or three separate depositional sequences will be identified in it. Sequences II and III make up the former Latorp sequence. Sequences IV, V, VI and VII agree with the Volkhov, Kunda, Tallinn and Kegel sequences of Dronov and Holmer (1999), respectively (Fig. 2). Sequence VIII comprises the Oandu and Rakvere regional stages. The Hirmuse Formation and lower parts of the Mossen and Variku formations can be interpreted as a transgressive systems tract while the shallow-water light-coloured micritic limestones of the Rägavere Formation, together with the upper parts of the Mossen and Variku formations seem to represent a highstand systems tract. Basal unconformity of this sequence is one of the best developed regional unconformities in the entire Ordovician succession of Baltoscandia (Lashkov and Paškevičius, 1989). Sequence IX comprises the Nabala Regional Stage. Argillaceous limestones of the Paekna and Möntu formations seem to represent a transgressive systems tract deposits. Similar to the limestones of Rägavere Formation, micritic limestones of the Saunja Formation are interpreted as a highstand systems tract deposits. Sequences VIII and IX have been included in the former Wesenberg sequence of Dronov and Holmer (1999). Sequence X comprises the Vormsi Regional Stage. It includes the Fjäckå Formation, a black shale, which seems to represent a condensed section in the transgressive systems tract and probably the lower part of the highstand systems tract in its upper calcareous transition interval. In more shallow-water settings the sequence is represented by the Tudulinna and Kõrgessaare formations. The lower sequence boundary possesses characteristics of a transgressive surface. The sequence coincides with the sequence 2 of Harris et al. (2004). Sequence XI comprises the lower part of the Pirgu Regional Stage (Moe Formation in a shallow-water setting and Jonstorp Formation in a deeper-water environment). In the shallow-water setting two cycles were identified within the sequence (comprising sequences 3 and 4 by Harris et al., 2004), but they can not be distinguished in deep-water red limestones and marls of the Jonstorp Formation. Sequences X and XI were assigned to the Fjäckå sequence by Dronov and Holmer (1999). Sequence XII corresponds to the sequences 5 and 6 of Harris et al. (2004) and the Jonstorp sequence of Dronov and Holmer (1999). Sequence XIII

comprises the lower part of the Porkuni Regional Stage. This sequence includes the Ärina Formation in a shallow-water setting and the Kuldiga Formation in a relatively deep-water setting, corresponding to sequence 7 of Harris et al. (2004). Basal unconformity is represented by an erosional surface with deep erosional valleys cutting the underlying deposits (Lashkov and Paškevičius, 1989). Sequence XIV corresponds to the upper part of the Porkuni Regional Stage. It is represented by the Saldus Formation which is mainly distributed in the central parts of the Livonian Tongue and corresponds to sequence 8 of Harris et al., (2004). The basal unconformity of the sequence represents an erosional surface that cuts different underlying deposits. Dronov and Holmer (1999) have attributed sequences XIII and XIV to the Tommarp sequence.

SEA-LEVEL CHANGES

The reconstructed sea-level curve (Fig. 2) is based on the following assumptions: 1) The main regional unconformities reflect substantial sea-level drops and forced regressions. More widespread unconformity and deeper erosion of the underlying beds means higher magnitude of a sea-level fall. 2) The main transgressions are recognized according to the widening of the deep water facies area. The absolute magnitude of individual sea-level events is difficult to estimate but the relative magnitude of each transgression and regression can be derived from the assumptions above. We followed approach of Haq and Schutter (2008) and classified each event semi-quantitatively as minor (<30m), medium (30-75m), or major (>75m). Analysis of the regional unconformities and associated gaps shows that the greatest unconformities and deepest erosion occur at the bases of the sequences II, VIII, XIII and XIV. Hirnantian sea-level drops seem to be comparable to that induced by Quaternary glaciation, which requires sea-level fall up to 100-125 m. Erosion at the base of sequence XIII was deeper than at the base of sequence XIV, so the relative magnitude of the sea-level drop at the base of sequence XIII was higher. We assume the magnitude of 120 m and 110 m for bases of sequences XIII and XIV, respectively. Comparable regional unconformity but without such a deep cutting, was reported from the base of sequence VIII. A major sea-level drop of can be assumed for this level. Deep erosion associated with the most distal shift of facies, was also reported from the base of sequence II. The shoreline migrated over a distance of more than 700km which mean a sea-level drop of a major magnitude. Less pronounced unconformities were recognized at the bases of sequences III, V, VI, VII and XII. All of them were connected with regressions of medium magnitude. Among these regressions the highest relative magnitude was attributed to the base of the sequence III, considering almost complete erosion of the Varangu deposits in the shallow-water settings. The regression at the base of the sequence V resulted in traces of sufficient erosion. The sea-level falls of a minor magnitude are reflected at the bases of the sequences IV, IX, X and XI more remarkable being the regression at the base of the sequence IV. Analysis of spatial distribution of the most deep-water facies allows distinction of transgressive events into those of a major magnitude (sequences I, II, III, IV, VIII, X and XII), medium magnitude (sequences V, VII, XI, XII and XIV) and minor magnitude (sequences VI and IX). The most remarkable sea-level occurs at the base of the sequence III. At that time deep-water red bed facies for the first time occupied wide areas in the Ordovician basin of Baltoscandia. Mass migration of a new fauna into the basin supports this interpretation. High magnitude transgression are indicated by wide distribution of black shale in shallow-water environments during the sequence I and returning of marine environments after high magnitude regression at the base of the sequence II. Magnitude of the Volkhovian transgression (sequence IV) was less than the magnitude of the Lower Ordovician transgressions. The

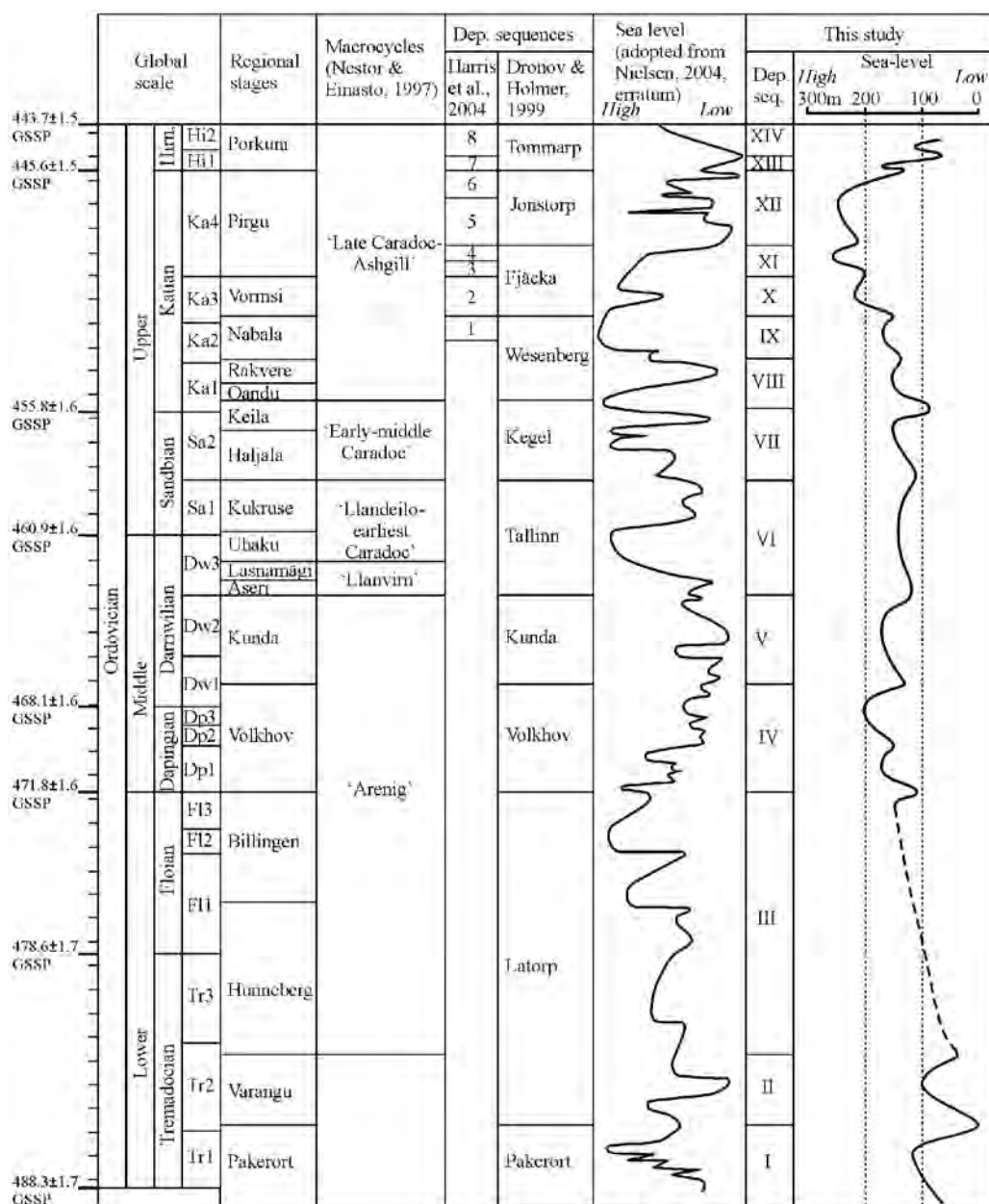


Figure 2. Comparison of sea level curves (Nielsen, 2004; this study), macrocycles (by Nestor and Einasto, 1997) and depositional sequences (by Harris et al., 2004; Dronov and Holmer, 1999 and this study) in the Ordovician Paleobasin of Baltoscandia.

regression at the basal Volkhov sequence boundary was of minor and the following sea-level rise enhanced the previous transgression which leads to a deepening of the basin. Depth of the basin and a territory occupied by deep-water red bed facies reaches its maximum in the Volkhov time. Major transgressions can be recorded also for the sequences XIII, VII and X. The two latter events are marked by invasion of black shale facies (lower Mossen and Fjäckå formations, respectively). The highest position of the sea-level is marked by appearance of marine red bed facies in the middle part of the Livonian basin (Jonstorp Formation). The Lower Jonstorp transgression inherited the Fjäckå sea-level rise and the depth of the basin reached its maximum at that time. It is worth to note that black shales occupied the most distal position on the Baltoscandian facies profile and they are the first to invade into shallow-water setting when the sea-level starts to rise rapidly after regression, but in case of a further sea level rise the black shales are replaced by marine red bed. This model is different from previous interpretations (see Nestor and Einasto, 1997). Moderate individual transgression events occurred in the sequences V, VII, XI, XII and XIV. Transgressions of a minor magnitude associated with the sequences VI and IX. Lack of accommodation space in a shallow-water setting can be recognized in these levels.

DISCUSSION

Comparison of the sea level curves published in the last decades (Nestor and Einasto, 1997; Dronov and Holmer, 2002; Nielsen, 2004) displays some disagreements between the curves reconstructed for the shallow-water part of the basin and those based on relatively deep-water sections. The main contradictions are in the Dapingian–Lower Darriwilian and in the Upper Katian intervals. For those intervals (the Volkhov, Kunda and Pirgu regional stages) we propose the highest sea-level stands whereas the deep-water model assumes the lowest sea level stands, termed the “Late Arenig–Early Llanvirn Lowstand Interval” and “Ashgill Lowstand Interval” (Nielsen, 2004), respectively. This contradiction reflects opposite opinions in the interpretation of limestone units within the Scanian Confacies belt. The invasion of carbonate facies into the black shale realm is interpreted as a shallowing event in the deep-water model, assuming that limestones represent more shallow-water facies than the black shales (Nielsen, 2004). On the other hand, the same episodes in shallow-water areas are characterized by the expansion of the relatively deep-water marine red bed facies into the shallow-water realm, suggesting deepening events. In our opinion the invasion of limestone facies into the deep-water black shale environment could be explained through the mechanism of “highstand shedding” (Schlager, 2007). According to this view carbonates were transported from a shallow-water environment into a deep-water setting only at the time of maximum carbonate production in the shallow-water environment, i.e. during sea-level highstand. In this case the limestones in a deep-water Scanian Confacies belt mark maximum flooding events, rather than regressive ones. The sea-level curve for the Ordovician of Baltoscandia elaborated in the framework of the deep-water model is thought to be in a good agreement with the sea-level curve suggested for the North American craton by Ross and Ross (1995). This coincidence was regarded as an evidence for the eustatic nature of the displayed sea-level fluctuations. Both curves demonstrate sufficient sea-level lowstand during almost the entire Middle Ordovician. In the Ordovician basin of Baltoscandia, however, this lowstand could be regarded as an artefact. In Laurentia the Middle Ordovician lowstand was apparently deduced from the big gap comprising this stratigraphic interval in the Ordovician succession of the New York State. This gap is maximal in the pericratonic basin of the New York State and becomes less pronounced towards the middle part of the continent. The Ordovician succession is almost complete in the Great Basin (Nevada and

western Utah). This pattern, accordingly, could be regarded as an evidence of tectonic uplift and erosion in the New York State basin during the Middle Ordovician, probably due to the Taconic orogeny.

CONCLUSIONS

Based on the distribution of facies and regional unconformities, 14 major 3rd-order depositional sequences were differentiated and correlated in the Ordovician succession of Baltoscandia. The presented sea-level curve for the Ordovician of Baltoscandia is based on the sequence stratigraphic approach and semi-quantitative estimations of the magnitudes of sea-level rise and falls mainly in the relatively shallow-water setting. The most prominent regressions, marked by unconformities and extensive erosion coincide with the top of the Ordovician, as well as with the base of the sequences VIII, XIII and XIV (early Katian, basal Hirnantian and middle Hirnantian). The most distinct sea-level highstands, marked by the widening of the relatively deep-water marine red bed facies, occurred during Volkhov–Kunda (Dapingian–early Darriwilian) and Pirgu (latest Katian) times.

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POSSIBLE REMAINS OF THE DIGESTIVE SYSTEM IN ORDOVICIAN TRILOBITES OF THE PRAGUE BASIN (BARRANDIAN AREA, CZECH REPUBLIC)

O. Fatka¹, P. Budil² and Š. Rak^{1,3}

¹ Charles University, Institute of Geology and Palaeontology, Albertov 6, 128 43 Praha 2, Czech Republic. fatka@natur.cuni.cz

² Czech Geological Survey, Klárov 3, 118 21 Prague 1, Czech Republic. petr.budil@geology.cz

³ Museum of the Czech Karst, Husovo náměstí 88/16, Beroun-Centrum, Czech Republic.
deiphon@geologist.c

Keywords: Czech Republic, trilobites, digestive system, Prague Basin, Upper Ordovician.

INTRODUCTION

Isolated parts and even complete articulated trilobite exoskeletons frequently occur in “Middle” Cambrian to Middle Devonian sediments in the Barrandian area (e.g. Šnajdr, 1990; Bruthansová et al., 2007), while rests of soft parts have been only rarely documented in several Cambrian and Ordovician genera (e.g. Beyrich, 1846; Barrande, 1852; Jaekel, 1901; Šnajdr, 1990, 1991; Budil and Fatka, 2008 and Fatka et al., 2008) and often have been considered as questionable. The Cambrian record includes the *Ptychoparia* (three specimens, see Jaekel, 1901; Šnajdr, 1958 and Kordule, 2006), *Conocoryphe* (one specimen, see Budil and Fatka, 2008 and Fatka and Budil, 2008), and several tens of recently collected specimens belonging to several genera (e.g. *Ptychoparioides*, *Ctenocephalus*, *Germanopyge*; Fatka et al., in prep.). This material comes from sandy greywackes to sandstones disclosed at several outcrops in diverse stratigraphical levels of the Middle Cambrian Buchava Formation of the Skryje-Týřovice Basin. In Ordovician of the Prague Basin, all trilobites containing parts of the supposed digestive system were collected from the generally poorly fossiliferous (common remains of fauna occur in several horizons only) sandstones of the Letná Formation (Upper Ordovician, Sandbian Stage = Berounian Regional Stage). All findings came in the area of the town Beroun (see Figs. 1, 2). All known specimens are preserved as internal moulds in quartzose sandstones.

SO FAR PUBLISHED MATERIAL

The known Ordovician material involves specimens of two genera only; the trinucleid *Deanaspis* Hughes et al., 1975 and the dalmanitid *Dalmanitina* Reed, 1905.

Occasional finds of *Deanaspis goldfussi* (Barrande, 1846) showing anatomical details interpreted as parts of the alimentary canal were figured by Beyrich (1846), Barrande (1852) and Příbyl and Vaněk (1969). They were described by Šnajdr (1990, 1991); some aspects were discussed by Shaw (1995).

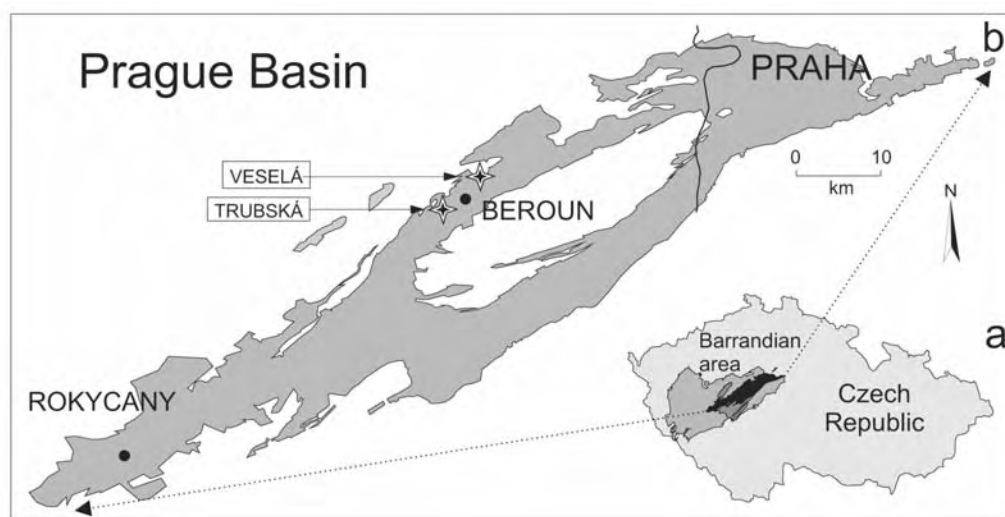


Figure 1. a, Map of the Czech Republic and the Barrandian area. b, Sketch map of the Prague Basin showing location of the Veselá and Trubská outcrops at which the described materials were collected.

In the last comprehensive study, Šnajdr (1991) summarized the earlier information and assembled all the accessible specimens. He detailed the existence of twenty two specimens of *D. goldfussi*, which he shortly described and eleven of them figured. Šnajdr (1991, pl. 3, fig. 14) also figured the only known articulated specimen of *Dalmanitina socialis* (Barrande, 1846) showing a simple intestine preserved under the axis of thoracic and pygidial parts. Recently, this material was exhaustively discussed and partly re-interpreted by Lerosey-Aubril et al. (in press).

NEW MATERIAL

Two new specimens with preserved rests of the digestive system in axial parts were discovered in a complete internal mould of the large exoskeleton of *Birmanites* Sheng, 1934 and in the internal and external moulds of cephalon associated with major part of thorax (very probably, there are remains of entire specimen with hypostome *in-situ*) of *Selenopeltis* Hawle and Corda, 1847.

The preserved remain of *Selenopeltis buchi* (Barrande, 1846) bears paired "gut diverticula" visible in the posteral part of glabella and in the axis of the anterior five thoracic segments. A simple "intestine" (= alimentary canal posterior to the crop *sensu* Lerosey-Aubril et al., in press) extends through the thoracic axis to the pygidial end in *Birmanites ingens* (Barrande, 1852).

Both specimens originate from quartzose sandstones of the Upper Ordovician Letná Formation, e.g. from the same stratigraphical level as the earlier described findings of *Deanaspis* and *Dalmanitina*.

GLOBAL						REGIONAL					
SYSTEM	SERIES	STAGES	STAGE SLICES	TS	TU	STAGES	FORMATIONS				
ORDOVICIAN	UPPER	HIRNANTIAN	Hi2	6c	21	KOSOVIAN	KOSOV	Selenopeltis	Dalmanitina	Deanaspis	Birmanites
			Hi1								
		KATIAN	Ka4	6b	20	KRALODVORIAN	KRÁLŮV DVŮR				
			Ka4	6a	19						
			Ka3		18	BEROUNIAN	BOHDALEC				
			Ka2	5d	17		ZAHOŘANY				
			Ka1	5c	16		VINICE				
			Sa2	5b	15		LETNÁ				
	MIDDLE	SANDBIAN	Sa1	5a	14		LIBEŇ				
			Dw3	4c	13	DOBROTIVIAN	DOBROTIVÁ				
		DARRIWILIAN	Dw2	4b	12	ORETANIAN	ŠÁRKA				
			Dw1	4a	11						
			Dp3		9	ARENIGIAN					
		DAPINGIAN	Dp2	3b	9						
			Dp1	3a	8						
	LOWER	FLOIAN	F3	2c	7						
			F2	2b	6						
			F1	2a	5						
			Tr3	1d	4						
		TREMADOCIAN	Tr3	1c	4						
			Tr2	1b	3	TREMADOCIAN	MÍLINA				
			Tr1	1a	2		TŘENICE				
					1						

Figure 2. Stratigraphy of Ordovician in the Prague Basin with the marked level at which the trilobites with remains of the digestive system were collected.

CONCLUSIONS

In the Ordovician of the Prague Basin, the exceptional preservation of rests of the alimentary canal is surprisingly exclusively restricted to quartzose sandstones of the Letná Formation only (localities Veselá and Trubská, Fig. 1). They are represented by the two earlier known types of the digestive system. Presence of the metamERICALLY paired digestive caeca was ascertained in the cephalon and thorax (cephalic and thoracic gut diverticulae) of the odontopleurid *Selenopeltis buchi* (Barrande, 1846) only. Simple tube-like intestine was recovered in the dalmanitid *Dalmanitina socialis* (Barrande, 1846) as well as in the asaphid *Birmanites ingens* (Barrande, 1852), e.g. in the probably unrelated taxa. Comparatively simple intestine associated with a crop is probably present also in the trinucleid *Deanaspis goldfussi* (Barrande, 1846). It is apparent that the type of the digestive system is only partially dependent on the phylogenetic position of the trilobites but probably corresponds with its mode of live only.

Acknowledgements

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LATE ORDOVICIAN-EARLY SILURIAN SELECTIVE EXTINCTION PATTERNS IN LAURENTIA AND THEIR RELATIONSHIP TO CLIMATE CHANGE

S. Finnegan¹, S. Peters² and W.W. Fischer¹

¹ Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, CA 91125.

² Department of Geoscience, University of Wisconsin-Madison, 1215 W Dayton St. Madison WI 53706.

Keywords: Late Ordovician mass extinction, selectivity, paleoclimate, temperature, glaciation.

INTRODUCTION

There is general agreement that the Late Ordovician mass extinction is causally related to climate change, but the precise mechanism of the relationship is not well established. A mechanistic understanding the relationship between climate change and extinction is inhibited by uncertainties about the timing, nature and magnitude of climate change and by the lack of a distinct selective extinction pattern. Here we summarize recent and ongoing work aimed at clarifying both of these uncertainties

PALEOCLIMATE RECONSTRUCTIONS

The recently developed clumped isotope proxy (Eiler, 2007) is a thermodynamically-based proxy for carbonate precipitation temperature that is independent of the isotopic composition of fluid from which the carbonate precipitated. Because it is therefore also independent of the growth and decay of continental ice sheets, it provides a means of untangling trends in local temperature from trends in global ice volume, a problem that has crippled deep-time paleoclimate reconstructions for decades. Recent application of this proxy to well-preserved Late Ordovician-Early Silurian biogenic carbonates from Anticosti Island, Quebec, Canada, and the U.S. midcontinent (Finnegan et al., 2011) suggests that the Laurentian tropics experienced ~5° C of cooling during the Late Ordovician but that most of the cooling was restricted to the Hirnantian Stage (here we consider only the Laframboise Member of the Ellis Bay Formation to be of Hirnantian age, but our substantive conclusions would be unaffected by assigning all of the Ellis Bay Formation to the Hirnantian as advocated by some (Copper and Long, 1989; Copper, 2001; Desrochers et al., 2010; Achab et al., 2011). In contrast to the Hirnantian temperature change seen in the tropics, isotopic evidence for moderate ice sheets spans a much longer interval, from the late Katian to at least the late Rhuddanian. These data support observations from sequence stratigraphy suggesting that the growth of Gondwanan ice sheets initiated prior to Hirnantian time. Clumped isotope data also reveal a

large Hirnantian peak in the $\delta^{18}\text{O}$ of seawater, suggesting that continental ice volumes at this time were very large and may have equaled or exceeded those of the last Pleistocene glacial maximum. Altogether our data supports aspects of both the protracted and short models of glaciation, with evidence for a substantial glacial maximum superimposed on a longer glacial interval that lasted ca. 10 myr. The coexistence of substantial polar ice sheets with tropical temperatures locally exceeding 35°C implies that the Late Ordovician-Early Silurian world may have exhibited steep meridional temperature gradient relative to subsequent “icehouse” modes.

EXTINCTION SELECTIVITY

Additional work on other well-preserved Late Ordovician-Early Silurian sections is required to confirm that temperature and ice volume trend from Anticosti are representative of the global tropical oceans. Such work is ongoing, but the Anticosti results provide a preliminary set of predictions about the nature and timing of environmental stresses on Laurentian marine ecosystems. Extinctions related to changes in temperature or its correlates should be limited to the Hirnantian, and those related to habitat losses and/or environmental shifts due to sea-level fall may occur throughout the later part of the Late Ordovician but should peak in the Hirnantian.

To evaluate these predictions, we combined fossil collections comprising more than 80,000 genus occurrences from the Paleobiology Database (Alroy et al., 2011) with data on the spatio-temporal fabric of the Laurentian rock record from the Macrostrat Database (Peters, 2005) to produce an integrated paleontological, environmental and stratigraphic framework for the Late Ordovician-Early Silurian of Laurentia. This framework allows us to map out patterns of faunal distribution, environmental distribution, and stratigraphic completeness for twelve late Middle Ordovician through Early Silurian time slices (see Fig. 1 for an example). Mapping genus occurrences onto sites of sedimentation allows us to determine geographic range (measured as site occupancy) for each genus in each interval. Geographic range has been shown to be one of the most consistent predictors of extinction risk both in the fossil record, and temporal changes in the buffering effect of wide geographic range on extinction risk may convey important information about extinction mechanism (Payne and Finnegan, 2007).

Because Hirnantian (or Gamachian) strata were not differentiated from Richmondian (late Katian) strata in the correlation charts on which Macrostrat is based (Childs, 1985), we have worked to refine the chronostratigraphic framework through these intervals. The refined

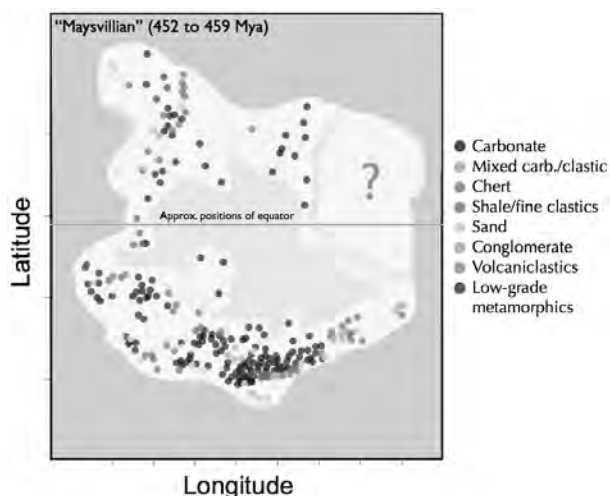


Figure 1. An example base map: distributions of sediments of definitely or likely Maysvillian (mid-late Katian) age in Laurentia. Each point represents a local stratigraphic column in the Macrostrat database, and the shading indicates lithology. Question mark indicates Greenland, which is not currently included in the Macrostrat database.

dataset confirms a very large regression in the Hirnantian, with only comparatively minor changes in continental flooding (measured by the number of sites with sedimentary rocks of a given age) through the Sandbian-Katian (Fig. 2A). As previously noted at coarser scales (Peters, 2005) trends in genus diversity bear a striking similarity to those in continental flooding (Fig. 2B).

Although the similarity of these trends is striking, it does not necessarily prove a causal relationship -both continental flooding and extinction could be responding to a common driver -climate change- without habitat losses having any direct influence on extinction risk. We can improve on the analysis by mapping changes in continental flooding and environmental distribution onto the geographic ranges of individual taxa. Accounting for local section truncations (sedimentation at that site ceases for at least

one interval) and environmental shifts (sedimentation style changes, for example from limestone to shale) permits us to calculate what proportion of each genus' range was affected by regression or environmental shifts in each time interval. Along with geographic range, these proportions can then be included as explanatory variables in a logistic regression with extinction/survival of genera as the response variable in order to determine how well habitat losses and/or environmental shifts predict not just the magnitude but the selectivity of extinction in each time interval. Within this framework we can also evaluate the explanatory power of a wide variety of other ecological variables that have been shown or suggested to be important determinants of extinction risk either during the Late Ordovician or at other time in Earth History. These include depth preference (whether the taxon tends to occur in relatively shallow or relatively deep facies), environmental preference (whether the taxon tends to occur in carbonate or clastic environments), trophic level, and life habit (benthic or not, infaunal or epifaunal).

A final variable we examined was whether or not a given genus had been sampled at high latitude ($>45^\circ$) during or prior to each time interval (because the analysis is limited to Laurentia, all examined genera have at least partially tropical ranges). This variable is related to both geographic range and taxon age, but also provides information on tolerance for variation in the correlates of latitude (temperature, seasonality, etc.) -a trait that may be expected to be important during times of rapid climate change.

We used a Bayesian model averaging approach to compare all possible combinations of explanatory variables and select the set of models that explain the most variation in extinction selectivity with the fewest predictors (e.g., complex models are penalized to reflect the fact that adding new parameters always results in some improvement in model fit). Preliminary results of these analyses are summarized in Fig. 3. A variety of interesting trends are apparent. As is true for many assessments of extinction risk, geographic range (global and/or Laurentian) is an important determinant of extinction risk in most intervals, with wider-ranging genera less likely to go at extinct in any given interval. Proportional range

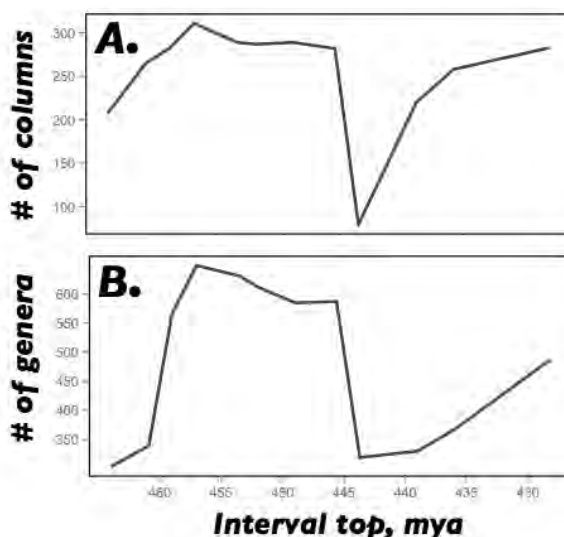


Figure 2. Time series of total number of stratigraphic columns with sediments of a given age (A) and total genus diversity (B) for Laurentia.

truncation, on the the other hand, has a major influence on extinction risk only during some intervals in the Katian and, especially, at the Katian-Hirnantian boundary. Slmilarly, the proportion of a genus' range that experiences an environmental shift has a significant influence on extinction risk primarily in the late Katian and at the Katian-Hirnantian boundary. Another striking trend relates to the preference for carbonate or clastic environments -carbonate-preferring genera are at lower risk of extinction than clastic-preferring genera throughout much of the Late Ordovician, but this trend reverses at in the Hirnantian and remains reversed throughout much of the Early Silurian. Finally, whether or not a genus has previously been sampled at high latitudes has a significant influence on extinction risk only at the Katian-Hirnantian boundary. This coincides with the major drop in tropical temperatures, and implies that in Laurentia extinctions driven by local climate change (as opposed to far-field eustatic effects) are largely limited to this interval. We emphasize that our conclusions are preliminary and may change as we continue to refine the chronostratigraphic, taxonomic, and paleoenvironmental framework of the dataset. However, these preliminary results are generally consistent with expectations from existing and emerging Late Ordovician-Early Silurian paleoclimatic datasets, and support a direct link between glaciation and mass extinction.

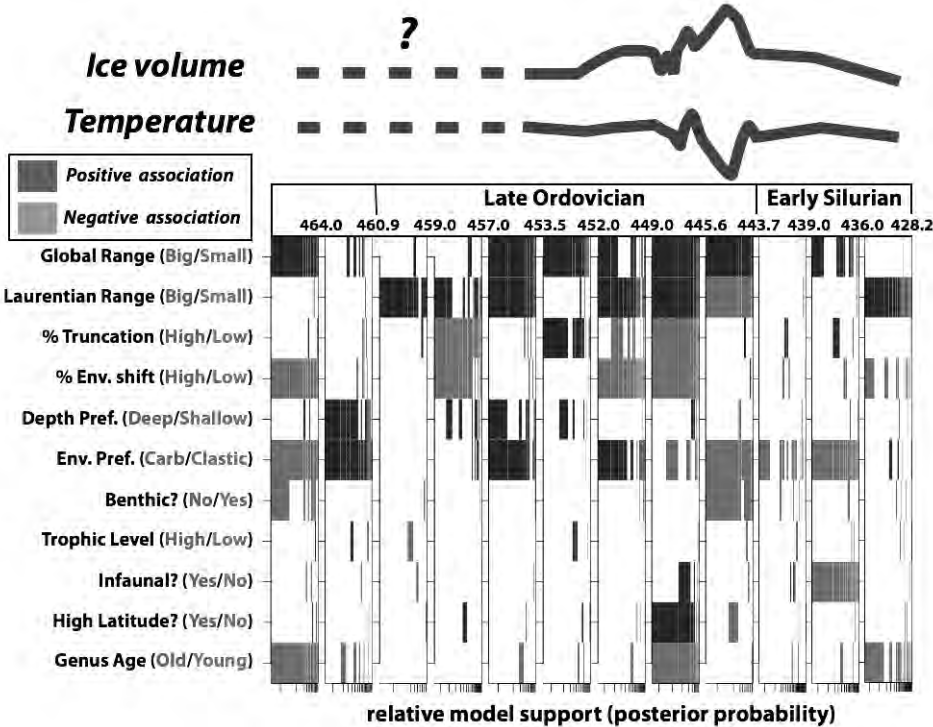


Figure 3. Results from Bayesian model averaging. All possible combinations of explanatory variables (variables indicated by text on left) are considered and the set of models that explain the most variation in extinction risk with the fewest model parameters are selected. The number of selected model is indicated by the hash marks along the x-axis for each interval; the x axis is scaled to the proportional support for each model (posterior probability) relative to the full set of most-likely models. Dark gray indicates a positive association between the predictor and extinction risk; light gray indicates a negative association, and white indicates that the variable is not included in the model in question. Ice volume and temperature trends are modified from Finnegan et al. (2011). Bottom and top ages of time intervals analyzed and indicated across below ice volume and temperature trends.

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GSSP BOUNDARY INTERVALS ARE CRITICAL FOR CHARACTERIZATION AND CORRELATION OF CHRONOHORIZONS THAT DEFINE GLOBAL STAGES, SERIES, AND SYSTEMS

S.C. Finney

Department of Geological Sciences, California State University-Long Beach,
Long Beach, CA, USA 90840, scfinney@csulb.edu

Global Boundary-Stratotype Sections and Points (GSSPs) define chronostratigraphic boundaries for a single set of global chronostratigraphic units (stages, series, and systems) and their corresponding geochronologic units (ages, epochs, and periods). The chronostratigraphic boundary is defined in a single section (stratotype) at a specific horizon or level that is marked by a single, distinct stratigraphic signal (bio-, chemo-, paleomagneto-, and/or sequence-stratigraphic). To be useful, this point that defines the lower boundary of a stage and also, in some instances, its encompassing series and system must be correlative as widely as possible geographically and between facies and with the highest level of resolution possible given the available correlation tools. However, to be correlated with confidence, the boundary marking stratigraphic signal must be evaluated for its consistent position with regard to other stratigraphic signals within the boundary interval. Excursions in stable isotopes, paleomagnetic polarity reversals and eustatic sea-level changes are not unique stratigraphic signals. Boundaries defined on them can only be correlated to sections away from the stratotype by first correlating the boundary interval with reasonable confidence. For boundaries of stages in the Phanerozoic, biostratigraphy provides the unique stratigraphic signals that allow for these initial correlations. Furthermore, even correlation of a boundary that was defined biostratigraphically (lowest or highest occurrence of a single taxon) can be made with confidence only if the biostratigraphic signal is characterized in the stratotype section by its position relative to all other stratigraphic signals in the boundary interval, particularly biostratigraphic ones. The precision and confidence in the correlation of a boundary away from the stratotype section is, of course, dependent on the stratigraphic signals available in the other section(s). The other section(s) may have a significant hiatus at the boundary level, and it (they) may have a very poor record of stratigraphic signals. The stratigraphic signal that was used to mark the boundary in the stratotype may be absent in other section(s). Nevertheless, if an extended boundary interval in the stratotype section has been well characterized by many, varied stratigraphic signals, then the approximate stratigraphic interval in which the boundary falls in the deficient section(s) can be identified and a useful chronocorrelation can be made.

The GSSP that defines the lower boundary of the Sandbian Stage and Lower Ordovician Series, i.e., the level of the lowest known occurrence (LO) of the graptolite *Nemagraptus gracilis* in the section at Fågelsång, Sweden, well illustrates the importance of characterizing the boundary interval. In black shale facies at Fågelsång, as well as at Calera, Alabama, USA and Dawangou, Xinjiang Province, China, the LO of *N. gracilis* is within a thin stratigraphic interval that includes the LO's of species of *Dicellograptus*. *D. geniculatus* has a relatively very short stratigraphic range in all three sections, and the LO of *N. gracilis* is within this stratigraphic

range. The highest occurrence of *N. subtilis*, interpreted as the direct ancestor of *N. gracilis* (Finney, 1986; Finney and Bergström, 1986), occurs directly below the LO of *N. gracilis* in the Fågelsång and Calera sections (Bergström and Finney, 2000). Also, in all three sections, the LO of *N. gracilis* is within a thin stratigraphic interval in which species of *Dicellograptus* and *Dicranograptus* follow directly above the LO of *D. geniculatus*. These include *D. vagus*, *D. gurlleyi*, *D. sextans*, *D. salopiensis*, *D. alabamensis*, *D. bispiralis*, and *Dicranograptus irregularis*. Furthermore, the LO of *N. gracilis* in all three sections is stratigraphically slightly above the evolutionary transition between the conodonts *Pygodus serra* and *P. anserinus*. Thus, many biostratigraphic signals, including the single one marking the boundary in the stratotype, are consistent in three sections on different continents and paleo-plates (Bergström and Finney, 2000). However, in south Wales and Shropshire, Britain, the LO of *N. gracilis* is considered to be much higher stratigraphically relative to the LO's of species of *Dicellograptus*, and, for this reason, the reliability of the LO of *N. gracilis* for chronocorrelation away from the stratotype is called into question (Bettley et al., 2001). Its proposed inconsistent position may be due to sampling failure. The sections in south Wales and Shropshire are relatively thick but poorly exposed and covered over thick stratigraphic intervals. Furthermore, sampling density is very low with intervals of 50 m or more between collection levels. Regardless, even if the LO of *N. gracilis* is atypically high in the British sections, which could naturally arise from its late migration into the area or to its not being preserved in lower stratigraphic intervals, the boundary can be correlated with confidence into the relatively thin stratigraphic interval that includes the LO's of several species of *Dicellograptus* and *Dicranograptus irregularis* and which is stratigraphically above the boundary between the conodont zones of *P. serra* and *P. anserinus*. The focus on correlation of the base of the Sandbian Stage and Upper Ordovician Series solely on the LO of *N. gracilis* leads to unwarranted criticism of the GSSP. It is based on the belief, in this instance, that the occurrence of such a distinct stratigraphic signal marking the boundary in the stratotype (the LO *N. gracilis*) should maintain the same exact same stratigraphic position in all sections where it is discovered, ignoring the expected, natural possibilities of collection failure, lack of preservation, and late migration into the basin of deposition. It is a belief that is exacerbated by publications that illustrate only the single stratigraphic signal that marks the boundary in the stratotype section (e.g., Ogg et al., 2008). This example for the base of the Sandbian Stage and Upper Ordovician Series demonstrates the importance of characterizing boundary intervals by the consistent relative positions of many, varied stratigraphic signals.

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THE LATE TREMADOCIAN–EARLY ARENIG 2ND ORDER SEQUENCE OF THE CADENAS IBÉRICAS (NE SPAIN) AND ITS COMPARISON WITH BALTICA

J.A. Gámez Vintaned¹ and U. Schmitz²

¹ Depto. de Geología, Universitat de València. Dr. Moliner 50, E-46100 Burjassot (Valencia), Spain.

j.antonio.gamez@uv.es, gamez@unizar.es

² LO & G Consultants, Baderweg 149, D-45259 Essen, Germany. ulrich.schmitz@lo-g.de

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INTRODUCTION

Lately, the authors of this study had subdivided the Late Vendian–Early Ordovician succession of the Cadenas Ibéricas (Fig. 1) into 2nd order sequences (Schmitz, 2006; Gámez Vintaned et al., 2009), with the aim to place the succession into the regional and/or global context and to correlate the sequences – subject to verification – with those of adjoining areas, particularly of Northwestern Gondwana.

At about the same time, global sea-level charts covering the entire Palaeozoic succession, were published (Haq and Schutter, 2008; Snedden and Liu, 2010). The basis of these global charts are analyses of successions predominantly from cratonic and peri-cratonic areas. Reference districts of the Ordovician interval under study are those of North America and Australasia, complemented in certain parts by Estonian sections (identified on Fig. 2). Presented are 3rd order sequences (“short-term”) and their “long-term” envelope (see Fig. 2). They are controlled by absolute age dating and specific biozones. At Ordovician level, index fossils of the biozones are conodonts and graptolites.

Obviously, the data base in the Cadenas Ibéricas differs from that available for the global curve. The fact that (1) in the Cadenas Ibéricas the sedimentary pattern of the investigated levels reflects fast subsidence under sag phase conditions (Gámez Vintaned et al., 2009), that (2) only 2nd order sequences have been established, and that (3) fundamentally only trace fossils, brachiopods and rarely trilobites are at hand for age determination, makes correlation with the 3rd order sequences established by Haq and Schutter (2008) difficult. Upon applying, however, the results from the analyses of Nielsen (2003) – who, for Baltica, at the levels of interest had established not only 2nd order sequences but had also underpinned those by 3rd order drowning events – it seems possible to recognize the pattern of the studied Cadenas Ibéricas succession in the equivalent succession of Baltica, and as a consequence to critically review earlier suggested allocations.

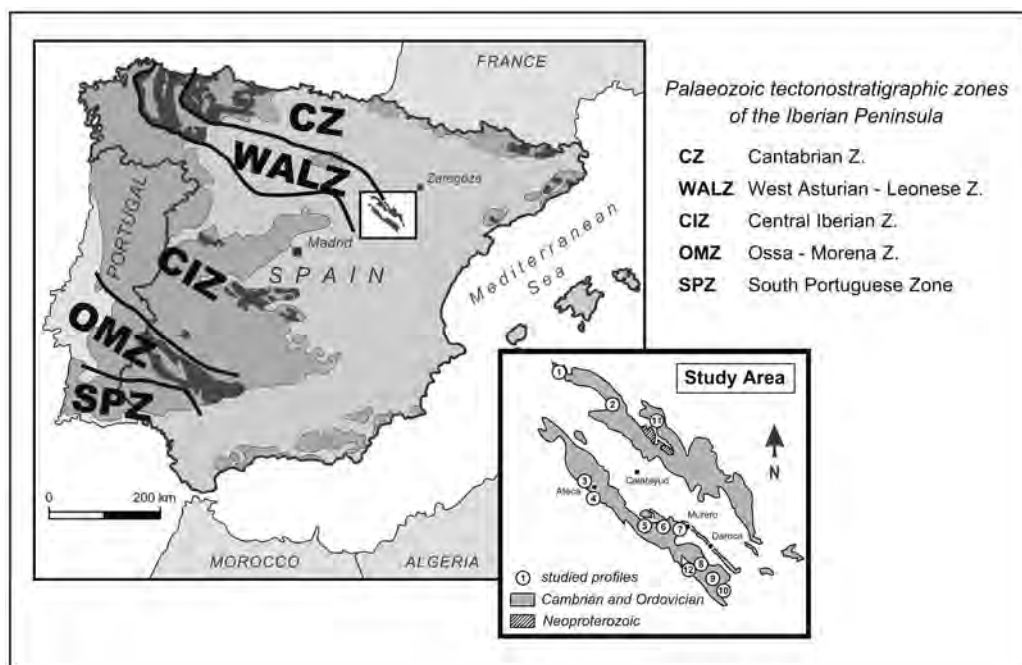
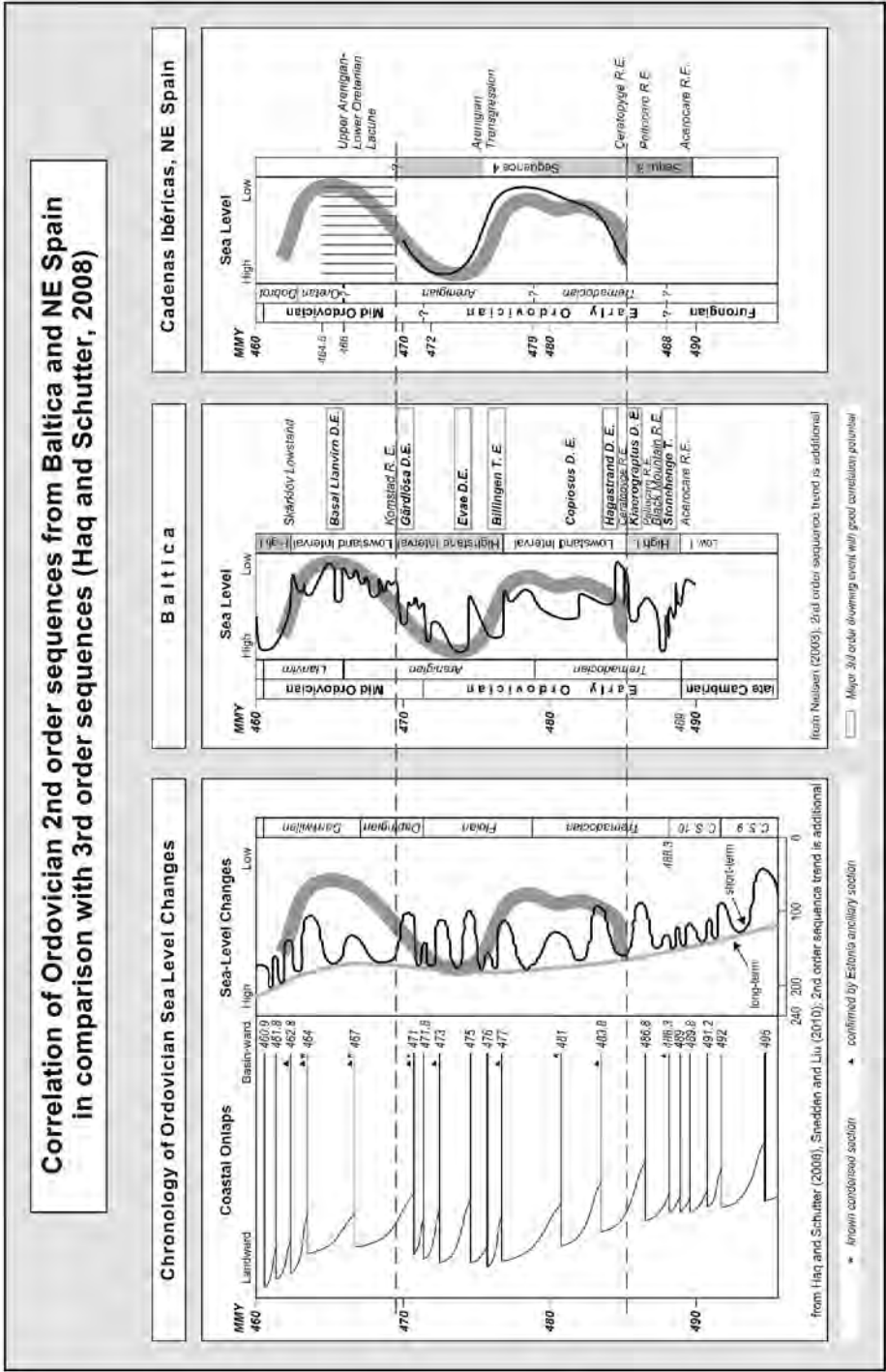


Figure 1. Geological setting of the Cadenas Ibéricas in the context of the Iberian Massif. General map: slightly modified from von Raumer et al. (2006). Inset map: from Schmitz (2006).

THE 2ND ORDER SEQUENCES OF THE BALTIC CRATON

The Baltic (Russian) Craton had been covered at early Palaeozoic times by an epicontinental sea (Nielsen, 2003). The Ordovician deposits reflect differentiated shelf conditions (Kanev et al., 2001), they are dominated by shales and limestones, with coarser clastics only in their basal part. The sedimentary succession of the Ordovician is being summarized by Nielsen (2003) as comprising three highstand intervals and three lowstand intervals. Subject subdivision is being accompanied by a detailed sea level curve (see Fig. 2).

According to Nielsen (2003), the Tremadocian–Arenig succession is covered by two high and low stand intervals each. The Ordovician starts with the early–mid Tremadocian High Stand interval which follows on the *Acerocare* Regression. Obviously that represents the younger part of a full 2nd order sequence, which in turn starts within the Furongian (not having been dealt with by Nielsen, 2003). Follow the intervals of primary interest in the context of this study, which are the Late Tremadocian–earliest Arenig Lowstand interval and the early Arenig Highstand interval. They form the Late Tremadocian–Early Arenig 2nd order sequence. It is bound by the *Ceratopyge* Regression at its base and by the Komstad Regression at its top, whilst the sequence's lowstand and highstand intervals are separated by the Billingen Transgression. The sequence, as a whole, is adequately controlled by index fossils and shows a good fit with the global curve (see Fig. 2). To facilitate comparison with the suggested sea level curve derived from the Cadenas Ibéricas, on Figure 2 a trend line is being superimposed on the detailed sea level curve (and duplicated on the



respective sections of the global and Cadenas Ibéricas curves). Comparing the Nielsen (2003) curve with the global one, it is noted that the lowstand intervals show similarities to a lesser degree than the highstand intervals. Upwards, the sequence passes into a conspicuous lowstand interval trend which in its lower part is of late Arenig age. That trend is used in the comparison with the sedimentary pattern interpreted from the Cadenas Ibéricas succession (see below).

THE LATE TREMADOCIAN–EARLY (?) ARENIG SEQUENCE OF THE CADENAS IBÉRICAS

This 2nd order sequence likely represents the closing succession of the Middle Cambrian–Early Ordovician sag phase, which had been preceded by a rift phase (Gámez Vintaned et al., 2009). Age identification within the sequence is poor (see Gutiérrez-Marco et al., 2002), in particular some uncertainty exists as to the exact position of the Tremadocian–Arenig boundary. The base of the sequence has been placed by Schmitz (2006) at the *Ceratopyge* Regression level, on the basis of circumstantial evidence. The sequence top is the top of the “Armorican Quartzite”, with no biozone control of its age allocation. (Middle Arenigian graptolites were recovered from Armorican facies of the Cantabrian Zone, N Spain, while the oldest graptolites found elsewhere in shaly levels above the Armorican formation are middle to late Arenigian in age; Gutiérrez-Alonso et al., 2007, and references therein.) With reference to fossil assemblages in the Castillejo Formation (the unit overlying the “Armorican Quartzite” in the Cadenas Ibéricas), a late Oretanian age was evidenced (Gutiérrez-Marco et al., 2002). From the biostratigraphic data mentioned, it is concluded that we face a gap between the “Armorican Quartzite” and the Castillejo Formation, referred to as “upper Arenigian-lower Oretanian lacune”.

The sequence (see Fig. 3) consists of a lower part, the Santed Formation, which is dominated by shales, with sandstone intercalations at various levels, and of an upper part, the “Armorican Quartzite”. The latter comprises sandstones and quartzites as well as shale intercalations and shale intervals. The lower parts of the Santed Formation represent a lowstand phase, with turbidites characterizing the slope phase and the coarsening upward trend characterizing the progradational wedge phase. Somewhat generalized, the succession in the Santed Formation below the “Armorican Quartzite” and in the lower part of the “Armorican Quartzite” is suggested to represent the transgressive systems tract, whilst highstand deposits seem to dominate its upper parts, as indicated by the listing of sedimentological and ichnological features (Schmitz, 2006).

The suggested sea level curve related with the sequence is shown on Figure 3. It is a generalized curve, to be modified as additional information becomes available. Likewise, the conversion of that Figure 3-curve into a time-controlled sea level curve involves uncertainties, mirroring the little age control along the 2nd order sequence. One way to assess the reliability of the conversion, under the given circumstances, is the comparison with an established, time-controlled curve. In the case of the Cadenas Ibéricas sequence, as will be noted on Figure 2, correlation with the Nielsen (2003) curve only fits if its highstand phase is being moved, ending at intra-late Arenigian times. Consequently, the match for the entire curve fits only if the belief is being abandoned that the top of the “Armorican Quartzite” equals the top of the Arenig (which is also supported by biostratigraphic data mentioned above; cf. Gutiérrez-Alonso et al., 2007).

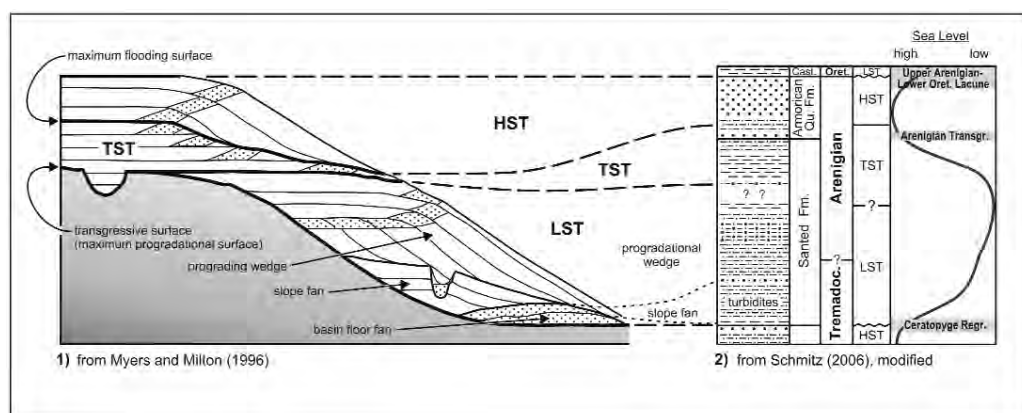


Figure 3. Suggested systems tracts and sea level curve of the Tremadocian–Arenig 2nd Order Sequence in the Cadenas Ibéricas. Cast., Castillejo Formation. HST, highstand systems tract. LST, lowstand systems tract. Oret., Oretanian. Qu., Quartzite. TST, transgressive systems tract.

SUMMARY AND CONCLUSIONS

It is suggested that the depositional history of the Tremadocian–Arenig clastic succession in the Cadenas Ibéricas was dominated by a 2nd order sequence. The exact stratigraphic positions of the sequence phases cannot be verified, owing to poor age-dating. To still enable correlation with globally established sequences, the succession is being compared with the equivalent succession of Baltica. That succession had been analyzed for 2nd order sequence subdivision, it is reliably controlled through age-dating and 3rd order sequence events (Nielsen, 2003). The comparison suggests that the correlation proves an effective tool in as much as it shows good agreement for the major part of the succession and that it points to a discrepancy which ought to be addressed, relative to the top part of the section: there, the conventional belief that top “Armorican Quartzite” and top Arenig are identical is incorrect and needs to be critically reviewed.

Acknowledgements

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STRATIGRAPHIC EVIDENCE FOR THE HIRNANTIAN (LATEST ORDOVICIAN) GLACIATION IN THE ZAGROS MOUNTAINS, IRAN

M. Ghavidel-syooki¹, J.J. Álvaro², L. Popov³, M. Ghobadi Pour⁴, M.H. Ehsani¹ and A. Suyarkova⁵

¹ Institute of Petroleum Engineering, Technical Faculty of Tehran University, P.O. Box 11365-4563, Tehran, Iran.

m_ghavidelsyooki@yahoo.com, mh_ehsani@hotmail.com

² Centre of Astrobiology (CSIC/INTA), Ctra. de Torrejón a Ajalvir km 4, 28850 Torrejón de Ardoz, Spain. jjalvaro@unizar.es

³ Department of Geology, Natural Museum of Wales, Cardiff, Cathays Park, Cardiff CF10 3NP, Wales, UK,

Leonid.Popov@museumwales.ac.uk

⁴ Department of Geology, Faculty of Sciences, Golestan University, Gorgan 49138-15739, Iran. m.ghobadipour@gu.ac.ir

⁵ Department of Stratigraphy and Palaeontology, Russian Geological Research Institute (VSEGEI), 74 Sredniy prospect, 199106 St. Petersburg, Russia. Anna_Suyarkova@vsegei.ru

High-latitude Hirnantian diamictites (Dargaz Formation) and lower-Silurian kerogenous black shales (Sarchahan Formation) are spotty exposed in the Zagros Mountains. The glaciogenic Dargaz deposits consist of three progradational/retrogradational cycles, each potentially controlled by the regional advance and retreat of the Hirnantian ice sheet. Glacial incisions of sandstone packages change laterally from

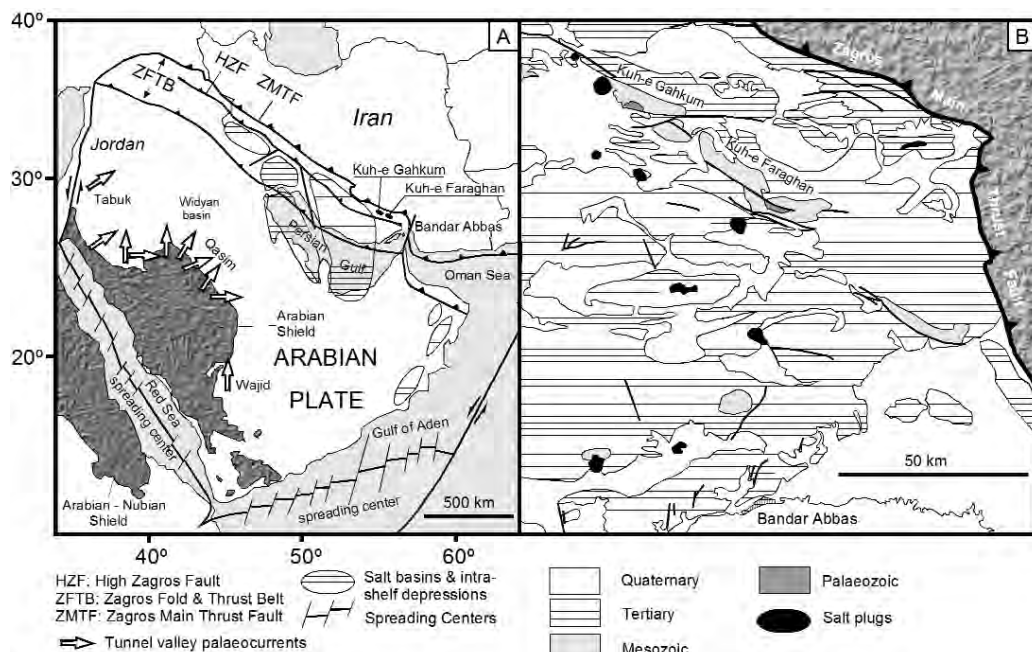


Figure 1. A, Major tectonic features of the Arabian Plate, the Zagros Mountains, and adjacent areas with setting of tunnel-valley palaeocurrents. B, Geological map of the study areas in the southeastern Zagros Fold and Thrust Belt, North of Bandar Abbas.

simple planar to high-relief (< 40 m deep) scalloped truncating surfaces that join laterally forming complex polyphase unconformities that scour into the underlying Seyahou Formation (Katian). The glaciated source area was to the present-day West, in the region of the Arabian Shield, where numerous tunnel valleys have been reported. Based on a study of palynomorphs and graptolites, the glaciomarine Dargaz diamictites are dated as Hirnantian, whereas the younger Sarchahan black shales are diachronous throughout the Zagros, ranging from the Hirnantian *persculptus* to the earliest Aeronian (Llandovery) *triangulatus* zones. The diachronism is related to onlapping geometries capping an inherited glaciogenic palaeorelief that preserved different depth incisions and source areas. Our data suggest the presence of Hirnantian satellite ice caps neighbouring the Zagros margin of Arabia and allow us to fill a gap in the present knowledge of the peripheral extension of the Late Ordovician ice sheet.

NEW DATA ON THE LATE ORDOVICIAN TRILOBITE FAUNAS OF KAZAKHSTAN: IMPLICATIONS FOR BIOGEOGRAPHY OF TROPICAL PERI-GONDWANA

M. Ghobadi Pour¹, L.E. Popov², L. McCobb² and I.G. Percival³

¹ Department of Geology, Faculty of Sciences, Golestan University, Gorgan, Iran.
mghobadipour@yahoo.co.uk. m.ghobadipour@gu.ac.ir

² Department of Geology, National Museum of Wales, Cardiff CF10 3NP, Wales, United Kingdom.
leonid.popov@museumwales.ac.uk, lucy.mccobb@museumwales.ac.uk

³ Geological Survey of NSW, 947-953 Londonderry Road, Londonderry 2753, New South Wales, Australia.
ian.percival@industry.nsw.gov.au

Keywords: Late Ordovician, Gondwana, Kazakhstan, trilobites, biogeography.

INTRODUCTION

During the Late Ordovician, microplates and volcanic arc systems presently incorporated into the Kazakhstanian orogen converged to form a huge archipelago, which extended far into the ocean along subequatorial latitudes west of the tropical Australasian sector of Gondwana (Popov et al., 2009). The shelves of volcanic islands and microcontinents within this archipelago supported diverse benthic faunas, with trilobites as one of the most important components. Kazakhstanian Late Ordovician trilobite faunas have been documented in a number of publications (Ghobadi Pour et al., 2011; Koroleva, 1982 and references therein). Koroleva (1982) gave up to date summaries with outlines of taxonomic diversity and geographical distributions of trilobite taxa throughout Kazakhstan. A total of about 110 genera and more than 200 species were counted, but their generic affiliation often requires revision. Apollonov (1975) published a brief review of the Kazakhstanian trilobite biofacies, whereas Fortey and Cocks (2003) gave a brief outline of biogeographic affinities of Kazakhstanian faunas throughout the Ordovician, mainly based on personal assessment of unpublished collections by Richard Fortey. Nevertheless, existing data on characters of trilobite faunas from individual terranes are still incomplete and there was little progress in their study during the last 25 years.

In spite of significant losses of collections and geological information after the collapse of the Soviet Union, there is a substantial amount of unpublished data which has been preserved and is available for study. It includes an enormous trilobite collection assembled by the late Michael K. Apollonov, which covers almost all areas in Kazakhstan where Ordovician deposits are present. In addition to published information, new data are presented in this paper, mainly based on a preliminary assessment of the samples available from Apollonov's collections, which are currently under study.

ORDOVICIAN PALAEOGEOGRAPHY OF KAZAKHSTANIAN TERRANES

Three major clusters of early Palaeozoic terranes can be recognised in the Kazakhstani orogen. The southern cluster includes three major crustal terranes (i.e. Chu-Ili, North Tien Shan and Karatau-Naryn), which were amalgamated together by the Late Silurian (Popov et al., 2009) (Fig. 1). The published record of Mid to Late Ordovician trilobite faunas of Chu-Ili is the most complete in comparison with other regions of Kazakhstan, whereas it is virtually nonexistent for North Tien Shan (Ghobadi Pour et al., 2009; Koroleva, 1982 and references here).

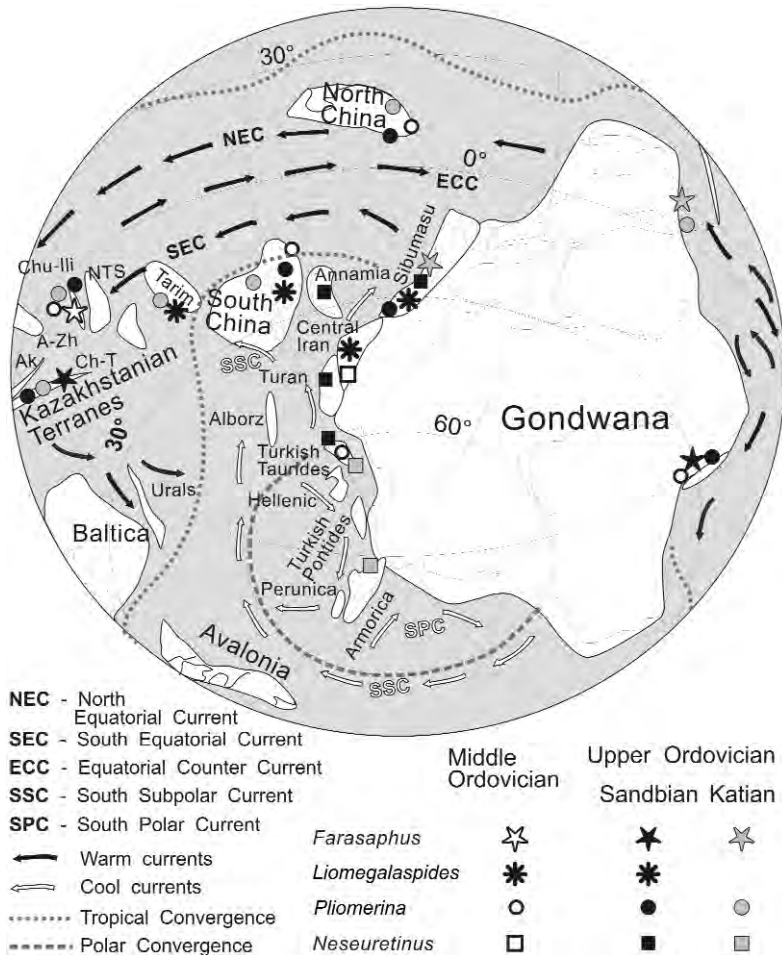


Figure 1. Palaeogeographical reconstruction for the Upper Ordovician (Katian) showing geographical distribution of selected biogeographically informative trilobite genera. Position of the major early Palaeozoic continents mainly after Fortey and Cocks (2003) with emendations after Popov et al. (2009). Surface water circulation for a Northern Hemisphere summer is mainly after Wilde (1991). Abbreviations for Kazakhstani island arcs and microplates are as follows: A-Zh – Atasu-Zhamshi, Ak – Akbastau, Ch-T – Chingiz-Tarbagatai, K-N – Karatau-Naryn, NTS – North Tien Shan.

The southern cluster of Kazakhstanian terranes is separated by an oceanic suture from the Atasu-Zhamshi microplate (Apollonov, 2000; Popov et al., 2009) (Fig. 1). The area north-east of Atasu-Zhamshi represents a complicated mosaic of island arc and continental fragments, separated by ophiolitic belts associated with sutures and often strongly reworked since the Early Palaeozoic (for summary, see Popov et al., 2009; Windley et al., 2007). At least three major island arc systems can be recognised, including Akbastau, Chingiz-Tarbagatai and Boshchekul. Published information on the Late Ordovician trilobite faunas of the Chingiz-Tarbagatai and Boshchekul regions was reviewed by Koroleva (1982 and references here) and recently by Ghobadi Pour et al. (2011).

Another group of early Palaeozoic terranes are those of north-central Kazakhstan, i.e. the Kalmyk Kokchetav unit of Şengör & Natal'in (1996) or Shatsk and Kokchetav microplates of Dobretsov et al. (2006) and adjacent island arcs. Data on the Neoproterozoic to Early Palaeozoic geological history of this north-central sector of the Kazakhstanian orogen, provided by Dobretsov et al. (2006), substantiates the idea that these units did not interact with the south Kazakhstanian cluster of terranes throughout the Cambrian-Ordovician. Koroleva (1982) published a detailed outline of Late Ordovician trilobite distributions in the Selety, Ishim and Stepnyak regions, based in a significant part on her earlier publications.

TRILOBITE BIOFACIES

In the late Darriwilian – Sandbian, asaphid-illaenid biofacies were characteristic for inshore environments in almost all Kazakhstanian terranes, but they are well documented only for Chu-Ili. A good example is the monotaxic '*Isotelus*' *romanovskyi* Association of Apollonov (1975), which spread widely on a shallow clastic shelf across Chu-Ili. It replaced the lingulid *Ectinoglossa* Association seaward and was confined to a sandy bottom, nearshore setting, inhabited mainly by gastropods and bivalved molluscs. '*Isotelus*' *romanovskyi* Weber, 1948 is probably assignable to *Damiraspis* (Fig. 2.11-13), but hypostome morphology in this species is as yet unknown. On the shallow carbonate shelf of Chu-Ili, the asaphids *Damiraspis* and *Farasaphus* formed oligotaxic communities, usually in association with the endemic illaenid *Alperillaenus* (Fig. 2.1, 6-7) as a second major component. Other minor components comprised *Ceraurina*?, *Eorobergia*, *Pliomerina* (Fig. 2.21) and *Sphaerexochus* (Ghobadi Pour et al., 2009).

The pliomerid-styginid biofacies first emerged during the Sandbian. At that time, it was most characteristic for silty bottom, nearshore settings and probably occupied a quiet environment, affected occasionally by seasonal storms. During the Sandbian, these biofacies were dominated by styginids, namely *Dulanaspis*, *Styginella* and *Bronteopsis*. Other common taxa are *Lonchodomas*, *Pliomerina*, *Remopleurides* and *Sinocybele* (Fig. 2.22), whereas asaphids are rare to almost absent. During the Katian, these associations gradually replaced asaphid-dominated associations nearshore, and there were changes in the taxonomic composition of the assemblages. *Pliomerina* and *Remopleurides* proliferated and became dominant by the mid-Katian, whereas the proportion of styginids gradually declined.

The illaenid-cheirurid biofacies was confined to the carbonate build-ups, which became widespread throughout Kazakhstanian island arcs and microcontinents in the Sandbian–Katian. This biofacies was characterised by rich generic diversity, but remains very poorly known. In addition to the nominative families, asaphids, lichids, pliomerids, remopleuridids, raphiophorids and styginids usually occur. Such genera as *Acrolichas* (Fig. 2.5, 10), *Eokosovopeltis*, *Glaphurina*, *Holotrachelus*, *Metopolichas* and *Sphaerexochus* (Fig. 2.8-9) are the most characteristic. The nileid biofacies is known from the offshore

environment of almost all major Kazakhstani terranes. Faunas characteristic of this biofacies usually lack distinct dominant taxa and may be rather diverse. For example, a nileid association recently described from the lower Katian Karagach Formation of the Tarbagatai Range (Ghobadi Pour et al., 2011) contains 15 different trilobite genera, including leiostegiids (*Aegirina*), asaphids (*Birmanites*), encrinurids (*Encrinuroides*, *Sinocybele*), remopleuridids, raphiophorids and shumardiids (Fig. 2.4, 17-19).

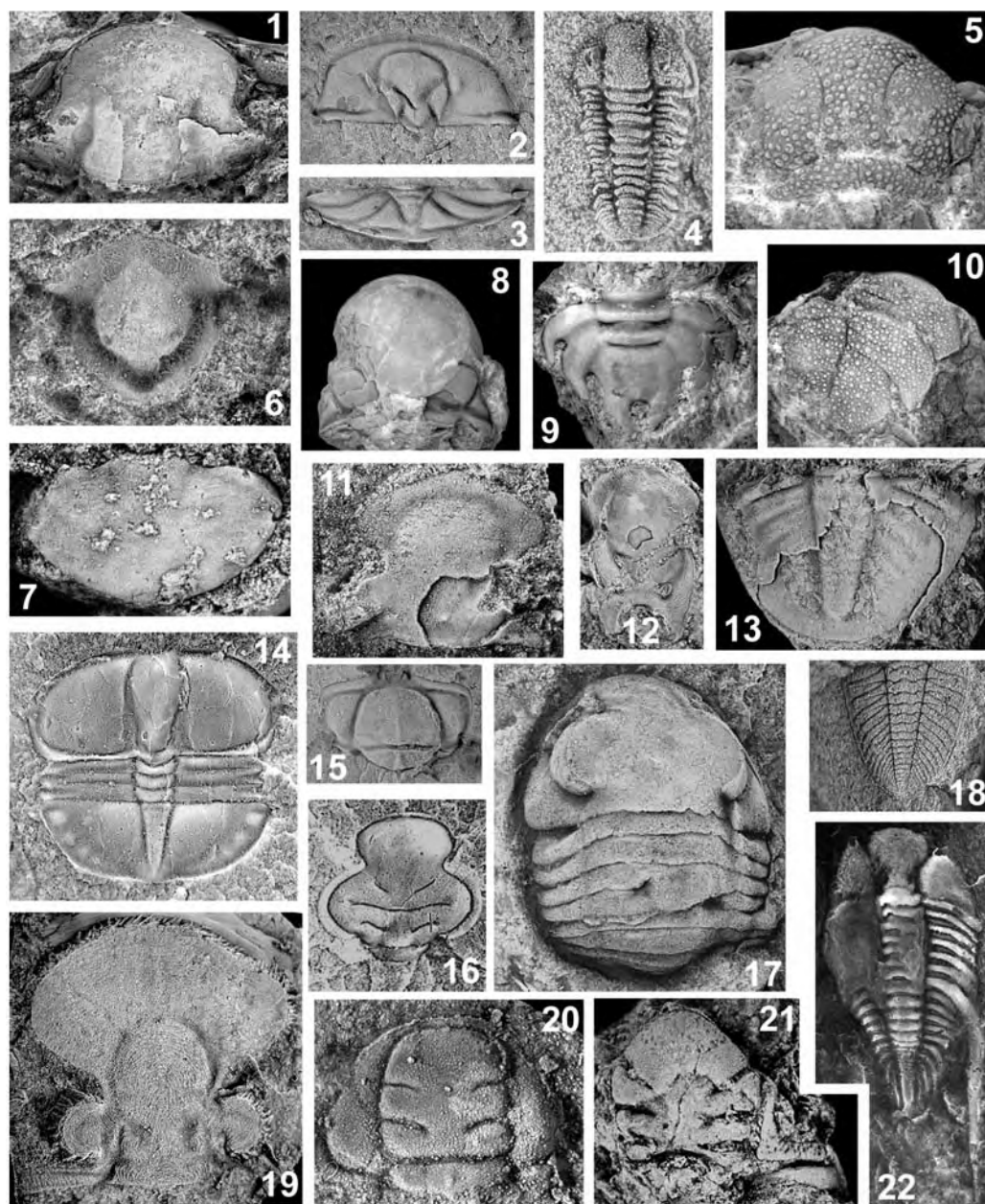
The raphiophorid biofacies probably occupied the disphotic zone in deeper water offshore. Trilobite taxa characteristic of this biofacies are often blind (many raphiophorid genera), or possess hypertrophic eyes (e.g. *Arator* and *Telephina*). Trilobite associations of this biofacies may be oligotaxic (e.g. *Bulbaspis* Association from the Dulankara Formation of Chu-Ili terrane), or display remarkable taxonomic diversity with more than 25 genera (*Caganaspis* Association from the Bestamak Formation of the Chingiz Range). The list of characteristic genera includes the three-segmented raphiophorid *Caganaspis* (Fig. 2.14) and remopleuridids (e.g. *Arator*, *Eorobergia*, *Robergia*?) (Fig. 2.16) that are widespread in the Chu-Ili, Boshchekul and Chingiz-Tarbagatai terranes, but are as yet unknown outside Kazakhstan. There are also more widespread taxa, e.g. *Ampyxinella*, *Bulbaspis*, *Endymionia*, *Birmanites*, *Dionide* and *Telephina* (Fig. 2.2, 3, 15), which are also documented from the Australasian sector of Gondwana. A significant proportion of Kazakhstani faunas characteristic of the raphiophorid biofacies remain formally undescribed.

The deepest water olenid biofacies was not previously documented from Kazakhstan. In Atasu-Zhamshi, the olenid *Porterfieldia* occurs in association with *Endymionia* in black limestones of the Shundy Formation (Sandbian). The only other fossils to occur at that locality are radiolarians. In Chu-Ili, olenid trilobites occur in black graptolitic shales of Katian age, exposed on the Akkermere Peninsula on the western coast of Balkhash Lake. In this locality, the olenid *Triarthrus* (Fig. 2.20) occurs in association with *Dionide*, *Caganaspis* and a new, as yet undescribed, harpetid genus.

IMPLICATIONS FOR BIOGEOGRAPHY

In spite of incomplete knowledge of Kazakhstani trilobite faunas, there is good evidence that during the Late Ordovician they exhibited similar biogeographical signatures, suggesting affinity to the *Eokosovopeltis-Pliomerina* Province of Webby et al. (2000). Indeed, *Eokosovopeltis* and *Pliomerina* proliferated on the shallow shelves of all major Kazakhstani terranes. However, significant work is still needed to establish the faunal signatures of individual Kazakhstani island arcs and microplates.

Figure 2. Selected Ordovician trilobites from Kazakhstan. Specimens deposited in the National Museum of Wales Cardiff (NMW), and F.N. Chernyshev Central Geological Scientific Research and Exploration Museum (CNIGR), St Petersburg. 1, 6, 7, *Alperillaenus intermedius* Ghobadi Pour and Popov, 2009; Darriwilian, Kypchak Limestone, northern Betpak-Dala; 1, NMW 2008.34G.3, cranium, x2; 6, NMW 2008.34G.11, hypostome, x5.5; 7, NMW 2008.34G.9, pygidium, x5.5. 2, 3, *Ampyxinella balashovae* Koroleva, 1965; Sandbian, Sarytuma, West Balkhash Region; 2, NMW2008.34G.150, internal mould of cranium, x3; 3, NMW2008.34G.151, internal mould of pygidium, x3. 4, *Agerina acutillimbata* Ghobadi Pour et al., 2011, Katian, Karagach Formation, east side of the Ayaguz River, about 7 km north of Akchii village, Tarbagatai Range; NMW 2005.32G.135, holotype, articulated exoskeleton, latex cast, x5. 5, 10, *Acrolichas clarus* Koroleva, 1959, Sandbian, Myatas Formation, northern coast of Atansor Lake; 5, NMW2008.34G.155, cranium, x4; 10, NMW2008.34G.156, incomplete pygidium, x4. 8, 9, *Sphaerexochus consuoides* Koroleva, 1959; age and locality as Fig. 2.5; 8, NMW2008.34G.157, cranium, x2.5, 9, NMW2008.34G.158, pygidium, x2.2. 11-13, *Damiraspis margiana* Ghobadi Pour and Popov, 2009, age and locality as Fig. 2.1; 11, NMW 2008.34G.46, cranium, x4. 12, NMW 2008.34G.42, holotype, hypostome, x1.1; 13, NMW 2008.34G.48, partly exfoliated pygidium, x2.5. 14, *Caganaspis unica* Kolobova, 1985, area about 7 km south-west of Alakul Lake, West Balkhash Region, NMW2008.34G.149, articulated exoskeleton, latex cast, x4.5. 15, *Telephina omega* Koroleva, 1982, age and locality the same as Fig. 2.2; NMW2008.34G.152, internal mould of cranium, x3.5. 16, *Robergia*? sp., age and locality the same as Fig. 2.14; NMW2008.34G.153, cranium, latex cast, x5. 17, *Nileus* sp., age and locality the same as Fig. 2.4;



NMW 2005.32G.193, cephalon with attached thoracic segments, latex cast, x6. 18, *Aethedionide* sp., Sandbian, Karagach Formation, locality as Fig. 2.4; NMW 2005.32G.191, pygidium, internal mould, x4. 19, *Birmanites akchiensis* Ghobadi Pour et al., 2011, age and locality the same as Fig. 2.4; NMW 2005.32G.181, holotype, cranidium, latex cast of external mould, x1.5. 20, *Triarthrus* sp., Katian, Ak-Kerme Peninsula, west coast of Balkhash Lake, NMW2008.34G.154, cranidium, x9. 21, *Pliomerina* aff. *sulcifrons* (Weber, 1948), age and locality as Fig. 2.1; NMW 2008.34G.25, cranidium, x5. 22, *Sinocybele weberi* (Kolova, 1936), Katian, Besharyk Formation, Dzhebagly Mountains; CNIGR 60/4263, lectotype, incomplete dorsal exoskeleton, latex cast, x1.8.

Recent studies also demonstrate that the asaphid trilobites *Basilicus* and *Basiliella* were probably confined to peri-lapetus settings, while the Kazakhstani and Australasian species traditionally assigned to these genera in fact belong to separate asaphid lineages (Ghobadi Pour et al., 2009), which evolved independently in tropical peri-Gondwana and should be assigned to different genera (i.e. *Damiraspis* and *Farasaphus*). Remarkably, although these asaphids commonly occur in the Australian sector of Gondwana, they are absent from the Darriwilian–Katian rocks of South China, where they are replaced by genera of the Subfamily Nobiliasaphinae (e.g. *Liomegalaspides*). A similar pattern was observed for *Eokosovopeltis*, which is absent from the Sandbian to early Katian of South China (Zhou and Zhen, 2009). Zhou and Zhen (2009) recently suggested that Australian trilobite faunas had closest affinities with those of North China during the Arenig–Caradoc interval (=Sandbian–early Katian). It is likely that there was a continuous belt of tropical peri-Gondwanan, shallow water faunas during the Sandbian–early Katian, which included Kazakhstani terranes, North China and the Australian sector of Gondwana.

The most likely explanation can be found in features of oceanic surface circulation along the western coast of Gondwana (Fig. 1). It is well established that the Australian sector of Gondwana, North China and Kazakhstani microplates and island arcs occupied a subequatorial position in the Ordovician, whereas a more temperate latitude is evident for South China during the Early to Mid Ordovician, based on palaeomagnetic data, characteristics of shallow marine benthic communities and sedimentation (Fortey and Cocks, 2003). In particular, the occurrence in South China of trilobites from the Family Reedocalymeninae and *Taihungshania* represents a distinct link with temperate to high latitude Gondwanan faunas (e.g. Armorica, Turkish Taurids, Iran), whereas they are virtually absent from Kazakhstani terranes and the Australian sector of Gondwana. It is probable that a cool water, South Subpolar Current, running along the western Gondwanan coast (Wilde, 1991), might have an effect on climate comparable to the present-day Humboldt Current. As a result, average annual temperatures of surface waters along the coasts of the South China continent during the Early to Mid Ordovician were considerably lower than in subequatorial peri-Gondwana, which prevented the immigration of some warm water taxa. Only in the Katian, when South China entered low latitudes, did affinity with the shallow shelf faunas of the Kazakhstani terranes become firmly established.

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A CONOP9 COMPOSITE-TAXON RANGE-CHART FOR ORDOVICIAN CONODONTS FROM BALTOSCANDIA: A FRAMEWORK FOR BIODIVERSITY ANALYSES

D. Goldman¹, S.M. Bergström², H.D. Sheets³ and C. Pantle¹

¹ Department of Geology, University of Dayton, Ohio, 45469. dan.goldman@notes.udayton.edu

² School of Earth Sciences, The Ohio State University, Columbus, Ohio, 43210. stig@geology.ohio-state.edu

³ Department of Physics, Canisius College, Buffalo, New York, 14208. sheets@canisius.edu

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INTRODUCTION

Epicontinental Ordovician strata form the surface bedrock in many parts of Baltoscandia. Although condensed, this succession is remarkably complete stratigraphically and richly fossiliferous. The fact that limestone lithofacies dominates in the East Baltic and in most parts of Sweden, and these strata are readily available in numerous outcrops and drill cores has made it possible to easily obtain large collections of conodont elements. Pander (1856) first described conodonts from the St. Petersburg region of western Russia, but the modern era of research on conodont taxonomy and biostratigraphy did not start until the 1950s (e.g. Lindström, 1955). Since then, the Baltoscandic Ordovician conodont faunas have been dealt with in numerous papers and monographs (~10 in Norway; >40 in Sweden and Denmark; and a similar number in the East Baltic). Apart from the Lower Tremadocian, where the dominating clastic lithofacies yields few conodonts, the taxonomy and biostratigraphy of the Ordovician conodonts in Baltoscandia are now better known than in any comparable region in the world. The vertical ranges of more than 150 taxonomically well understood multielement species are now known in detail and we believe that the taxonomic and biostratigraphic data base now available is of a magnitude large enough to be appropriate for the studies presented here and in Sheets et al. (this volume). In this paper we discuss the construction of a high resolution correlation model and composite range chart from the stratigraphic range data of 159 conodont species in 24 boreholes and outcrops around Baltoscandia (Fig. 1). In a complementary paper, Sheets et al. (this volume) uses this correlation model and composite range chart to examine the patterns of biodiversity, origination and extinction in Ordovician Baltoscandian conodonts.

METHODOLOGY

Measuring biodiversity through geological time and across different geographic regions presents a number of difficulties that need to be taken into consideration. Some of the problems stem from sampling biases and a lack of taxonomic consistency in data sets compiled by different workers, whereas others

result from the process of converting stratigraphic range data derived from biostratigraphic studies into diversity measures (Cooper, 2004) or an inability to correlate fossiliferous successions with enough precision to be sure that diversity scores are compiled from coeval intervals. In an attempt to eliminate taxonomic inconsistencies, one of us (SMB) has carefully reviewed all the individual range charts and occurrence tables from which this study is conducted, and revised and updated the nomenclature as necessary. Our methods for evaluating and minimizing sampling biases, and for converting stratigraphic range data into diversity measures is comprehensively discussed in Sheets et al. (this volume). Herein we focus on stratigraphic correlation, the construction of a composite range chart, and its conversion to a timescale that can be used in calculating origination and extinction rates.

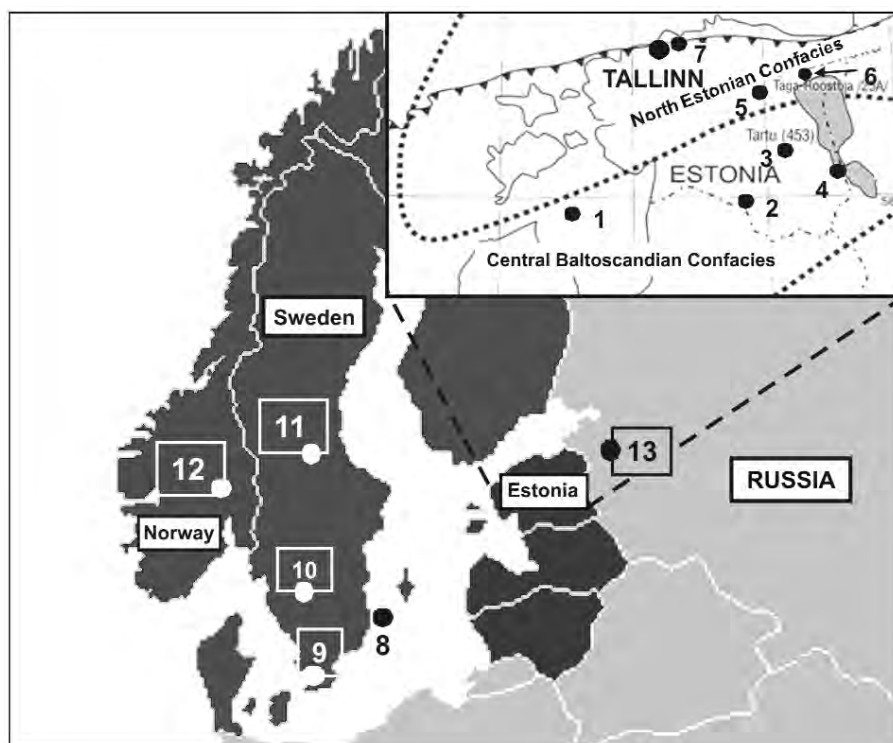


Figure 1. Locality map for outcrops and boreholes in Baltoscandia. Closed circles on the Baltic States inset map are borehole locations; saw-toothed line delineates the present extent of Ordovician carbonates; and the dotted lines represent boundaries of the confacies belts. Numbered boreholes are: 1) Ruhnu; 2) Valga; 3) Tartu; 4) Mehikoorma; 5) Kerguta; 6) Taga-Roostoja; 7) Mäekalda. Numbered outcrop localities are: 8) Öland; 9) Scania, southern Sweden; 10); Västergötland; 11) Siljan Region of Sweden; 12) South-central Norway; 13) Putilovo Quarry and Lava River, Russia. Estonia inset map modified from Modliński et al. (2002).

Stratigraphic correlation requires three separate tasks, namely establishing a temporal sequence of events, determining the relative interval length between those events, and locating the horizons that match in age with each event in every section (Kemple et al., 1995). Unfortunately, due to sampling inconsistencies, partial preservation of taxon ranges, and missing taxa the sequence of events (particularly taxon first and last appearance datums or FAD's and LAD's) is often contradictory among stratigraphic

sections (Sadler et al., 2009). Traditional graphic correlation solves a correlation problem by plotting the location of stratigraphic datums that two sections have in common on an X-Y plot. A possible solution to the problem of correlating individual horizons between the two sections is represented by the Line of Correlation (LOC) (Miller, 1977; Edwards, 1995). The best solution to the correlation problem is the one where the LOC requires the minimum net range extension necessary to make all local ranges fit a single sequence and spacing of events. This is called "economy of fit" (Shaw, 1964), and as well as defining a best solution, it can be used to examine levels of "misfit" and define a penalty function that can rank alternate solutions (Kemple et al., 1995).

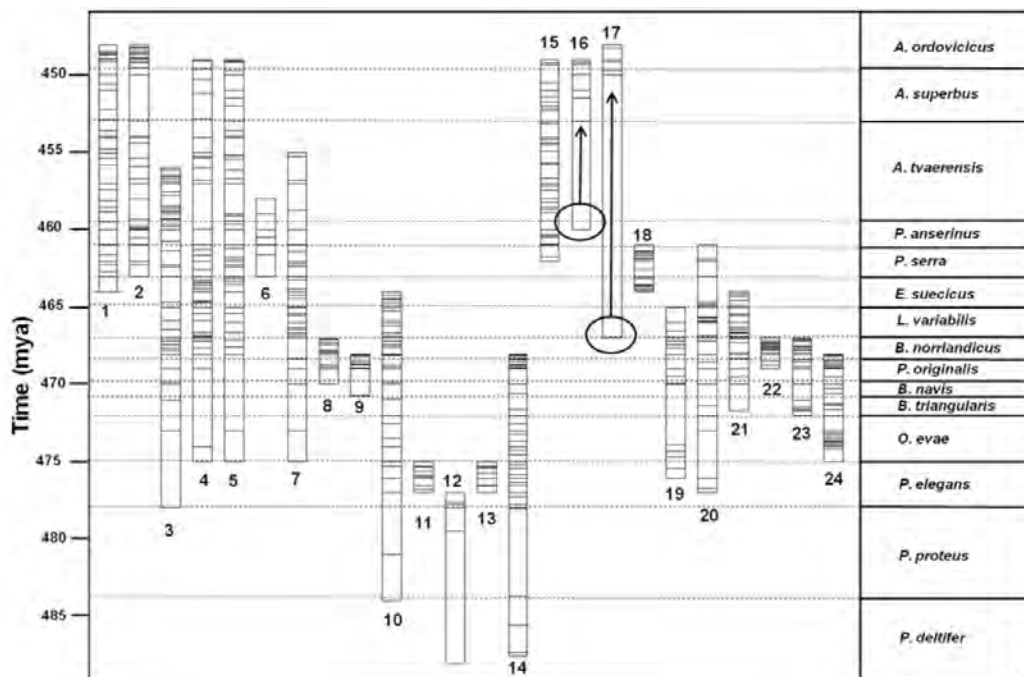
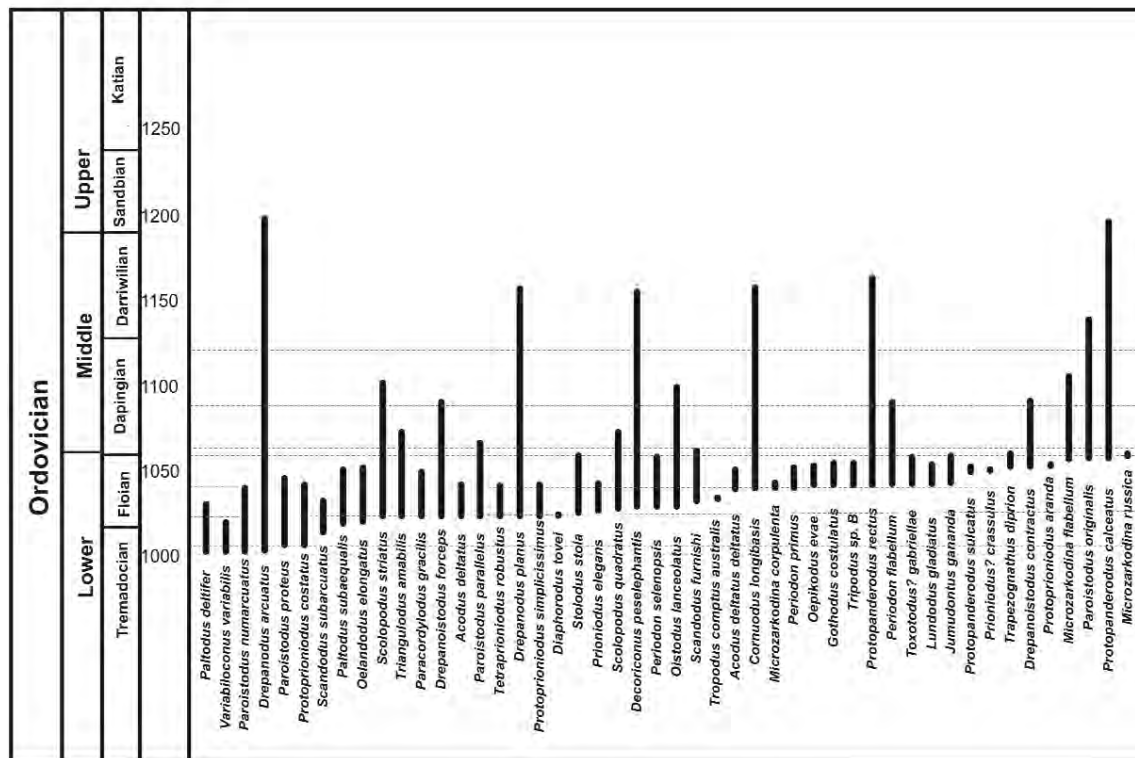
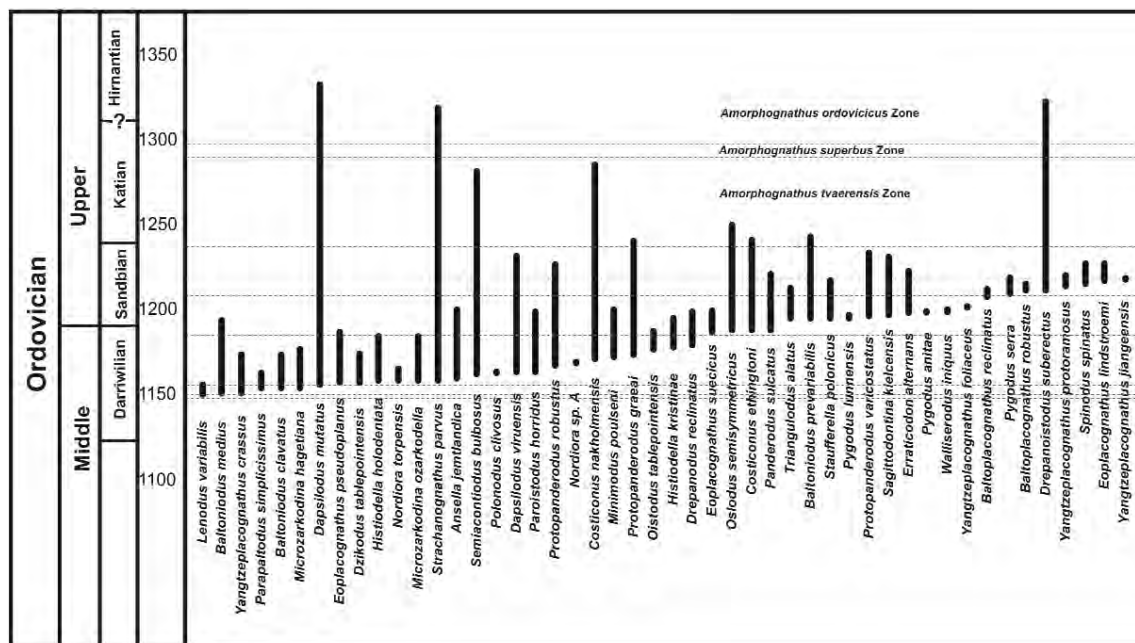
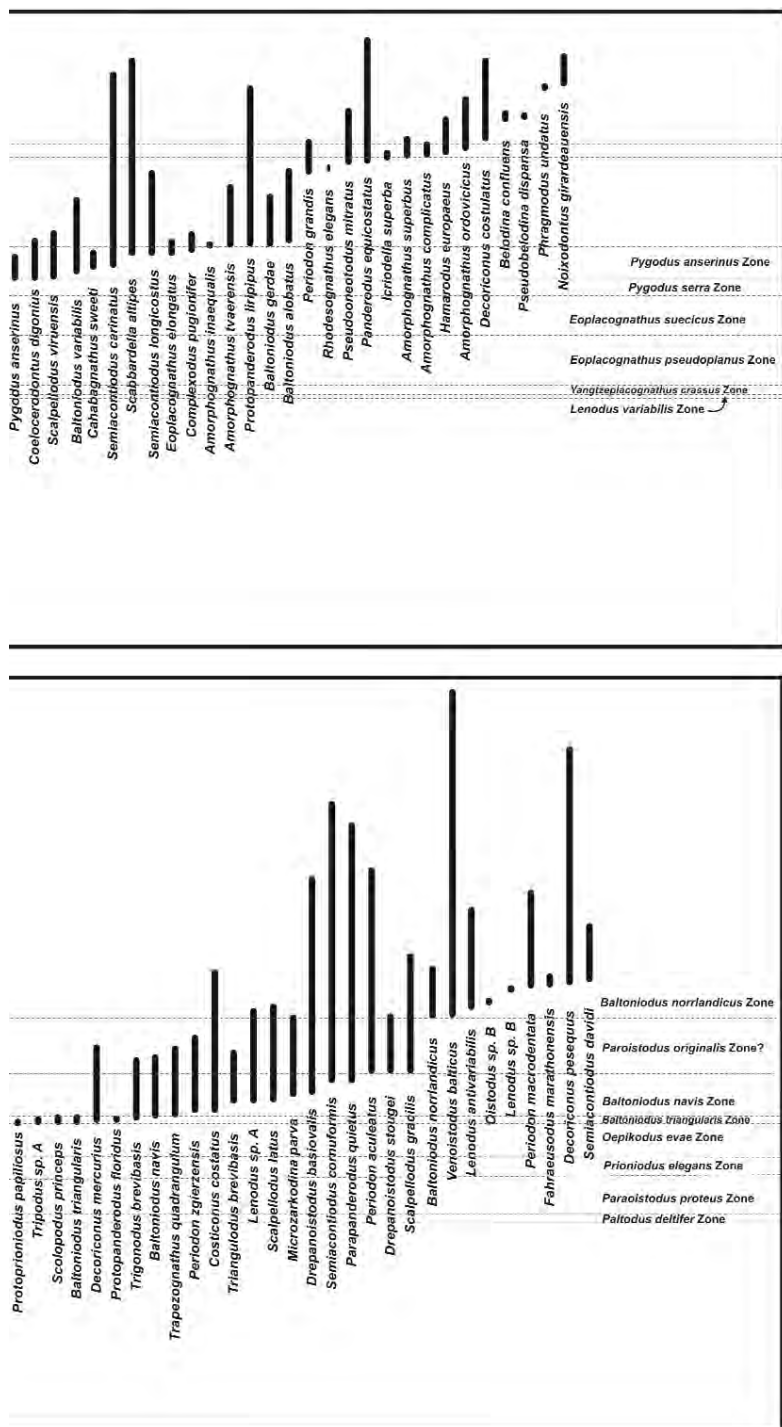


Figure 2. Chronostratigraphic correlation chart for 24 Ordovician boreholes and outcrops in Baltoscandia. Conodont biozones are listed on the right of the diagram and the temporally scaled composite on the left. Individual sections are represented by rectangles and the black horizontal lines within each rectangle are event-rich portions of the sections. Note that the unconstrained bases of sections 16 and 17 exhibit downward "sinking". Arrows indicate the generally accepted age for the section bases. Section numbers represent the following localities, 1) Ruhnu (Männik, 2003); 2) Valga (Männik, 2001); 3) Tartu (Stouge, 1999); 4) Taga-Roostojav (Männik and Viira, 1999); 5) Mehikoorma (Männik and Viira, 2005); 6) Fågelsång Outcrop (Bergström et al., 2000; Bergström, 2007b); 7) Kerguta (Viira et al., 2006); 8) Gillberga (Löfgren, 1995, 2000, 2004); 9) Putilivo Quarry (Tolmacheva et al., 2003); 10) Mäekalda (Viira, et al., 2001); 11) Mossebo (Löfgren, 1993); 12) Storeklev (Löfgren, 1993); 13) Hunneburg (Löfgren, 1993; Bergström et al., 2004); 14) Lava River (Tolmacheva, 2001); 15) Fjäckä (Bergström, 2007a); 16) Amtjarn (Bergström, 2007a); 17) Kullberg (Bergström, 2007a); 18) Kårgårde (Löfgren, 2004); 19) Andersön – A (Rasmussen, 2001); 20) Andersön – B (Rasmussen, 2001); 21) Steinsodden (Rasmussen, 2001); 22) Haggudden, Öland (Stouge and Bagnoli, 1990); 23) Horns Udde Quarry, Öland (Bagnoli and Stouge, 1996); 24) Horns Udde, Öland (Bagnoli and Stouge, 1996).

Unlike graphic correlation, which integrates ranges into a composite time scale one section at a time, Constrained Optimization (CONOP9, Sadler et al., 2003) is multi-dimensional - it works with observations from "n" number of sections simultaneously. CONOP rejects impossible solutions (constraint) and then





searches for the best possible solution (optimization) (Kemple et al., 1995). The best correlation solutions are those that require the minimum net adjustment of observed ranges in local sections. Thus, a penalty function based on the sum of range extension for all taxa in all sections can be calculated and used to rank the various possible solutions. In sum, CONOP9 eliminates impossible correlation solutions, those that contain last before first appearance datums for any individual species and/or miss known taxon co-existences; and chooses the "best" among possible solutions, one that has the minimum net adjustment of observed ranges in local sections and smallest number of unobserved taxon co-existences.

The Middle and Upper Ordovician rocks of Baltoscandia have been divided into spatially distinct, composite litho- and biofacies units called confacies belts (Jaanusson, 1976, 1995). A precise regional correlation of outcrops and boreholes in different confacies belts has always been problematic due to the pronounced biogeographical and lithofacies differentiation. Even within the carbonate-rich North Estonian and Central Baltoscandian confacies belts (Figure 1), local unconformities

Plate 1. Composite Ordovician conodont range chart produced using CONOP9.

Main sources of data for individual sections are listed in the explanation to Figure 2.

and the problems mentioned above have made it difficult to always produce the precise correlations needed for diversity analyses. Hence, we used CONOP9 (Sadler et al., 2003) to overcome these problems and construct a high resolution correlation model (Fig. 2) and composite conodont range chart (Plate 1).

As with all quantitative correlation techniques, CONOP – generated taxon ranges and correlation solutions must be carefully checked for errors. In particular, conflicts with expert observation and opinion should be carefully evaluated. A full discussion of typical CONOP errors is beyond the scope of this paper, but it is important to note that taxon range ends and section tops or bottoms can often float or sink well beyond any reasonable age assignment (sections 16 and 17, Fig. 2). Generally, this occurs when there is no constraint on a range or section end – for example, the last collection in a section may contain a single taxon LAD that is artificially truncated at the section top. With no other constraint, CONOP may let the section top float upward to the actual LAD of that taxon (as it occurs in another section). In any section, any LAD's that are above the last FAD, or any FAD's below the first LAD are essentially unconstrained and range extension would not generate additional penalty.

After using CONOP9 to construct a composite section, we converted it into a timescale by assigning the absolute ages of conodont biozone bases (taken from Webby et al., 2004, figure 2) to the FAD's of the key conodont index taxa in the composite (name-bearers for each zone), and then scaled the composite appropriately. The key index taxa used to establish the temporal scaling, their height in the composite section and their FAD absolute ages are listed in Table 1. We correlated each stratigraphic section with the new timescale, producing a chronostratigraphic correlation chart (Fig. 2) We also tabulated presence/absence data for each species at every collection horizon in all 24 sections – data needed for the Capture, Mark, Recapture analysis discussed in Sheets et al. (this volume), and projected each of those collection horizons back into the composite timescale. Finally, we subdivided the timescale into 60 685 ky intervals (a temporal resolution approximately one fourth that of the median conodont zone duration) spanning the *Paltodus deltifer* through *Amorphognathus ordovicicus* conodont zones. Within this binned timescale Sheets et al. (this volume) calculated conodont biodiversity, origination rates, and extinction rates from the middle Tremadocian to the Hirnantian.

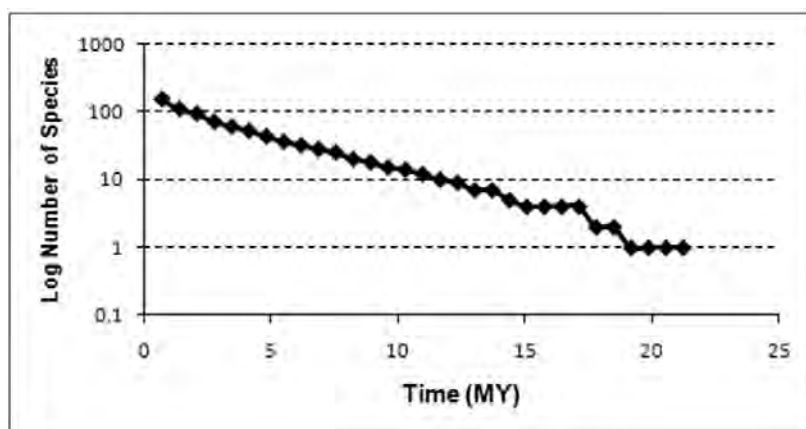


Figure 3. Graph of Ordovician conodont survivorship. The log plot shows a relatively constant risk of species loss over time, indicating that extinction risk is independent of species duration.

Species	Composite Level	Age (my)
<i>Paltodus deltifer</i>	1000.00	488.0
<i>Paroistodus proteus</i>	1002.69	484.0
<i>Prioniodus elegans</i>	1015.97	477.8
<i>Oepikodus evae</i>	1042.80	475.0
<i>Baltoniodus triangularis</i>	1053.08	472.0
<i>Baltoniodus navis</i>	1056.08	470.5
<i>Baltoniodus norrlandicus</i>	1079.00	468.4
<i>Lenodus variabilis</i>	1105.58	467.0
<i>Eoplacognathus suecicus</i>	1137.78	465.0
<i>Pygodus serra</i>	1173.98	463.1
<i>Pygodus anserinus</i>	1184.48	461.2
<i>Amorphognathus tvaerensis</i>	1204.81	459.5
<i>Baltoniodus gerdae</i>	1209.22	457.0
<i>Amorphognathus superbus</i>	1267.12	453.0
<i>Amorphognathus ordovicicus</i>	1272.90	449.6

Table 1. The key conodont index taxa used to establish the temporal scaling for converting the CONOP composite section into a timescale, their FAD position in the original composite section and their absolute age from Webby et al. (2004, text-figure 2).

DISCUSSION

A range chart for Ordovician conodonts from Baltoscandia is presented in Plate 1. The y-axis represents thickness in the CONOP9 composite. An inspection of the range chart reveals no major unconformities, although somewhat data poor intervals exist in the early Tremadocian, early Katian (the upper *Amorphognathus tvaerensis* Zone) and Hirnantian. Relatively minor problems with range end floating occur in the upper Tremadocian and Floian. For example, *Drepanodus planus*, *Paroistodus parallelus*, *Oistodus lanceolatus*, and *Protopanderodus calceatus* all exhibit LAD's that are younger in the composite than in any individual section. These LAD's apparently occur at section tops where other species LAD's are artificially truncated by the end of the section. As CONOP9 extends the composite ranges of the artificially truncated ranges, it drags these other range tops (e.g., *D. planus*) with them. Similarly, some artifactual overlap also occurs in the ranges of *Amorphognathus superbus* and *A. ordovicicus*. This is evidently due to the highly inconsistent ranges of coeval taxa (e.g., *Hamarodus europaeus* and *Belodina confluens*) in certain sections, a situation that causes CONOP9 to minimize range extension penalty in those taxa by extending the range of *A. superbus* too high.

Clark (1983) presented survivorship curves for conodont genera and families, and also calculated average taxon durations (30 million years for genera and 40 my for families). Similar to many phyla (Van Valen, 1979) conodont cohorts exhibited a constant rate of extinction over time (Clark, 1983, text-figure 7). We used the CONOP9 composite range chart and timescale to calculate average conodont species durations in geological time. Species that only occurred in one section and in one collection (range of 0 meters in the composite) were arbitrarily given a species duration of 0.1 my, and species that occurred in

the first and or last collections (488 and 443 mya, respectively) were removed to eliminate edge effects (Foote, 2000). The mean duration for all species is 4.0 million years, although durations are highly variable, with a standard deviation also of 4.0 my. Dividing species into groups based on the Ordovician series in which their FAD occurs results in species durations of 4.7, 3.0, and 3.4 million years for Lower, Middle, and Upper Ordovician species, respectively. The longer duration of species that originate in the Lower Ordovician may be partly an artifact of CONOP methodology - range tops having more space to float – but this requires further investigation. We also examined species survivorship and found that similar to Clark's (1983) results, extinction risk was independent of species duration. A species survivorship curve is illustrated in Figure 3. These results are corroborated in the biodiversity, origination and extinction rate analyses conducted by Sheets et al. (this volume).

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BIOSTRATIGRAPHY OF THE GENUS *CALIX* (ECHINODERMATA, DIPLOPORITA) IN THE MIDDLE ORDOVICIAN OF THE SOUTHERN CENTRAL IBERIAN ZONE (SPAIN)

J.C. Gutiérrez-Marco¹ and J. Colmenar²

¹ Instituto de Geociencias (CSIC-UCM), Facultad CC. Geológicas, José Antonio Novais 2, E-28040 Madrid, Spain.
jcgrapto@geo.ucm.es

² Área de Paleontología, Dpto. CC. de la Tierra, Universidad de Zaragoza, Pedro Cerbuna 12, E-50009 Zaragoza, Spain.
colmenar@unizar.es

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INTRODUCTION

Diploporite cystoids are relatively common in the Middle Ordovician formations of SW Europe, where they are represented by the genera *Calix* Rouault, *Aristocystites* Barrande, *Codiacystis* Jaekel, *Phlyctocystis* Chauvel, *Batalleria* Chauvel and Meléndez and *Oretanocalix* Gutiérrez-Marco (Chauvel, 1941, 1973, 1977, 1980; Meléndez, 1951, 1958; Chauvel and Meléndez, 1978, 1986; Gutiérrez-Marco et al., 1984, 1986; Gutiérrez-Marco and Baeza Chico, 1996; Couto and Gutiérrez-Marco, 1999; Gutiérrez-Marco and Aceñolaza, 1999; Gutiérrez-Marco, 2000; Gutiérrez-Marco and Bernárdez, 2003). However, the identification of most species included in these genera poses a significant problem, due to the fact that the available and published material commonly corresponds to internal moulds of the aboral region of the thecae, where only *Codiacystis*, and to a lesser extent *Oretanocalix* and *Aristocystites*, are recognizable. Thus, almost all the described species need a deep taxonomic review in terms of modern diploporite taxonomy, which is based on a number of structural details of the theca and its openings. These details are totally unknown in most of the taxa described from the Middle Ordovician shales and sandstones of the Ibero-Armorican and North African parts of the Gondwana margin.

In spite of the taxonomical problems regarding the generic affiliation of many of these diploporite echinoderms, the vertical distribution of some species of the genus *Calix* have a significant biostratigraphic interest. This note focuses on the proposal of some regional biozones based on this genus, that are complementary of those derived from other fossil groups (Fig. 1). These *Calix* biozones can be recognized over an area covering the southern part of the Central Iberian Zone, and with correlation potential with other areas of NW Spain, the Iberian Cordillera and, to some extent, the Armorican Massif of western France and the Moroccan Anti-Atlas.

TAXONOMIC NOTE

The diploporid “cystoid” *Calix* Rouault, 1851 (= *Dorycystites* Klouček, 1917; *Lepidocalix* Termier and Termier, 1950) ranges from the early Oretanian to the late Berounian (earliest Darriwilian 2 to latest Katian 2 in terms of global stages and substages: Bergström et al., 1999) from SW and central Europe to North Africa, in a paleogeographic setting of high Gondwanan paleolatitudes. This typical member of the family Arystocystididae Neumayr is characterized by an elongate conical to cylindrical theca, provided with an aboral terminal tubercle and composed of numerous plates, mostly of irregular shape. The plates corresponding to the aboral region bear a central tubercle or prominence, and the tubercles are irregularly arranged or forming definite cycles, in this case showing great intraspecific variability. Mouth elongate, tetra- or radiate, with scarcely developed and unbranched ambulacra, ended in articular facets for brachiole insertion adjacent to mouth. Diplopores with simple oval or slightly curved pits covered over with epitheca, when the latter is preserved.

The genus *Calix* (redescribed by Rouault, 1878, 1883) comprises the following valid species: *Calix sedgwicki* Rouault, 1851 (type species), *C. purkynei* (Klouček, 1917) [= *C. rouaulti buchoti* Chauvel, 1936; *C. rouaulti* Chauvel, 1936 p.p.], *C. pulchra* (Termier and Termier, 1950b) and *C. gutierrezzi* Chauvel and Meléndez, 1986, bearing all of them aboral tubercle and tuberculiferous plates (the tetra- or radiate peristome is fully known only from the type species). Other species incompletely characterized and probably related to the genus are: *Calix? inornatus* Meléndez, 1958 (with tetra- or radiate peristome but without evidence of tubercles: exterior aboral region unknown); *C.? rotundipora* Chauvel, 1980 (with small tubercles and circular diplopores, remaining theca unknown); *C.? cornuta* Chauvel, 1978 (horn-shaped aboral annulated interior, remaining details unknown); *C.? segaudi* (Termier and Termier, 1950a) (tuberculiferous plates replaced by primary and secondary cycles of strongly domed plates; remaining theca unknown); and *C.? hajraensis* Chauvel, 1978, a rare Upper Ordovician species densely ornamented by conical tubercles, apparently with a tetra- or radiate peristome but resembling other diploporite genera or even a Moroccan specimen of *C.? segaudi* (see Chauvel, 1978, pl. 2, fig. 1).

The species *Calix rouaulti* Chauvel, 1936, one of the most commonly cited among all the echinoderm literature from the Ordovician of SW Europe, is very poorly known and was regarded as highly polymorphic by Chauvel (1980). As the holotype of *C. rouaulti* (the “morphotype c” of Chauvel) is clearly conspecific with *C. purkynei* (Klouček, 1917), the name “*rouaulti*” becomes a junior synonym of the Czech species. However, the usage of *Calix “rouaulti”* s.l. is maintained provisionally here in order to refer to the remaining morphotypes (other than *C. purkynei*) designated by Chauvel (1980), some of which deserve biostratigraphic potential but that are impossible to characterize taxonomically until complete specimens are found.

Other highly questionable species of *Calix* are “*C. barrandei* Rouault” and “*C. davidsoni* Rouault” both proposed by Lebesconte (in Rouault, 1883, note infra to pl. 8) based on poorly preserved specimens either of *C. rouaulti* or *C. sedgwicki* s.l. (Chauvel, 1941); “*Calix halli*” Rouault, 1851 (type species of the genus *Pachycalix* Chauvel, 1936), which is only known from poorly preserved specimens most probably related to the genera *Aristocystites* or *Phlyctocystis*; “*Calix lebescontei*” Chauvel, 1936, an Upper Ordovician minute form, with some tubercles, but of dubious generic assignment (Chauvel, 1941, p. 84); “*C. munchisoni*” (Verneuil and Barrande, 1855) sensu Meléndez (1958), often synonymized with *Calix “rouaulti”* s.l. (starting from Chauvel, 1980), sometimes considered as a separate species of *Calix* (Meléndez and Chauvel, 1983) and lately re-evaluated as the type species of the genus *Oretanocalix* (Gutiérrez-Marco, 2000); “*C. sampelayoi*” (Meléndez in Bouyx, 1962), never described and finally

CHRONOSTRATIGRAPHY		GRAPTOLITES		CHIITINOZOANS	TRILOBITES	BRACHIOPODS	"CYSTOIDS"
GLOBAL	MEDITERRANEAN REGIONAL (POLAR GONDWANA)						
UPPER ORDOVICIAN	HIRNANTIAN (= KOSOVIAN)			<i>S. oulesis</i>	trilobites almost absent	Hirnantia Fauna (rare)	
	KRALODVORIAN	graptolites absent in Iberia		<i>T. elongata</i>			
				<i>An. merga</i>	<i>H. insculpta</i>	<i>L. pranli</i>	
				<i>Ar. nigerica</i>	<i>S. cf. obliata</i>	<i>G. luesmae</i>	
	BEROUNIAN	upper	" <i>Dp. "salgatus" - Oc. lingulitheca</i> "	?	<i>D. malladai</i>	<i>A. meneghiniana</i>	
		middle	graptolites almost absent (in study) without zonation)	<i>Ar. barbata</i>		<i>S. hayliceki</i>	
				<i>T. fistulosa</i>			
	SANDBIAN			<i>B. robusta</i>	<i>Ce. dijordani</i> - <i>D. seanesi</i>	<i>S. armoricana</i> - <i>G. melout</i>	
		lower		<i>E. tanvillensis</i>			
				<i>L. dalbyensis</i>			
MIDDLE ORDOVICIAN	DOBROTIVIAN			<i>L. denoffi</i>	<i>Ce. chauveli</i>	<i>T. loi</i>	
		upper	<i>O. bekkeri</i>	<i>L. poncetii</i>	<i>Ma. burgati</i>	FAD Jézerceta	
			<i>C. aff. tricornis</i> (Bobemian)		<i>P. (C.) borni</i>	<i>H. ksfornici</i>	<i>Calix "rouaulti" s.l.</i>
		lower	<i>H. terpsichorus</i>	<i>L. paxosensis</i>	<i>M. luyet</i> (FAD P. borni)	<i>A. maritima</i>	<i>E. muscolosa</i>
			<i>G. timarssani</i>		<i>P. (C.) M. primitiva</i>		
					<i>bioherminal</i>		
	ORETANIAN	upper	<i>P. dist.</i>	<i>L. paxosensis</i>	<i>Z. toledana</i> - <i>E. macrocephala</i>	<i>H. morganensis</i>	
			<i>D. marchisani</i> - <i>D. clavatus</i> Previgi	<i>C. renkasi</i>	<i>E. deambrosi</i> nava	<i>C. ribetini</i>	
				<i>Ar. armoricana</i>			
	"ARENIGIAN"	lower	<i>D. spinulosus</i> - <i>D. artus</i>	<i>S. formosa</i>			
				<i>C. calix</i>		Orth.-Siv. novitillo	<i>Calix? inornatus</i>
		upper	FAD <i>Co. retroflexus</i>	<i>H. protocallix</i>	<i>P. (P.) cambresensis</i>		

Figure 1. Correlation chart of the main Middle and Upper Ordovician biostratigraphical units defined in SW Europe, redrawn and updated from Gutiérrez-Marco et al. (2002), to which a right column ("Cystoids") has been added to show the diploporite biozones considered in this work.

synonymized with morphotype "f" of *C. "rouaulti"* s.l. (Chauvel and Meléndez, 1978; Chauvel, 1980); "*C. termieri*" Chauvel, 1966, a problematic taxon described from a very incomplete specimen, reported as an "ambiguous species" by Chauvel (1980, p. 8); and "*C. toledensis*" Chauvel and Meléndez, 1978, a taxon restricted to its inconclusive holotype by Gutiérrez-Marco and Aceñolaza (1999).

BIOSTRATIGRAPHY

The studied material comes from fourteen Ordovician sections extending across the southern Central-Iberian Zone of the Iberian Massif, located in the synclines of Los Yébenes, Navas de Estena, La Chorrera, Guadarranque, Piedrabuena, Corral de Calatrava, Valdepeñas, Herrera del Duque, Almadén, Puertollano-Almuradiel and Guadalmez, plus the areas of El Centenillo and Eastern Sierra Morena (see San José et al.,

1992 and Gutiérrez-Marco et al., 2002, for location and summary of the main lithostratigraphic units). The vertical distribution of 12 diploporite species belonging to 6 genera has been studied, and their relative ranges plotted with reference to other trilobite and brachiopod biozones (Fig. 1), and dated by graptolites occurring in the assemblages. Our results show that there are some diploporite species related with *Calix* that are widespread in the studied area and show a restricted vertical distribution, being therefore of biostratigraphic interest. Five regional biozones are here proposed and named according to the respective diploporite species, and the contribution of *Calix gutierrezzi* is analyzed in the frame of the rhombiferan-dominated assemblages of the Upper Ordovician “Heliocrinites Fauna”. The new units are described below in ascending biostratigraphical order.

Calix? inornatus Biozone

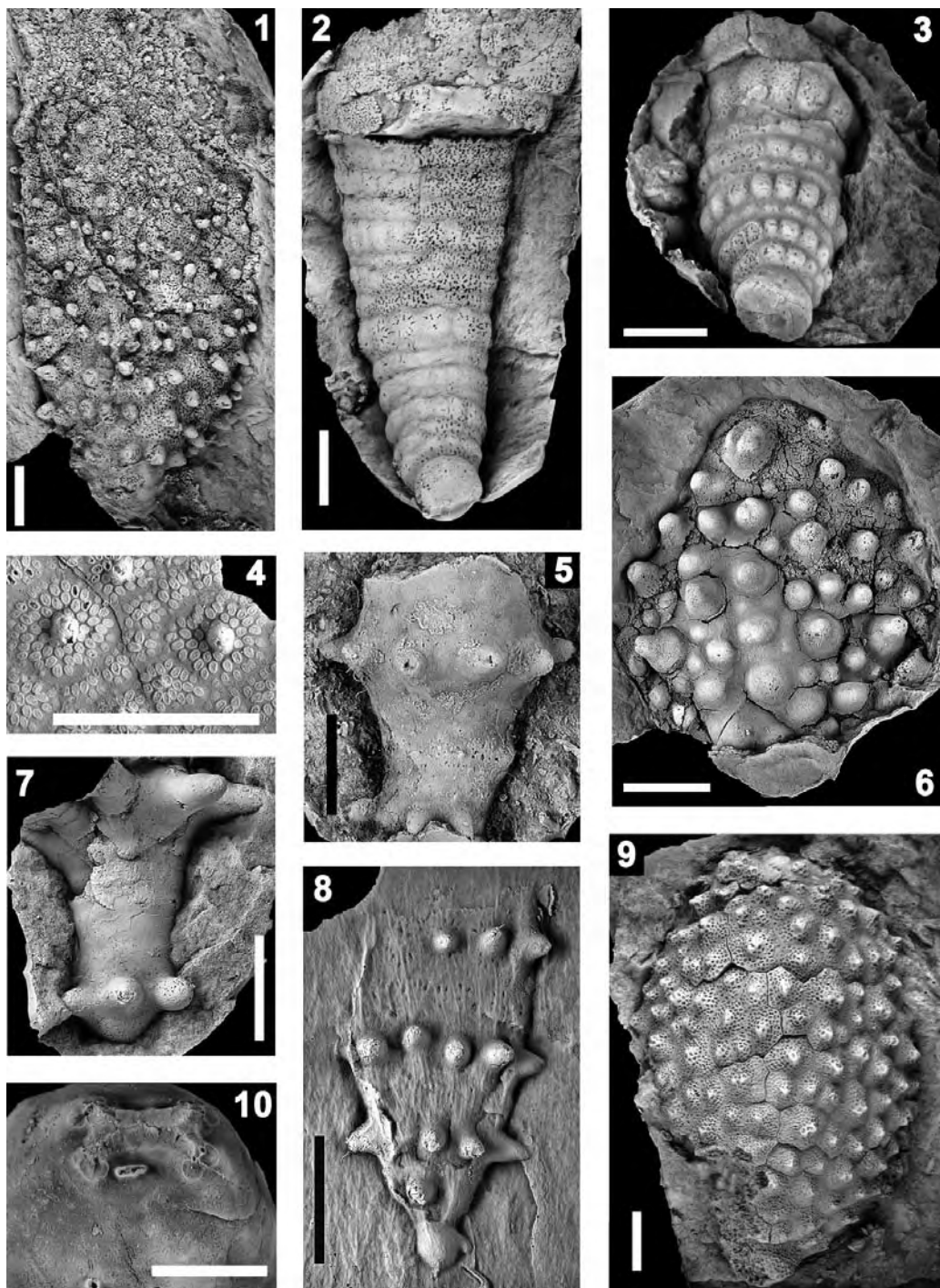
Defined by the entire vertical range of *Calix? inornatus* Meléndez, 1958 (Pl. 1, fig. 10), a species very easily recognizable by its carrot-shaped thecae, with a smooth and inflated oral region, that spans through the range of the *Orthambonites-Sivorthis noctilio* brachiopod Zone (see Reyes Abril et al., 2010, 2011) and is also recorded abundantly with trilobites of the upper part of the *Placoparia cambriensis* Zone, especially in the Montes de Toledo area. The *C.? inornatus* biozone can be dated as early Oretanian (earliest mid Darriwilian), as indicated by concurrent graptolite fauna (Fig. 1). Other valid species of diploporids recorded from this biozone are *Calix? rotundipora* Chauvel and *C.? cf. cornuta* Chauvel, present in some localities with a single specimen (e.g. *C.? rotundipora* from Ventas con Peña Aguilera: Chauvel, 1980).

Calix sedgwicki Biozone

This biozone is defined by the appearance and vertical extent of *C. sedgwicki* Rouault, 1851, a species of elongate morphology with numerous small tubercles irregularly distributed over the whole theca, and where most of the diplopores have a characteristic rim (Pl. 1, figs. 1 and 4). The FAD of the nominal species is clearly below the base of the *Cacemia ribeiroi* and *Placoparia tournemini* brachiopod and trilobite zones, respectively, and their total range is paralleled by that of the *Didymograptus murchisoni* graptolite Zone, which indicates a late Oretanian age (late mid–early late Darriwilian 2 substage).

The species *C. sedgwicki* was defined in the French Armorican Massif, where its detailed biostratigraphic position within the Oretanian-Dobrotivian range is still unknown. In Morocco, *C. sedgwicki* has been recorded in the Bou-Zeroual Formation of the First Bani Group, that according to Gutiérrez-Marco et al. (2003) is of late Oretanian age.

Plate 1. Some diploporite echinoderms/Arystocystitid “cystoids” with biostratigraphic interest from the Ordovician of the Central Iberian Zone, Spain. 1 and 4, *Calix sedgwicki* Rouault, 1851, lower Oretanian of Navas de Estena: 1, latex cast from the external mould of an almost complete theca in lateral view, JLC-102; 4, detail of tubercles and diplopores with preserved epitheca, latex cast from specimen JLC-103. Lateral view.-- 2-3, *Calix? segaudi* (Termier and Termier, 1950a), lower Dobrotivian of Navas de Estena: latex cast of two thecae with partly preserved epitheca showing details of aboral tubercle, specimens JLC-128 and JLC-127, respectively.-- 5, 7 and 8, *Calix purkynei* (Klouček, 1917). Dobrotivian from Czech Republic, Retuerta del Bullaque and Alía, respectively: 5, lateral view (latex cast) of holotype specimen; 7, aboral portion of a theca with widely spaced cycles of tubercles, MT-82; 8, latex cast of a flattened specimen in shale, lateral view of JLC-121, showing isolated tubercle among the aboral tubercle and first cycle.-- 6, *Calix “rouaulti”* Chauvel, 1936 s.l., terminal lower Dobrotivian from El Viso del Marqués, latex cast of a fragmentary specimen showing irregularly arranged conical tubercles.-- 9, *Calix gutierrezzi* Chauvel and Meléndez, 1986, uppermost Berounian of Almadén, latex cast of holotype specimen MT-227 showing tubercles and “polygonal” diplopores.-- *Calix? inornatus* Meléndez, 1958, lower Oretanian, Ventas con Peña Aguilera. Latex cast of the oral region showing thecal apertures in oblique-lateral view, MGM-2000-O). Scale bars, 10 mm.



Calix purkynei Biozone

This biozone is defined by the total range of *C. purkynei* (Klouček, 1917) [= "*C. rouaulti* Chauvel forme c"], a species easily recognizable by its elongated conical theca ornamented by tubercles of variable length, which are arranged in regular cycles separated by smooth areas, corresponding to constrictions in the internal mold (Pl. 1, figs. 5, 7 and 8). On the same beds, the species is locally associated with rare specimens of *C.? cornuta* Chauvel and also to *Calix? segaudi* (Termier and Termier), which makes its first appearance in this biozone. In the studied area, the first record of *C. purkynei* preceded the tempestitic sedimentation generalized in the southern part of the Central Iberian Zone during the early Dobrotivian, and is dated by the record of graptolites of the *Hustedograptus teretiusculus* Zone and their association to *Placoparia tournemini* (trilobite) and *Heterorthina morgatensis* (brachiopod) as early Dobrotivian (early Darriwilian 2 age of the global scale).

The species has also been recorded from Bohemia (Klouček, 1917; Prokop, 1964), represented by a single fragmentary specimen (Pl. 1, fig. 5) found in the Skalka quartzite (Dobrotivá Formation), and also from Dobrotivian shales in the French Armorican Massif (= "*C. rouaulti*", morphotypes "a" and "c" of Chauvel, 1980) and possibly also in Morocco. In Spain, *Calix purkynei* was also found in lower Dobrotivian shales from NW Spain (Gutiérrez-Marco and Bernárdez, 2003) and from the Iberian Cordillera (Gutiérrez-Marco et al., 1996), in both cases misidentified as "*C. rouaulti*".

Calix? segaudi Biozone

This biozone is based on the local abundance, in the Montes de Toledo, of *C.? segaudi* (Termier and Termier, 1950), unknown in coeval beds of the remaining Central Iberian Zone because the development of thick sandy tempestites that do not show recognizable diploporid remains. In its laterally-equivalent strata in the north of the region, these sandy tempestites change into a distal tempestite facies developed as lutitic alternations very rich in cystoids. Besides the highly characteristic *C.? segaudi* (Pl. 1, figs. 2-3), *C.? cornuta* Chauvel, *C. "rouaulti"* s.l. and several forms of the genera *Oretanocalix*, *Codiacystis* and *Phlyctocystis* have been recognized (Gutiérrez-Marco et al., 1984).

Calix "rouaulti" s.l. Biozone

This is an informal zone based on an incorrectly named taxon, due to the fact that *C. rouaulti* Chauvel, 1936 sensu stricto (its holotype specimen) is a junior synonym of *C. purkynei* (Klouček, 1917). With the exception of morphotypes "a" and "c" (= *C. purkynei*), morphotypes "f" and "g" of *Calix "rouaulti"* sensu Chauvel are usually restricted to beds of latest-early to late Dobrotivian age, as indicated by the remaining fossil groups of stratigraphical value (Fig. 1). In the terminal lower Dobrotivian shales, *C. rouaulti* s. l. may be locally accompanied by *Aristocystites metroi* Parsley and Prokop, and in higher upper Dobrotivian beds by rare *C.? cornuta* Chauvel and representatives of the genera *Batalleria* and *Phlyctocystis*, the latter involving specimens of giant size with thecas formed by more than 2,000 plates. As indicated in the taxonomic note above, the name for this biozone is provisional, and should be changed when the involved *Calix* taxa are accurately reviewed after complete specimens are found.

Biostratigraphic potential of *Calix gutierrezii*

Diploporite echinoderms became rare in Upper Ordovician strata from Ibero-Armorica, where they were replaced in number and diversity by the rombiferans that characterize the "*Heliocrinites* Fauna" (genera *Heliocrinites*, *Caryocrinites*, *Hemicosmites*, *Rhombifera* and *Echinosphaerites*?; for references see Chauvel and Le Menn, 1979). The few diploporids recorded from the Kralodvorian (Katian 3-4 substages) belong to the sphaeronitid genus *Eucystis* Angelin (= *Proteocystites* Barrande), but in Berounian beds some indeterminate aristocystidids still persisted together with the last representatives of the genus *Calix*. Two of them ("*C. lebescontei*" Chauvel and "*C. hajraensis*" Chauvel) are questionable forms (see taxonomic note above), but *C. gutierrezii* Chauvel and Meléndez, 1986 is a distinct form, characterized by its closely-set diplopores arranged in a polygonal pattern (Pl. 1, fig. 9). The type material of this species comes from the late Berounian sandstones in the Central Iberian zone, but probably the species is already present in the mid Berounian shales from the Iberian Cordillera (Gutiérrez-Marco et al., 1996). If so, in absence of other fossils, *C. gutierrezii* can be used provisionally to estimate a biostratigraphic range comprised between the basal Middle Berounian until the topmost Berounian (from uppermost Sandbian to topmost Katian 2 substage of the global scale), which cannot be regarded as a biozone owing to its scattered occurrences, limited some Spanish areas.

CONCLUSIONS

Despite their abundance in Ordovician rocks from Ibero-Armorica and North Africa, *Calix* is a poorly known genus represented by four valid species and six other taxa probably related to it, but left in open nomenclature at present.

The study of the vertical distribution of all these taxa in fourteen Ordovician sections representative of the southern Central-Iberian Zone of the Iberian Massif, allow the definition of five regional biozones based on the distribution of diploporite echinoderms, that are paralleled with those of brachiopods and trilobites previously recognized from the same area. These biozones display potential value for correlating fossiliferous strata in absence of better biostratigraphical markers, as in this case, where the applicability of some of them could extended to other areas of NW and NE Spain, as well as Morocco and western France. In terms of Mediterranean regional chronostratigraphy (see Gutiérrez-Marco et al., 2008 and Bergström et al., 2009 for their equivalence with the global scale), the *C. ? inornatus* Biozone is restricted to the lower Oretanian, the *C. sedgwicki* Biozone to the upper Oretanian, the *C. purkynei* biozone to the lowermost Dobrotivian, the acme of *C. ? segaudi* with the lower Dobrotivian s.l., and the *C. "rouaulti"* s.l. Biozone to the uppermost lower Dobrotivian and to the upper Dobrotivian. Finally, the range of *Calix gutierrezii* extends from middle to upper Berounian strata in the frame of the Upper Ordovician "*Heliocrinites* Fauna".

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A PRELIMINARY STUDY OF SOME SANDBIAN (UPPER ORDOVICIAN) GRAPTOLITES FROM VENEZUELA

J.C. Gutiérrez-Marco¹, D. Goldman², J. Reyes-Abril³ and J. Gómez³

¹ Instituto de Geociencias (CSIC-UCM), Facultad CC. Geológicas, José Antonio Novais 2, 28040 Madrid, Spain.

jcgripto@geo.ucm.es

² Department of Geology, University of Dayton, 300 College Park, Dayton, OH 45469, USA. Dan.Goldman@notes.udayton.edu

³ Departamento de Geología General, Facultad de Ingeniería, Universidad de Los Andes, Núcleo La Hechicera, Edificio Ingeniería, Piso 2, Ala Este, 5101 Mérida, Venezuela. jreyes@ula.ve, jessicagomez21@gmail.com

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INTRODUCTION

The Lower Sandbian *Nemagraptus gracilis* Zone comprises one of the most widespread, and easily recognizable graptolite faunas in the Ordovician System. The base of the *N. gracilis* Zone also marks the base of the Upper Ordovician Series, and is internationally defined by the FAD of the eponymous species, with the Global Stratotype Section and Point (GSSP) located at Fågelsång in Scania, southern Sweden (Bergström et al., 2000, 2009). Finney and Bergström (1986) provide a general account of the widespread record of this biozone in Europe, America, Australasia and China. In South America, graptolites of the *N. gracilis* are best known from the Argentine Precordillera (Cuyania Terrane), within the Portezuelo del Tontal, Las Aguaditas, Los Azules and Sierra de la Invernada formations in the central Precordillera, in the Yerba Loca Formation of the western Precordillera, and in the La Cantera Formation of the eastern Precordillera (see for example Borrello and Gareca, 1951; Blasco and Ramos, 1976; Brussa, 1996, 1997; Peralta, 1998; Ortega and Albanesi, 1998; Ortega et al., 2008 and references therein). *Nemagraptus gracilis* Zone faunas are rare in the Central Andean Basin, where single occurrences of only *N. gracilis* itself have been reported from three localities in Bolivia and Peru (Laubacher, 1974; Brussa et al., 2007).

In northern South America, the single known occurrence of Sandbian age graptolites, including possible specimens of *Nemagraptus* is restricted to the Caparo Formation, which crops out in the southern Mérida Andes of Venezuela, close to its tectonic boundary with the Barinas-Apure basin (Leith, 1938; Pierce et al., 1961; Shell and Creole, 1964). Recent new collecting by some of the authors (JCGM, JR, JG) has provided additional material that confirms the identification of a Sandbian graptolite fauna in the region that can be assigned to the *N. gracilis* Biozone, a fauna that is described and illustrated for the first time in this part of South America.

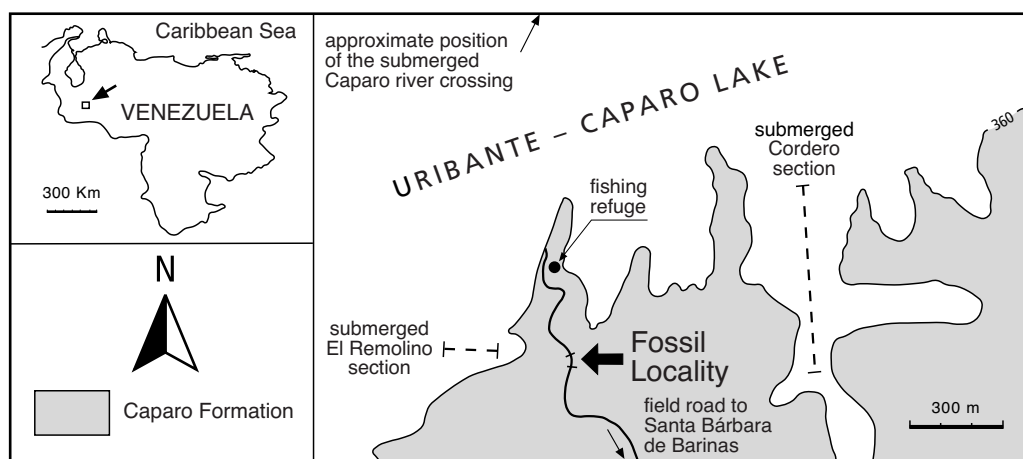


Figure 1. Sketch map of a sector south of the former Caparo River crossing, along the field road to Santa Bárbara de Barinas near the southern margin of the Uribante-Caparo Lake. Also shown are the positions of the sections (now submerged) with Ordovician fossils that were listed by Shell and Creole (1964).

PREVIOUS DATA AND LOCATION OF THE STUDIED MATERIAL

The discovery of fossils in the old "Caparro-Bellavista series" (Christ, 1927) within the Caparo Block of the Mérida Andes of Venezuela, was described by Terry (1935, p. 692) as occurring near the Caparo River crossing (currently submerged under Uribante-Caparo Lake), along an abandoned field road from Mucuchachí to Santa Bárbara de Barinas. His fossil collection, belonging to the Sinclair Exploration Company, was later studied by Leith (1938), who described three new fossil species including the graptolite *Dicranograptus "caparroensis"* (a junior synonym of *D. ramosus* Hall), a trilobite ("*Cryptolithus*" *terryi*) and a bivalve (*Allonychia? brevirostris*). Leith (1938) also listed an "?Orthoid brachiopod" and an "undetermined pelecypod".

Four additional fossiliferous sections were found by Pierce et al. (1961, fig. 8) and Shell and Creole (1964, figs. 2 and 3): two of them recorded as adjacent to the Caparo River crossing (El Remolino and Cordero creeks, neither in existence today), a third along the Lirán creek (about 17.4 km northeast of the old Caparo River crossing), and the fourth in the upper valley of the Caparo River, about 19.5 km northeast from the Lirán creek. These authors also listed the occurrence of Ordovician trilobites, graptolites, brachiopods, bryozoans, crinoids and questionable corals in several beds within the Caparo Formation, and considered all the fossil localities to be of late Mohawkian (early "Caradocian") age. A partial review of the original trinucleid trilobite material from these collections reassigned specimens of *Cryptolithus terryi* to the genera ?*Salterolithus* (Dean in Shell and Creole, 1964), *Onnia* (see the redescription of *O. terryi* by Whittington, 1954) or *Reuscholithus* (Hughes et al., 1975; Hughes, 1980). More recent research at the type section of the Caparo Formation was presented by Benedetto and Ramírez Puig (1982) and Gutiérrez-Marco et al. (1992), with few new paleontological discoveries.

Shell and Creole (1964) reported two separate lists of graptolites collected from locality no. 5 at the Lirán creek, which formerly yielded *Dicranograptus ramosus* and *Didymograptus* sp. (Pierce et al., 1961, p. 358). A sample taken at the same section by the Shell de Venezuela company yielded the following taxa (identifications by I. Strachan): cf. *Nemagraptus* sp., *Climacograptus peltifer*, *C. cf. parvus*, *C. aff. antiquus*,

Climacograptus sp., *Orthograptus* sp., *Glyptograptus* cf. *teretiusculus*, *Cryptograptus* sp., *Dicranograptus* cf. *caparroensis*, *Dicranograptus* sp., cf. *Dicranograptus* sp., cf. *Didymograptus* sp., *Amphigraptus* cf. *divergens* and cf. *Thamnograptus* sp., in association with some trilobites, brachiopods and bryozoans. Additionally, a sample collected by the Creole Petroleum company from the same section, yielded *Dicranograptus caparroensis*, *D. nicholsoni*, *Dicranograptus* sp. and some trilobites (identifications by A. Boucot and G.A. Cooper). The age in both cases was established as "Middle" Ordovician (Caradoc). The original graptolite material collected by Shell and Creole (1964) was briefly reviewed by Rickards (in Hughes, 1980, p. 11) who recognized the taxa *Dicranograptus caparroensis* Leith, *Dicranograptus* sp., *Orthograptus amplexicaulis* (Hall), *O. ?quadrimucronatus* (Hall), *Corynoides* sp. and *Acanthograptus* sp., assigning the assemblage to "Caradoc age, probably Longvillian or younger."

However, according to the maps associated with the above mentioned data, there are two clearly separated graptolite collections coming from Lirán creek: one apparently made by H.C. Arnold for the Company Shell de Venezuela, and the other was probably made by W.R. Smidht for the Creole Petroleum Corporation. Moreover, the detailed map of Pierce et al. (1961, fig. 8) shows that the Lirán creek section comprises two distinct fossiliferous localities separated by more than two hundred meters. As a consequence of these statements, we cannot be sure that all the graptolite data mentioned in the two previous papers came from a single locality and horizon, and thus Rickards's review and age designation (in Hughes, 1980) of the Lirán creek graptolite fauna is based on separate collections that could be of different ages. Unfortunately, due to the low water level present during the field research of January 2011, navigation to the mouth of the Lirán Creek, a tributary of the Caparo River proved impossible. Thus, we were unable to recollect that section and clarify its age relationships.

The construction of the La Honda dam in 1986 flooded the area in 2003, producing the Uribante-Caparo Lake and submerging the sections of the Caparo Formation located south (El Remolino) and west (Cordero creek) of the former Caparo River crossing (after the latter was proposed as the best reference section for the unit by Shell and Creole, 1964). A recent review of the existing outcrops above the water level along the former Cordero creek, provided some brachiopod and trilobite finds in sandstone and weathered ironstone, but no graptolites. However, a careful examination of original type section of the Caparo Formation, placed along the trail from the Uribante-Caparo Lake to Santa Bárbara de Barinas, led to the rediscovery of several fossiliferous beds, partially listed by Benedetto and Ramírez Puig (1982) and Gutiérrez-Marco et al. (1992). In addition to badly preserved remains of *Dicranograptus*, and uncommon specimens of *Amphigraptus* and dendroids, which occur through more than 30 m of strata, a reasonably well preserved graptolite assemblage was discovered in a 20 cm thick bed of laminated shale located in the trail itself (geographic coordinates S7° 52' 56"; W71° 16' 13"; H 392 m). This bed yielded a fairly abundant fauna of *Archiclimacograptus* specimens, along with uncommon *Hustedograptus*, *Nemagraptus*, and the same *Dicranograptus* species found in stratigraphically lower horizons. A preliminary description of this assemblage is presented below.

THE GRAPTOLITE ASSEMBLAGE

Occurrence

The Caparo Formation graptolites are preserved as organic films on dark argillaceous shales frequently weathered to yellow to grey colours in the section. These shales are intercalated with lighter colored

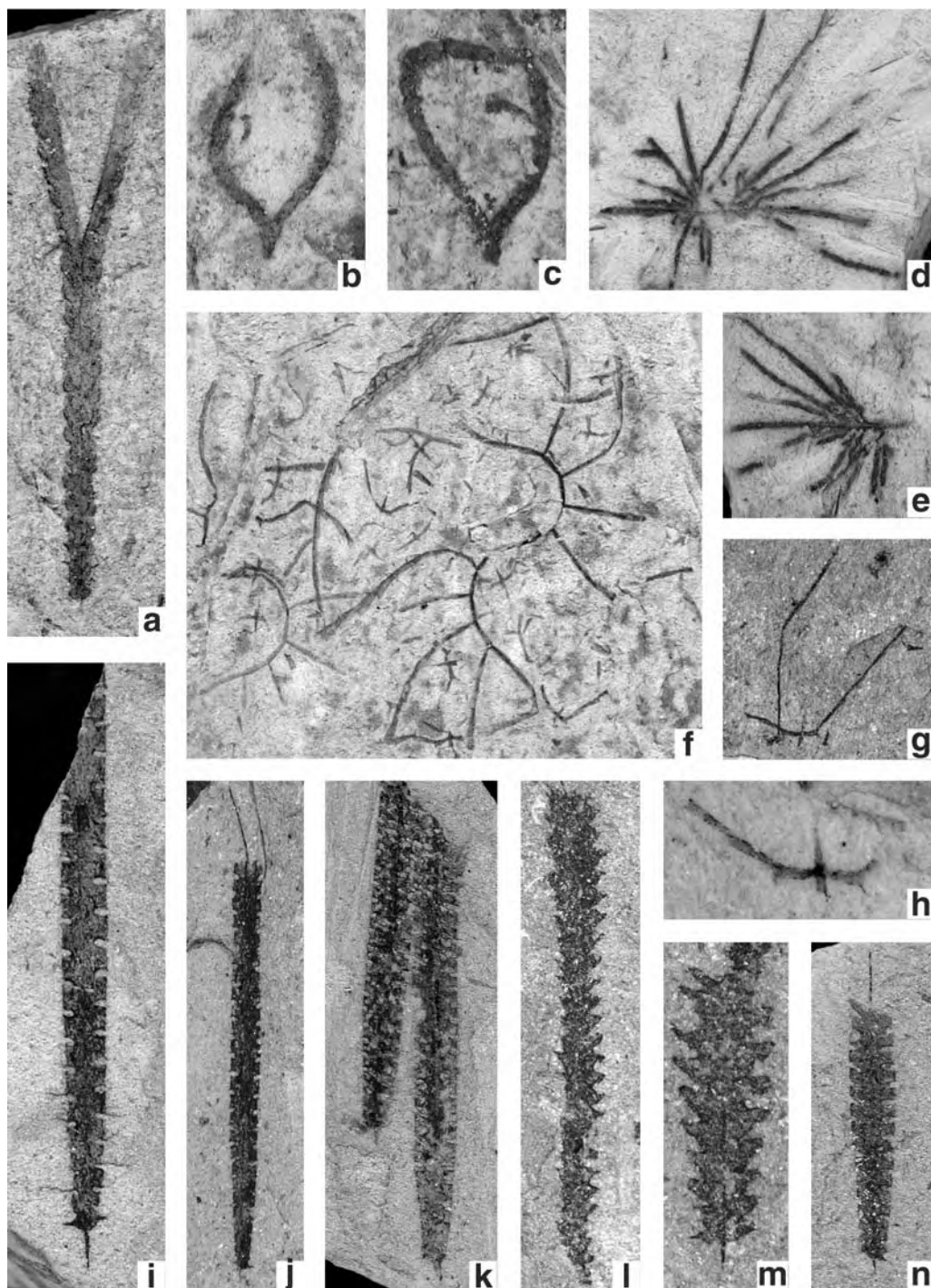
laminae consisting of thin bands with a sandy texture that are very rich in transported fossils, such as dissociated sclerites of trinucleid and calymenacean trilobites, isolated valves of orthid and organophosfatic brachiopods, small fragments of ramose bryozoans and graptolites, crinoid columnals, and a few smooth ostracods. Our sedimentological interpretation is that the graptolite-bearing beds represent distal turbidites, an analysis in agreement with the occurrence of some “deep-water” olenid trilobites (*Porterfieldia*, *Triarthrus*?) with the graptolites at the Lirán creek (Shell and Creole, 1964; Hughes, 1980).

Transported fragments of benthic graptolites belong to the genera *Dictyonema*, *Desmograptus* (both recorded from the section by Gutiérrez-Marco et al., 1992), and another undetermined form resembling *Callograptus* or *Dendrograptus*. All the specimens are too fragmented for species level identification.

Taxonomic notes

1. *Nemagraptus gracilis* (Hall) was fully re-described and illustrated from both flattened and isolated specimens by Finney (1985). Our specimens (Figs. 1f–h) fully agree with Finney’s (1985) description.
2. *Dicranograptus ramosus* (Hall). Ruedemann (1947) provided a full description of Hall’s (1847) species, noting that it is characterized by a very long biserial portion (13 to 18 thecal pairs) and a narrow axial angle between the uniserial stipes. Topotypical and other specimens collected by one of the junior author (DG) exhibit similar variability in the length of the biserial portion and also have mesial spines on the first 2 to 5 thecal pairs. Leith (1938) differentiated *Dicranograptus caparroensis* n. sp. from *D. ramosus* based on the former having greater sigmoidal curvature to the ventral thecal walls, a slightly longer biserial portion (17 as opposed to 15 mm), and a slightly larger axial angle (40 as opposed to 30 degrees). An examination of Leith’s (1938) figures and our new specimens (Fig. 1a) indicates that all the Venezuelan material falls within the range of variation exhibited by other specimens of *D. ramosus*. In South America, *D. ramosus* has also been recorded from the *C. bicornis* Zone of the Argentine Precordillera (Cuerda et al., 1998; Toro and Brussa, 2003).
3. *Dicranograptus furcatus* (Hall). Several small species of *Dicranograptus* that have short, spinose, biserial portions (3 – 8 thecal pairs) and exhibit pronounced torsion in the uniserial stipes have been described from Sandbian strata in the eastern United States and Great Britain (e.g., *D. contortus* Ruedemann, *D. furcatus* (Hall), and *D. ziczac* Lapworth). Our specimens exhibit a very short (3 – 4 thecal pairs), spinose, biserial portion, and short, curved, uniserial stipes that form the start of spiral loops (Figs. 1b–c). The Venezuelan specimens best fit the descriptions for *D. furcatus* (Hall), which is also the name that maintains priority if future studies demonstrate that any of these taxa are synonymous with one another. The Venezuelan material confirms earlier but questionable records of this species from South America that were listed as *D. cf. furcatus* from the *N. gracilis* Zone of the central Precordillera, Argentina (Ortega et al., 2008).
4. *Amphigraptus divergens* (Hall). Specimens of *Amphigraptus* exhibit two stipes that diverge from the sicula at approximately 180 degrees from each other, and also bear distinctive paired cladia. The Venezuelan specimens (Figs. 1d–e) agree with Ruedemann’s (1947) description and no other species of *Amphigraptus* are known from Sandbian age strata. This rare but characteristic graptolite was

Figure 2. Sandbian graptolites from the Caparo Formation, Venezuelan Andes. a, *Dicranograptus ramosus* (Hall), x 3.3; b–c, *Dicranograptus furcatus* (Hall), both x 3; d–e, *Amphigraptus divergens* (Hall), x 0.7 and x 1.6, respectively; f–h, *Nemagraptus gracilis* (Hall), x 3.5, x 3.8 and x 8, respectively; i–k and n, *Archiclimacograptus meridionalis* (Ruedemann), x 5.2 (i), x 3 (j–k) and x 3.6 (n); l–m, *Hustedograptus vikarbyensis* (Jaanusson), x 3.3 and x 6.2, respectively.



previously recorded in South America only from the Upper Ordovician of the Argentine Precordillera (Cuerda, 1979).

5. *Hustedograptus vikarbyensis* (Jaanusson). Jaanusson (1960) described a new species of "*Glyptograptus*" (now *Hustedograptus* Mitchell, 1987), "*G.* vikarbyensis, from the Furudal Limestone (*Hustedograptus teretiusculus* Zone) on Öland, Sweden. *Hustedograptus vikarbyensis* was differentiated from the more commonly cited *H. teretiusculus* by its narrower rhabdosome and more symmetrical proximal end – i.e., the first two thecae form a symmetric "w" shape with their upward facing apertures occurring at approximately the same level (Jaanusson, 1960; Maletz, 1997). Our specimens (Fig. 1l–m) agree completely with Jaanusson's (1960) description of the specimens from Sweden. In South America, *H. vikarbyensis* has also been recorded from the *H. teretiusculus* Zone of the central Precordillera, Argentina (Ortega et al., 2008).
6. *Archiclimacograptus meridionalis* (Ruedemann). This genus currently comprises two distinct sets of species (Mitchell, 2007), a more primitive group that has strongly introverted apertures (e.g., *Archiclimacograptus decoratus* and *A. sebyensis*) and a derived group with nearly horizontal, semi-circular apertures (e.g., *A. meridionalis* and *A. antiquus*). The Venezuelan specimens have thecae with straight ventral walls and relatively shallow, horizontal, semi-circular apertures (Figs. 1i–k, n), and clearly belong to the derived group. Their dimensions (rhabdosomes widen from about 0.8 mm at the second thecal pair to 1.3 – 1.5 mm distally, and having 11 – 13 thecae in 10mm proximally) fit most closely to *Archiclimacograptus meridionalis* (Ruedemann). The slightly fusiform shape of the rhabdosome also agrees with the morphology of *A. meridionalis*. Our specimens also resemble *A. antiquus* (Lapworth), but tend to be narrower with shorter thecae than the latter species. In South America, another possible record of *A. meridionalis* comes from the *N. gracilis* Zone of the central Precordillera, Argentina (Ortega et al., 2008).
7. *Cryptograptus* sp. Several fragmentary specimens of *Cryptograptus* occur in our collection. These are not well enough preserved for a species level identification.

Biostratigraphy

Our new collections from the Caparo Formation along the trail from the Uribante-Caparo Lake to Santa Bárbara de Barinas contain a fauna that is referable to the *Nemagraptus gracilis* Zone. The presence of the eponymous species along with *Dicranograptus ramosus*, *D. furcatus*, and *Archiclimacograptus meridionalis* clearly indicate a Sandbian age for the strata. Although many of the taxa range up into the upper Sandbian, the complete absence of *Climacograptus bicornis* or any astogenetic Pattern G orthograptids (e.g., *Orthograptus calcaratus* group species) indicate that a lower Sandbian (*N. gracilis* Zone) age assignment is most appropriate.

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ORDOVICIAN BRACHIOPOD DIVERSITY REVISITED: PATTERNS AND TRENDS IN THE OSLO REGION

J.W. Hansen, D.A.T. Harper and A.T. Nielsen

Natural History Museum of Denmark, Geological Museum, Øster Voldgade 5-7, DK-1350 Copenhagen K, Denmark.
jwhansen@snm.ku.dk, dharper@snm.ku.dk, arnet@snm.ku.dk

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INTRODUCTION

The Great Ordovician Biodiversification Event is the single most significant marine radiation in the history of planet Earth and led to a major increase of taxa at species through to family level within many of the animal groups that already was established in the Cambrian, including the brachiopods (e.g. Harper, 2006, 2010; Rasmussen et al., 2007; Servais et al., 2008; Servais et al., 2010). The radiation was associated with an extensive continental breakup, high sea levels together with the plankton revolution and has also been linked to extraterrestrial factors, such as widespread asteroid impacts (Schmitz et al., 2008; Servais et al., 2009; Parnell, 2009). This study re-evaluates the regional diversity curve from the Oslo Region, placing these data in a global context.

During the Early Paleozoic the Oslo Region formed part of a cratonic basin, today confined within a Permian graben structure (Owen et al., 1990). The sediments of the Oslo Region are predominantly shale (some places mudstone) interbedded with limestone that are more dominant in the upper parts of the succession. The Oslo Basin was, during the Ordovician, positioned closer to the developing Caledonian Orogen than the facies belts of East Baltica (Bruton and Harper, 1988). This is indicated by the greater thicknesses and significantly higher content of siliciclastic material (Jaanusson, 1972), than apparent in contemporaneous sediments elsewhere in the Baltic Region (Fig. 1). The Oslo Region with its tight stratigraphical constraints, rich brachiopod faunas and detailed palaeoenvironmental data is ideally suited to monitor regional changes in diversity through the Ordovician Period. Data from the Oslo Region indicate that the first phases of the GOBE here and elsewhere probably involved endemic taxa in shallow-water environments, whereas the later Ordovician radiations, during the late Sandbian and late Katian, were characterized by firstly a move into deeper-water environments and the engagement of more cosmopolitan taxa and secondly by a greater occupation of carbonate environments during the late Katian.

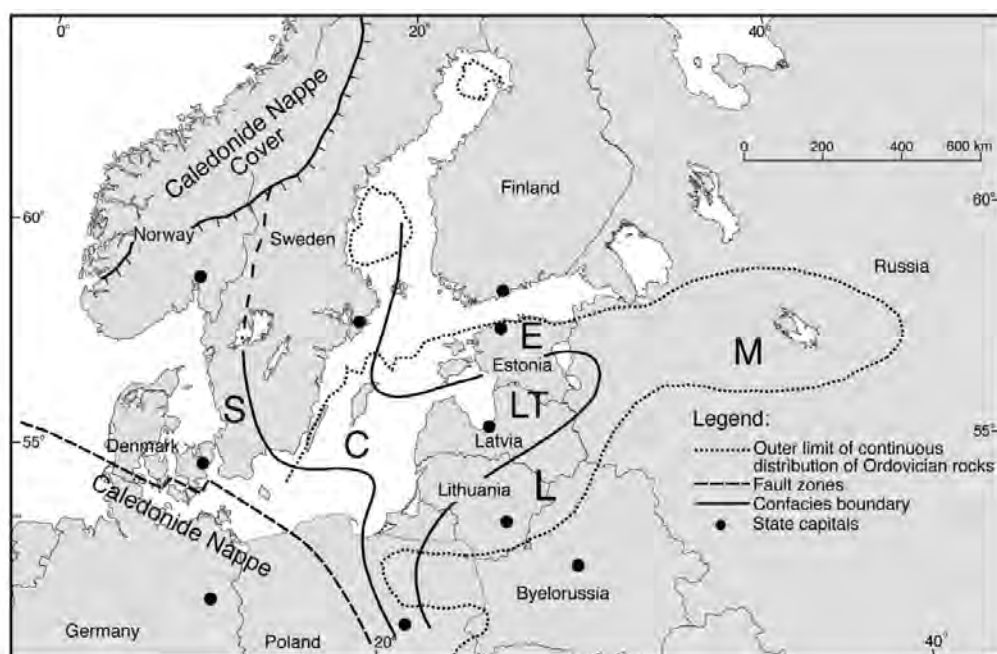


Figure 1. The Baltoscandian area, showing positions of Confacies Belts and other important geological features (Jaanusson, 1982). Abbreviations for Confacies Belts. C: Central Baltoscandian, E: North Estonian, L: Lithuanian, LT: Livonian Tongue, M: Moscow Basin, S: Scanian. (Modified from various sources including Nielsen, 2004 and Rasmussen et al., 2007).

GLOBAL DATA

The statistical analyses of global marine diversity patterns trends was pioneered in the late 1970s (Sepkoski, 1979) based initially on family data (see also Benton, 1993) but later extending to generic data sets (Sepkoski, 1993). These extensive databases were extracted from the literature, in particular the *Treatise on Invertebrate Paleontology* usually accurate to the nearest stratigraphical stage. The reality and interpretations of such global marine biodiversity trends have been debated for years (e.g. Miller and Foote, 1996), partly due to the incompleteness of the fossil record and sampling bias especially in older stratigraphical units (Alroy et al., 2008). The data, however, for the Ordovician radiation are now substantial, the trends well established and stratigraphical precision refined. Harper et al. (2004) demonstrated three clear peaks for the typical “modern” articulated forms, the Orthida; the group diversified during the Darriwilian (late Arenig–early Llanvirn), late Sandbian (mid-Caradoc), and late Katian (mid-Ashgill), coincident with the disparate continental and microcontinental configuration of that time, and a subsequent expansion into deeper-water habitats; and finally a radiation in carbonate buildups. The pattern for the Strophomenida is different; the group expanded first in the Dapingian (mid-Arenig) but did not peak until the early Katian (mid-Caradoc), with a less marked late Katian (mid-Ashgill) spike. The patterns of the other, more minor groups—the atrypides, pentamerides, and rhynchonellides—differ in detail, diversifying later in the lower Darriwilian (upper Arenig), with maximum levels in the later Katian (Ashgill); the late Katian (mid-Ashgill) diversifications may have been associated with carbonate

environments during the later Ordovician. The last three groups, in particular, dominated the Silurian benthos following the end-Ordovician extinction event (Harper and Rong, 2001), when carbonate facies were more prevalent. Nevertheless such global databases are by their nature isolated from local environmental factors that may be the initial triggers for diversifications.

Regional data sets, constructed through locality-based collections may be more complete, being a product of meticulous sampling, yielding more precise stratigraphical data together with environmental and palaeogeographic information. However, global diversity trends are difficult to correlate from the curves alone, since the controlling parameters can be highly complex and not fully understood (Servais et al., 2008). Thus regional studies reflect more the reality of diversification but nonetheless can act as a proxy for more global trends. In this study we dissect data sets from the Oslo Region and the East Baltic Region (Hints and Harper, 2003), both of which have been sampled thoroughly during the last 3 decades, and therefore contain a large compilation of detailed data both in the field and from museum collections.

The global data indicate that the threefold global brachiopod radiation initiated at the transition from the Floian to the Dapingian with its first major peak in the Darriwilian. Preliminary results from the Oslo Region suggests that the radiation here took place in the Dapingian, somewhat earlier than in the East Baltic Region, implying that the brachiopod assemblages in the Oslo Region were influenced by local factors such as a higher influx of siliciclastic material from the Caledonide Orogen. In broad terms the majority of the Middle Ordovician brachiopod faunas are endemic to Baltoscandia, which retained its

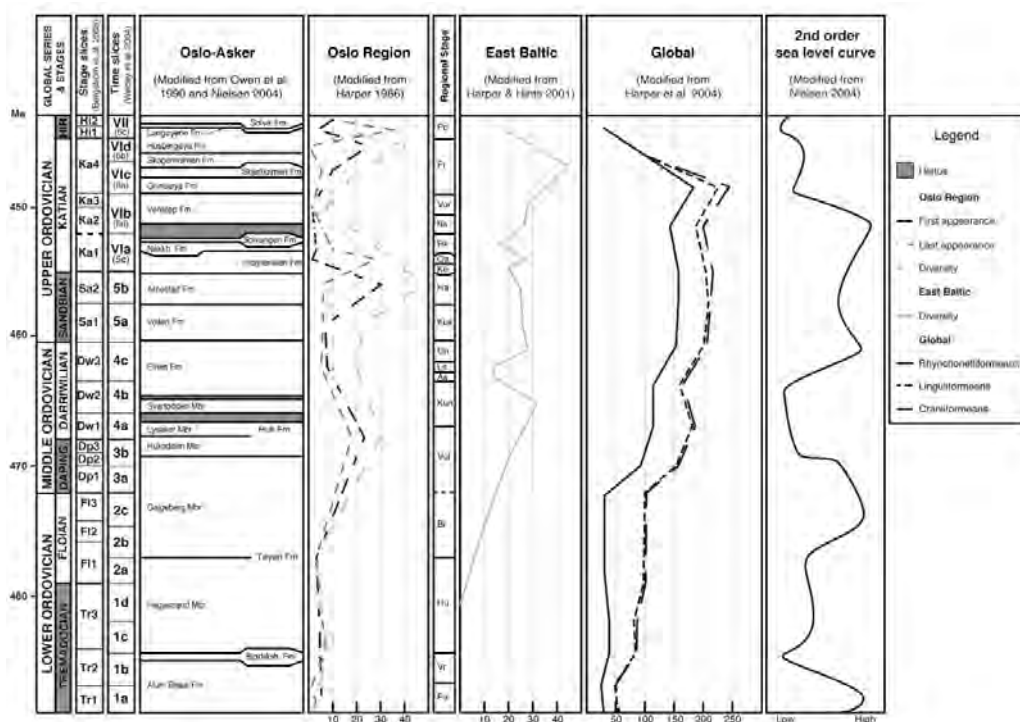


Figure 2. Brachiopod diversity curves through the Ordovician of the Oslo Region, East Baltic area and at a global level. Column to the right shows the 2nd order sea level curve. (Modified from sources noted on the figure).

insularity during this interval (Rasmussen et al., 2007). As Baltica drifted towards Equator during the period, modifying restructuring depositional environments against a background of sea level change, little or no immigration contributed to the Dapingian/Darriwilian peak (Rasmussen et al., 2007). In the Sandbian, however, migration took place into deeper water settings, and immigrations into the Baltica area were more frequent, culminating in the abundant brachiopod faunas of the Katian, commonly associated with carbonate facies. Here we investigate these signals in data from the Oslo Region and discuss briefly some of the regional factors that may have driven the diversifications.

REGIONAL DATA

The brachiopod biodiversity profile from the Oslo Region was first assembled by Harper (1986), who provided a diversity curve together with appearance and disappearance data, based on 'bag samples' from the literature and extensive collections in the then Palaeontologisk Museum, Universitetet i Oslo (now the Geology Department of the Natural History Museum in Oslo). The curve has been enhanced by including a range of new data but the overall trends in peaks and troughs remain roughly the same. The curve can be dissected into three main peaks: Dapingian/Darriwilian boundary, late Sandbian and late Katian. The initial Dapingian/Darriwilian boundary peak is roughly coeval with data from western Russia but earlier than the hike apparent in Estonia. It postdates the early radiations during the Tremadocian on the South China Plate (Zhan and Harper, 2006) but predates diversifications around on and around Laurentia during the Darriwilian (Droser and Finnegan, 2003).

This initial biodiversification apparently involved the diachronous expansion of largely endemic faunas, where radiations were regional phenomena, controlled by the effects of both local and global factors on indigenous populations. In the Oslo Region the faunas were dominated by characteristic endemic orthides and clitambonitides with rarer strophomenides (Öpik, 1939); the diversification was associated with a siliciclastic substrate, sourced from the west in the emerging Caledonian mountain chain. These diversifications were generally associated with shallow-water environments in the Huk Formation. Despite a marked deepening in the subsequent Elnes Formation, much of the brachiopod remained endemic dominated by the plectambonitoids *Alwynella*, *Cathyrina* and *Wandaasella* (Candela and Hansen, 2010). Low-diversity faunas continued in the deep-water facies of the lower Arnestad Formation (Hansen and Harper, 2007). The late Sandbian diversifications, however, were linked to deeper-water conditions and marked migrations of more cosmopolitan taxa into the Oslo Basin, above the thick bentonites of the Arnestad Formation (Hansen and Harper, 2008). Here the faunas were punctuated by the appearance of a number of amphirotic taxa and more diverse deep-shelf assemblages dominated by more cosmopolitan taxa together with a few Baltic endemics. The deep-water environments of the Nakkholmen and Venstøp formations yield sparse, low-diversity faunas dominated by nonarticulates, small dalmanelloids and plectambonitoids. Brachiopod diversity peaked again during the late Katian associated with moderately high sea levels, the Boda Warming Event and widespread development of carbonate facies. Elsewhere on Baltica, the Boda Limestone formed a centre for brachiopod endemism associated with the availability of a range niches developed within and around carbonate mudmound facies (Rasmussen et al., 2010). In the Oslo basin a wide range of mainly carbonate facies hosted diverse orthide-strophomenide assemblages developed across a spectrum of shelfal depths (Brenchley and Cocks, 1982), and prior to the immigration of, first the typical *Hirnantia* brachiopod fauna of the Kosov Province followed by elements of the Edgewood Province (Rong and Harper, 1988).

CONCLUSIONS

Global databases, based largely on the literature and more specifically on the *Treatise*, have indicated three main peaks during the Ordovician biodiversification of the Brachiopoda. The timing of these peaks is relatively precise and while the first two peaks are dominated by the orthides and strophomenides, the third peak is characterized by a significant number of taxa such as the atrypides, athyrides, pentamerides and rhynchonellides, more common in the Silurian. Nevertheless the regional dynamics of these peaks are far from clear. Here the dissection of a regional dataset from the Oslo Basin demonstrates some clearer patterns: An initial diversification in shallow-water environments of the Baltic fauna, a second biodiversification triggered by an expansion into deeper-water environments together with immigrations and thirdly the increased exploitation of carbonate environments. In this respect the Oslo Region has contributed initially to an increase in α -diversity during all three phases and to β -diversity with the occupation of deeper-water environments and carbonate facies during the late Sandbian and late Katian. In broad terms the data from the Oslo Region conforms to large-scale models for the GOBE (Harper, 2010) but dissection of these regional trends provides a direct opportunity to relate diversification to more local factors that may have driven at least the early stages of the event.

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ORDOVICIAN ON THE ROOF OF THE WORLD: MACRO- AND MICROFAUNAS FROM TROPICAL CARBONATES IN TIBET

D. A.T. Harper¹, R. Zhan², L. Stemmerik³, J. Liu⁴, S.K. Donovan⁵ and S. Stouge¹

¹ Natural History Museum of Denmark (Geological Museum), University of Copenhagen, Øster Voldgade 5-7, DK-1350 Copenhagen K, Denmark. Dharper@snm.ku.dk, Svends@snm.ku.dk

² State Key Laboratory of Palaeobiology and Stratigraphy, Nanjing Institute of Geology and Palaeontology, Chinese Academy of Sciences, 39 East Beijing Road, Nanjing 210008, China. rbzhan@nigpas.ac.cn

³ Department of Geography and Geology, University of Copenhagen, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark. Ls@geo.ku.dk

⁴ Institute of Paleontology and Palaeoenvironment, School of Earth and Space Sciences, Peking University, China. jbliu@pku.edu.cn

⁵ Department of Geology, Netherlands Centrum voor Biodiversiteit - Naturalis, Postbus 9517, NL-2300 RA Leiden, The Netherlands. Steve.Donovan@ncbnaturalis.nl

Keywords: Tibet, Ordovician, Darriwilian, palaeoenvironments, palaeogeography.

INTRODUCTION

The highest rocks on Earth, marking the summit of Mount Everest, are Ordovician limestones, deposited in a warm, shallow-water sea some 450 million years ago. More remarkably, these rocks still contain the fossils of marine animals such as brachiopods and crinoids that occupied tropical habitats during one of the most important intervals in Earth history, the Great Ordovician Biodiversification Event (GOBE). Some of the first groups of mountaineers to attempt the summit in the 1930s correlated these strata with the Carboniferous or Permian systems. These correlations were revised by Chinese expeditions to the region in the 1970s, who assigned the rocks to the Ordovician and described some of the macrofaunas collected from these rocks Yin et al. (1978). The Ordovician unit we investigated crops out at lower altitudes, for example, near Jiacun, where it is exposed adjacent to the Lhasa - Kathmandu highway, north of Nyalam, at some 4.5 km altitude. These strata have yielded new shallow-water faunas dominated by brachiopods and crinoids while new conodont data precisely correlate part of the succession with the *P. serra* conodont Zone (upper Darriwilian); the Colour Alteration Index of the conodonts indicate a finite temperature of 350°C to 550°C. The shallow-water shelly faunas were dominated by suspension feeders including orthide and strophomenide brachiopods, and a robust pentameride crinoid, '*Pentagonopentagonalis*' (col.) sp. Multivariate statistical analyses of the distributional patterns of the Brachiopoda, place the fauna within the Toquima-Table Head realm, a circum equatorial province contrasting against the higher latitude Celtic and Gondwanan faunas during the late Mid Ordovician.

HISTORICAL BACKGROUND

Geologist Noel Odell noted, with some excitement, the occurrence of fossiliferous limestones overlying metamorphic rocks at some 25,500 feet near the summit of Everest during the 1922 British expedition (Odell, 1924), but it was not until the 1933 expedition that the rocks were systematically sampled. Lawrence Wager collected a suite of over 200 representative rock samples during his ascent together with Percy Wyn Harris along the NE ridge of the mountain. He collected a grey, nodular limestone from a band forming the first step at some 27,890 feet, a lithology that probably composed most of the subsequent summit of Everest. Fragments of fossils were noted in the rock and both Wager (1934) and, previously, Odell (1925) considered the rocks to be of Permian age. During the mid 1970s, Chinese mapping expeditions to the high Himalayas including the Everest Region developed maps and cross sections for this part of the mountain belt (Yin et al., 1978). Palaeontological investigations established faunal lists for the main unit, the Chiatsum (Jiacun) Group, assigning an Ordovician age, while the brachiopod faunas were described for the first time. Thus, the entire upper part of Everest, above the distinctive Yellow Band and the Qomolangma detachment structure, consists of Ordovician carbonates over 200 m in thickness and these rocks can be traced westwards where they crop out at lower, more accessible altitudes in the Nyalam Region, near Jiacun. Thus, the highest rocks on Earth are Ordovician in age (see also Ross, 1984) and amongst the highest fossils on the planet are brachiopods, conodonts and crinoids.

A Sino-Danish expedition visited the southern part of the Tibet plateau during June-July 2009, developing a focus on past and recent climate. The Tibetan plateau offers an ideal setting to study biodiversity and climate change while providing geologists an opportunity to map and resample the Ordovician limestones in the Nyalam region near the uninhabited village of Jiacun (Fig. 1).

GEOLOGICAL SETTING AND REGIONAL STRATIGRAPHY

The Tibetan plateau, variously called the 'Roof of the World' or the 'Third Pole', is an extensive, elevated area covering most of the Tibet Autonomous Region occupying an area of 1.2 km² at an average elevation of over 4,000 m. The Ordovician limestones exposed adjacent to and north of Nyalam form part of the southern margin of the Tethyan Himalaya zone, delimited to the south by the Southern Tibetan detachment system that separates the tectonized rocks, including the Rouqiecu Group and Yellow Band, from the lower-grade carbonate rocks of the Chiatsum Group (Myrow et al., 2009). This succession can be tracked eastwards to the summit of Everest

where the Chiatsum Group, locally assigned to the Qomolangma Formation, is rich in brachiopods and pelmatozoan debris (Gansser, 1964; Harutaka et al., 2005). The stratigraphy of the Nyalam outcrop has been recently described in a comprehensive synthesis of the Cambrian-Ordovician rocks along the Himalayan belt with particular focus on the Everest Region (Myrow et al., 2009). The fossiliferous limestones above the Qomolangma Detachment System were assigned to the Lower Chiatsum Group, estimated to be some 450 m thick, and correlated with the Mt Qomolangma Formation at Everest. These units are correlated with the Lower and Middle Ordovician based on their shelly faunas, including brachiopods and cephalopods, together with conodonts. New conodont data precisely correlate part of the succession with the *P. serra* conodont Zone (upper Darriwilian); the conodonts have a Colour Alteration Index (CAI) of 6, indicating a finite temperature of 350°C to 550°C.

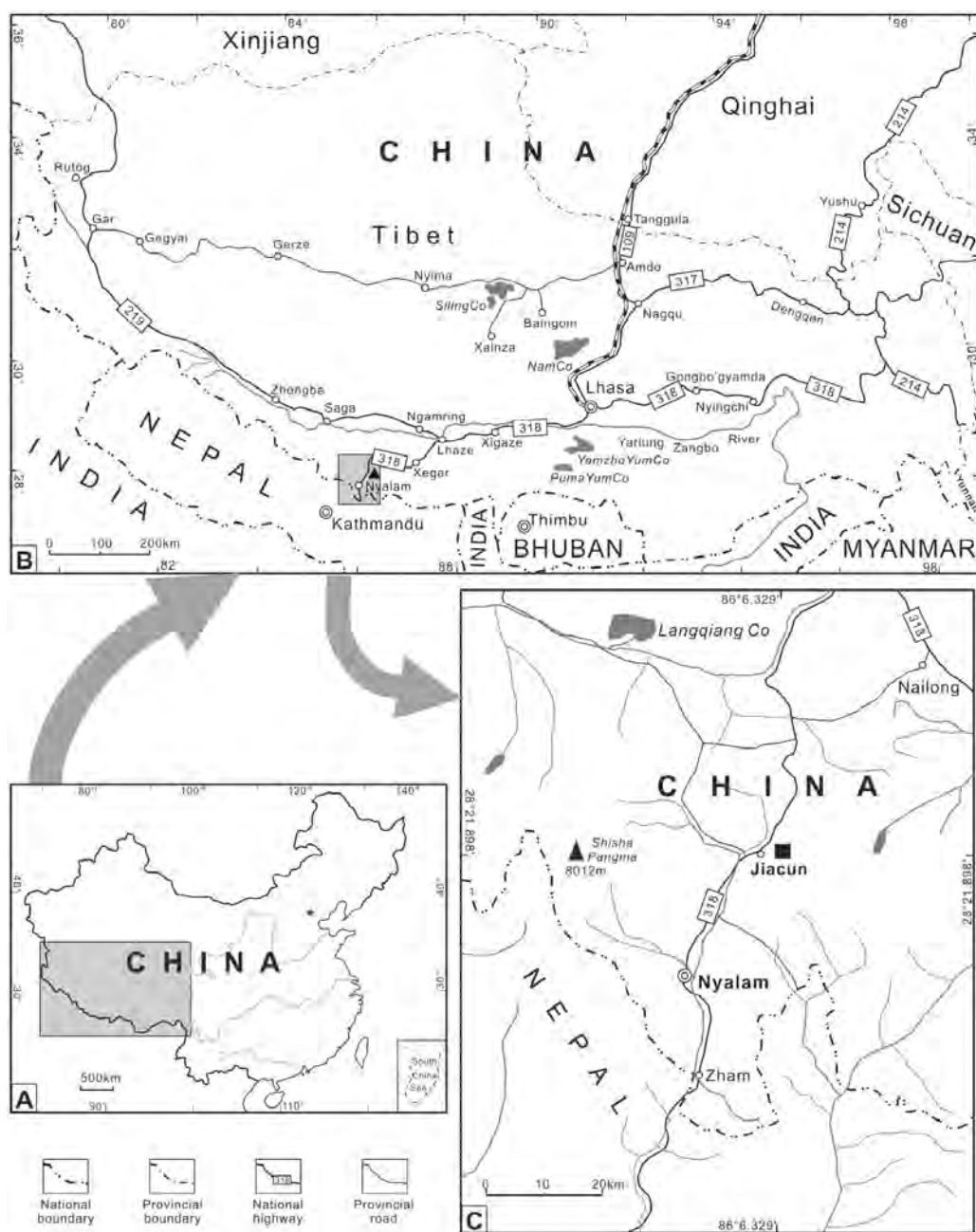


Figure 1. Location of the study area within China (A), along the Nepal-Tibet border area (B) and the site of the study area adjacent to Jiacun (C).

DEPOSITIONAL SETTING AND FAUNAS

The Lower Chiatsum Group comprises cycles of bioclastic shelf limestones and peritidal dolomites, ranging in depth from shallow to deep subtidal zones and suggesting warm, subtropical environments. Against this background of cyclicity, the succession is transgressive; the crinoids including a robust cladid crinoid, '*Pentagonopentagonalis*' (col.) were recovered from a unit of skeletal packstones in the deeper-water part of the succession supplementing previous data from the group (Mu and Wu, 1975). Preliminary palaeontological study shows that the brachiopod fauna is dominated by the plectambonitoids *Aporthophyla*, *Leptellina*, *Aporthophylina*, *Nanambonites* and *Spanodonta*, the orthoid *Orthambonites* (probably *Paralenorthis*) together with the syntrophoid *Xizangostrophia* (Liu, 1976). Three genera, *Aporthophylina*, *Nanambonites* and *Xizangostrophia*, were new. Irrespective of the endemic taxa, the fauna as a whole was compared with the Whiterock, Toquima-Table Head faunas of the Laurentian margins (Fig. 2).

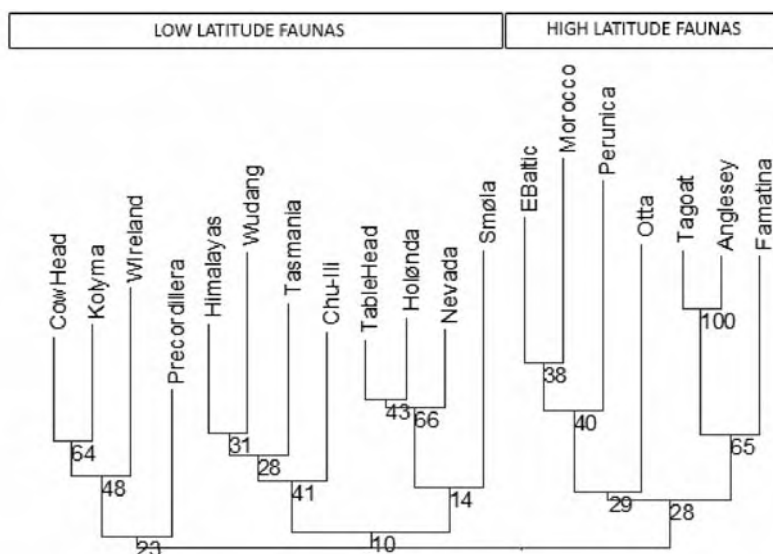


Figure 2. Cluster Analysis based on the dataset in Rong et al. (2005), but modified with additional data from Tibet and Australia (Laurie, 1991). The data from the Tibetan Himalayas plots within a cluster including South China (Wudang), Australia (Tasmania) and Chu-Ili (Kazakhstan) adjacent to the main group of Toquima-Table Head faunas. Values attached to nodes are based on 100 bootstraps.

PALAEOENVIRONMENTAL AND PALAEOGEOGRAPHICAL IMPLICATIONS

Initial studies of the brachiopod fauna from the Nyalam region suggested a Whiterock (early Darriwilian) age and biogeographical similarities with those of the Toquima-Table Head province (Fig. 2). The Toquima-Table Head Realm was first established for peri-cratonic faunas that developed around the margins of the Laurentian continent during the Mid Ordovician. Available palaeomagnetic data demonstrate that the Tethyan Himalaya was probably located in proximity to the Indian craton, adjacent

to 30°S during the Early Ordovician and forming part of a continuous west-facing Gondwanan margin at that time (Torsvik et al., 2009). This subtropical setting is consistent with the new sedimentological and palaeontological data.

CONCLUSIONS

Our new investigations have revealed a diverse fauna dominated by brachiopods, commonly in shell concentrations and deposited within midshelf carbonate environments (corresponding to lower BA2 to upper BA3 according to our preliminary analysis). The new fossil data confirm some of the previous Middle Ordovician correlations for this unit, but may allow more detailed monographic description of the brachiopods and interpretation of their palaeoecology and permit a more precise correlation of this part of the Chiatsun Group. In addition, some stratigraphically higher units within the group were also sampled that may provide a greater age range for the group than previously reported. Preliminary biogeographical analysis of the fauna suggests it may be related to the low-latitude Toquima-Table Head province that extended around the margins of the ancient continent of Laurentia (North America). This implies that a similar tropical belt was developed on the Tibetan margin of the continent of Gondwana during the Mid Ordovician.

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STRATIGRAPHIC RELATIONS OF QUARTZ ARENITES AND K-BENTONITES IN THE ORDOVICIAN BLOUNT MOLASSE, ALABAMA TO VIRGINIA, SOUTHERN APPALACHIANS, USA

J.T. Haynes¹ and K.E. Goggin²

¹ Department of Geology & Environmental Science, James Madison University, MSC 6903, Harrisonburg VA 22807.
haynesjx@jmu.edu

² Weatherford Laboratories, 200 North Sam Houston Parkway West Suite 500, Houston TX 77086.
Keith.Goggin@weatherfordlabs.com

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ABSTRACT

Quartz arenites and granule and pebble conglomerates in the Ordovician Blount molasse of the Taconic foredeep in the southern Appalachians occur at different time-stratigraphic intervals as shown by their stratigraphic relations with correlateable K-bentonites (altered tephtras) most notably the Deicke and Millbrig K-bentonite Beds) that are also in the molasse. The K-bentonites are isochrons; the sandstones and the discon-formities beneath them are not. Consideration of stratigraphy shows unequivocally that (1) the Walker Mountain Sandstone at 29 sections in southwestern Virginia, and the "middle sandstone member" (and the several thinner quartz arenites downsection) at 6 sections in northeastern Tennessee are older than the K-bentonites; (2) the Colvin Mountain Sandstone at 7 sections in Alabama and Georgia is contemporaneous with the K-bentonites; and (3) the unnamed thin granule and pebble conglomerates and pebbly sandstones at 2 sections near Dalton, Georgia, are younger than the K-bentonites. The diachroneity of these gravels and coarse sands is probably evidence of geographic changes in the fluvial networks draining the Taconic highlands, with the result being that pulses of gravel and coarse sand were delivered episodically into the basin over several million years and at different depositional loci as a result of primarily tectonic activity rather than eustatic changes.

Were it not for the presence of the K-bentonites, a likely interpretation would be that these quartzose units would be correlated with each other from Alabama to Virginia, and the prominent disconformity that exists beneath them at most exposures would likely be given local, regional, or maybe even global sequence stratigraphic significance. The unconformity beneath the Walker Mountain Sandstone in Virginia has in fact already been given just such significance, being labeled as either the "M4-M5" or M5-M6" sequence boundary in the Mohawkian. Yet regional stratigraphic relations show that other equally prominent disconformities occur beneath other equally coarse or coarser sandstones that are younger (Georgia and Alabama) or older (NE Tennessee) than the Walker Mountain. This confusion points to the importance of always needing to consider the possibility that not every unconformity is of glacioeustatic origin or has a eustatic component, especially in settings (such as the Taconic foredeep) where tectonic

activity governs and dominates the depositional system. Even though there may be an as-yet unrecognized eustatic signal in this Paleozoic stratigraphic interval, it was probably masked or even obliterated in this region by depositional events that were driven by tectonic events in the Taconic highlands, in much the same way that Cenozoic deposition on the Ganges Plain has been dominated by episodic influx of sands and gravels from the Himalayan highlands.

INTRODUCTION AND BACKGROUND

The coarsest sediments in redbeds of the Bays Formation and related units of the Ordovician Blount molasse in the southern Appalachians are non-red quartz arenites and granule and pebble conglomerates deposited ~460 – 450 Ma, during the early Taconic Orogeny. They include the Walker Mountain Sandstone at 29 sections in Virginia and West Virginia (Hergenroder, 1966; Haynes, 1992, 1994; Haynes and Goggin, 1993, 1994), the “middle sandstone member” and unnamed thinner and older units at 6 sections in northeast Tennessee (Hergenroder, 1966; Haynes, 1994; Haynes and Goggin, 1994), unnamed conglomerates at 2 sections near Dalton, Georgia (Allen and Lester, 1957; Hergenroder 1966; Bayona and Thomas 2003), and the Colvin Mountain Sandstone at 7 sections in Georgia and Alabama (Carter and Chowns, 1989; Haynes, 1994; Bayona and Thomas, 2003) (Fig. 1). Although the Bays Formation is up to 300 m thick in Tennessee, the thickest area of the Blount molasse, thicknesses of the non-red quartzose beds vary from only < 1 m to > 20 m for individual units, and < 1 m to > 50 m for the total thickness of arenites plus conglomerates in a single outcrop.

These molasse sediments were derived from the rising Taconic highlands that were forming as the Laurentian margin buckled and flooded behind its leading edge, where a large continental landmass was being deformed and uplifted as subduction occurred beneath it. Petrographic study (Kellberg and Grant, 1956; Hergenroder, 1966; Mack, 1985) indicates that these sediments were sourced by a terrane of older Paleozoic sedimentary rocks, moderate to high grade metamorphic rocks, and hydrothermally altered granitic or pegmatitic rocks, all of which were above or forelandward of a subduction zone that was generating tephra from explosive volcanic eruptions. A modern analog for this Ordovician setting is the Australia – New Guinea system (Fig. 1 inset). There, sediments are accumulating in a foredeep developed on foundering and downwarped Australian continental crust (Coney, 1973). Seaward of this developing foredeep is the uplifted and eroding tectonic highlands terrane of New Guinea, which includes upturned and eroding sedimentary strata on the deformed margin of the Australian plate, volcanics (both volcanoclastics and tephra) from the magmatic arc, various metamorphics, and unroofed and now eroding plutons of continental character (Hamilton, 1979).

SEDIMENTOLOGY AND PALEOGEOGRAPHY

Sedimentary structures in these non-red arenites and conglomerates of the Blount molasse include planar bedding, tabular and trough crossbedding with reactivation surfaces, current and oscillation ripples, rill marks, load casts, pebble lags and normal and reverse grading of pebbly zones, adhesion ripples, slump structures on oversteepened ripple crests and troughs, and channel structures. Body fossils are limited to laminae of broken and abraded brachiopod and bivalve fragments in some of the well-sorted beach sands. Trace fossils include both subvertical branching (*Lingulichnites?*) and vertical nonbranching (*Skolithos*) burrows.

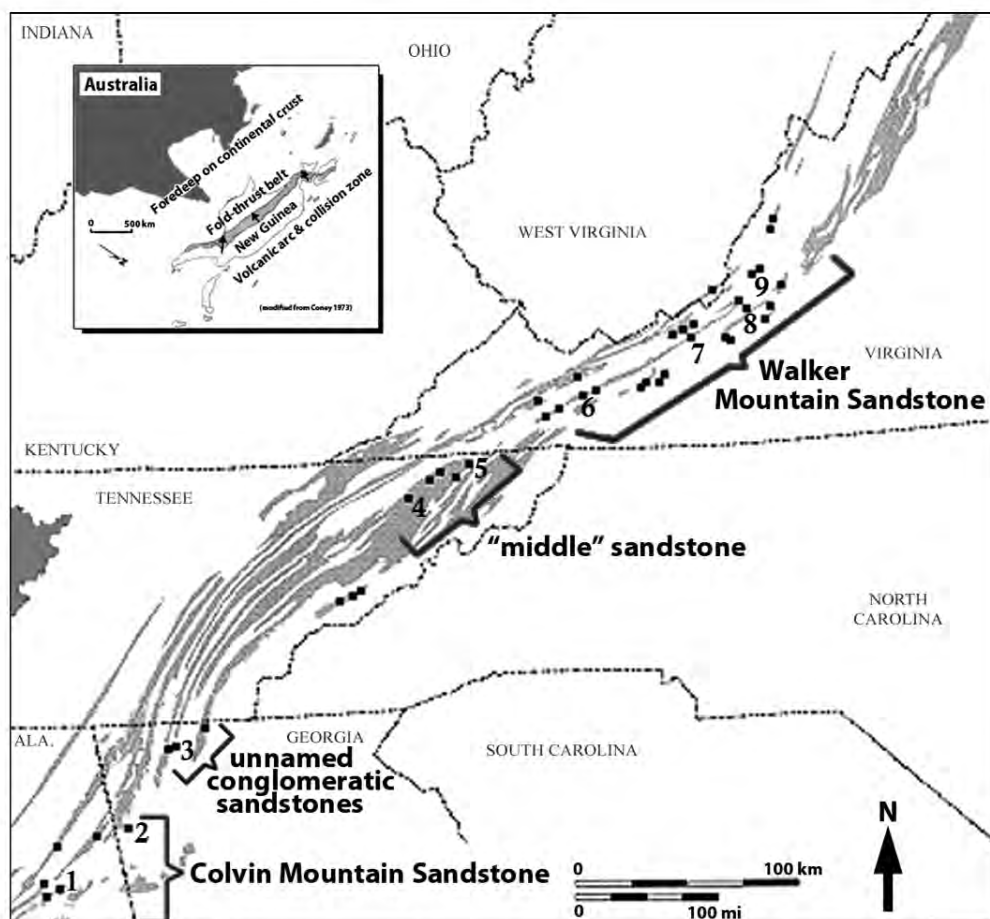


Figure 1. Location of sections with quartzose units in the Blount molasses from Alabama to Virginia. Inset shows modern tectonic setting that is analogous to the Ordovician Laurentian margin.

Sedimentary sequences include both coarsening- and fining-upward sequences, with both normal and reverse grading present. These sediments were likely deposited in nearshore, beach, and coastal plain settings along a current- and wave-dominated coast where tidal and fluvial influences were subordinate. Based on paleogeographic reconstructions and associated atmospheric circulation patterns for the Ordovician (McKerrow et al., 1991), the sands and gravels that are now these sandstones and sandy conglomerates probably accumulated on the leeward side of the Taconic highlands. In this rain shadow, the coastal environment would very likely have been drier and maybe even semi-arid. The lack of terrestrial vegetation in the Ordovician combined with the probable dearth of moisture in this setting would have meant that windblown sediments would have been more common, and in fact the bimodal texture in samples from several sections throughout the region is likely evidence of eolian sorting of pebbly lags in interdune areas (Folk 1968). Several of these bimodal sands have shell fragments associated with them, so these eolian sands were evidently later modified by waves and currents during subsequent transgression, and final deposition occurred in a nearshore or beach environment.

Regional sediment dispersal patterns were likely influenced by episodic tectonic activity in the Taconic highlands and orogenic zone. These gravels and medium- to coarse-grained sands were delivered to the coastal region at different times and places by a depositional complex of braided streams and fan deltas, with associated tidal, beach, and fluvial facies including some with significant eolian influence.

STRATIGRAPHIC RELATIONS WITH THE DEICKE AND MILLBRIG K-BENTONITES

These units not only share the petrographic characteristics of compositional maturity and textural maturity to submaturity, with distinctive framework grains being vein quartz with probable vermicular chlorite, stretched polycrystalline quartz, and common black, gray, and red chert, but they also share important stratigraphic characteristics as well. They are areally restricted to exposures in the eastern Valley and Ridge from Virginia to Alabama. They are also temporally restricted to a narrow stratigraphic interval that for each is nearly isochronous because of their association with uplift resulting from Taconic tectonism, but which for the group of sandstones as a whole from Alabama to Virginia is diachronous and non-eustatic, as evident from their stratigraphic position relative to the Deicke and Millbrig K-bentonites, and their wedge-shaped geometry (Fig. 2).

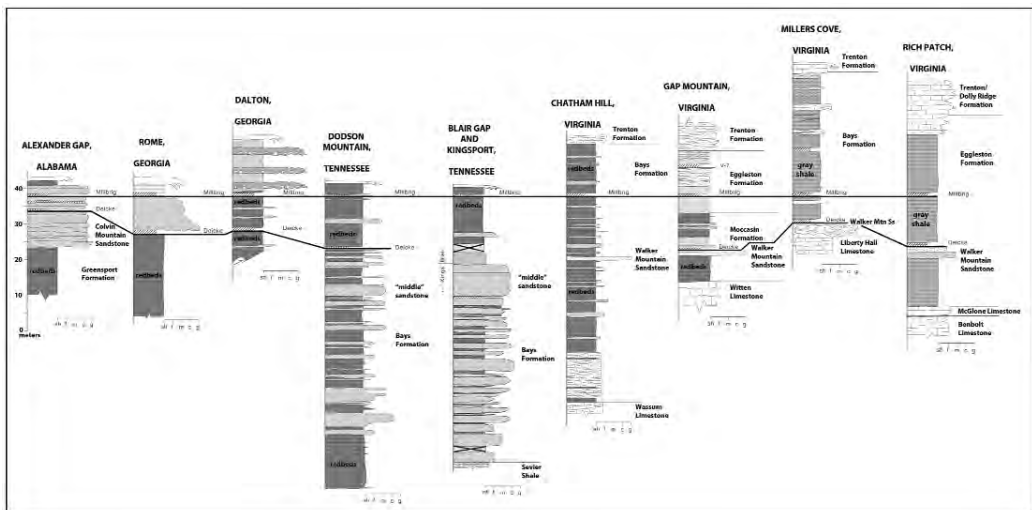


Figure 2. Correlation of the Deicke and Millbrig K-bentonite Beds from Alabama to Virginia showing their stratigraphic position relative to the quartzose units of the Blount molasse.

Stratigraphic relations with the K-bentonites show that the oldest non-red sandstones of the Blount molasse are in northeastern Tennessee and include the "middle sandstone member," as at the Dodson Mountain section (Fig. 2). At the Kingsport and Blair Gap sections (Fig. 2), there are likewise many sandstones downsection from one or both of the K-bentonites. The next youngest non-red sandstone is the Walker Mountain Sandstone of Virginia and West Virginia. It too is downsection from the Deicke K-bentonite, as at the Rich Patch section (Fig. 2). In some exposures, the Walker Mountain is actually directly and immediately beneath the Deicke, as at the Millers Cove section (Fig. 2). In exposures of the Bays

Formation in Virginia, the Deicke is absent, perhaps because the tephra was deposited coevally with the sands of the Walker Mountain, but in those sections the Millbrig is several meters above the Walker Mountain sandstone (Fig. 2). The Colvin Mountain Sandstone in Alabama and Georgia is younger still, as the Deicke and Millbrig K-bentonites are completely within that sandstone at Alexander Gap (Fig. 2), the Deicke is immediately beneath that sandstone at Horseleg Mountain (Fig. 2), and the Millbrig is immediately beneath the Colvin Mountain at the Dirtsellar Mountain section. The unnamed granule to pebble conglomerates in the Bays Formation near Dalton (Fig. 2) are the youngest of these quartzose units in the Blount molasse, as they occur several meters above the Millbrig.

DISCUSSION

Timing of the progradation of these sands and gravels

Figure 3 shows three cross-sections based on stratigraphic information obtained from study of over 40 exposures of the Blount molasse from Alabama to Virginia, including some of those in Figure 2, and from study of over 30 sections of more distal facies (Haynes, 1994). It is obvious that coarse sands and gravels entered the foredeep at different times and places, and subsequently prograded partway across the basin.

In the foreland basin at the southern edge of the Himalayan uplift, a succession of proximal gravels and sands derived from weathering and erosion of the rising Himalayan massif interfingers with finer siliciclastics deposited in more medial and distal settings, and these coarse deposits were deposited out-of-sync with the episodic uplift of the mountains (Heller et al., 1988). This sequence (tectonism being followed by a later pulse of coarse sediment into the foredeep) is recognized at other locations (Blair and Biladeau, 1988; Burbank et al., 1988), and it is a viable explanation for the difference in timing of delivery of the coarse sands and gravels into the Blount foredeep. This implies that the earliest Taconic tectonism was oceanward of present-day northeast Tennessee, and was followed by deposition of the "middle sandstone" and associated non-red sandstones at some time well before eruption of the tephra that became the Deicke K-bentonite. The next significant tectonic activity was to the north, resulting in deposition of the Walker Mountain Sandstone. Then tectonic activity shifted to oceanward of present-day west Georgia and Alabama, with subsequent deposition of the Colvin Mountain Sandstone. The last significant progradation of coarse-grained sediments was oceanward of present-day north Georgia; there the pebble conglomerates near Dalton were deposited after the Millbrig K-bentonite.

Implications for sequence stratigraphy

The ability to parse out differences in timing of progradation of sands and gravels from the Taconic highlands across the proximal Blount foredeep is significant for sequence stratigraphic models. At the over 40 exposures studied, there is a subtle to pronounced disconformity beneath the sandstone or conglomerate. If the K-bentonites were not present and there were no other useful isochrons, the petrography and stratigraphy of these sections would likely lead an investigator to correlate the Walker Mountain Sandstone with the "middle sandstone" – as we originally did – and then with the unnamed conglomerates near Dalton, and with the Colvin Mountain Sandstone. This might lead one to correlate the underlying unconformities with each other and with others including the sub-Walker Mountain Sandstone unconformity that in distal areas of the Taconic foredeep and in cratonic sequences has been given

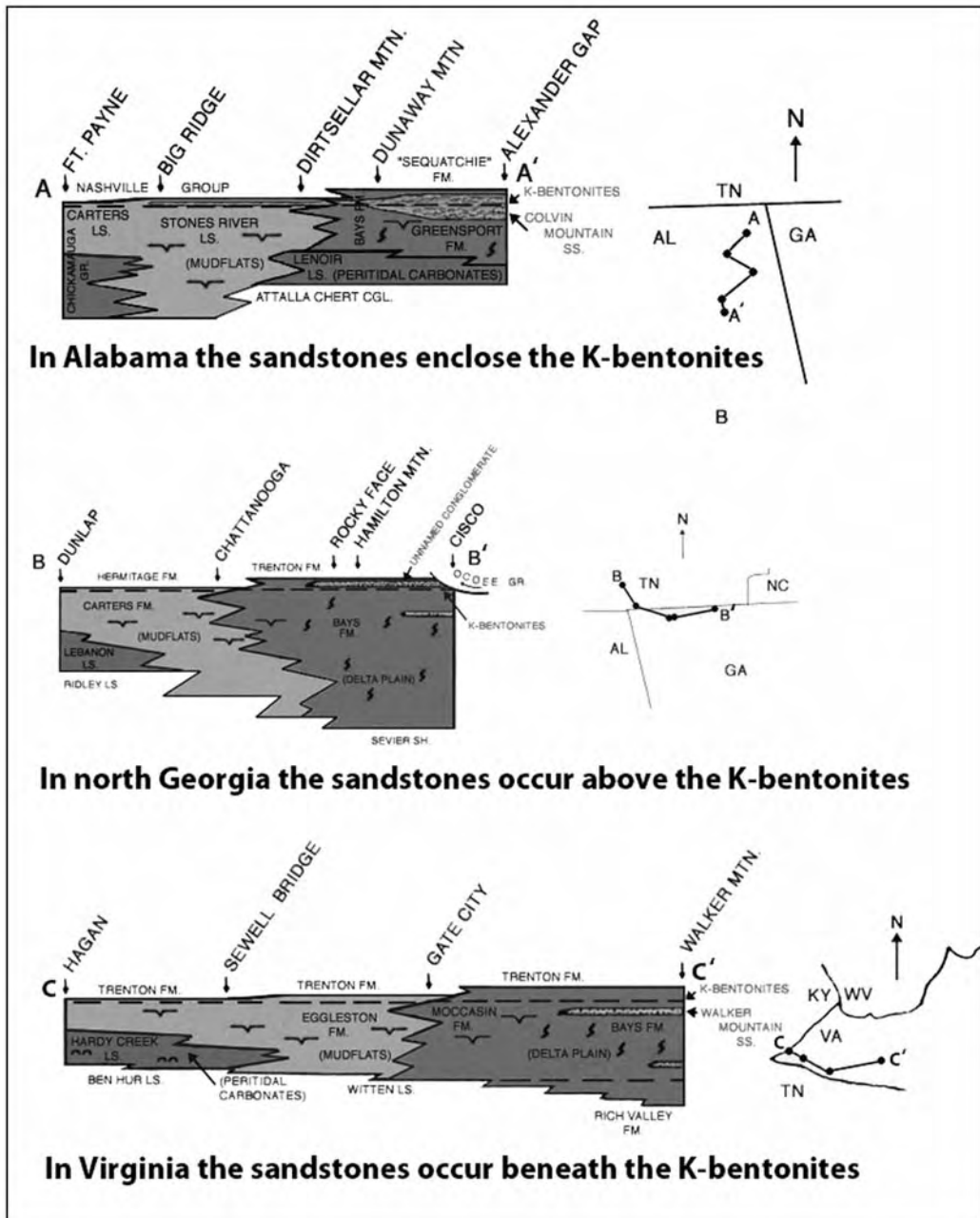


Figure 3. Lithofacies of the upper Ordovician in the southern Appalachians showing that the timing of quartzose sediments entering the basin was diachronous across the Taconic foredeep, and that the associated unconformities beneath these siliciclastics must also be diachronous.

sequence stratigraphic significance as the "M4-M5" boundary (Holland and Patzkowsky, 1996, 1997), a boundary that is crossed by conodont biofacies (Leslie and Bergström, 1994, 1997; Leslie, 2009) as well as by K-bentonites (Kolata et al., 1998). These sequence boundaries have been correlated between the foreland basin of New York and the shelf sequence of central Kentucky (Brett et al., 2004), but unlike the Blount foredeep, the New York foredeep, with its carbonates and shales, is a distal not a proximal part of the Taconic foreland basin, and eustatic signals are recognizable, whereas eustatic signals in the gravels and coarse sands of the proximal Blount foredeep have not yet been recognized. The diachroneity of the foredeep clastics discussed herein shows that correlation of sequences based on unconformities of supposed eustatic origin in the absence of true isochrons (e.g. tephra layers) in the sequence should be done with caution until it can be shown that the unconformities and the overlying coarse sands and gravels are demonstrably eustatic in origin, and are demonstrably isochronous, not diachronous, across the basin from the proximal areas governed by tectonic signals into distal areas governed more by eustatic signals.

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MAJOR ORDOVICIAN TEPHRAS GENERATED BY CALDERA-FORMING EXPLOSIVE VOLCANISM ON CONTINENTAL CRUST: EVIDENCE FROM BIOTITE COMPOSITIONS

J.T. Haynes¹, W.D. Huff² and W.G. Melson³

¹ Department of Geology & Environmental Science, James Madison University, MSC 6903, Harrisonburg VA 22807.
haynesjx@jmu.edu

² Department of Geology, University of Cincinnati, ML-13, Cincinnati OH 45221. huffwd@ucmail.uc.edu

³ Department of Mineral Sciences, NHB-119, Smithsonian Institution, Washington DC 20560. melsonw@si.edu

Keywords: Ordovician Biotite, Deicke, Millbrig, Kinnekulle, Ragland, caldera, explosive volcanism.

ABSTRACT

Compositional variability in $\text{SiO}_2\text{-Al}_2\text{O}_3\text{-FeO-MgO-TiO}_2$ space as determined by microprobe analyses of biotite phenocrysts and quartz-hosted melt inclusions from four altered Ordovician tephra (the Deicke, Ragland, and Millbrig K-bentonites from eastern North America and the Kinnekulle K-bentonite from northern Europe) is compared with compositional data obtained from analyses of many Cenozoic lavas and tephra to constrain tectonic setting and eruptive styles of the volcanoes that produced these beds. The biotites separate into two compositional groups, Mg and Ti-rich biotites (Deicke and Ragland), and Fe-rich biotites (Millbrig and Kinnekulle). Compositionally, the best Cenozoic matches for the Mg-Ti-rich biotites are biotites in calcalkaline lavas that are saturated to slightly oversaturated (quartz phenocrysts are present but rare), metaluminous to weakly peraluminous, and of rhyolitic to dacitic composition produced by large single vent to caldera forming eruptions that include post-caldera rhyolites of Yellowstone in Wyoming, dacites of the Mezitler area in Turkey, ignimbrites of the La Pacana caldera in Chile, rhyolites of the Cerro Chascun complex in Bolivia, and rhyodacites of the Toquima caldera in Nevada. The best Cenozoic matches for the Kinnekulle and Millbrig Fe-rich biotites, are biotites in mildly peraluminous calcalkaline lavas and tephra that are oversaturated, and of rhyodacitic to rhyolitic composition produced by very large caldera-forming Cenozoic eruptions. These include the Toba Tuff of Sumatra, the Bishop Tuff of California, the cordierite-bearing lavas ("ambonites") erupted by volcanoes in northern Ambon along the Banda Arc of Indonesia, the Whakamaru Ignimbrite of New Zealand, and the Cerro Panizos Ignimbrite of the Bolivian tin belt.

ORDOVICIAN TEPHRAS IN EUROPE AND EASTERN NORTH AMERICA

Significance of remnant primary igneous phases

Three of the thickest ($>0.5\text{m}$) and most widespread (regional to subcontinental extent) Ordovician K-bentonites are the Deicke and Millbrig beds of eastern North America and the Kinnekulle bed of northern Europe (Haynes, 1994; Haynes et al., 1995, 1996; Bergström et al., 1995; Kolata et al., 1996). A fourth, the Ragland K-bentonite, is nearly as thick as the other three but is not definitively known outside of some quarries at Ragland, Alabama (Haynes et al., 1996). "Ragland" is an informal stratigraphic name but "Deicke", "Millbrig", and "Kinnekulle" are formal names (Bergström et al., 1995). Figure 1 shows the location where samples of these four beds were obtained, with additional location details in Haynes (1994), and Haynes et al. (1995, 1996).

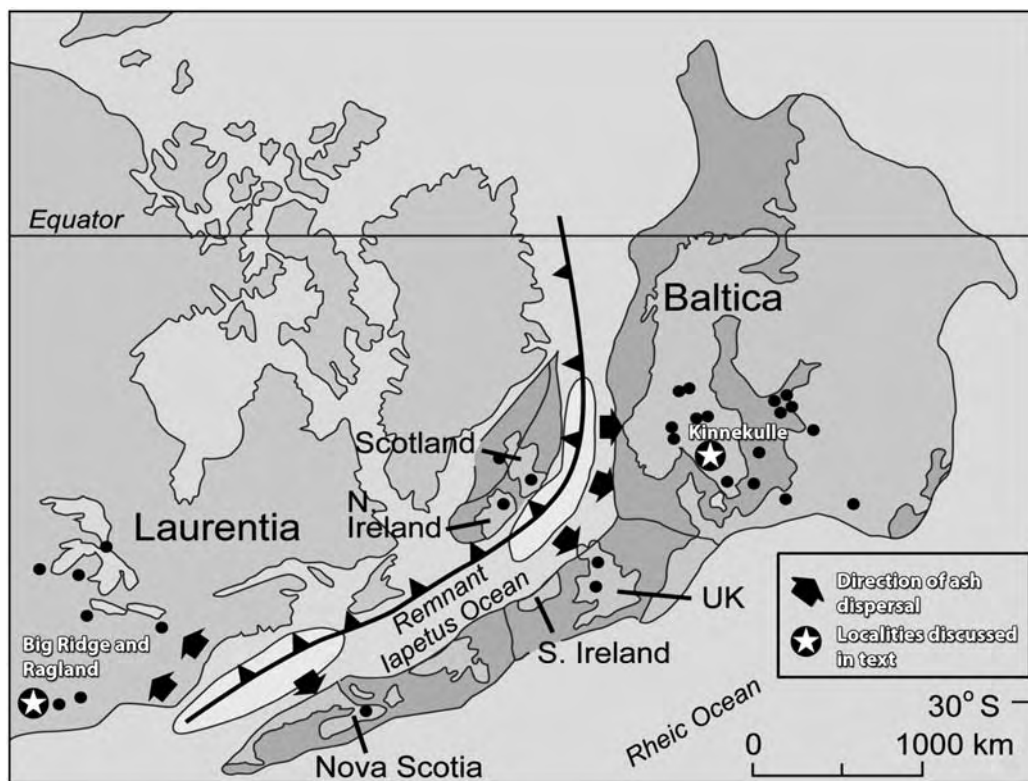


Figure 1. Location and paleogeographic setting of the Big Ridge and Ragland sections (Alabama) and the Kinnekulle section (Sweden) during the Ordovician.

Each of the four K-bentonites has one or more tuffaceous zones containing macroscopic primary phenocrysts including some or all of Qtz + Bt + Pl + Ilm + Kfs + Zrn + Ap + Hbl, with the most common being quartz, plagioclase (especially zoned and twinned andesine and labradorite), ilmenite, and biotite (Haynes, 1994; Haynes et al., 1995), some of which encapsulate melt inclusions (Verhoeckx-Briggs et al., 2001; Mitchell et al., 2004). The petrologic significance of biotite in igneous rocks is as the most common

mineralogic sink for excess alumina, and as an important sink for iron, magnesium, and titanium (Clarke, 1981). Petrogenetically, the remnant primary phenocrysts indicate that each of the four beds was generated by explosive volcanism, and the areal extent of the Deicke, Millbrig, and Kinnekulle indicates that these tephra may have been generated by the largest volcanic eruptions of the Phanerozoic (Kolata et al., 1996). Magmatic compositions can be inferred to a degree from comparison with Cenozoic lavas that have compositionally similar phenocrysts, and this information can constrain tectonomagmatic setting.

The problem of determining tectonic setting of an altered tephra

Although prior work suggests that these Ordovician tephra came from a fractionated magma that included assimilated continental crust (Samson et al., 1989, 1995), the actual setting of the volcanoes is unknown. They may have been part of a continental arc on continental crust above a subduction zone, as occurs in parts of Indonesia (Honthaasa et al., 1999), or part of an intraplate volcanic center, but because the source volcanoes no longer exist, their nature and setting cannot be directly determined. K_2O content of lavas is one indicator of tectonic setting (Le Bas et al., 1986), and the presence or absence of abundant biotite phenocrysts is an indicator of magma series and tectonomagmatic setting (Izett, 1981). Ewart and Le Maitre (1980) found that volcanic rocks with pyroxene and/or olivine (i.e., the most abundant non- K_2O -bearing phenocrysts) are approximately four times as abundant as those rocks with hornblende and/or biotite, and they found that the frequency of biotite occurrence peaks in dacitic rocks (those with 63-69 % SiO_2), with abundant biotite in many samples of lower (andesites) and higher (rhyolites) SiO_2 lavas. From these findings, we hypothesized that the Ordovician tephra were generated by eruption of hydrous metaluminous to peraluminous magmas of an evolved, continental character, a conclusion reached independently by other researchers on the basis of trace element and isotopic analyses (Samson et al., 1989, 1995; Kolata et al., 1996).

Standard petrologic methods for investigations of unaltered volcanic rocks commonly include a determination of pre-eruptive intensive variables (e.g. temperature, fugacity of water, oxygen, and sulfur, and the water content of the magma) through investigation of relationships between crystals and co-existing liquids via chemical analysis of phenocrystic, groundmass, and whole rock samples. Even though this comparative process can be quite involved when the phenocryst of interest is biotite (Conrad et al., 1988; Puziewicz and Johannes, 1990), much information about magmatic conditions can be acquired via this approach. With the Ordovician tephra, however, such an approach is greatly complicated by the devitrification and illitization of the groundmass, and by the moderate to significant alteration of many of the phenocrysts. Thus, direct compositional comparison between biotite phenocrysts and the host-rock or groundmass (e.g. Jezek, 1976; Jezek and Hutchison, 1978; De Pieri et al., 1978; Clemens and Wall, 1984; Boden, 1994) is not an option in these devitrified Ordovician tephra. Instead, compositional analysis of unaltered phenocrystic minerals (Samson et al., 1989; Min et al., 2001) and melt inclusions therein (Delano et al., 1990, 1994; Verhoeckx-Briggs et al., 2001) is the best means for investigating the petrogenesis of these altered tephra.

METHODS AND PROCEDURES

We generated a biotite database by a literature review and compilation of 420 biotite analyses and by carrying out 1038 electron microprobe analyses of biotite from 14 ignimbrites, tuffs, and pumices of Tertiary

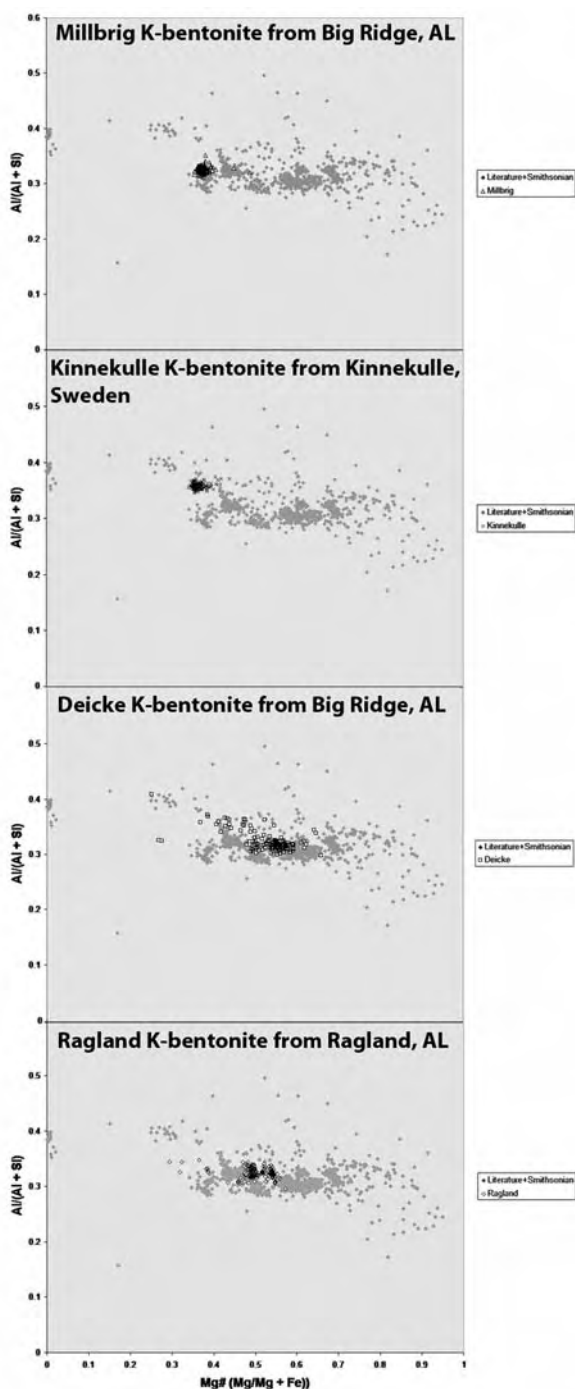
and Quaternary age in the Smithsonian's Petrology collection. These supplement our 450 microprobe analyses of Kinnekulle K-bentonite biotites, 120 analyses of from the Ragland K-bentonite biotites, 300 analyses of Deicke K-bentonite biotites, and 450 analyses of Millbrig K-bentonite biotites. Biotite grains were mounted and polished on standard petrographic slides and carbon-coated following the methods of Haynes (1994) and Haynes et al. (1995, 1996). A hornblende standard (USNM 143965, Kakanui hornblende of Jarosewich et al., 1980) was analyzed repeatedly to monitor precision during all analytical runs.

RESULTS

Comparisons were made using atomic ratios, specifically $Al/(Al + Si)$ vs. $Mg/(Mg + Fe)$ (the Mg number, an indicator of changing crystallization conditions in magmas) (Fig. 2) and Ti vs. $Al/(Al + Si)$ (Fig. 3). Use of these ratios provides information about the distribution and variability of biotite compositions across the spectrum of volcanic rocks. The compositional variability of magmatic biotites has fundamental limitations placed on it by the composition of the host magma (Abdel-Rahman, 1994; Stussi and Cuney, 1996; Righter et al., 2002), and the magmatic composition is itself governed to a certain extent by tectonic setting.

In Figure 2, the Millbrig biotites match best with biotites in the Toba Tuff rhyolites and rhyodacites of Sumatra, the Obsidian Dome rhyolites at Inyo Craters, California, the cordierite-bearing dacites ("ambonites")

Figure 2. Variation in Si, Al, Mg, and Fe in biotites from the four Ordovician K-bentonites compared with analyses of biotites from Cenozoic lavas in the literature and from the Smithsonian Petrology collection.



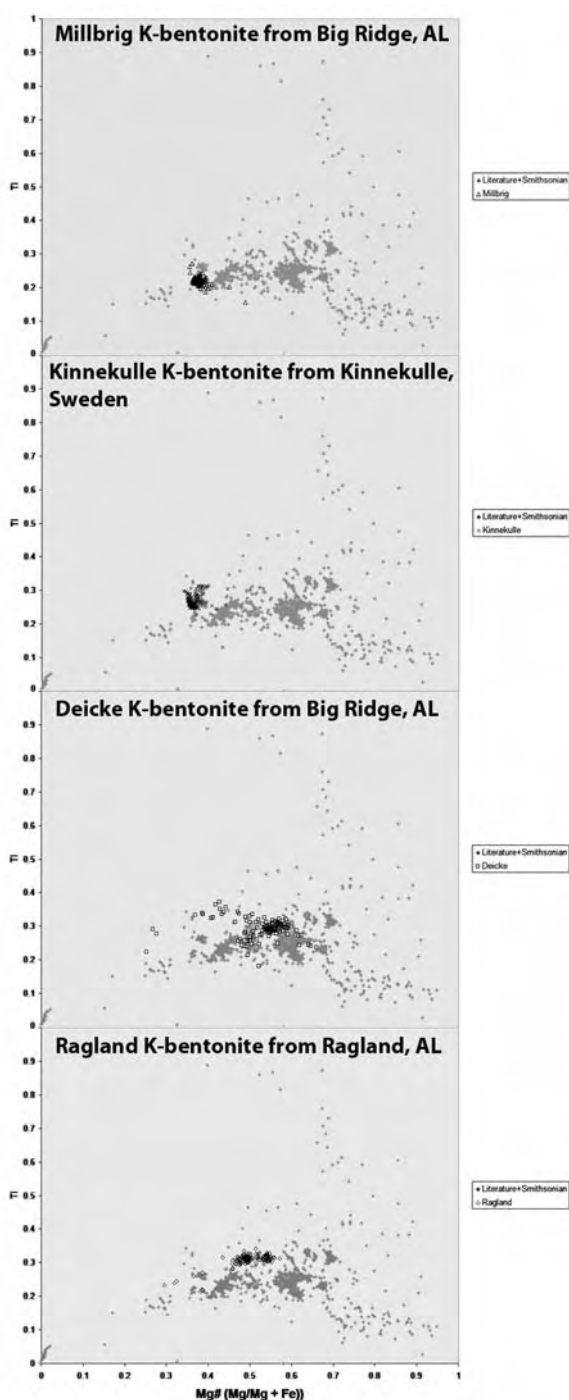
of Ambon, Indonesia, and the Bishop Tuff of California. The Kinnekulle biotites match best with biotites of the Toba Tuff. The Deicke biotites match best with Yellowstone post caldera rhyolites from Wyoming, the felsic ignimbrites of the La Pacana caldera, Chile, and rhyolites of the Cerro Chascun complex, Bolivia. The Ragland biotites match best with the Toquima caldera high-K rhyodacites from Nevada.

In Figure 3, the Millbrig biotites match best with the Whakamaru ignimbrite from the Taupo volcanic zone of New Zealand, the Valles Caldera rhyodacites of New Mexico, the Macusani ignimbrites of Peru, and the Toba Tuff. The Kinnekulle biotites match best with the Bishop Tuff, the Obsidian Dome rhyolites, and the Toba Tuff. The Deicke biotites match best with the Mezitler area dacites of Turkey, the La Pacana ignimbrites, the Cerro Chascun rhyolites, and the Bishop Tuff. The Ragland biotites match best with the Bishop Tuff.

CONCLUSIONS

A comparison of elemental variation in biotites from Ordovician K-bentonites with biotites from various Cenozoic volcanics suggests that the Ordovician K-bentonites are almost certainly (as has long been suspected) the product of explosive volcanism, with the source magmas having passed through continental crust on their way to eruption. Permissible volcanic models favored by these results would be volcanism that results in very large single vent eruptions and/or post-eruption caldera formation, and which is associated with subduction along continental

Figure 3. Variation in Ti, Mg, and Fe in biotites from the four Ordovician K-bentonites compared with analyses of biotites from Cenozoic lavas in the literature and from the Smithsonian Petrology collection.



margins (e.g. Bolivia and Chile), subduction beneath very large islands or microplates (e.g. Sumatra), or intraplate hot spots (e.g. Yellowstone). Cenozoic volcanic systems that are likely analogs for the Ordovician tephra include the Toba system of Sumatra, the Long Valley system of California, the La Pacana system of Chile, the Yellowstone system of Wyoming, the Banda Arc volcanics of Ambon, and the Toquima complex of Nevada, all caldera complexes on continental crust.

Although continental and intraplate hotspot settings cannot be ruled out, paleogeographic reconstructions as presently understood make it more likely that the Iapetan magmatic arc(s) that existed during the mid-Ordovician between Laurentia and Baltica-Scandia (Fig. 1) was located on a large Sumatra-like island or Ambon type microplate that had a basement of continental crust, but was located above a subduction zone.

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MIDDLE DARRIWILIAN CONODONT BIOSTRATIGRAPHY IN THE ARGENTINE PRECORDILLERA

S. Heredia and A. Mestre

CONICET- Universidad Nacional de San Juan, Instituto de Investigaciones Mineras,
Av. Libertador y Urquiza, 5400 San Juan, Argentina.
sheredia@unsj.edu.ar, amestre@unsj.edu.ar

Keywords: Conodonts, Ordovician, Darriwilian, biostratigraphy, Precordillera, Argentina.

INTRODUCTION

The Lower-Middle Ordovician carbonate succession of the Precordillera is developed along a meridional length of 400 km with a latitudinal width of 150 km in western Argentina. Several localities at the Precordillera are well studied but the Las Chacritas river (LCHA) section, and Cerro La Chilca (LCHI) section are considered here as the most complete and well exposed for detailed analysis of the Middle Ordovician conodont biostratigraphy (Fig. 1).

Albanesi and Astini (1994) reported conodonts of the *Eoplacognathus suecicus* Zone at the top of the San Juan Formation in the LCHA section, and Lehnert (1995) identified also the *E. suecicus* and *Pygodus serra* zones from the uppermost levels of the San Juan Formation and the “Transfacies” (in the sense of Baldis and Beresi, 1981). The occurrence of the *Lenodus variabilis* Zone in the carbonate succession was first mentioned by Peralta et al. (1999a) and was documented by Peralta et al. (1999b). Albanesi and Astini (2000) reported the occurrence of the *Eoplacognathus pseudoplanus* Zone in the LCHA section. Heredia et al. (2005) documented the distribution of the conodont taxa and analyzed the relationship between lithostratigraphy and biostratigraphy within the LCHA section.

Lehnert (1995) mentioned the first conodont fauna from LCHI, and Mestre (2010) defined the *E. pseudoplanus* Zone for the last meter of the San Juan Formation in this section.

Our stratigraphical and biostratigraphical study focuses on the upper part of the San Juan Formation and the lower member of Las Aguaditas (LCHA section) and Los Azules (LCHI section) formations (Fig. 1). In this contribution we propose the use of the Darriwilian Baltic conodont chart in the Argentine Precordillera due to the reliable stratigraphical distribution of Baltic conodonts.

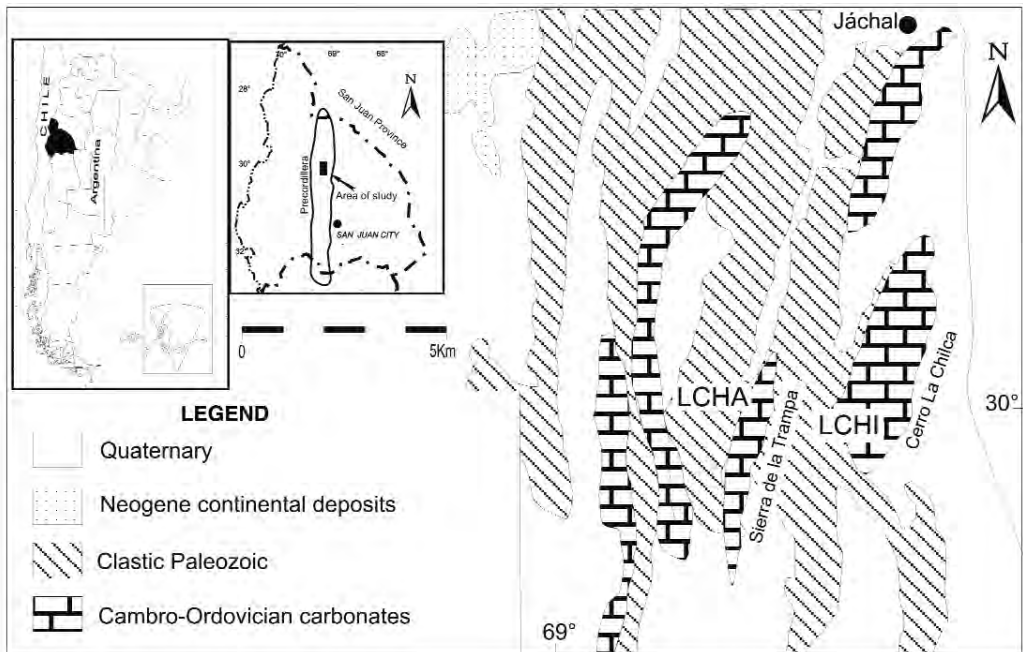


Figure 1. Location map of studied sections: La Chilca section (LChi) and Las Chacritas section (LCha).

GEOLOGIC SETTING AND STRATIGRAPHY

The Ordovician carbonates exposed in the LCHA and LCHI sections are composed of grey to dark grey limestone, marls and mixed carbonate/siliciclastic sediments deposited in a ramp setting (Peralta and Baldis, 1995; Carrera, 1997; Mestre, 2010). Each section begins with the Lower–Middle Ordovician San Juan Formation, composed mainly of fossiliferous limestone and marly limestone. Its base is not exposed because of faulting but the preserved part is 325 m thick in the LCHI section and 270 m on the LCHA section. The San Juan Formation is conformably overlain by 55 m of thin- to medium-bedded marly limestone and black shale of the Las Aguaditas/Los Azules Formation of the Middle to Late Ordovician age. The latter units consist of tabular, thin- to medium-bedded, dark mudstone, nodular fossiliferous wackestone to packstone, black shale and rare thin beds of bentonite. The contact between the San Juan and Las Aguaditas/Los Azules formations is transitional; the first level of black shale being used as the arbitrary boundary between these units (Fig. 2).

METHODOLOGY

Conodont samples were collected from limestone beds at random intervals (3 m to 1 m); however, the collecting interval was about 10 to 15 cm towards the top of the upper part of the San Juan Formation at the Las Chacritas river and La Chilca sections (Fig. 2). Almost 10,100 identifiable conodont elements were recovered from both sections. All elements have a color alteration index of 2–3 (60–200 °C) (Epstein et

al., 1977). The conodonts are housed in the collection of the INGEO at Universidad Nacional de San Juan, under the code-MP and in the collection of the INSUGEO at the Instituto Miguel Lillo de Tucumán, under the code ML-C.

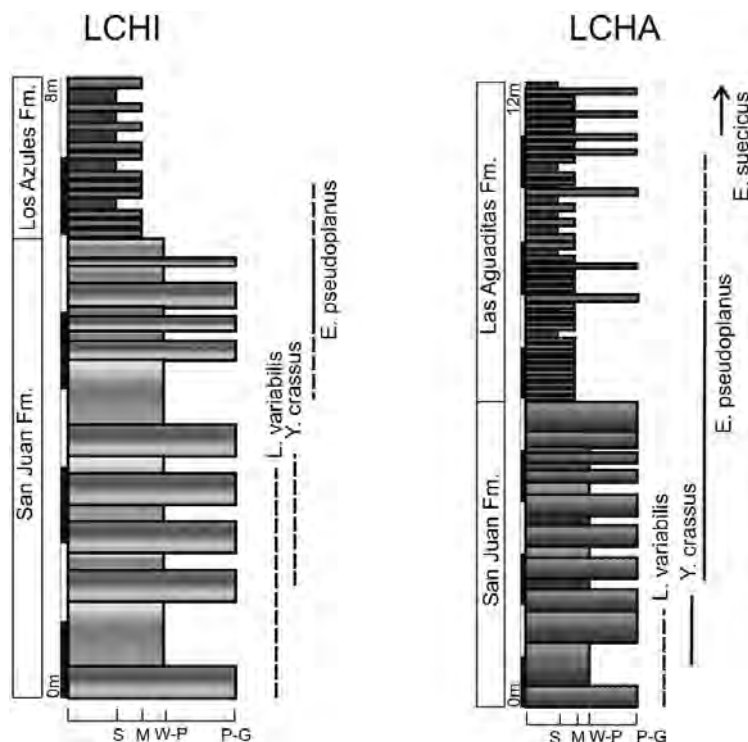


Figure 2. Stratigraphic column of La Chilca (LCHI) and Las Chacritas (LCHA) sections showing vertical distribution of Baltic index conodonts.

CONODONTS

The main purpose of this contribution is focused in the vertical distribution of the *Lenodus variabilis* (Sergeeva), *Yangtzeplacognathus crassus* Zhang, *Eoplacognathus pseudoplanus* (Viira) and *Eoplacognathus suecicus* Bergström in the studied section in the Precordillera, however results of great interest to mentioning the middle Darriwilian conodont fauna associated to these conodonts on the LCHA and LCHI sections: *Ansella jemtlandica* Löfgren, *Baltoniodus medius* Dzik, *Bryantodina* aff. *typicalis* (Stauffer), *Drepanodus gracilis* Branson and Mehl, *Drepanoistodus basiovalis* (Sergeeva), *Drepanoistodus bellburnensis* Stouge, *Drepanoistodus pitjanti* Cooper, *Dzikodus hunanensis* Zhang, *Dzikodus tablepointensis* Stouge, *Erraticodon balticus* (Dzik), *Fahraeusodus marathonsensis* (Bradshaw), *Histiodela kristinae* Stouge, *Histiodela holodentata* Ethington and Clark, *Microzarkodina ozarkodella* Lindström, *Paltodus? jemtlandicus* Löfgren, *Parapaltodus simplissimus* Stouge, *Paroistodus horridus* Barnes and Poplawski, *Periodon aculeatus zgierzensis* (Dzik), *Polonodus clivosus* Viira, *Polonodus galerus* Albanesi,

Polonodus magnum Albanesi, *Protopanderodus calceatus* Bagnoli and Stouge, *Protopanderodus gradatus* Serpagli, *Protopanderodus graeai* (Hamar), *Pygodus anitae* Bergström, *Rossodus barnesi* Albanesi, *Scolopodus oldstockensis* Stouge, and *Spinodus spinatus* (Hadding).

DISCUSSION

The middle Darriwilian conodont fauna of the LCHA and LCHI sections is very similar at species level to correlative faunas of the Baltic and South China regions (Heredia et al., 2005; Mestre, 2010). Nevertheless, the Darriwilian conodont zonation of the Precordillera is not the same as correlations in Baltica and South China (Bagnoli and Stouge, 1996; Albanesi and Ortega, 2002). Albanesi et al. (1998) and Albanesi and Ortega (2002) proposed that the middle Darriwilian of the Argentine Precordillera is subdivided into two conodont zones and four subzones: the *L. variabilis* Zone, composed by the lower *Periodon gladysi* Subzone and the upper *Paroistodus horridus* Subzone, and the *E. suecicus* Zone comprising the lower *Histiodella kristinae* Subzone and the upper *Pygodus anitae* Subzone (Fig. 3).

In Baltica, the middle Darriwilian conodont zonation includes four successive zones (Löfgren, 2000, 2004; Löfgren and Zhang, 2003): *Lenodus variabilis*, *Yangtzeplacognathus crassus*, *Eoplacognathus pseudoplanus* (*M. hagetiana* and *M. ozarkodella* Subzones) and *Eoplacognathus suecicus* (*P. lunnensis* and *P. anitae* Subzones) (Fig. 3). Zhang (1998) erected the *Dzikodus tablepointensis* Zone in South China, divided into the *M. hagetiana* and *M. ozarkodella* Subzones (Fig. 3). This zone and its constituent subzones span almost the same interval as the *E. pseudoplanus* Zone.

Sistem	Series			Stages			Baltic Conodonts zones & subzones		Conod. south China Zones & Subzones	Conodonts Precordillera	Conodonts Precordillera			
	Global	Britain	N.Amer.	Global	Austr.	China	Löfgren (1978)	Zhang(1997,1998) Bagnoli & Stouge (1997)	Zhang(1998)	Albanesi & Ortega(2002)	This study			
Ordovician	Middle	Llanvirnian	Whiterockian	Darriwilian	Darriwilian	Zhejiangian			<i>Pygodus anserinus</i>	<i>Y. jianyeensis-P. anserinus</i>	<i>Pygodus anserinus</i>			
							<i>P. serra</i>	<i>li.</i>	<i>P. serra</i>	<i>li.</i>	<i>Y. protoramosus</i>		<i>P. serra</i>	<i>li.</i>
								<i>ro.</i>		<i>ro.</i>				<i>ro.</i>
								<i>re.</i>		<i>re.</i>				<i>re.</i>
								<i>fo.</i>		<i>fo.</i>				<i>fo.</i>
							<i>E. suecicus</i>	<i>E. suecicus</i>	<i>E. suecicus</i>	<i>P. anitae</i>	<i>E. suecicus</i>		<i>P. anitae</i>	<i>H. kristinae</i>
	<i>P. sulcatus</i>	<i>P. lunnensis</i>	<i>?</i>											
	<i>E. suecicus</i>	<i>M. ozarkodella</i>	<i>M. ozarkodella</i>											
	Arenig.						<i>E. ? variabilis</i>	<i>E. ?va.-M. ozarkodella</i>	<i>Dzikodus tablepointensis</i>	<i>M. hagetiana</i>	<i>L. variabilis</i>	<i>P. horridus</i>	<i>Eoplacognathus pseudoplanus</i>	
								<i>E. ? variabilis</i>	<i>M. hagetiana</i>	<i>M. hagetiana</i>		<i>Y. crassus</i>	<i>Y. crassus</i>	<i>Yangtzeplacognathus crassus</i>
								<i>M. flabellum</i>	<i>Y. crassus</i>	<i>Lenodus variabilis</i>		<i>Lenodus variabilis</i>	<i>P. gladysi</i>	<i>Lenodus variabilis</i>

Figure 3. Global lower-middle Darriwilian conodont biostratigraphic chart comparing Baltica, South central China and the Argentine Precordillera.

The record of *L. variabilis*, *Y. crassus*, *E. pseudoplanus* and *E. suecicus* in the LCHA and LCHI sections provides a much better knowledge of the conodont faunas in the Precordillera, and is of great interest due to the signification on conodont provincialism of the area. The Ordovician biostratigraphic chart of the Precordillera turns out as very similar to the Baltic chart (Figs. 2, 3).

CONCLUSIONS

New data presented in this study allows a modification of the Middle Ordovician conodont biostratigraphic chart of the Argentine Precordillera. Darriwilian Baltic conodonts occur in this region as important components of the conodont association in the *L. variabilis*, *Y. crassus*, *E. pseudoplanus* and *E. suecicus* Zones. We thus suggest their biostratigraphical use in the Middle Ordovician chart of the Argentine Precordillera (Fig. 3).

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CONVENTIONAL AND CONOP9 APPROACHES TO BIODIVERSITY OF BALTIC ORDOVICIAN CHITINOZOANS

O. Hints, J. Nõlvak, L. Paluveer and M. Tammekänd

Institute of Geology at Tallinn University of Technology, Ehitajate tee 5, 19086 Tallinn, Estonia.
Olle.Hints@gi.ee

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INTRODUCTION

Chitinozoans are organic-walled microfossils, probably eggs of cryptic marine metazoans that were common and diverse from the Early Ordovician through Devonian times (Grahns and Paris, 2011). Chitinozoans have proved to be among the most useful index fossils for this time span (e.g., Nõlvak and Grahns, 1993; Webby et al., 2004) and their diversification history has been discussed by several authors (e.g., Paris and Nõlvak, 1999; Paris et al., 2004; Grahns and Paris, 2011).

With respect to Ordovician chitinozoans, the Baltic region stands out with excellent preservation, good stratigraphical coverage and some of the largest collections in the world (Paris et al., 2004). The first diversity curves of Baltic Ordovician chitinozoans were published by Kaljo et al. (1996), studying material from a single drill core. Based on the entire Baltic collection, the diversity patterns of Ordovician chitinozoans were summarised by Nõlvak in Paris et al. (2004). Using the latter Baltic curve, alongside with those from other regions, Achab and Paris (2007) argued about possible driving mechanisms behind the chitinozoan diversification, highlighting climatic, paleogeographic and paleo-oceanographic factors. More recently the diversity of Baltic Ordovician chitinozoans was discussed by Hints et al. (2010).

All these paleobiodiversity studies have been based on a temporal framework of regional stages, subdivisions thereof, or time slices such as those defined by Webby et al. (2004). Diversity curves using different time scales and data sets are often difficult to compare. A time scale that is too coarse, may also obscure the details of biodiversity patterns and the underlying environmental, climatic and paleogeographic signals.

In order to increase stratigraphical resolution of the hitherto available chitinozoan biodiversity curve and get more reliable estimation of the standing diversity in the Baltic Ordovician, we herein use quantitative stratigraphic approach based on CONOP9 software (Sadler and Cooper, 2003 and references therein). This tool has proved very efficient in reconstructing successions of biostratigraphical events for a large number of taxa and sections. The resulting best fit composite sequence can be used both as a time-scale, and as a basis for biodiversity curves. We aim to compare the CONOP9-derived results with those produced by a more conventional stage-based approach. As a lot of new material on Ordovician

chitinozoans has emerged from Estonia since the compilation of the data base for the IGCP410 compendium (Paris et al., 2004), we also aim to improve the previously published diversity curves.

MATERIALS AND METHODS

This study is based on collections of Mid to Late Ordovician chitinozoans from nine localities in Estonia (Fig. 1): the Kerguta, Männamaa, Mehikoorma, Ruhnu, Taga-Roostoja, Tartu, Valga and Viki drill cores (see Nõlvak 2010 and references therein), and the Uuga cliff section (see Tammekänd et al., 2010). These localities represent near-shore to deeper shelf carbonate facies of the eastern part of the Baltoscandian Palaeobasin (Fig. 1).

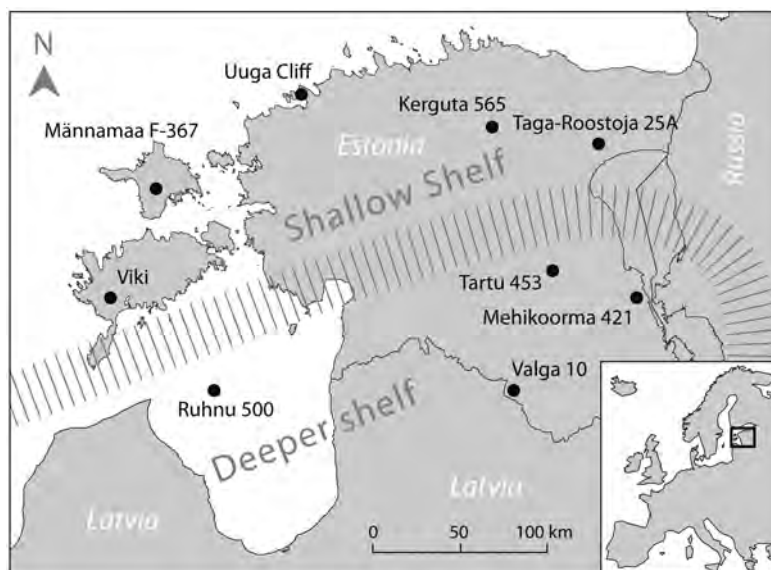


Figure 1. Locality map and broad scale facies patterns in the eastern Baltic.

The chitinozoans were extracted from limestone and marl samples, usually 100-500 g in size (depending on average yield), using digestion in acetic acid. Altogether, the data set consists of 1079 productive samples and 8565 occurrence records of 166 taxa, of which 145 species were included in the analysis (the other 21 being only genus-level or doubtful identifications). Species currently under open nomenclature were included in the data set. The vast majority of hitherto known Baltoscandian chitinozoan species were identified in the localities studied. Moreover, as the differences between chitinozoan faunas of Estonia, Latvia, Lithuania, Poland and Sweden are small, the current data set can be considered representative for the entire region. The Lower Ordovician, where the first chitinozoans are recorded, is not included in the current analysis, and lower Middle Ordovician and uppermost Ordovician are less completely covered, leaving possibilities for future improvement of the data set.

The general stratigraphical framework is based on Baltic regional stages (Nõlvak et al., 2006) with reference to time slices of Webby et al. (2004). Usage of diversity measures follows Cooper (2004). Total diversity (TD) is the number of species recorded from a time interval. Normalised diversity (ND) is the sum

of species that range from the interval below and above, plus half the number of species that appear and/or disappear within the time interval. Additionally the balanced total diversity (BTD) of Paris et al. (2004) is discussed, which is similar to ND, except that a full score is given to species that are confined to the time slice. All these measures are used to estimate the mean standing diversity (MSD).

For the CONOP9 analysis, as well as for the stage-based approach, all occurrences of all species were entered into Excel spreadsheets and then transferred to a custom-built SQL database. The database records were carefully checked for taxonomic inconsistencies and other potential errors. In order to enhance the CONOP9 composite sequence, the Kinnekulle K-bentonite at the base of the Keila Regional Stage was included where present. From the database, the data files in CONOP9 format were generated automatically, then sorted using the CONSORT utility and analysed with CONOP9 program (version 7.61 of July 5, 2009, courtesy of P. Sadler). Tests were run with different configuration options; rather consistent results were produced with 1500 steps and 500 trials using level penalty. The diversity curve is derived from the running FADs minus LADs along the composite sequence. For full explanation and examples of using CONOP9 software see Sadler and Cooper (2003).

RESULTS

Conventional approach

The results of the conventional stage-based approach to the diversity of Baltic Ordovician chitinozoans are illustrated in Fig. 2. The curves of TD and ND generally run parallel, the TD showing on average 7 more species per stage, and the BTD running in between. The highest diversity increase is observed in the Kunda and Aseri stages, lower to middle Darriwilian, where 31 and 36 species are recorded, respectively. It should be added, however, that the Volkhov Stage (most of the Dapingian) is poorly represented in the sections studied due to dolomitisation or redbeds and thus the increase from Volkhov to Kunda may, in fact, be more gradual. The increase continues in the Lasnamägi Stage, where the peak TD value of 41 and ND of 33.5 are recorded. Slightly lower diversity is observed in the Uhaku Stage, followed by a diversity peak in the lowermost Upper Ordovician Kukruse-Haljala interval, where a TD of 46 and a ND of 34 are recorded — the highest values for the Baltic Ordovician. It is worth noting that in this interval the discrepancy between TD, ND and BTD is the highest and TD shows increasing trend whilst ND and BTD reflect a slight diversity drop from Kukruse to Haljala. A significant decline begins in the Keila Stage, close to the Sandbian-Katian boundary, where 37 species are recorded (ND 27.5). A decreasing trend is characteristic of the rest of the Ordovician, with two minor positive shifts in the TD curve in the Nabala and Pirgu stages. The ND curve is slightly different with a low in the Oandu Stage and a minor peak in the Rakvere-Nabala interval (Fig. 2). The Hirnantian extinction is marked by the decrease of TD by 13 species, corresponding to 50% loss, from the Pirgu to Porkuni Stage. Only a low diversity assemblage crossed the Ordovician-Silurian boundary. Few data available from the lowermost Silurian reveal a very low diversity chitinozoan fauna, which is in agreement with Nestor (2009).

CONOP9 model

The diversity curve based on CONOP9 composite sequence has much higher resolving power and probably can be considered as the best achievable approximation of the standing diversity (SD), without

the usual binning problems of the conventional approach. It should also be noted that the CONOP9 curve is independent of the stage-based time scale and dating problems. This, in turn, means that the two curves presented on Fig. 2 cannot always be correlated precisely. With different model runs the maximum species richness estimate was between 35 and 37. Small fluctuations of 1-2 species, appearing at different levels, represent methodological uncertainty rather than true events.

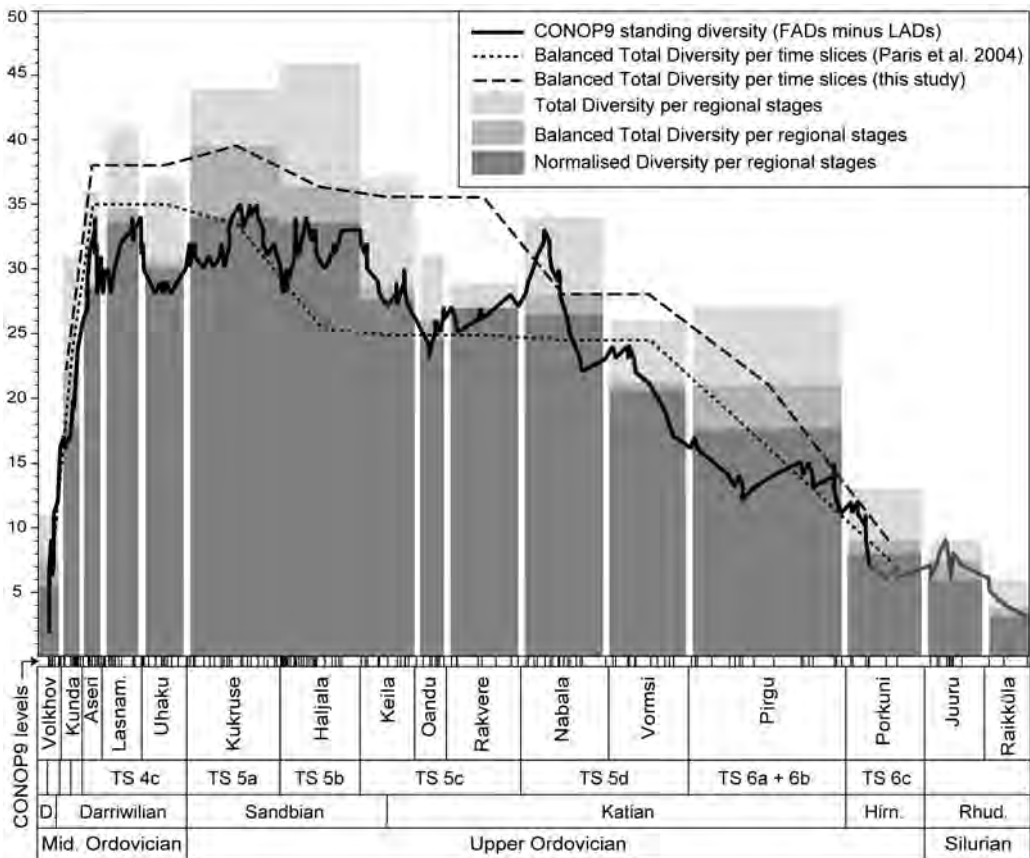


Figure 2. Diversity of Baltic Ordovician chitinozoans as revealed by the conventional stage-based approaches (TD, ND and LTD) and CONOP9 model. The CONOP9 composite sequence was fitted to regional time scale using ranges of selected chitinozoan species. White areas between the "stage bars" indicate that precise calibration of the CONOP9 composite against the stages was not possible. Two LTD curves based on the time slices of Webby (2004) are included: one using the current data set and the other redrawn from Nölvak in Paris et al. (2004). The horizontal scale corresponds to the average composite sequence of the CONOP9 model reflecting thickness rather than a regular time scale. Abbreviations: TS, time slice; D., Dapingian, Hirn., Hirnantian; Rhud., Rhuddanian.

A rapid diversity increase occurs in the Volkhov to Aseri stages, where the values reach to about 34 (Fig. 2). A slight decline is observed close to the Aseri-Lasnamägi boundary, possibly resulting from insufficient data. This is followed by a high diversity interval and another decline, both within the *Conochitina clavaherculi* range, which corresponds to the upper Lasnamägi and lower Uhaku strata. The Uhaku Stage is characterised by a relatively lower diversity, below 30 species, followed by an increasing

trend in the succeeding Kukruse Stage, where the maximum of about 35 species is met. In the upper Kukruse and lower Haljala strata another diversity low with less than 30 species is observed. However, the Haljala Stage, in general, has a rather high diversity. The lower boundary of the Keila Stage is precisely dated based on the widespread Kinnekulle K-bentonite. Starting from this level, the diversity starts to decline, and only about 25 co-existing species are recorded in the Oandu stage. The Rakvere Stage is characterised by a slightly increasing trend and a rather conspicuous diversity peak is recorded in the Nabala Stage, coinciding with the lower part of *Armoricochitina reticulifera* range. Subsequently further lowering of the diversity is observed, falling below 20 in the Vormsi, below 15 in the Pirgu, and below 10 in the Porkuni Stage. In the topmost Ordovician and lower Silurian the model is not well-constrained due to too few overlapping species and insufficient number of sections studied.

DISCUSSION AND CONCLUSIONS

The CONOP9 modeled diversity curve reflects the same general trends as the conventional stage-based approach (Fig. 2), but reveals also some differences and some features that were not evident in the latter.

From the methodological point of view it should be noted that the CONOP9 curve runs closest to the ND curve of the stage-based approach. The TD, on the other hand, clearly overestimates the MSD in most cases, as shown also by Cooper (2004). Similarly, the balanced total diversity (BTD) of Paris et al. (2004) tends to overestimate MSD, especially in longer time slices (Fig. 2).

A rapid diversification of chitinozoans from the Volkhov to Aseri stages is unveiled by both approaches. According to Achab and Paris (2007), a similar radiation event is recorded on other paleocontinents, probably driven by intrinsic factors, as suggested by the great number of morphological innovations that appeared during the Darriwilian.

The following biodiversity pattern appears slightly differently in conventional and CONOP9 curves. The stage-based approach suggests that the chitinozoan fauna reached the highest diversity during the Kukruse-Haljala interval. The CONOP9 curve, on the other hand, shows that rather similar maximum values were characteristic to the entire Aseri-Haljala interval. According to estimations by Kaljo et al. (1996) and Webby et al. (2004), the Aseri, Lasnamägi and Uhaku stages are notably shorter in duration than Kukruse and Haljala stages. Thus, the TD and ND peaks in the latter stages may merely represent "binning bias" and the CONOP9 curve likely provides more appropriate MSD estimation here. Several small scale diversity fluctuations revealed in the CONOP9 curve need further examination. However, the diversity low in the Uhaku Stage is documented also by the stage-based approach as well as by Kaljo et al. (1996).

A general diversity decline established by both approaches starts in the Keila Stage, at the Sandbian-Katian boundary. This interval coincides with the beginning of changes in regional environmental settings evidenced by first tropical carbonates and reefs, increased facies differentiation, increased variation in carbon isotope composition, and a general biotic change (Kaljo et al., 2011). The chitinozoan diversity was particularly low in the Oandu and Rakvere stages (the "Oandu crisis" according to Kaljo et al., 1996). Here it is important to stress good correspondence between the CONOP9-modeled and stage-based curves indicating that certain correlation problems have not affected per-stage calculations (but note that occurrences with ambiguous stratigraphy were omitted from the stage-based curves).

Following the Oandu crisis, a conspicuous peak in the CONOP9 curve, reaching 33 species in the lower Nabala Stage (Fig. 2), deserves further attention. In the binned TD curve this peak is less prominent, and entirely absent in the ND curve (Fig. 2). This short-lived diversity peak on the generally falling Katian trend

is probably related to temporarily improved environmental conditions for chitinozoans. Such interpretation is supported by the elevated concentration of phosphorus in the lower part of the Nabala Stage (Kiipli et al., 2010), which might have had positive effect on bioproduction and food supply for chitinozoans. The deeper shelf Mõntu Formation (lower Nabala Stage) is also rich in glauconite, which, together with elevated phosphorus concentration, may imply a regional upwelling event. An upwelling of presumably colder water masses might have had positive effect also through dropped water temperatures on the shelf – it has been shown by Vandenbroucke et al. (2010) that chitinozoans seem to thrive in high latitude (i.e. colder) regions.

Subsequently, the chitinozoan diversity continued to decline, with small positive peaks in the Vormsi and upper Pirgu stages revealed by the CONOP9 approach. By the Hirnantian (Porkuni Stage), the chitinozoan fauna was already strongly impoverished in Baltica and few species continued into the Silurian. According to Nestor (2009), the diversity of Silurian chitinozoans started to increase only in the late Aeronian.

The previous analyses of biodiversity of Baltic Ordovician chitinozoans are limited in number. As discussed above, the currently revealed trends fit well with those of Kaljo et al. (1996), even though the latter authors reported lower total numbers (maximum TD value of 29). The BTD data of Nõlvak in Paris et al. (2004; reproduced by Achab and Paris, 2007 and Hints et al., 2010), show a different pattern, which is partly due to the use of longer time bins. In order to facilitate comparison, the current data were also recalculated into time slices of Webby et al. (2004). The resulting curve (Fig. 2) still shows some differences from Nõlvak in Paris et al. (2004) curve, particularly in time slices 5b and 5c, where notably higher diversity is now recorded. To some extent this discrepancy could be explained by improved data and inclusion of open nomenclature species. Nonetheless, the diversity decline from the Kukruse (TS 5a) to Haljala (TS 5b) indicated by Nõlvak in Paris et al. (2004) seems to gain little support from the current data set – the new TS-based curve shows a major decrease in the Nabala Stage.

It should be stressed, however, that both curves based on time slices fail to resolve the Keila decline, the "Oandu crisis" and the peak in the Nabala Stage, which are prominent on the CONOP9 curve and evident on regional time scale. It follows that a global stratigraphic framework, such as that of Webby et al. (2004), is too generic to document at least regionally important bioevents. Although the new TS-based BTD curve is more accurate than the one discussed by previous authors (Paris et al. 2004, Achab and Paris 2007, Hints et al. 2010), higher temporal resolution is needed to reveal timing and driving factors of the diversification process.

In summary we conclude that the presented data set, albeit only from nine sections, is currently the best coherent data source for assessing biodiversity of Baltic late Mid to Late Ordovician chitinozoans. The CONOP9 model proved to fit well with the empirical data on chitinozoan distribution. The resulting composite sequence provides a valuable addition to the conventional paleobiodiversity approach and represents probably the best possible proxy for standing diversity.

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ORDOVICIAN ROCKS IN JAPAN

Y. Isozaki

Department of Earth Science & Astronomy, The University of Tokyo
Komaba, Meguro Tokyo 153-8902, Japan. isozaki@ea.c.u-tokyo.ac.jp

The Japanese Islands have ca. 700 million year history that started at the breakup of the Rodinia supercontinent (Isozaki et al., 2010, 2011). Proto-Japan originally formed a passive margin along South China (Yangtze) block detached from Rodinia, and was later converted tectonically into an active continental margin around ca. 520 Ma (Middle Cambrian) as marked by the oldest arc-granite formation in Japan. To date, the oldest non-metamorphosed sedimentary unit in Japan is represented by the late Early to Middle Ordovician fore-arc deposits, whereas the oldest high-P/T metamorphic rocks have Early Ordovician ages up to 480 Ma.

Although the Ordovician rocks in Japan are vital pieces of information concerning the tectono-sedimentary history of Japan and East Asia, their distribution is highly limited and their sizes are extremely small (some hundreds meter thick). The Ordovician sedimentary rocks occur solely in two areas; i.e. the Hida Mountains in central Japan and the Kitakami Mountains in NE Japan. In both areas, they are composed of terrigenous clastics with felsic volcanoclastics that overlie ophiolitic rocks. 1) Hida mountains: the latest Early to Middle Ordovician (472 Ma) Hitoegane Fm. (Tsukada and Koike, 1997; Nakama et al., 2010); 2) Kitakami mountains: Late Ordovician (457 Ma) Koguro Fm. overlying Middle Ordovician (466 Ma) trondhjemite of the Kagura ophiolitic complex (Shimojo et al., 2010). The ages were given by U-Pb dating of igneous zircons and partly by conodont biostratigraphy of intercalated felsic tuffs.

These Ordovician volcani-clastic rocks of calc-alkaline nature represent the sedimentary cover of the proto-Japan that already formed a matured arc-trench system. Together with the contemporary high-P/T metamorphic rocks, they clearly prove that the oceanic subduction has continued during the entire Ordovician period to build a thick juvenile arc crust off South China. The highly limited preservation/occurrence of these orogenic elements of the Ordovician arc-trench system was likely related to the severe tectonic erosion during the Late Paleozoic to Triassic (Isozaki et al., 2010, 2011).

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DARRIWILIAN BIOSTRATIGRAPHY AND PALAEOECOLOGY DURING THE GREAT ORDOVICIAN BIODIVERSIFICATION EVENT – A NORTHERN GONDWANAN PERSPECTIVE

K.G. Jakobsen¹, D.A.T. Harper¹, A.T. Nielsen¹ and G.A. Brock²

¹ Geological Museum, Natural History Museum of Denmark, University of Copenhagen, Øster Voldgade 5-7, DK-1350 Copenhagen K, Denmark. kgjakobsen@snm.ku.dk

² Palaeobiology, Department of Biological Sciences, Macquarie University, NSW 2109, Australia.

Keywords: Biodiversity, Mid Ordovician, benthic faunas, siliciclastics.

INTRODUCTION

The marine biosphere underwent a profound transformation during the Great Ordovician Biodiversification Event (GOBE), generally recognized as the longest Phanerozoic interval of sustained biodiversification with a three- to fourfold increase in numbers of families and genera.

During the Mid Ordovician (Darrivilian) marine sand- and siltstones were deposited in an epicontinental sea occupying part of central Australia (Webby, 1978). The local stratigraphy is shown in Figure 1. The Middle Darrivilian Stairway Sandstone Formation has, for the first time, been sampled stratigraphically to track the GOBE on this part of Northern Gondwana. Relatively little work has previously been aimed at describing the fossil fauna from this interval (Shergold, 1986), with the exception of studies of the bivalves and rostroconchs by Pojeta et al. (1977a, 1977b).

The newly sampled fauna is dominated by bivalves (around 25 species), whereas the remaining fauna comprises trilobites, brachiopods (both rhy-

AGE	FORMATION			MAXIMUM THICKNESS
ORDOVICIAN	LARAPINTA GROUP	CARMICHAEL SANDSTONE		150 m
		STOKES SILTSTONE		650 m
		STAIRWAY SANDSTONE		600 m
		HORN VALLEY SILTSTONE		120 m
		PACOOTA SANDSTONE (700-800 m)	S4	110 m
S3	120 m			
S2	320 m			
S1	160 m			
CAMBRIAN		GOYDER FORMATION	U	300 m
			L	300 m

Figure 1. Formations of the Larapinta Group including the Stairway Sandstone investigated herein. From Laurie et al. (1991).

chonelliforms and nonarticulates), rostroconchs, gastropods, bryozoans, cephalopods, sponges? and monoplacophorans in addition to abundant trace fossils.

In the present study, two sections located 40 km apart (Tempe Downs and Areyonga) in the Amadeus Basin, Central Australia, have been investigated for fossils and the lithology has been logged. The Stairway Sandstone is about 250 m thick in each section. Preliminary taxonomic identifications of the fauna are complete, and taxonomic descriptions are now being undertaken in order to develop an abundance-chart facilitating interpretation of the palaeoecology, local extinctions and originations together with the overall diversification.

AIMS OF STUDY

The primary focus is the record of marine benthic diversity in clastic shallow water settings (Fig. 2, area 1). Coeval carbonates in Tasmania (Fig. 2, area 2), will be sampled in 2011 and compared in terms of diversity patterns in these different settings.

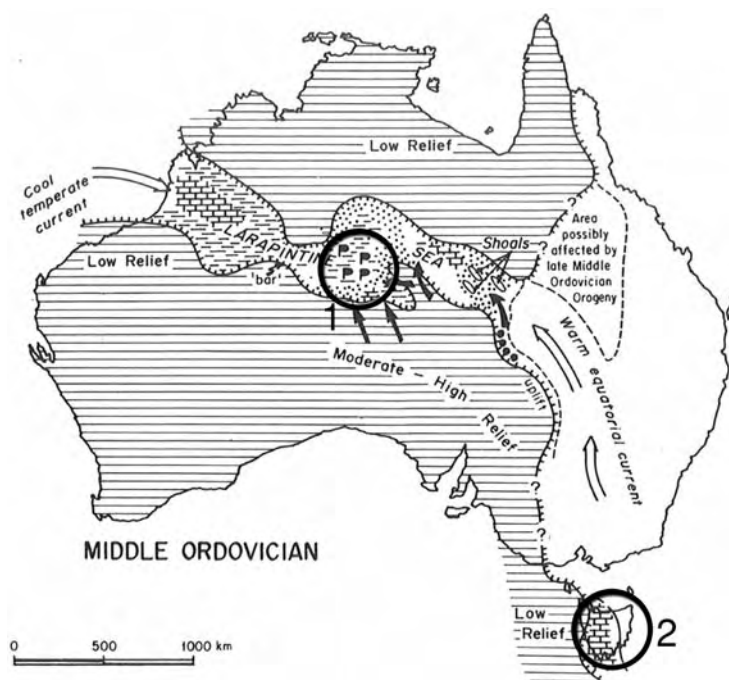


Figure 2. 1, (circled area) indicates the setting for the sampled Stairway Sandstone in Central Australia. 2, (circled area) indicates the setting for future sampling in Tasmania. From Webby (1978).

Webby (2000, 2004) noted that there were three diversity maxima during the Ordovician radiation based on a compilation of global data of all major fossil groups, such as the trilobites, brachiopods, bryozoans, echinoderms and bivalves. The onset of the first major biodiversity surge occurred during the Mid Ordovician with an extensive expansion of the benthos. The investigated stratigraphical interval

corresponds to the peak of the first maximum in the early Darriwilian (i.e. transition from Arenig to Llanvirn in terms of the British Series, see Fig. 3).

Biodiversity during the first diversity maximum in the Mid-Ordovician has been examined comparing Australia (this study), South China and Baltica in order to test diversity patterns on a global scale. The main focus is the brachiopod assemblages.

Webby (1978) suggested that a warm equatorial current flowed by Tasmania between the late Early Ordovician and Mid Ordovician (eventually flowing into the epicontinental sea investigated in Central Australia, see Fig. 2). A test of this model can be provided by interrogation of ocean chemistry using stable carbon and oxygen isotopes on whole rock and brachiopod shells, from the Tasmanian carbonates in order to examine palaeoenvironmental conditions (especially water temperature).

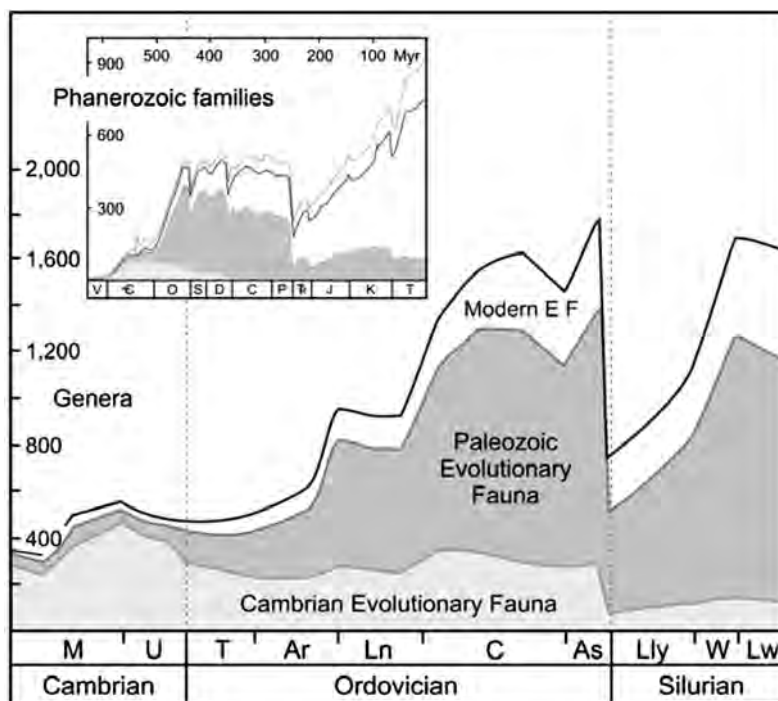


Figure 3. From Trotter et al. (2008). Modified from Sepkoski (1995). Cambrian: M, Middle; U, Upper. Ordovician: T, Tremadoc; Ar, Arenig; Ln, Llanvirn; C, Caradoc; As, Ashgill. Silurian: Lly, Llandovery; W, Wenlock; Lw, Ludlow.

DISCUSSION

The warm equatorial current suggested by Webby (1978), flowing past Tasmania and entering the siliciclastic belt in Central Australia is not supported by more recent studies (e.g. Haines et al., 2008) that suggest that the Larapintine Seaway (Fig. 2) probably did not exist during the Mid Ordovician. Macrofossils from the Amadeus Basin display a high degree of endemism compared to those of the Canning Basin (and vice versa) farther to the west where a supposed cool temperate current entered the basin (Fig. 2). The rare

species in common between these basins are generally cosmopolitan taxa. Stable oxygen isotopes from these sections will provide important information on relevant aspects of ocean chemistry. This will modify estimates of the pattern of marine currents and the sea temperature during the Darriwilian in Australia.

Bivalves are the most abundant and diverse group in the Areyonga and Tempe Downs sections. The majority of the bivalves belong to the suborder Nuculoidea. Bivalves occur in all types of lithologies within the Stairway Sandstone, varying from siltstone, calc-arenite, quartzite to dolomitic calc-sandstone.

Brachiopods and trilobites typically occur in the silt dominated facies intervals (deeper water) compared to the sandy dominated intervals (shallow water). These observations indicate that bivalves dominated the benthos in the epicontinental sea apparently regardless of the facies type. Maximum depth of deposition in the sea was likely less than 40 m (probably close to storm wave base).

The fauna from the Stairway Sandstone is dominated by filter feeders. Bivalves, brachiopods and rostroconchs constitute a large portion of the fauna. Filter feeders are better adapted in the Ordovician faunas than the Cambrian faunas, where detritus feeders are more dominant. As the Stairway Sandstone probably was deposited in near shore environments, relatively high energy conditions would be expected. The filter feeders in the Stairway fauna therefore have adapted to these palaeoenvironmental conditions, whereas many other filter feeder communities would prefer deeper- and calmer water conditions. Oceanic currents carrying micro-plankton into the epicontinental sea could be one of the explanations for this. On a global perspective the increase of micro-plankton in the Ordovician is one of the key explanations for the sudden increase in number and diversity of filter feeders (Servais et al., 2008).

Biodiversity of the benthic faunas during the first diversity maxima in the Mid-Ordovician (see Fig. 3) have been examined by comparing data from Australia (this study), South China and Baltica in order to test diversity patterns on a global scale. Data presented in Zhan et. al. 2009 (South China) and Rasmussen et. al. 2007 (Baltica) are compared below with the preliminary results from this study.

Previous collections at Tempe Downs in Central Australia (undertaken by Nielsen, 1990) from the very upper part of the Horn Valley Siltstone underlying the Stairway Sandstone (Fig. 1), correspond to the upper Arenig. The diversity of the bivalves is slightly higher in the Stairway Sandstone, while the brachiopod fauna is more diverse in the upper part of the Horn Valley Siltstone. Here there are six rhynchonelliform (orthoid) taxa compared to only two orthoid taxa in the entire Stairway Sandstone; no brachiopods have been found in the very lower part of the Stairway Sandstone. The higher diversity of the brachiopods in the siltstone and the higher diversity of bivalves in the sandstone suggest facies preferences in the two groups.

This study of regional palaeobiogeographic patterns in Northern Gondwana will form the basis for a global palaeobiogeographic analysis of Middle Ordovician shelly assemblages. For example, the brachiopod fauna investigated in this study is strongly dominated by orthoids. Rasmussen et al. (2007) demonstrated, based on a detailed study of Eastern Baltic assemblages, that the orthoid brachiopod fauna displays a maximum diversity at roughly the same time in Baltica. New data from the shelf carbonates in Tasmania will provide a further test for some of the patterns and trends in brachiopod diversity – for example, if they are facies related or truly represent part of a global biodiversity signal.

The diversity signal among the orthoids in the Stairway Sandstone in Australia and the pattern seen in the inner shelf carbonates in Baltica are quite different. Thus the increase in diversity in Baltica seems to be coincident with a decrease in Australia. The diversity at both localities could be facies controlled, but another aspect is that the first diversity maximum in Ordovician (maybe independent of the facies) was not necessarily coeval on all the palaeocontinents.

In South China for example, the first diversity maximum in the Early Ordovician brachiopods was apparently earlier than that of other marine invertebrates, such as trilobites, graptolites and bivalves. Here

brachiopods occurred initially in normal marine, shallow water environments and then gradually moved to both nearer-shore and offshore locations (Zhan et al., 2009). The reason why the Ordovician brachiopod radiation of South China apparently was earlier than the global trends, together with data available from other palaeoplates or terranes, may be related to its unique palaeogeographic position (peri-Gondwanan terrane gradually moving to equatorial latitudes). In the Early and Mid Ordovician Australia was already situated in equatorial latitudes. Therefore it is possible that the first Ordovician diversity maximum in Australia had occurred even earlier than in South China and Baltica because of its optimal (in terms of diversity) palaeogeographic position.

Taxonomic analysis of the first Ordovician diversity maximum in South China, Baltica and Australia indicates that the main contributors to the Early Ordovician brachiopod radiation were the Orthida and Lingulida.

Therefore even though the geological settings for the three comparative studies are quite different, the palaeolatitude varies a lot from Australia to Baltica and South China and the timing of the peak of diversity varies in time it is still the same types of brachiopods that dominate. More detailed analysis and comparison of diversity curves are required to draw any final conclusions, whether for example this diversification of orthoids and lingulids is relatively local and merely facies controlled.

CONCLUSIONS

Based on preliminary taxonomic identifications there is good correlation between the Areyonga and Tempe Downs sections and their faunas investigated from Central Australia.

Detailed lithostratigraphic logs and detailed stratigraphic sampling of the Mid Ordovician fossil fauna from Areyonga and Tempe Downs have been carried out for the first time.

Bivalves are the most abundant and diverse group in the Areyonga and Tempe Downs sections, in all facies types.

Orthoids and lingulids are apparently the dominant brachiopods in the Early Ordovician brachiopod radiation in the Amadeus Basin, Central Australia. Examination of sections from South China and Baltica show the same pattern.

Many different groups are represented in the sections investigated and the diverse Stairway Sandstone fauna includes brachiopods, bivalves, rostroconchs, gastropods, monoplacophorans, trilobites, cephalopods, bryozoans, sponges? and abundant trace fossils (*Skolithos*, *Diplocraterion* and *Cruziana*).

The Stairway Sandstone fauna is dominated by filter feeders.

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THE UPPER KATIAN (UPPER ORDOVICIAN) BRYOZOANS FROM THE IBERIAN CHAINS (NE SPAIN): A REVIEW

A. Jiménez-Sánchez

Departamento de Ciencias de la Tierra, Instituto Universitario de Investigación en Ciencias Ambientales de Aragón (IUCA),
Facultad de Ciencias, Universidad de Zaragoza, Pedro Cerbuna 12, 50009, Zaragoza. andrea.j@unizar.es

Keywords: Bryozoans, Cystoid Limestone Formation, Ordovician, upper Katian, palaeobiogeography, palaeotemperatures indicators, Spain.

INTRODUCTION

The Phylum Bryozoa is composed of colonial organisms that appeared in marine environments at the end of the Cambrian period (Landing et al., 2010). They subsequently diversified throughout the Palaeozoic, the Mesozoic and two of its groups have successfully reached the present day. More than 90% of living species and almost 100% of fossil bryozoans have a carbonate skeleton, which has allowed for a relatively continuous fossil record since Cambrian to present. This phylum is divided into three classes: Stenolaemata, Gymnolaemata and Phylactolaemata. The Stenolaemata was dominant throughout the Palaeozoic, with more than 500 genera recorded, but its importance decreased in the Mesozoic and today the Order Cyclostomata is its only surviving branch. The Gymnolaemata have been the most abundant since the late Mesozoic, with the Order Cheilostomata as its main representative. Finally, the class Phylactolaemata, recorded back to the Late Permian or Triassic, is the smallest of the three classes and is found exclusively in freshwater environments.

The Phylum Bryozoa has adapted to living in waters of all temperatures, from the warmth of the tropics to the cold of the Arctic and Antarctic. However, whereas Palaeozoic bryozoans mainly colonized the warm waters of the tropic, today they are more or less restricted to temperate and polar waters of middle-high and high latitudes.

The preference of bryozoans for warm water in the Palaeozoic make them one of the main and most abundant invertebrate groups in the tropical and middle-low carbonate platforms during the Ordovician, delivering a continuous fossil record in that period in the palaeocontinents of Baltica, Laurentia and Siberia and in several other small terrains also located in these latitudes. In spite of this Palaeozoic bryozoans' preference for warm water, during the upper Katian (Upper Ordovician) large carbonate platforms were established in the North Gondwana margin [whose latitude was 55°-70° S (Jiménez-Sánchez and Villas (2010))], and bryozoans for the first time colonized the cold waters of middle-high and high latitudes, becoming one of the most abundant and diversified groups inhabiting these platforms.

The carbonate platforms developed during the upper Katian in the North Gondwana margin crop out today in central and southern Europe and also in north Africa in the form of widespread carbonate formations. These formations have faunistic and sedimentologic features that make them different from the tropical carbonate platforms developed during the same period. The aim of this work is to review briefly the bryozoan fauna of the Cystoid Limestone Fm, the representative unit in the Iberian Chains of those Katian North Gondwana carbonates, as well as to put this fauna into a global context and to set out the new research lines that this knowledge is opening.

The work carried out on Cystoid Limestone bryozoans by Jiménez-Sánchez and co-workers (Jiménez-Sánchez et al., 2007; Jiménez-Sánchez, 2009, 2010; Jiménez-Sánchez and Villas, 2010; Jiménez-Sánchez et al., 2010) adds to the already existing in-depth studies of Conti (1990), Buttler et al. (2007) and Ernst and Key (2007) about Upper Ordovician bryozoans from Sardinia, Libya and Montagne Noire, respectively, and has notably contributed to a better understanding of the Upper Ordovician bryozoans from the Mediterranean region. Thanks to the study of this region's bryozoans has been possible to deepen the knowledge of high-latitude bryozoans, much less known than their low latitudes counterparts.

GEOGRAPHICAL AND GEOLOGICAL SETTINGS

The best sections of the Cystoid Limestone Fm are located in the northeast and southwest of the Eastern Iberian Chain. However, in the northeast sections diagenetic alteration has deleted most of the morphological features needed to identify bryozoans even to family level. Thus, all taxa mentioned in this work come exclusively from the southwest sections. These sections are: Valdelaparra and La Peña del Tormo, near the village of Fombuena, and Luesma 2 and Luesma 3, near the village of Luesma; all of them are located in the province of Zaragoza (Jiménez-Sánchez, 2010, fig. 1).

Using lateral and vertical facies changes, Hammann (1992) divided this formation into four members. In the Valdelaparra and La Peña del Tormo sections, located westward, the La Peña and Rebollarejo members can be distinguished, whereas in the Luesma 2 and Luesma 3 sections, located eastward, the Ocino and Rebosilla members can be distinguished (Jiménez-Sánchez, 2009, fig. 1). The distribution of the species in the studied sections and lithostratigraphic units is not homogenous (Fig. 1). The La Peña Member is the most productive, having yielded the 29 known species; in the Rebollarejo Member only three species are represented, the same as in the Rebosilla Member; and the Ocino Member has not yet yielded any taxa. Valdelaparra is the section with the highest bryozoan diversity, with 26 recognized species.

A detailed description of the geographical and geological setting of the studied sections, as well as complete information regarding the stratigraphic and sedimentary characteristics of the Cystoid Limestone Formation, can be found in Villas (1985), Vennin et al. (1998), and Jiménez-Sánchez (2010).

SYSTEMATIC SUMMARY

A total of 29 species (Fig. 1) have been described in the Cystoid Limestone Fm. These species belong to 25 genera and 13 families, which are assigned to the five Stenolaemata orders already present since the Middle Ordovician. The order Trepostomata, with 15 species, and the order Cryptostomata, with 9 species, are the most diverse. The other 5 species are assigned to the orders Cystoporata and Fenestrata (two species per order) and Cyclostomata (one species only). A detailed description of these species can be found in Jiménez-Sánchez (2009, 2010), and Jiménez-Sánchez et al. (2010).

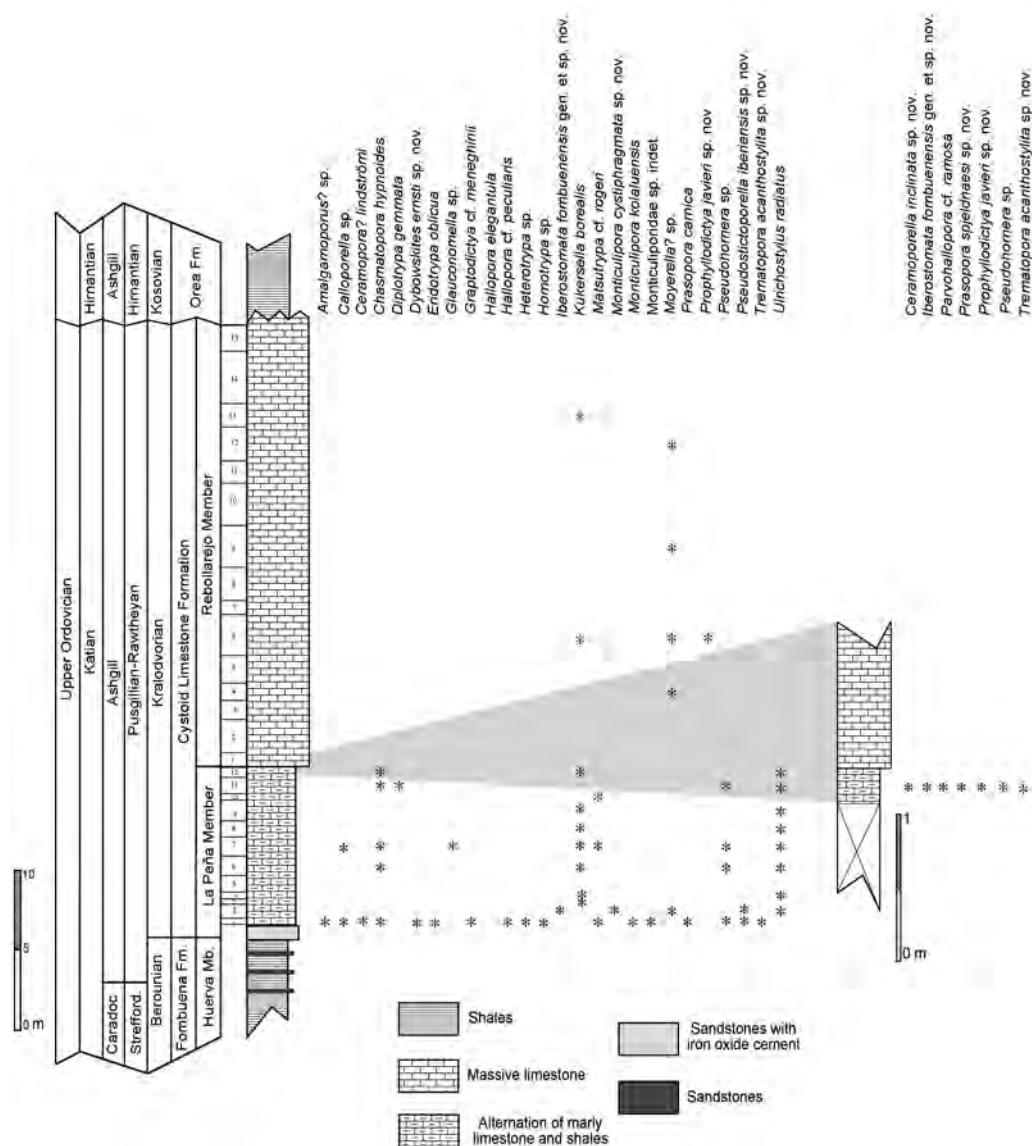


Figure 1. Bryozoans from the Cystoid Limestone Fm and their stratigraphic distribution in the sections of Valdelaparra (on the left) and La Peña del Tormo (on the right).

The species *Ceramoporella inclinata* Jiménez-Sánchez, 2009, *Dybowskiites ernsti* Jiménez-Sánchez, 2009, *Prophyllodictya javieri* Jiménez-Sánchez, 2009, *Pseudodictoporella iberiensis* Jiménez-Sánchez, 2009, *Trematopora acanthostylita* Jiménez-Sánchez, 2009, *Monticulipora cystiphragmata* Jiménez-Sánchez, 2010, *Prasopora spjeldnaesi* Jiménez-Sánchez, 2010 and *Iberostomata fombuenensis* Jiménez-Sánchez and Anstey 2010 were only known from the Cystoid Limestone Fm.

Also the genus *Iberostomata* Jiménez-Sánchez and Anstey, 2010 (in Jiménez-Sánchez et al., 2010), is not known out of this formation. The complexity of the diagnostic features in this genus made cladistic methodology the only feasible alternative for its inclusion in a superior level group. In this way, Jiménez-Sánchez et al. (2010) used for the first time the cladistic methodology to generic and familiar level in Palaeozoic bryozoans, assigning the new genus to the family Rhinidictyidae of the suborder Ptilodictyina (order Cryptostomata) and clarifying the phylogenetic relationships between the families of that suborder.

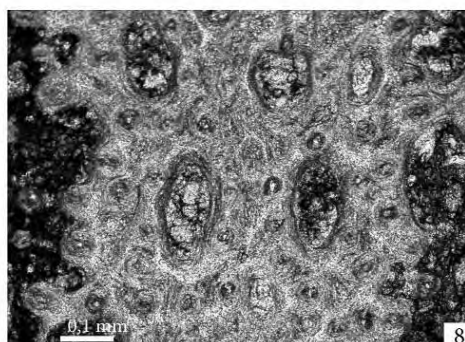
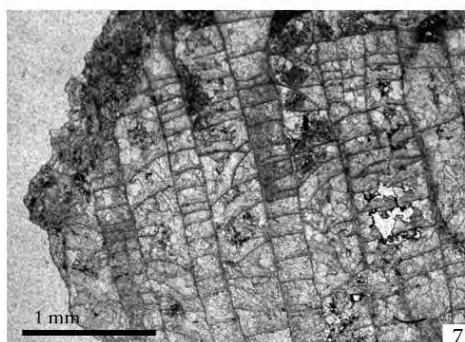
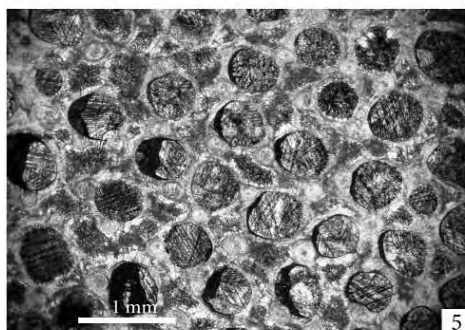
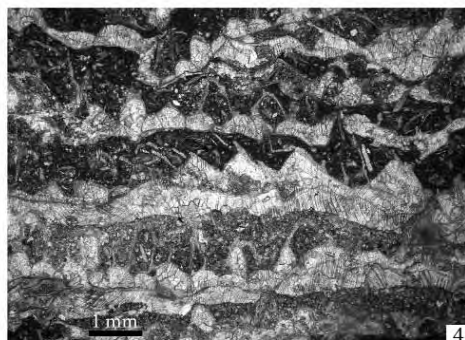
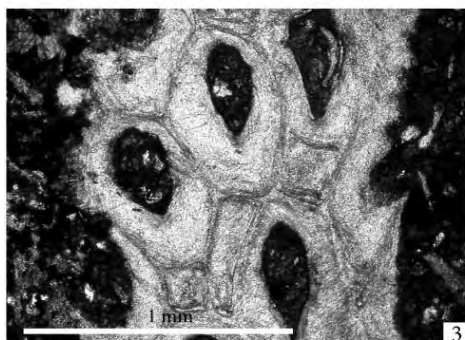
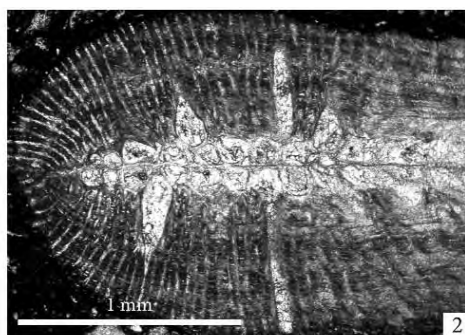
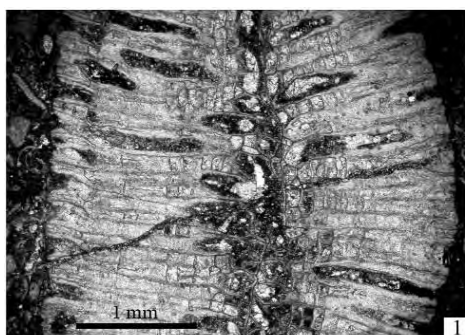
PALAEOBIOGEOGRAPHY

The systematics study of the bryozoans from the Cystoid Limestone Fm has been a key step in the assessment of the palaeobiogeographic relationships of the group during the Upper Ordovician helping to clarify the origin of high-latitude bryozoans. It has also contributed to clarify the palaeogeographic relationships between palaeocontinents in this period. The identification of Cystoid Limestone bryozoans and its addition to the list of North Gondwana margin bryozoans (Carnic Alps, Libya, Montagne Noire, Morocco and Sardinia), whose knowledge has been greatly improved in the latest 20 years, has allowed to built a data-base with the palaeogeographic distribution of all genera present in the upper Katian (Jiménez-Sánchez and Villas, 2010). This data-base collects for the first time the distributions of all upper Katian bryozoans, from tropical latitudes to near polar high latitudes.

Jiménez-Sánchez and Villas (2010) designed a presence/absence matrix where 136 upper Katian genera are registered and their geographic distribution assigned to 45 localities. These localities, except for Libya and Morocco, have a diversity of more than 8 genera and belong to the palaeocontinents of: Laurentia (24 localities), Baltica (5 localities), Siberia (5 localities), Avalonia (2 localities) and North Gondwana margin (6 localities); as well as one locality in each of the terrains of Altai Sayan, South China and India. The matrix was analyzed with two multivariant methods: Detrended Corresponding Analysis (DCA) and Principal Coordinates Analysis (PCO), using the Dice and Simpson similarity indices. The analyses show that in the upper Katian the bryozoan fauna was distributed in the Laurentia-Siberian province (including Altai Sayan and South China), the Baltic province (including Avalonia), and the Mediterranean province, composed by all the localities from the North Gondwana margin plus India. The Laurentia-Siberian province occupied tropical latitudes, the Baltic province spanned from tropical to middle-low latitudes and the Mediterranean province was placed in middle-high to high latitudes (Fig. 2). The analyses also show that the Mediterranean province is the one with the most defined faunistic identity.

During the upper Katian 68 genera thrived on the high latitude carbonate platforms of the North Gondwana margin, characterising the Mediterranean province. Forty-six of these genera are also present in low to middle-low latitude provinces, although the analysis show that the Mediterranean province has a higher faunistic resemblance with the Baltic province than with the Laurentia-Siberian province. Twenty-two genera of the Mediterranean province are endemic. However, within them, the genera *Amalgamoporous*, *Moorephylloporina*, *Nematotrypa*, *Orbipora*, *Prophyllodictya*, *Pseudostictoporella* and *Ralfinella* inhabited tropical and middle-low latitude platforms before the upper Katian. Based on the stenothermic character of most bryozoans, Jiménez-Sánchez and Villas (2010) proposed that the migration

Plate 1. The new species described in the Cystoid Limestone Fm by Jiménez-Sánchez et al. (2010: 1) and Jiménez-Sánchez (2009: 2-8). 1, *Iberostomata fombuenensis*; 2, *Prophyllodictya javieri*; 3, *Pseudostictoporella iberiensis*; 4, *Ceramoporella inclinata*; 5, *Dybowskites ernstii*; 6, *Monticulipora cystiphragmata*; 7, *Prasopora spjeldnaesi*; 8, *Trematopora acanthostylita*.



of these genera from the tropical and middle-low latitudes (warm water) of the Laurentia-Siberian and Baltic provinces to the high and middle-high latitudes (cold water) of the Mediterranean province, could be linked to a temperature increase in the tropics. The extinction of these genera in tropical latitudes, simultaneously with their migration to higher latitudes, agrees with the hypothesis of global warming [Boda event, Fortey and Cocks (2005)] previous to the Hirnantian glaciation which ended the Ordovician period.

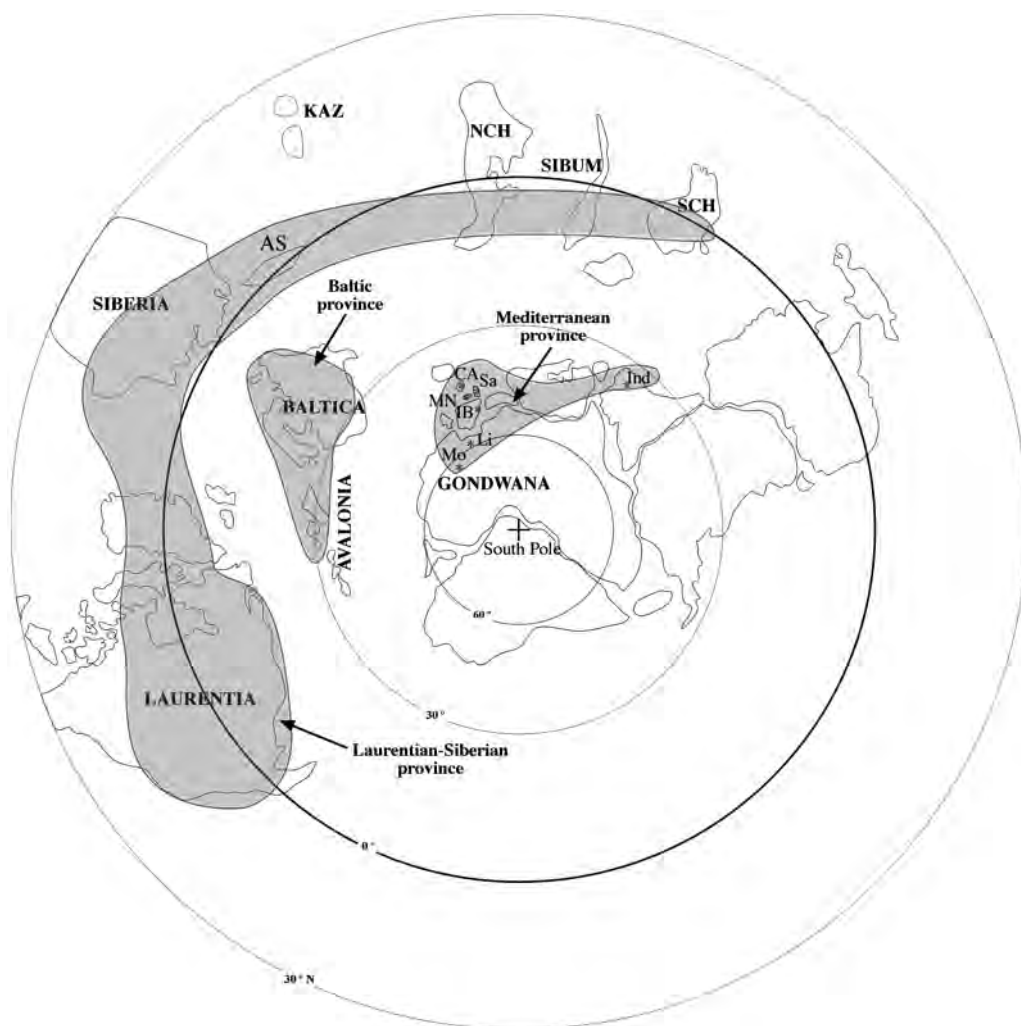


Figure 2. Palaeogeographic distribution of the Laurentia-Siberian, Baltic and Mediterranean provinces. AS, Altai Sayan; KAZ, Kazakhstan; NCH, North China; SCH, South China; SIBUM, Sibumasu; CA, Carnic Alps; IB, Iberian Chains; Ind, India; Li, Libya; MN, Montagne Noire; Mo, Morocco; Sa, Sardinia. Modified from Jiménez-Sánchez and Villas (2010).

OPEN QUESTIONS: THE FUTURE

The existence during the Upper Ordovician of bryozoans that inhabited both warm and cold waters opens up the possibility of using bryozoans as palaeotemperature indicators. The question is: is there any morphological, mineralogical or chemical feature in bryozoans which depends on temperature in a systematic way?

Studies carried out so far with recent species of the orders Cheilostomata and Cyclostomata point to some common patterns (Kuklinski and Taylor, 2008, 2009). Carbonate skeletons of bryozoans living in warm waters are made up of calcite, aragonite or are bimineralic (calcite and aragonite in superposed layers), and, when the skeleton is of calcite it has a high percentage of Mg. On the other hand, in cold water species, aragonite and bimineralic skeletons are rare and the Mg content in the calcite is low. These mineralogical and chemical differences in the bryozoan skeletons have been linked to environmental factors because they have also been found to occur within species that inhabit different temperatures. Other studies of recent bryozoans show that differences between cold and warm water species are reflected in morphology, including the development of polymorphic zooids and other features adapted to completely different environments varying in seasonality, predator pressure, availability of food, etc. In addition, there is an inverse relationship between the size of the zooids in bryozoan colonies and the temperature at which they were budded. This relationship parallels Bergmanns Rule and may be due to surface area/volume and correlated metabolic factors. For example, oxygen is less soluble in warm water so that zooids have to be smaller, and consequently have a larger relative surface, if the same amount of oxygen is to be acquired.

In summary, it seems clear that several characteristics of recent bryozoans correlate with sea temperature and therefore the latitude at which they live. However, nothing is known about the differences, if any, between cold and warm water bryozoans in the Mesozoic and Palaeozoic. During the Late Ordovician 46 of the 136 existing genera and some of their species were able to survive both on temperate to tropical platforms from the palaeocontinents of Baltica, Laurentia and Siberia and on the high-mid and high latitude platforms of North Gondwana (Jiménez-Sánchez and Villas, 2010). The good knowledge of tropical bryozoans and the recent advances made in high-latitude bryozoans with the study of Cystoid Limestone bryozoans and with the ongoing study of the upper Katian bryozoans from Morocco, will help to answer the questions of what are the differences between stenothermic genera and eurythermic genera and what distinguishes the species found in cold waters from those found in warm waters. Answers to these questions will help in the study of Ordovician palaeogeography and palaeoclimatology by providing additional temperature proxies.

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CARBON ISOTOPE TREND IN THE MIRNY CREEK AREA, NE RUSSIA, ITS SPECIFIC FEATURES AND POSSIBLE IMPLICATIONS OF THE UPPERMOST ORDOVICIAN STRATIGRAPHY

D. Kaljo and T. Martma

Institute of Geology, Tallinn University of Technology, Ehitajate tee 5, Tallinn 19086, Estonia.
kaljo@gi.ee, martma@gi.ee

Keywords: Carbon isotopes, chemostratigraphy, types of isotope trends, uppermost Ordovician, NE Russia.

INTRODUCTION

The Hirnantian section at Mirny Creek located in the Omulev Mountain area, NE Russia (Fig. 1), played an important role in deliberations concerning the boundary between the Ordovician and Silurian systems some three decades ago. The history of studies and modern state of knowledge about the section were recently reviewed by Koren' and Sobolevskaya (2008). Taking this paper as a basis in the sense of local geology and stratigraphy, we wish to report about our new carbon isotope data and to discuss a couple of more general topics of the uppermost Ordovician chemostratigraphy. Those include differences in patterns of the corresponding $\delta^{13}\text{C}$ trends in Avalonia, Baltica, Laurentia, the Kolyma terrain and South China plate and the dating of carbon isotope excursions in terms of biostratigraphy.

The studied Mirny Creek and Neznakomka River bank sections embracing the upper Katian through the lower Rhuddanian, represent a rather thick succession (ca 450 m) of carbonate sediments that have accumulated rapidly and alternate with thinner argillaceous and siltstone packets (Koren' et al. 1983). Among these the Hirnantian rocks (89 m) constitute the upper part (Unit Q) of the Tirekhtyakh Formation having a rather specific fauna, especially due to scarcity of microfossils (Zhang and Barnes, 2007). Shelly fossils like brachiopods, corals and trilobites are quite common at some levels, however, graptolites are of greatest significance, making a general biozonal



Figure 1. Location of the Mirny Creek and Neznakomka River outcrops in Northeastern Russia.

framework of the Mirny section highly clear (Koren' and Sobolevskaya, 2008). Despite the occurrence of mixed shelly-graptolite fauna, detailed correlation with the Baltic and some other, prevalingly shelly faunal sections, is problematic and therefore chemostratigraphic correlation criteria become most decisive. Considering recent advancements in many areas, including Anticosti (Achab et al. 2011), Baltica (Schmitz and Bergström, 2007; Kaljo et al., 2008; Ainsaar et al., 2010; Hints et al., 2010) and China (Chen et al., 2006; Zhang et al., 2009), but also older data on Dob's Linn (Underwood et al., 1997) and Nevada (Finney et al., 1999), we wish to refine the understanding of the Hirnantian $\delta^{13}\text{C}$ trend and its global utility as a chronostratigraphic tool. Due to remoteness of the Mirny Creek area, we had to limit our analysis to samples collected for Koren' et al. (1983) monograph, not specifically for isotope studies. The sampling depths referred to below (Fig. 2) were calculated on the basis of thickness values in the measured sections. Carbon isotope data were obtained by using conventional methods of whole rock analysis.

CARBON ISOTOPE EXCURSIONS

Mirny Creek

The Katian part of the $\delta^{13}\text{C}$ excursion (not shown in Fig. 2), except the uppermost portion (Unit P), is substantiated by too widely spaced samples, but still a general trend is rather stable, varying close to 1‰ (0.6–1.3‰). The two topmost samples of Unit P and the first two samples of Unit Q show a rather quick increase in $\delta^{13}\text{C}$ values from 0.3 to 3.0‰ and 2.8‰. They form the first clear medium-sized peak of values identified in the uppermost Ordovician of the Mirny Creek section. Higher along the section carbon isotope values slightly decrease and form a relatively even (1.4–1.7‰) long plateau, which reaches, after a 13 m sampling gap, a slightly increased value (1.9‰) at 53 m depth. The next four samples show rather variable $\delta^{13}\text{C}$ values (0.8–1.6‰), which could be considered as the end of the above noted long plateau.

The next three samples demonstrate the highest $\delta^{13}\text{C}$ values (3.7–5.4‰), which surely mark the peak of the well-known HICE (Hirnantian carbon isotope excursion) at 34 m depth. The peak is followed by a steep decrease in values (about 4‰) down to nearly 1.2‰. Then the excursion continues as a nearly horizontal or very slightly falling plateau remaining close to 1‰ at 20 m depth.

The topmost 3 m of the Tirekhtyakh Formation shows very variable $\delta^{13}\text{C}$ values (–0.28 to 1.1‰, mean value 0.4‰ from 5 measurements). The Silurian Maut Formation begins with even lower values (mean 0.1‰ from 2 samples), but a low positive $\delta^{13}\text{C}$ shift (1.5‰) occurs slightly higher, followed by a steep drop of values down to –2.1‰.

Key levels of the above Mirny carbon isotope data curve are reliably dated by graptolites (Koren' and Sobolevskaya, 2008). The upper Katian part of the curve belongs to the *Appendispinograptus supernus* Biozone. Identification of the Katian-Hirnantian boundary level within the Tirekhtyakh Formation is more problematic. Without going into details, we quote *Normalograptus extraordinarius* occurring at 95 m depth just above a pack- and grainstone bed at the very bottom of Unit Q. This bed shows a typical "rising limb" of the HICE. Kaljo et al. (2008) pointed out that the actual increase in the HICE values began in several sections of China slightly before the first appearance datum (FAD) of *N. extraordinarius*, i.e. in the *Diceratograptus mirus* Biozone (Chen et al., 2006). This seems to be a case also in the Mirny Creek section, but the pack- and grainstone bed noted is not prospective for finding graptolites. The upper boundary of the HICE is well defined by the FAD of *Akidograptus ascensus* in the lowermost part of the Maut Formation (Silurian).

Summarizing the above data, we observe the following patterns of the $\delta^{13}\text{C}$ trend (Fig. 2): (1) A Katian plateau of low $\delta^{13}\text{C}$ values varying close to 1‰ in Unit O and ending with a brief interval of increasing

values in the top of Unit P. (2) The first Hirnantian medium-size increase in values (3‰) at the very beginning of Unit Q. (3) A plateau of slightly higher values varying close to 1.5‰ in the lower part of Unit Q. (4) A narrow peak of the HICE (maximum value 5.4‰) occurs in the middle of Unit Q. (5) A short falling plateau and a low of values at the Ordovician–Silurian boundary, followed by a minor positive excursion (1.5‰) in the lowermost Silurian.

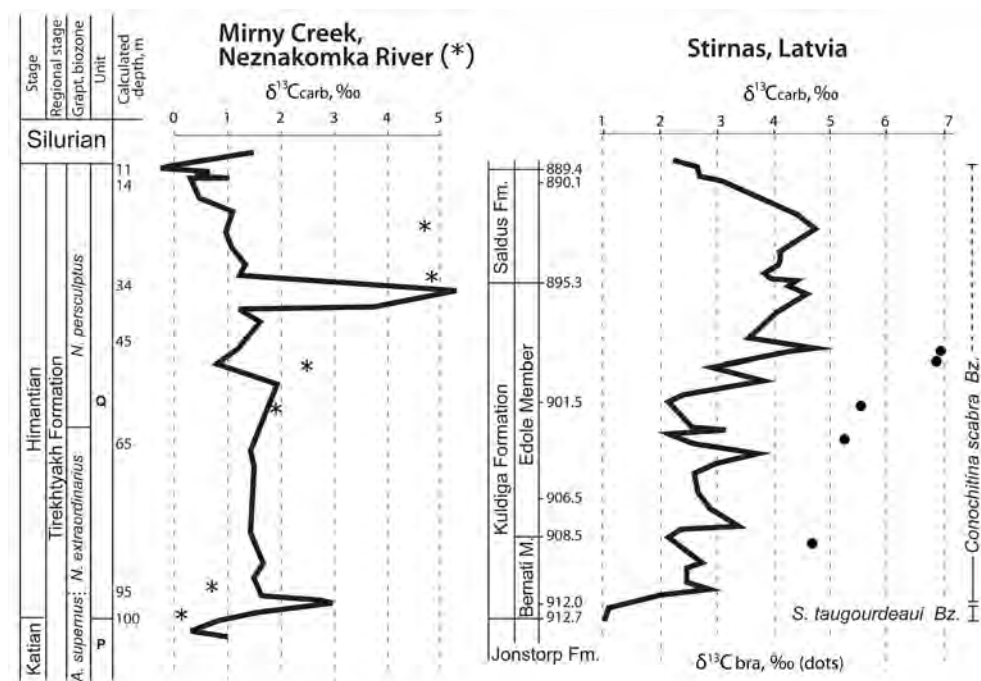


Figure 2. Hirnantian carbon isotope trends compared. Mirny Creek and Neznakomka River (*) – new data, stratigraphy simplified from Koren' and Sobolevskaya (2008); Stirnas – modified from Hints et al. (2010). Note that the Hirnantian column for all curves pictured is shown with equal longitude despite of huge difference in actual thickness of corresponding beds. The latter can be deduced by depth values or vertical scale given.

Neznakomka River banks

Having seen a rather specific carbon isotope trend of the Mirny Creek section, we analysed samples from the Neznakomka River. However, the number of available rock samples (6) was too small for compiling a normal $\delta^{13}\text{C}$ curve, but the data are shown in Fig. 2 together with the curve from the Mirny section. Due to possible correlation errors, we cannot be sure that the Neznakomka analyses are shown in fully correct positions. Nevertheless these samples provided valuable information for interpreting the Mirny Creek section.

The first two samples at the bottom of Bed 3 near the first occurrences of *N. extraordinarius* gave very low $\delta^{13}\text{C}$ values close to 0 but still showed certain increase upwards. Several samples higher in the section evidence a clear rising trend of $\delta^{13}\text{C}$ values, e.g. the value 1.9‰ was measured 30 m above the FAD of *N. extraordinarius* in Bed 4 and 2.5‰ 10 m higher. The trend was continued in Bed 5 (constitutes the

uppermost 35 m of the Tirekhtyakh Formation) - a sample 20 m higher gave the value 4.9‰ and another, 18 m higher, the value 4.8‰. Three samples from the Maut Formation (Silurian) are irrelevant to our discussion topic. All Unit Q samples represent a clear major $\delta^{13}\text{C}$ excursion, but because of the scarcity of samples we cannot be sure about the real shape of the isotope curve. Anyway, it is obvious that the HICE peak at the Neznakomka River is much wider than that in the Mirny section. If two samples in Bed 5 constitute the main peak of the HICE, the peak occurs very high in the Neznakomka section and may suggest a gap within the top of the Tirekhtyakh Formation or complications in the sedimentary process and correlation.

COMPARISON OF TRENDS

The Mirny Creek carbon isotope trend described above is surely the widely known HICE, but a highly specific one differing from the others in several aspects. Actually, there are two peaks: a medium-sized peak at the very beginning and a much higher major one, separated from the former by a slightly elevated plateau. The HICE ends with another plateau. Such plateaus, especially the rather long one between the peaks, were observed for the first time. A small set-back of $\delta^{13}\text{C}$ values, revealed by a couple of analyses after the initial increase in the HICE, can be seen in curves presented by several authors (Finney et al., 1999; Kaljo et al., 2001; Bergström et al., 2006; Chen et al., 2006). Based on this set-back, Fan et al. (2009) even distinguished peaks 1 and 2. Data from the Neznakomka River (although insufficiently precisely positioned) soften to some extent the Mirny $\delta^{13}\text{C}$ excursion pattern, but the presence of the wide plateau still needs explanation.

The first striking circumstance that may affect the curve is the thickness of Hirnantian rocks and the very high accumulation rate of sediments. This might have had some influence if the isotope trend had a punctuated character instead of continuous one. The thickness of the Hirnantian is ca 90 m in Mirny and reaches ca 70 m in Copenhagen Canyon, Nevada (rather close to the figure at Mirny), but no plateau pattern is observed in the latter region (Finney et al., 1999). No such plateaus has been recorded also in much thinner sections in Baltic area and elsewhere (Kaljo et al., 2001, 2008; Brenchley et al., 2003; Bergström et al., 2006). It seems that a high sedimentation rate cannot be the only reason for the origination of the plateau pattern and another local reason should be looked for. The Neznakomka data refer to facies differences – much higher values are tied to purer carbonate rocks (limestones) with a lesser content of the argillaceous and silty component. The same is obvious from Fig. 3 where the Dob's Linn and Wangjiawan curves document much smaller $\delta^{13}\text{C}$ changes (2-3‰) in deep-water terrigenous rocks than in the Kardia carbonate-rich mid- to shallow shelf section (relative amplitude of values 4-5‰). Such a pattern has been observed in several occasions (Kaljo et al., 1998; Munnecke et al., 2010).

The isotope excursions of the Mirny and Stirnas sections (Fig. 2) show certain similarity, when leaving aside a stronger variability of the first third of the latter curve. The beginning of both excursions is identical, follow a long plateau in Mirny and a variable interval in Stirnas, where the mean value reaches the 2.7‰ level, i.e. ca 1.2‰ higher than mean of the plateau. Both the plateau and the variable interval end with a pronounced low, where the values begin to rise stepwise up to the highest peak of the trend. These main peaks are rather similar - both close to 5‰, but reaching 7‰ in Stirnas by $\delta^{13}\text{C}_{\text{bra}}$. Both peaks occur in the upper half of the Hirnantian, but a little higher in Mirny (74% from the bottom) in the middle of the *N. persculptus* Biozone. In Mirny the peak is followed by a steep drop of values (ca 4‰) and then by a smoother decline of the curve. In Stirnas another variable plateau with a mean value of 4.1‰ occurs after

the peak within the next 8 m and only then a clearly declining limb follows. The two uppermost analyses from the Neznakomka River shown in Fig. 2 suggest an analogous wider excursion peak also in that area. At the same time these samples raise a question why the main peak is so narrow (only 3-4 m) and steep at Mirny Creek. Having in mind several truncated sections demonstrated by Brenchley et al. (2003) from the Baltic and Anticosti (also Achab et al., 2011), we may suggest that a part of the Mirny section just above the peak is missing.

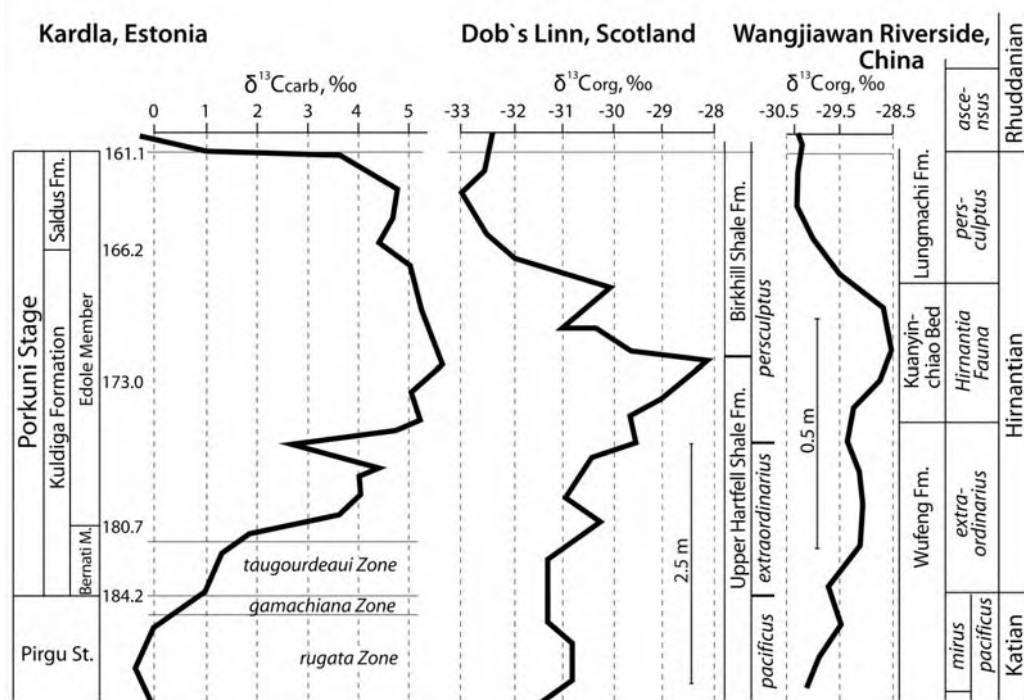


Figure 3. Hirnantian carbon isotope trends compared. All parts modified from as follows: Kardla – Brenchley et al. (2003); Dob's Linn – Underwood et al. (1997); Wangjiawan – Chen et al. (2006). For thickness see Fig. 2.

Another type of the shape of the $\delta^{13}\text{C}_{\text{carb}}$ curve is represented here by the Kardla section (Fig. 3), and is widely known in the Monitor Range, Nevada and elsewhere (Finney et al., 1999; Saltzman & Young, 2005; Bergström et al., 2006; Kaljo et al., 2007). This type of excursion is biostratigraphically most convincingly constrained by graptolites in the Central Nevada sections. Organic carbon data are often more variable (Melchin and Holmden, 2006; Fan et al., 2009; Zhang et al., 2009). We do not go into details, because according to conodont colour alteration index (CAI 4-5), the Mirny Creek sections are heated up to 400-500°C (Zhang and Barnes, 2007; P. Männik, pers. comm., 2010) and therefore not suitable for C_{org} analysis. On the other hand, graptolite-bearing rocks, usually analysed for C_{org} , do not cause normally serious dating problems and can help through cautious chemostratigraphic correlation improve dating of events in shelly faunal sections.

CONCLUSIONS

1. The longest Hirnantian $\delta^{13}\text{C}$ trend in the Mirny Creek section has a highly specific shape but is well constrained by graptolite biostratigraphy. The beginning of the trend is dated by the FAD of *N. extraordinarius*, but it might commence slightly below this level. The main peak occurs nearly in the middle of the *N. persculptus* Biozone. A few additional samples from the Neznakomka River suggest a somewhat wider peak interval than at Mirny Creek.

2. Detailed comparison of the Mirny and Stirnas (Latvia) $\delta^{13}\text{C}$ curves shows their general similarity despite great specifics of both trends. This correlation facilitates the linking of the Baltic chitinozoan and conodont biostratigraphy with the global graptolite biozonal standard.

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FOSSIL ASSEMBLAGES REFLECTING PROCESSES OF THE EARLY DEVELOPMENT OF THE PRAGUE BASIN (BOHEMIAN MASSIF, CZECH REPUBLIC)

P. Kraft¹, T. Hroch^{1, 2} and M. Rajchl²

¹ Institute of Geology and Palaeontology, Faculty of Science, Charles University in Prague, Albertov 6, 128 43 Praha 2, Czech Republic.; kraft@natur.cuni.cz, tom.hroch@gmail.com

² Czech Geological Survey, Klárov 3, 118 21 Praha 1, Czech Republic.; michal.rajchl@geology.cz

The history of the Prague Basin began during the Tremadocian with deposition of the basal unit of the Třenice Formation. These initial deposits of the basin infill are represented by coarse-grained siliciclastics of marine origin. Their thickness ranges from a few to 70 meters. The Třenice Formation passes without interruption into the Mílina or Klabava Formation with sediments continuously fining upwards reaching the prevalence of shales.

Three main facies associations have been recognized during the sedimentological studies: (i) graded conglomerates, (ii) cross-bedded and massive sandstones, and (iii) volcanogenic clastics. Graded conglomerates are interpreted as transgressive coastal lag that was accumulated during marine transgression into the area of the Prague Basin. Cross-bedded and massive sandstones are interpreted as a record of dunes with significantly sinuous crests in the shoreface environment. The massive sandstones are considered to be proximal tempestites. Volcanogenic clastics are interpreted as a record of cohesive debris flows and turbidity currents. The latter association contains a significant volcanogenic admixture of vesicular rhyolitic clasts and volcanic glasses intercalated within the shoreface deposits. However, the volcanogenic material is very common even in both of the other facies associations. It documents a syndimentary volcanic activity. Preservation of non-resistant volcanic glasses argues for short-distance transport. The absence of hyaloclastics and lithological similarity to products of subaerial volcanic complexes adjacent to the Prague Basin supports a subaerial origin of the volcanoclastic material, that was subsequently resedimented by subaerial and shoal-marine processes. Finally, the resedimented volcanic material was deposited due to debris flows or turbidity currents in shore face environment. Thus, the Třenice Formation represents the sedimentary record of interaction between shallow-marine processes in the embayment setting and syndimentary subaerial volcanic activity. The described environment fundamentally influenced biotic colonization of the Prague Basin, the character and diversity of communities and their distribution patterns.

The fossil content of the Třenice Formation is generally poor with low species diversity. Linguliformean brachiopods are strongly dominant in number of specimens as well as species. The principal exception is the locality Holoubkov – V Ouzkém where an extraordinarily diversified assemblage occurs, composed of trilobites, echinoderms, linguliformean and rhynchonelliformean brachiopods, amongst others. Essentially all recorded macrofossil species of the Třenice Formation are demonstrably benthic. The absence of planktic

forms could, however, be a taphonomic or diagenetic bias. On the other hand, a primary prevalence of benthos, and especially linguliformean brachiopods, can be assumed among shelly fauna.

The assemblages from all fossiliferous localities in the south-western part of the Prague Basin were studied in detail. Species distribution and comparison of locality assemblages were analyzed to trace any distribution pattern of fossils. As a result three regions, characterized by specific fossil associations, were distinguished. These regions correspond to distribution of prevailing facies and lithological associations only very approximately. It seems that all facies associations occur in each region. The distribution pattern of fossil assemblages was more controlled by other environmental features. Chemistry of the environment, especially water acidity that was directly influenced by coeval subaeric volcanism, is considered to be responsible for low primary diversity, the prevalence of animals with organo-phosphatic shells, their spatial distribution and subsequent secondary taphonomic processes reducing diversity of the fossil record in affected sediment. Thus, initial colonization of the Prague Basin and its early communities reflected stressful conditions. Inhospitable conditions were locally improved by hydrothermal fluids accompanying the volcanism. They partly neutralized water around the vents and constituted "hydrothermal oases". The above mentioned locality Holoubkov – V Ouzkém is such a case.

The overlying Mílina and Klabava formations, characterized by different facies, reflect a gradual deepening of the south-western part of the Prague Basin, its differentiation and changes in character and sources of volcanigenic material. Subaerial rhyolitic and andesitic volcanism decreased and was replaced by submarine basaltic extrusion, resulting in changes in communities. Fossil assemblages display increasing diversity and more complicated distribution patterns. Graptolites and rhynchonelliformean brachiopods dominate, and linguliformean brachiopods remain very abundant. In the latter group, some opportunistic taxa survived in changed conditions and occur commonly in the fossil assemblages while others disappeared or became scarce.

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LATE KATIAN STRATIGRAPHY IN THE PRAGUE BASIN (CZECH REPUBLIC)

P. Kraft¹, J. Bartošová², T. Hroch^{1,3}, L. Koptíková⁴ and J. Frýda³

¹ Institute of Geology and Palaeontology, Faculty of Science, Charles University in Prague, Albertov 6, 128 43 Praha 2, Czech Republic.; kraft@natur.cuni.cz, tom.hroch@gmail.com

² K Petrově komoře 1415/3, 143 00 Praha 4, Czech Republic.; AjaBart@seznam.cz

³ Czech Geological Survey, Klárov 3, 118 21 Praha 1, Czech Republic.; bellerophon@seznam.cz

⁴ Institute of Geology v.v.i., Academy of Sciences of the Czech Republic, Rozvojová 269, Praha 6, Czech Republic.; koptikova@gli.cas.cz

The Králův Dvůr Formation is a distinct unit in the Upper Ordovician of the Prague Basin. It is of late Katian age and begins with a prominent change in sedimentation, traceable throughout the whole "Mediterranean Province". The black shale lithofacies of the underlying Bohdalec Formation was succeeded by fine greenish mudstones with carbonate nodules. Change in lithology is associated with a prominent faunal change. The *Aegiomena-Drabovia* fauna of underlying units was replaced by low diversified associations assigned to the *Foliomena* Fauna, the diversity of which sharply increased in the uppermost part of the formation. This peak of diversity is, however, followed by a dramatic impoverishment in response to global climatic changes.

The biostratigraphic subdivision of the Králův Dvůr is imprecise. Almost the whole thickness is characterized by occurrence of *Normalograptus angustus*. In the uppermost part, there is a very thin level with *Normalograptus ojsuensis*. Auxiliary biostratigraphic subdivision is based on typical faunal assemblages of brachiopods and trilobites. In summary, a detailed stratigraphy based on traditional biostratigraphic measures is unobtainable. Alternative methods such as chemostratigraphical and magnetic susceptibility stratigraphy were tested to provide a better understanding of the unit which was deposited during global climatic changes.

A model study was made on the drill core from Ořech near Prague. The well-studied exposure at Levín near Beroun was used as a reference section. Sedimentological features were documented and described in detail at both sections. Samples for stratigraphic studies were taken from the non-weathered 80 m of the drill core and from the upper 20 m Králův Dvůr Formation at Levín. Values of $\delta^{13}\text{C}_{\text{org}}$ and TOC and rock magnetic susceptibility were measured. The available data show that the section at Ořech represents the lower and middle part of the Kralodvorian strata and that the whole sampled thickness includes the only positive excursion of $\delta^{13}\text{C}_{\text{org}}$. A precise regional stratigraphic subdivision is based on inhomogenities caused by different materials or their sources transported to the basin. These differences affect a curve of the rock magnetic susceptibility and correspond to quantities of the Fe minerals in the basin.

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A GIANT *RUSOPHYCUS* FROM THE MIDDLE ORDOVICIAN OF SIBERIA

V.B. Kushlina¹ and A.V. Dronov²

¹ Boryssiak Paleontological Institute of Russian Academy of Sciences, Profsoouz'naya ul. 123, 117997, Moscow, Russia.
vkush@paleo.ru

² Geological Institute of Russian Academy of Sciences, Pyzhevsky per.7, 119017, Moscow, Russia.
dronov@ginras.ru

Keywords: Trace fossils, *Rusophycus*, Middle Ordovician, Siberia.

INTRODUCTION

The trace fossil *Rusophycus* (Hall, 1852) is an ichnogenus usually attributed to trilobites and formed as a result of the producing organism resting, hunting or seeking protection (Osgood, 1970; Bergström, 1973). It ranges in age from Cambrian to Triassic and has been commonly reported from shallow marine and non-marine predominantly siliciclastic deposits throughout the world (Osgood, 1970; Seilacher, 2007). From the Ordovician of Siberia however it has never been reported before. The purpose of this paper is to record an extremely large specimen of the ichnogenus discovered in the Middle Ordovician Baykit Sandstone of the Siberian Platform. This specimen is the largest *Rusophycus* yet recorded in Russia and can be truly regarded as "giant". The specimen is currently housed in the Paleontological Museum of the Russian Academy of Sciences in Moscow.

LOCATION AND STRATIGRAPHY

The Middle Ordovician Baykit Sandstone of the Siberian Platform is exposed on the south-western margin of the Tungus Basin (Tungus Sineclise) mainly along the Podkamennaya Tunguska River and its tributaries (Fig.1). It constitutes a distinctive sedimentary body extending for over 600km along the river valley. The succession consists of monotonous light grey and yellowish, sometimes pink or reddish coarsely bedded and frequently massive quartz sandstones. At certain levels a well developed cross-stratification and locally, especially near the base, conglomerates are typical. Thickness of the unit varies from 12m to 80-100m (Markov, 1970). The Baykit Sandstone (Baykit Formation) includes deposits of the Vkhorevian and Mukteian regional stages which correspond to the mid-Darriwilian of the Global Scale (Bergström et al., 2009). Baykit Sandstone is bounded at the base and at the top by regional unconformities and represents a complete depositional sequence. The monotonous composition of the sporadically exposed Baykit Sandstone prevents identification of depositional systems tracts. The Baykit depositional sequence

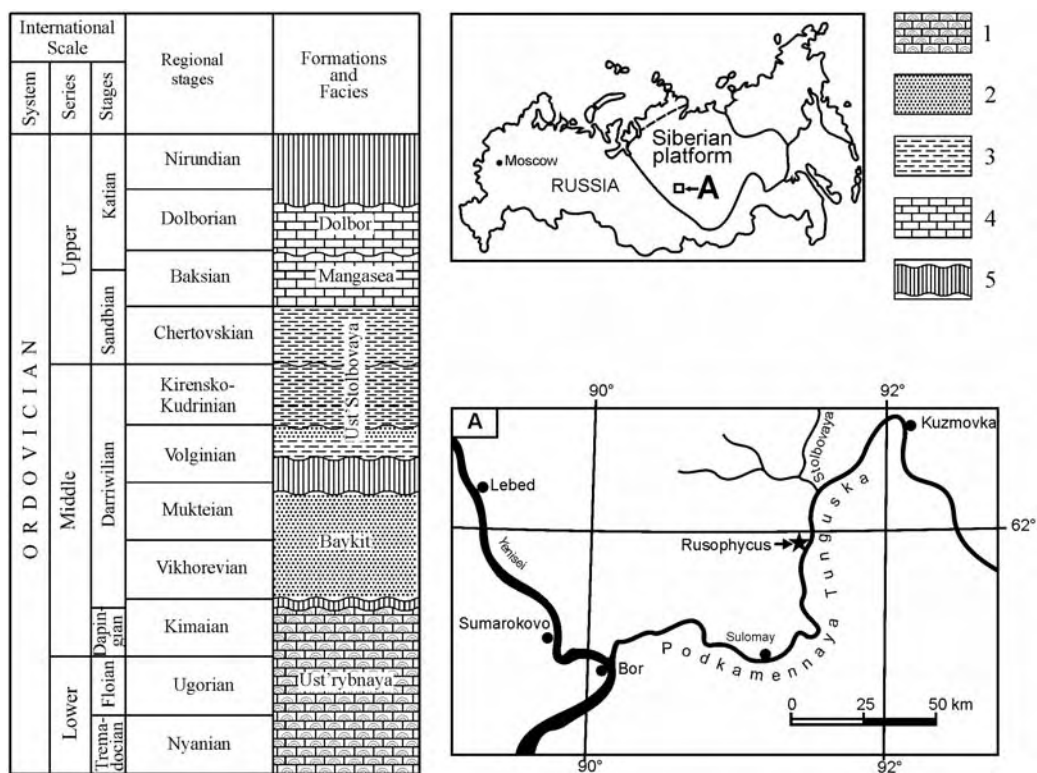


Figure 1. Location of the study area and stratigraphy. 1, warm-water carbonates; 2, quartz sandstones; 3, variegated (green and red) shales; 4, cool-water carbonates; 5, stratigraphic gaps.

roughly correlates with the Kunda depositional sequence of the Russian Platform (Dronov et al., 2009; Kanygin et al., 2010).

The *Rusophycus* trace fossils were found on the basal surface of the overturned fallen blocks of quartz sandstone in the locality on the right bank of the Podkamennaya Tunguska River about 3 km downstream from the mouth of the Stolbovaya River (Fig. 1). Unfortunately the *Rusophycus* specimens were not found *in situ* but were present in talus material located at the base of the outcrop. Judging from lithology the block of rocks fell down approximately from the level about 10 m from the base of the Baykit Sandstone.

DEPOSITIONAL ENVIRONMENT

Markov (1970) studied the Baykit Sandstone almost in all the territory of its exposure concluding that it was shallow marine in origin. Remnants of fauna are scarce but at some levels phosphatic shells of *Angarella* and lingulid brachiopods as well as usually very poorly preserved nautiloids and gastropods could be found. Trace fossils assemblage within the formation includes *Skolithos*, *Rusophycus*, *Planolites* and *Kouphichnium*. The later one seems to demonstrate tracks on subaerial exposed surfaces. Polygonal desiccation cracks are usual on some levels. Cross-stratification is very common. Sometimes herringbone

cross-stratification could be found indicating bidirectional current orientation, although one current direction is usually dominant. Different types of ripples and current lineations provide an evidence of frequently active bottom currents. Integrating palaeontological, ichnological and sedimentological data one can conclude that Baykit Sandstone was formed in near shore tide-dominated environment. It contains intertidal, supratidal and shallow subtidal strata.

DESCRIPTION

The two best preserved specimens are located very close to each other on the sole of an 20 cm thick massive fine-grained quartz sandstone as a large positive feature (positive hyporelief) of broadly convex outline (Fig. 2C). The length of the left one of them (Fig. 2C) is 32 cm while width is 20 cm. The right one

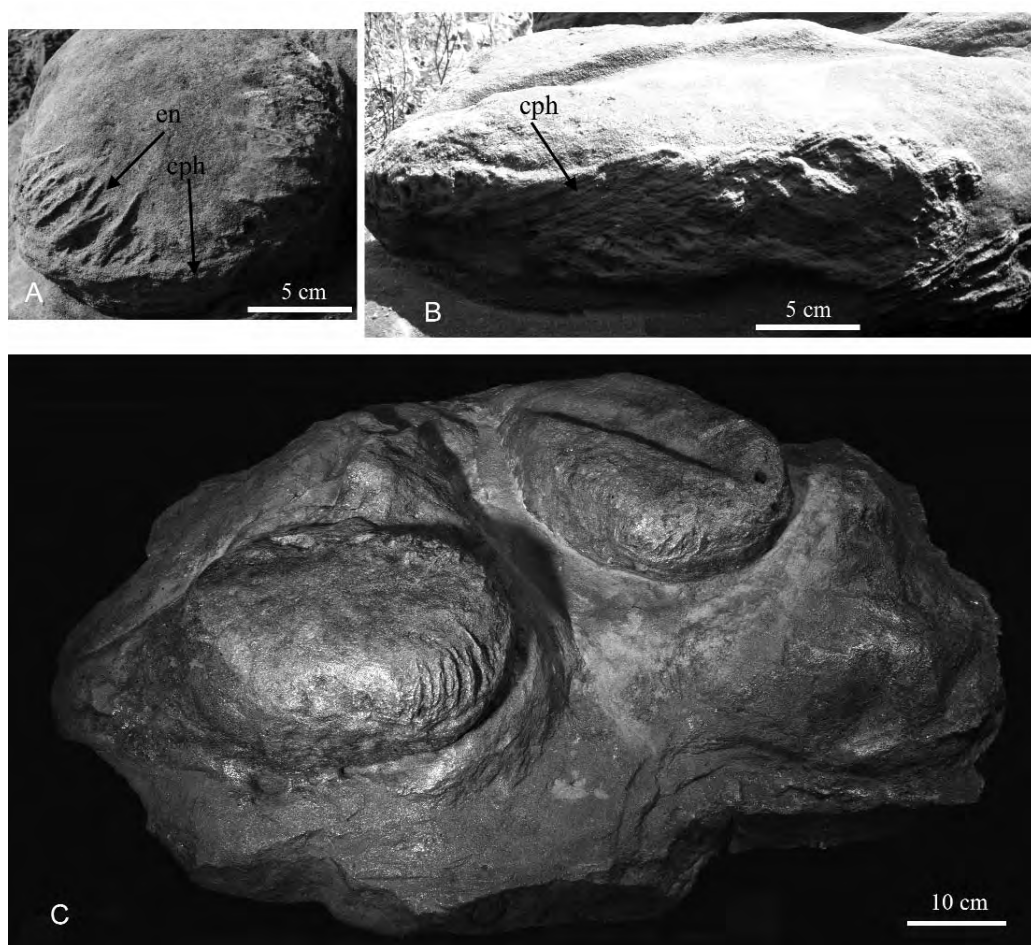


Figure 2. Giant *Rusophycus* from the Middle Ordovician of the Siberian Platform. A, front (anterior) view with 3-clawed endopodite scratch marks (en) and marks made by the edge of the cephalon (cph); B, lateral view of the *Rusophycus* showing marks made by the edge of the cephalon (cph); C, general view on the two closely spaced *Rusophycus* that are interpreted as trilobite nests.

has length 31 cm and width 21 cm. Consequently the trace fossils were with a length:width ratio of approximately 3:2. Depth is about 12 cm. Unlike most *Cruziana* and *Rusophycus* our specimens do not demonstrate clear bilobate structure with a median (groove) axis. Endopodal scratches up to 4-5 cm long are clearly visible on the front side of the burrows. They are represented by bundles of 3-clawed scratches in which the posterior scratches are stronger than the others (Fig. 2A). Distance between lateral claw scratch marks of one bundle is 4-15 mm. Impressions of the cephalon edge are also well preserved on the steep or undercut front slope (Fig. 2A) and on the lateral slopes of the burrow (Fig. 2B). No imprints of segments, pygidium, pleural spines or other parts of trilobite have been detected.

On the sole of the other big (3 x 9 m) fallen block of the quartz sandstone from the same locality 11 more poorly preserved *Rusophycus* burrows have been found. They are slightly different in morphology and represented mainly by bilobate horseshoe-like structures with different orientation (Fig. 3). No scratch marks preserved on these burrows probably due to softer consistency of the sediment at the time of burrowing. Length of the structures varies from 36 to 53 cm with width variations from 19 to 24 cm. Depth does not usually exceed 6 cm. Despite their slightly different morphology length and width of the *Rusophycus* with scratch marks and the ones without them (horseshoe-like modifications) are very close to each other. That means they could be made by the same animals. Some more examples of the horseshoe-like *Rusophycus* can be found on the basal bedding planes of the fallen blocks of quartz sandstones about 2 km upstream the River from the main locality.

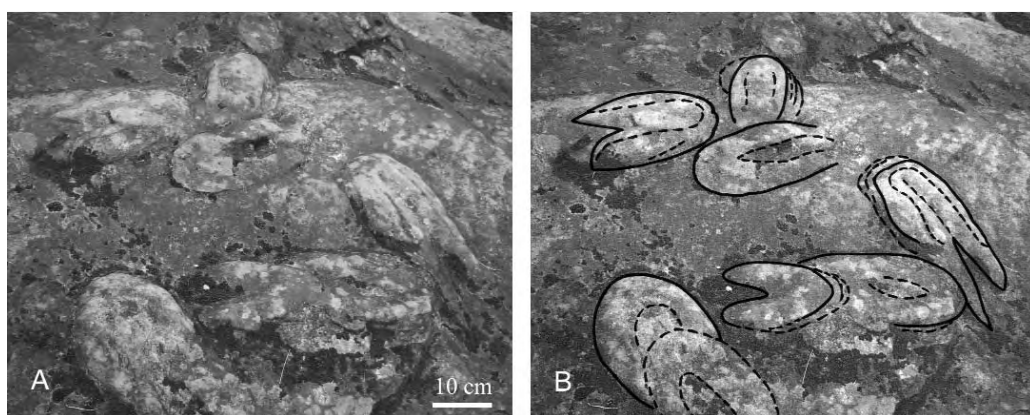


Figure 3. Horseshoe-like specimens *Rusophycus*. A, general view on the sole of an overturned block of quartz sandstone; B, the same photo with *Rusophycus* outlined.

REMARKS

Although Seilacher (1970, 2007) suggested unification under the name *Cruziana* both long furrows (= *Cruziana*) and shorter impressions (= *Rusophycus*) most subsequent authors have preferred to retain the two as distinctive ichnogenera (Crimes et al., 1977). We also incline to the later alternative especially because *Cruziana* is assigned to repichnia (crawling trace) while *Rusophycus* represents cubichnia (resting trace) according to ethological classification of Seilacher (1964) himself. But we do admit that both these traces sometimes could be produced by the same animal and these ichnogenera could include the same ichnospecies. For the distinction of *Rusophycus* and *Cruziana* ichnospecies scratch morphology has become

a most important database (Seilacher, 2007). In our case a well preserved 3-clawed scratch marks demonstrates similarity with *C. semiplicata*, *C. omanica* and *C. petraea*. None of these ichnospecies however has a width of 19 cm. All these traces were made by much smaller animals. Our specimens resemble also representatives of *Cruziana almadensis* group of Seilacher (1970), especially by multiple front lag scratches in the deepest part of the trace. But all species of *C. almadensis* group do not exceed 15 cm in width. Based on this observation we assume that giant *Rusophycus* from the Ordovician of Siberian Platform should be assigned to a new ichnospecies.

DISCUSSION

Despite much palaeontological and biostratigraphical research on the Ordovician of the Siberian Platform (Kanygin et al., 2007 and reference herein) the ichnology has been relatively overlooked. This report therefore represents the first account of large *Rusophycus* from the Ordovician of this region. According to Seilacher (1970) the largest *Cruziana* and *Rusophycus* are typical for the Cambro-Ordovician strata and decrease in size from the Silurian onwards. The majority of previous recordings of large *Rusophycus* confirm this observation. The Ordovician large *Rusophycus* are known from Canada and Australia. Hoffman (1979) has recorded *Rusophycus carleyi* from the Middle Ordovician Chazy Group 31 cm in length and 21 cm in width which is exactly the size of one of our specimens. Draper (1980) has recorded forms resembling both *Cruziana* (= *Rusophycus*) *dilatata* and *C.* (= *Rusophycus*) *carleyi* from the Early Ordovician of Mithaka Formation of the Georgina Basin (Australia) up to 31 cm in length. The largest Silurian recordings are by Osgood (1970) who noted *Rusophycus* up to 25 cm in length from the Clinton Group in Cincinnati and Tansathien and Pickerill (1987) who reported about *Rusophycus* 35 cm in length and 18 cm in width from the Arisaig Group of Nova Scotia.

While there is still controversy as to whether trilobites were responsible for producing all marine *Cruziana* and *Rusophycus* (see Whittington, 1980) it is almost universally accepted that in most cases they were responsible for that. The discovery of trilobites preserved *in situ* within *Rusophycus* (Osgood, 1970; Draper, 1980) together with closely comparable morphological features preserved in some *Rusophycus* when compared to the ventral morphology of trilobites leaves little doubt that trilobites were responsible for their production. Since the *Rusophycus* impressions correspond closely to the dimension of the trilobite which made them one can deduce that large trilobites at least 30 cm in length and 20 cm in width were inhabitants of the Siberian epicontinental seas in the Middle Ordovician. The problem however is that no such a big trilobites have been reported from the Ordovician of Siberian Platform. Judging from the broken fragments the largest exemplars rarely exceeded 20 cm (maximum 24 cm) in length and no more than 10–12 cm in width. These trilobites are from the family Asaphidae (Maksimova, 1962). It is of course dangerous to speculate on producers of trace fossils when no positive evidence is preserved. But asaphid trilobites seem to be a reasonable guess.

According to morphological analysis of trilobite skeletons the largest trilobites most probably were predators (Fortey and Owens, 1999). The *Rusophycus* trace fossil attributed to trilobites usually interpreted as a result of the producing organism resting, hunting or seeking protection (Osgood, 1970; Bergström, 1973). But the Siberian large *Rusophycus* (Fig. 2) seems to represent seem to represent deep resting burrows or “nests”, dug in a slightly head-down position for the reception of eggs. Similar interpretation was suggested by Fenton and Fenton (1937) for the Lower Cambrian burrow “*Cruziana*” (= *Rusophycus*) *jenningsi*. The front (anterior) portions of each of the two traces bears horizontal ridges which seem to

represent impressions made by a cephalon pushed forward and from side to side (Fig. 2 A, B). The regularity, symmetry and depth of the burrows are inconsistent with functions of feeding or hunting. The fact that there are two burrows indicates that they are not accidental. They may be compared with the burrows that modern *Limulus* digs in sand on a beach as receptacles for its eggs. The horseshoe-like morphological type *Rusophycus* (Fig. 3) represents a different function. These trace fossils seem to be dug out by the trilobites seeking shelter from the strong currents during a tide activity.

Rusophycus and *Cruziana* with 3-clawed scratch marks are known from the Upper Cambrian–Upper Ordovician strata (Seilacher, 2007). This seems to be a maximum precision for global *Cruziana* stratigraphy nowadays. Regional *Cruziana* stratigraphy could be more precise but on Siberian Platform we still do not have enough findings of these trace fossils to establish a regional scale. As for trilobite body fossils in the Siberian Ordovician, they are mainly endemics (Maksimova, 1962). Up to now the Lower Paleozoic trilobite burrows have been reported only from the fragments of ancient Gondwana continent (Seilacher, 2007). This fact have been even used for palinspastic purposes in order to identify terranes of Gondwanan origin that happen to dock at other paleocontinents (Seilacher and Crimes, 1969). The giant specimens of *Rusophycus* documented herein suggest that either trilobite burrows existed also outside Gondwana in the Ordovician or the trace makers were not trilobites.

CONCLUSIONS

Giant *Rusophycus* 32 cm in length and 20 cm in width have been found in the Baykit Sandstone (Middle Ordovician) of the Siberian Platform. It is for the first time when the Lower Paleozoic giant *Rusophycus/Cruziana* reported from the outside of the ancient Gondwana continent.

Well preserved 3-clawed scratch marks allow identification of the specimens on the ichnospecies level. Morphology of the burrows, their size and a claw formula suggests that Siberian giant *Rusophycus* should be attributed to a new ichnospecies.

The pair of large *Rusophycus* (Fig. 2) seems to represent burrows dug by trilobites for the reception of eggs in subaerially exposed sandy supratidal environment. The numerous horseshoe-like *Rusophycus* (Fig. 3) are interpreted to be digging out by the trilobites seeking shelter from the strong currents during a tide activity.

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ON THE KATIAN/HIRNANTIAN BOUNDARY. APPLICATION ON THE NORTH-AFRICAN BORDER OF GONDWANA

Ph. Legrand

Tauzia, 216 cours Général de Gaulle, 33170 Gradignan, France.
legrandblain@wanadoo.fr

Keywords : Ordovician, Hirnantian, Graptolites, Glaciation, North African Gondwana.

INTRODUCTION

Since Collomb (1962), forty years ago, suggested a glacial origin for some Ordovician rocks in Libya, the concept of glacial origin has been extended to many Late Ordovician sediments of the north African border of Gondwana (Sougy and Lecorché, 1963; Debyser et al., 1965; Beuf et al., 1966). This shows the significance of the Late Ordovician chronostratigraphy in understanding this glacial episode in these regions, especially the Katian-Hirnantian boundary.

THE BASE OF THE HIRNANTIAN, ITS DEFINITION

"The Global boundary Stratotype Section and Point (GSSP) for the base of the Hirnantian stage is defined at a point 0.39 m below the base of the Kuanyinchiao Bed in the Wangjiawan North Section.... The GSSP level coincides with the first appearance of the graptolite species *Normalograptus extraordinarius* (Sobolevskaya). Secondary markers include the onset of a positive carbon-isotope excursion and a lightly earlier first appearance of *Normalograptus ojsuensis* (Koren' and Mikhailova)" (Chen et al., 2006). Obviously, this definition must accurately reflect the proposal voted on by the International Subcommission in Ordovician Stratigraphy, approved by the International commission in Stratigraphy and ratified by the International Union of Geological Sciences. Consequently, this boundary is defined by the first order marker and less accurately by the secondary markers. In the case of the secondary markers, no reference is made to other groups of fossils or to an important climatic event such as a glacial episode, but only to the onset of a positive carbon-isotope excursion to which it is linked by a way that is none too clear.

USE OF THE DEFINITION OF THE BASE OF THE HIRNANTIAN IN THE WORLD

Chen et al. (2000, 2006) have shown “a global correlation of the Hirnantian Stage and its underlying and overlying strata” in several countries. They have not made clear, however, that correlations are by zones, whose concept changes from school to school and does not necessarily imply the presence of the typical species. Thus, at Dob’s Linn, Williams (1983), though he recognizes a *persculptus* Zone in the upper half of the Hirnantian, does not record *Gl. persculptus* but a smaller form referred to as *Gl. cf. persculptus* in agreement with our own collections. It follows that the base of the Hirnantian is known precisely in fewer areas of the world than the tables of Chen et al. (2000, 2006) would lead to believe (Fig. 1). Moreover, the generic attributions of the more commonly cited graptolite species have changed. Thus, the species *ojsuensis* is attributed to *Diplograptus*, *Glyptograptus* or *Normalograptus*, the species *extraordinarius* to *Climacograptus* or *Normalograptus*, and the species *persculptus* to *Glyptograptus*, *Persculptograptus* or *Normalograptus*. Problems of synonymy may further complicate this problem.

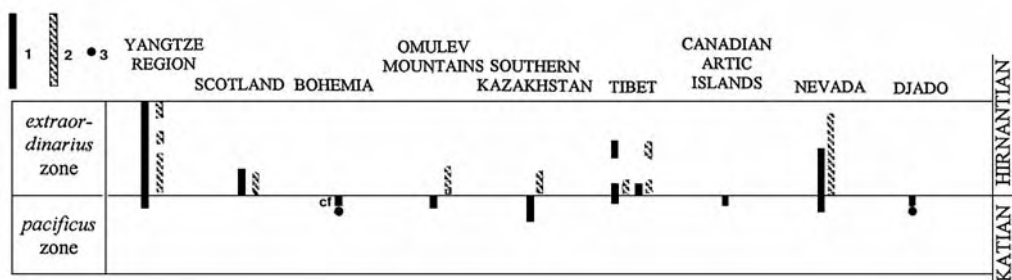


Figure 1. Distribution of *Normalograptus ojsuensis* and *N. extraordinarius* near the boundary Katian/Hirnantian in the most important sections in the world. 1, range of *N. ojsuensis*; 2, range of *N. extraordinarius*; 3, Trinucleidae. Yangtze Region from Chen et al. (2000, 2005, 2006); Scotland from Williams (1982), Melchin et al. (2003); Bohemia from Štorch and Mergl (1989), Štorch (1989), Shaw (1995), Owen (2007); Omulev Mountains from Oradovskaya and Sobolevskaya (in Sokolov et al., 1983), Koren' et al. (1988); Southern Kazakhstan (from Apollonov et al., 1980, 1988); Tibet from Mu and Ni (1983); Canadian Arctic Islands from Melchin and Holden (2006); Nevada from Finney (1999); Djado from Legrand (1993).

Yangtze valley. This is the stratotype area of the Hirnantian. *N. ojsuensis* first appears 4 cm below *N. extraordinarius* (Chen et al., 2006) which is important, considering the small thickness of the Hirnantian strata (some meters only). The species ranges above in association with *N. extraordinarius* and even in the (?)*persculptus* Zone (Chen et al., 2005).

Scotland. Dob’s Linn: According to Melchin et al. (2003), *N. ojsuensis* and *N. extraordinarius* occur together in the *anceps* Band E of Williams (1982) about 3.80 m below the Ordovician-Silurian boundary. Therefore, the *anceps* Band E can be assumed as the base of the Hirnantian stage. However, since barren beds occur below the Band E, it is impossible to say whether *N. ojsuensis* is absent below *N. extraordinarius*, or that the Band E is the base of the Hirnantian. Otherwise, *N. persculptus* would be collected in the *extraordinarius* Band. Thus, the *extraordinarius* Zone would be the lower part of the *persculptus* Zone as it was proposed by Štorch and Loydell (1996).

Bohemia. *Gl. cf. ojsuensis* is recognized in some sections (Karlík, Zadní Třebañ, Liteň, Zličín) of the Upper Králův Dvůr Formation (Štorch, 1989). Just above, the last trinucleid *Marekolithus(?) kosoviensis* (Marek) (Shaw, 1995; Owen, 2007) makes its last appearance. The Rawtheyan (or Katian)-Hirnantian

boundary has been supposed to pass between this level with *Gl. cf. ojsuensis* and the top of the formation (Štorch and Mergl, 1989). If the synonymy "*Gl. persculptus* - "*Gl. bohemicus*" (Štorch and Loydell, 1996) is accepted, this species appears in Bohemia only at the top of the beds attributed to the Hirnantian.

South-western Sardinia. The determination of *N. ojsuensis* (Štorch and Leone, 2003) seems doubtful because the material is tectonically deformed. *N. extraordinarius* has not been found.

Omulev Mountains (northeastern Siberia), Mirny Creek Section. According to Oradovskaya and Sobolevskaya (in Sokolov et al., 1983) and Koren et al. (1988), *Glyptograptus? ojsuensis* has been reported at the top of the *supernus* Zone (*pacificus* Subzone) marking the top of Rawtheyan. *N. extraordinarius* appears 2 m above indicating the base of Hirnantian. It has been written (Chen et al., 2000) that in the neighbouring section (the Ina River) the occurrence of *N. ojsuensis* coincides with the appearance of *N. extraordinarius*. However, in the description of the section, the two species are not reported as occurring at the same level.

Southern Kazakhstan. At its type locality *N. ojsuensis* first appears at the top of the *pacificus* Zone, a little below the appearance of *N. extraordinarius* according to Apollonov et al. (1980, 1988).

Tibet (Xainza area of Xizang). If the synonymy *Diplograptus bohemicus* - *Normalograptus extraordinarius* (Chen et al., 2005) is accepted in this region, *N. ojsuensis* appears at the same level (Riajue section) or a little below (Zhiwazuagu Section) *N. extraordinarius* (Mu and Ni, 1983).

Canadian Arctic Islands. *N. ojsuensis* seems to occur in the Truro Island Cominco T-89 borehole (Melchin and Holden, 2006) very likely at the top of an equivalent of the *pacificus* Zone above the occurrence of "*Climacograptus? pogrebovi*" Koren' and Sobolevskaya. The beds above (without graptolites) would be the equivalent of the *extraordinarius* zone.

Nevada. The Vinini Creek section is one of the most prolific Upper Ordovician section, particularly in the abundance of graptolites and the possibility to study conodonts and chitinozoans associated, only brachiopods are lacking (Finney et al., 1999). Unhappily, there is no description, neither figure of the graptolites listed. In this section, *N. ojsuensis* would appear near the top of the *pacificus* Zone just below the appearance of *N. extraordinarius*. *N. ojsuensis* would still occur above in the lower to middle part of the *extraordinarius* Zone.

To conclude, the sections in the world, where *N. ojsuensis* and *N. extraordinarius* have been collected, are not numerous, especially if the sections where the determinations are still doubtful are not taken into account. Several cases can be met where *N. ojsuensis* appears few meters below or at the same level as *N. extraordinarius* and is still present above or not with *N. extraordinarius*. It seems that these various situations follow a certain geographic distribution. In all cases, it is the appearance of *N. extraordinarius* that allows one to place the base of the Hirnantian according to the definition of the GSSP. The occurrence of *N. ojsuensis* can only suggest that one is just below the equivalent of this GSSP though the lack of *N. extraordinarius* can be accidental. The first appearance of *N. extraordinarius* just above *N. ojsuensis* shows that the base of the Hirnantian has been reached.

APPLICATION TO THE NORTH-AFRICAN BORDER OF GONDWANA

On the north-African border of Gondwana, a "disconformity" separates more often the formations associated to the glacial episode from the formations preceding this episode. More or less important gaps corresponding to the partial or complete disappearance of the Ordovician and reaching even the Precambrian accompanies this "disconformity". The dating of the formations below the "disconformity"

can only give a lower limit on the beginning of this glacial episode, as the reworked fauna in the first glacial sediments. Besides, the dating of these formations is generally bad because fossils are either rare (Taoudeni basin, border of Ahaggar) or, if more abundant, endemic and unreliable (Anti-Atlas, Ougarta Mountains). The first beds associated with the glacial event are generally unfossiliferous and all direct dating is out of the question. The periglacial sediments above them will help date the glacial episode in progress but not its beginning. Gaps are common. Nevertheless, a precise chronology is essential to writing the history of the glacial episode (Figs. 2-3).

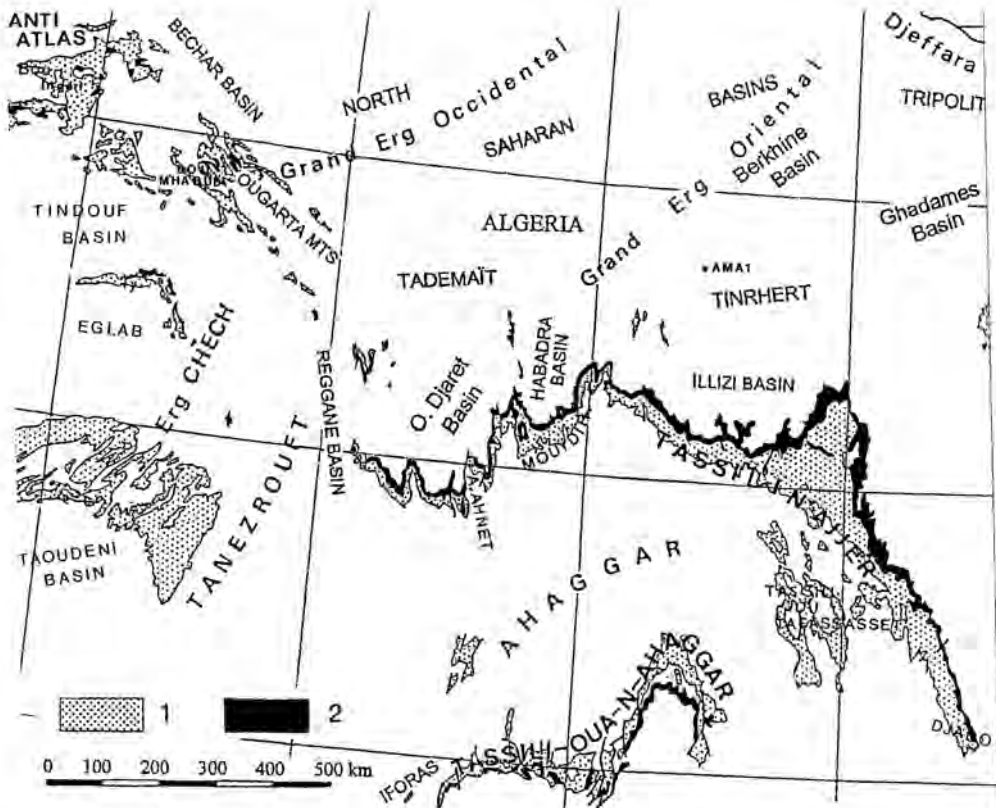


Figure 2. Location map of the North-African countries cited in the text. 1, Cambro-Ordovician outcrops; 2, Silurian outcrops. Redrawn from *The Geological Word Atlas*, Sheet 6.

Anti-Atlas (Morocco). In spite of the basic works of Destombes et al. (1985 and references therein), there is a lack of the age of the beds preserved under the glacial “disconformity” that seems multiform. However, in several points of Central Anti-Atlas, the lower Second Bani Formation shows the at least partially preserved Upper Katian that can be absent in other parts of the Eastern Anti-Atlas (Villas et al., 2006). The lack of graptolites (at least cited) and the lack of fauna in the lower beds of the glacial episode avoid all precision. A particular case would be the Bou Ingarf section (Loi et al., 2010), where the

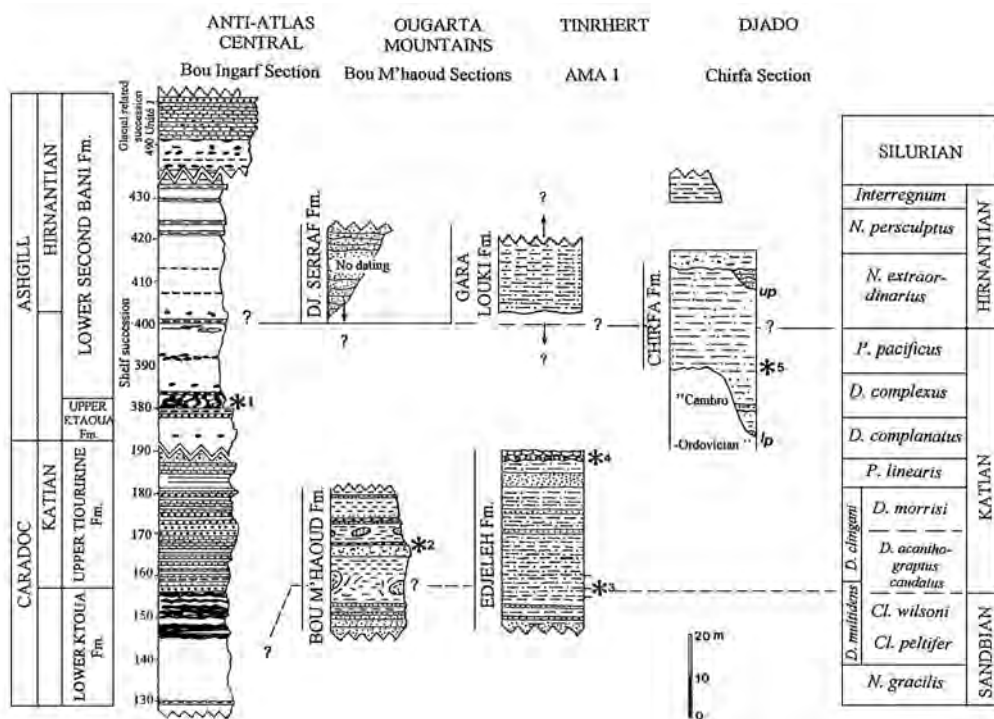


Figure 3. The Katian-Hirnantian boundary on the North-African Border of Gondwana. Fossils *1, *Flexicalymene ouzregui* Destombes, *Brongniartella platynota? marocana* Destombes, *Stenopareia aff. oblita* (Barrande) and trinucleids; *2, *Drabovinella maxima* Mergl; *3, *Diplograptus foliaceus tinrherti* Legrand (in Kichou-Braïk et al., 2006); *4, Bryozoans; *5, *Normalograptus ojsuensis* (Koren' and Mikhailova). Bou Ingarf Section from Loi et al., 2010, simplified; Bou M'haoud Sections from Legrand, 1968-1986 (unpublished). Borehole AMA 1 from Kichou Braïk et al., (2006); Chirfa Section from Denis et al. (2007), simplified: *lp*: lower glacial valley, *up*: upper glacial valley; location of fauna after the Prepa-Petropar reports in Legrand (1993).

sedimentation related to the glacial episode only begins during the Hirnantian and where the Katian-Hirnantian boundary would be stratigraphically below within a classical sedimentary sequence. Its position deduced from the study of chitinozoans (Bourahrouh et al., 2004) must be confirmed.

Ougarta Mountains (Algeria). The highest beds of the Bou M'haoud Formation, below the last Ordovician "disconformity" are at least middle Caradocian in age, i.e. early Katian (Mergl, 1983). The occurrence of younger lower Ashgill beds (i.e. upper Katian) is still uncertain in our opinion. The lower member of the unconformably overlying Djebel Serraf Formation has yielded no fossils.

Illizi Basin, Tinhert Threshold (Algeria). The occurrence of *Diplograptus foliaceus tinrherti* Legrand in several boreholes of Tinhert (Kichou Braïk et al., 2006) just below the "disconformity" marking the base of the Late Ordovician glacial episode in the region, gives a middle Caradocian lower limit (Sandbian-Katian boundary) age for the beginning of the glacial episode. A carbonate level yielding many bryozoans is intercalated between the beds with graptolites and the disconformity. The age according to Spjeldnaes could be late Caradocian or early Ashgillian (i.e. middle Katian) and, therefore, possibly not related to the Boda event.

Central and Eastern Tassili N'Ajjer (Algeria and Western Libya). These countries have yielded the first elements for dating the glaciation (Borocco and Nyssen, 1959). These fauna, supposed no reworked, have suggested a first Caradocian glaciation (Havlíček and Massa, 1973). In fact, these faunas are reworked or intercalated between two disconformities and their age, middle or late Caradocian, is only indicative of a lower limit of the beginning of the glacial episode (Legrand, 1962, 1995 and references therein).

Djado (Niger). In the Chirfa country, the Chirfa Formation lies disconformably on the Ordovician Ajers Sandstone Formation. Its periglacial features were underlined from the origin. Since then, an accurate sedimentologic description has been given (Denis et al., 2007). Near the base of the formation *Glyptograptus* (*Glyptograptus?*) *ojsuensis* has been found (Legrand, 1993). In the absence of *N. extraordinarius*, this species indicates the top of the Katian (*pacificus* Zone), confirmed by a piece of Trinucleidae. This dates the glacial events that marks the base of the Chirfa Formation as latest Katian. Without an accurate biostratigraphic study of the Chirfa Formation, it would not be possible today, to place the base of the Hirnantian that can be a little higher, though the sedimentation of glacio-marine clays offer many examples of unexpected acceleration. Otherwise one may well wonder if this latest Katian glacial pulse is the last of a "Katian" glacial episode as inferred from the study of the lower part of the Bou Ingarf Section (Loi et al., 2010), or the beginning of a "Hirnantian" glacial episode. This last way of seeing things has our preference and accords with a diachronism of the glacial pulses defended for a long time.

CONCLUSIONS

The timing of the Katian/Hirnantian boundary is very important in the North-African border of Gondwana since it is close to the time when the first sign of the end-Ordovician glacial episodes or of one of them if more than one are considered. The position of this limit, and more generally, the biostratigraphy before and after it, require exceptional care because of different interpretations as it has a bearing on the adjustment of the carbon-isotopic curves.

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CONODONT BIOSTRATIGRAPHY FROM SHALLOW WATER UPPER ORDOVICIAN PLATFORM ROCKS IN THE SUBSURFACE OF SOUTH TEXAS

S.A. Leslie¹, J.E. Barrick², J. Mosley³ and S.M. Bergström⁴

¹ Department of Geology and Environmental Science, James Madison University, USA. lesliesa@jmu.edu

² Department of Geosciences, Texas Tech University, Lubbock, Texas, 79409. jim.Barrick@ttu.edu

³ Blinn College, 2423 Blinn Blvd. Bryan, TX 77802. john.mosley@blinn.edu

⁴ School of Earth Sciences, The Ohio State University, Columbus, Ohio, 43210. stig@geology.ohio-state.edu

Keywords: Conodont, Ordovician, Texas.

INTRODUCTION

Little information has been published on conodont biostratigraphy and biofacies from Upper Ordovician platform strata from southwestern Laurentia because these strata lie deeply buried in the subsurface of south Texas. The Magnolia Brown-Bassett #1 well in Terrell County, Texas, penetrated Upper Ordovician shallow water rocks in the Val Verde Basin, a complex fault-bounded subsurface structural province along the continental margin of Laurentia that was overridden by the Ouachita fold-thrust belt to form a deep foreland basin during the Late Carboniferous (Montgomery, 1996). Leslie et al. (2002) reported the occurrences of conodonts from the Brown-Bassett #1 well in an abstract for the ECOS VIII meeting, and Lehnert et al. (2005) referred to some of these conodont data.

Strata of the Ouachita facies exposed in the Marathon fold and thrust belt adjacent to the Val Verde Basin comprise a classic allochthonous deep-water graptolitic section that is possibly in part coeval with the cored interval in the Brown-Basset #1 (Berry 1960). Bergström (1978) described the conodont biostratigraphy of the Middle and Upper Ordovician deep-water succession of the Woods Hollow Shale and Maravillas Formation, and Goldman et al. (1995) revised the graptolite biostratigraphy of part of this same interval. Graptolites place the top of the Woods Hollow in the lower *Climacograptus bicornis* graptolite Zone (= lower-middle *D. foliaceus* Zone) and the base of the Maravillas in the *Dicellograptus gravis* graptolite Zone (= upper *A. manitoulinensis* Zone) (Goldman et al., 1995). This suggests that at least the North American *C. lanceolatus*, *O. ruedemanni*, *C. spiniferus*, and *G. pygmaeus* zones are missing. In terms of conodont biostratigraphy, the top of the Woods Hollow is in the upper *Pygodus anserinus* conodont Zone, and the base of the Maravillas is in the *Amorphognathus ordovicianus* conodont Zone (Bergström, 1978; Goldman et al., 1995). This indicates that the unconformity separating these formations in the Ouachita facies corresponds to at least the *A. tvaerensis* and *A. superbus* conodont zones. In contrast, the Laurentian Brown-Bassett #1 shallow water platform succession represents largely an interval not known from the Marathon deep-water succession. In this short paper we expand on previous brief reports (Leslie, 2002; Lehnert et al., 2005) and provide more detailed conodont biostratigraphic information. The Brown

Basset #1 conodont fauna is particularly interesting in view of the fact that the occurrence of a conodont fauna bearing *Scyphiodus primus* in south Texas is also quite unexpected, because it has not been recorded previously south of the Upper Mississippi Valley of the Midcontinent region.

CONODONT BIOSTRATIGRAPHY OF THE BROWN BASSETT WELL

In the uppermost studied sample of Ordovician rocks penetrated by the Brown-Basset #1 well *Noixodontus* is present at 12,437-12,438 ft. indicating a Hirnantian (late Cincinnati, late Ashgillian) age (Fig. 1A). No conodonts are present in the interval between 12,438 and 12,461 ft. Below this barren interval the conodont fauna between 12,461 and 12,575 ft. consists of *Phragmodus undatus*, *Belodina compressa*, *Drepanostodus suberectus*, *Panderodus gracilis*, *Erismodus radicans*, *Aphelognathus* cf. *A. gigas*, *Plectrodina aculeata*, "*Oistodus*" sp., *Curtognathus* spp., and *Scyphiodus primus* indicating a late Turinian - Chatfieldian (late Sandbian to early Katian; middle to late Mohawkian; middle Caradocian) age. Figure 3 shows selected elements of the fauna and documents the collection level in the core.

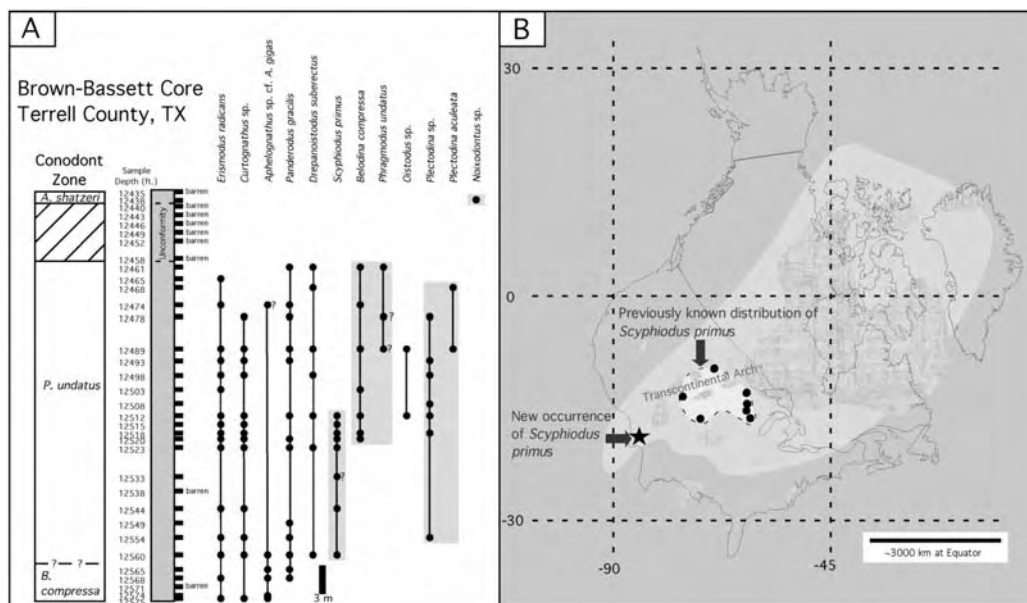


Figure 1. A. Conodont occurrences within the Magnolia Brown-Basset #1 core (Sec. 218, Blk. Y, TC Surv., 1980 feet FNL; 1980 feet FWL, Terrell County, TX). Key species in determining the conodont zone are shaded. B. Distribution of *Scyphiodus primus* (dots) showing the location of the Brown-Basset #1 (star) and how it increases its known geographic distribution.

Leslie (2000) revised the systematics of *S. primus*, and described this species in terms of apparatus-based taxonomy. The presence of *S. primus* is of particular interest because it greatly expands the known geographic range of this characteristic western Midcontinent Fauna species (Fig. 1B). The recovery of conodonts from the Brown Bassett #1 indicates how dependent we often are on sparse subsurface data when trying to establish the complete geographic ranges of species.

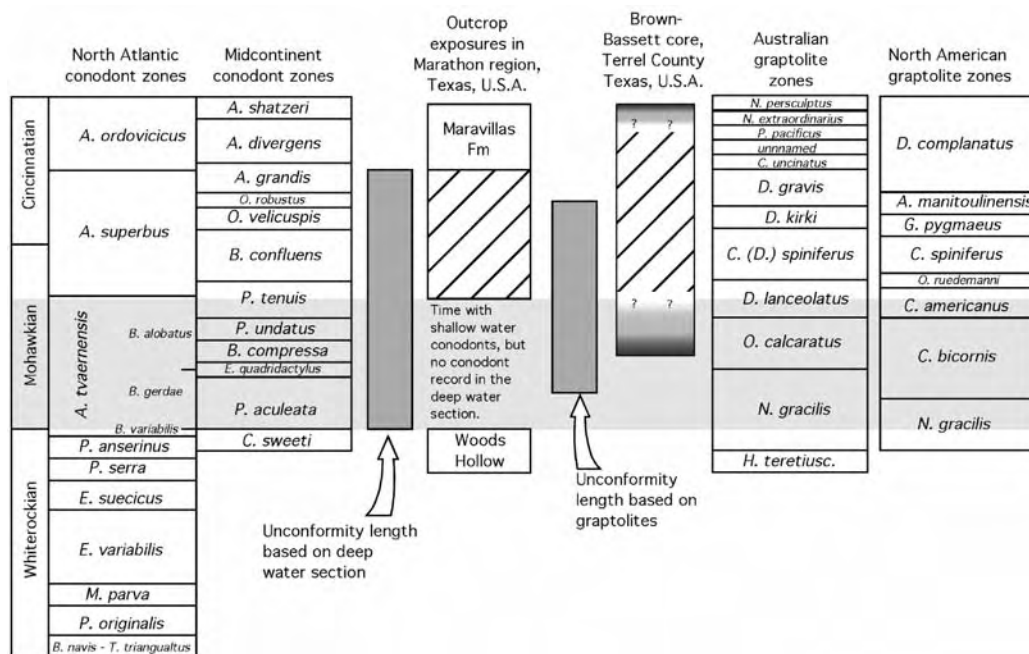


Figure 2. Conodont biostratigraphy of the Marathon region and the Brown-Bassett #1 core (Modified from Lehnert et al., 2005). Note that there is shallow water platform deposition recorded in the Brown-Bassett well corresponding to an unconformity in the deep-water section exposed in the Marathon region.

INTERPRETATION AND CONCLUSIONS

The shallow water Midcontinent conodont fauna recovered from the core is difficult to place within the temporal context of the deeper water Ouachita Facies conodont faunas recovered from surface exposures approximately 150 km away, none of the key species in the core samples being known from the Marathon outcrops. Apparently, at least a part of the section that is missing in the deeper water rock succession (corresponding to the Mohawkian Series) is preserved on the platform. Using the graptolite ages together with the conodonts from the Brown Bassett #1 core, three alternative explanations for this unusual stratigraphical problem are: (1) The upper Woods Hollow is of Mohawkian age and some of the elements in the conodont fauna recovered from the carbonate beds in the Woods Hollow may be a "zombie" fauna that was redeposited in carbonate turbidites as sea-level fell and incision of the platform-slope carbonates began. (2) Tectonic subsidence occurred on the platform at the same time as shallowing in the deep-water environment in an aulacogen or foreland basin. (3) Strata equivalent to those in the *Scyphiodus* succession in the Brown-Bassett #1 well were deposited in the Marathon region but subsequently eroded away during the emersion prior to the deposition of the Maravillas Formation. Explanation 1 is highly unlikely because many of the Woods Hollow samples collected by Bergström (1978) came from lenses and interbeds of fine-grained limestone that showed no turbidite structures. Furthermore, there is excellent agreement between the biostratigraphic evidence provided by conodonts and that from graptolites. Explanation 3 is also discarded because no clasts of strata from the missing interval has been found in the fossiliferous debris flows present in the basal most Maravillas Formation that include Cambrian and Early Ordovician

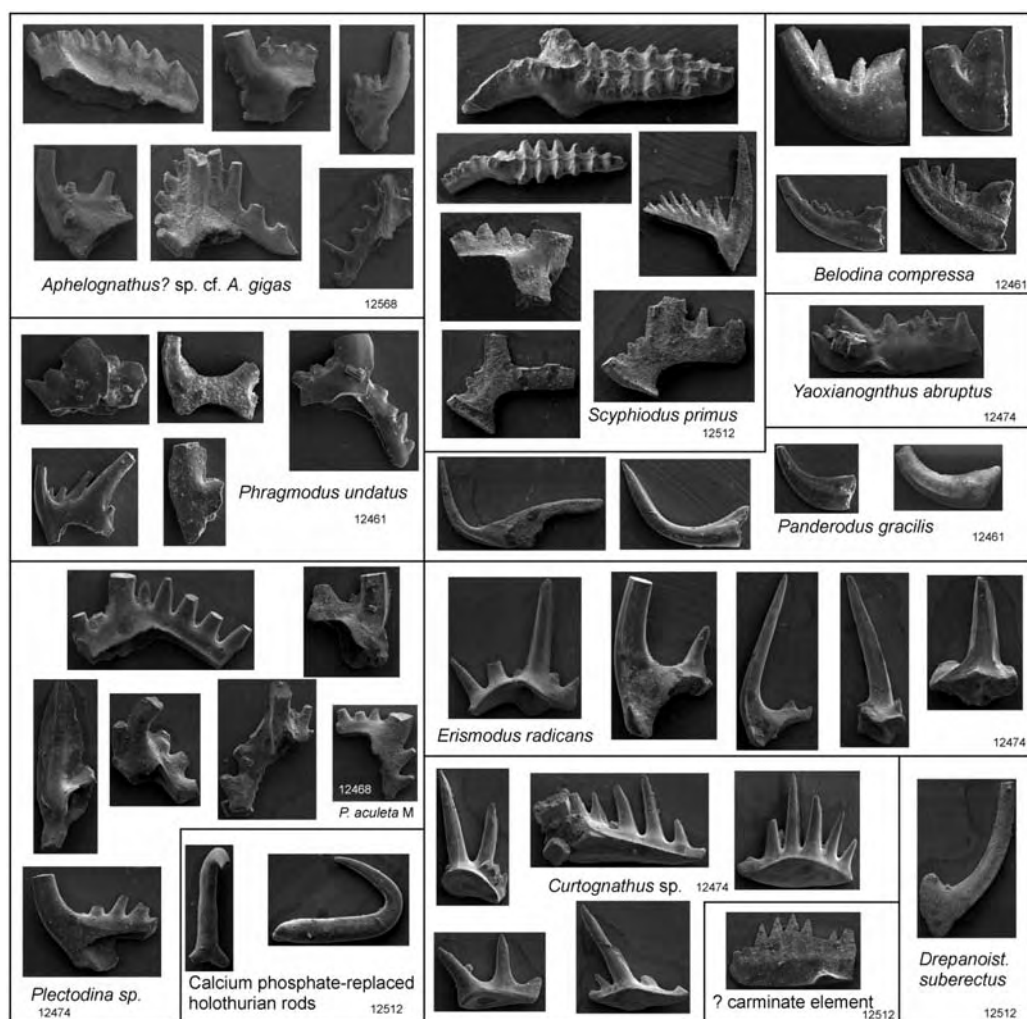


Figure 3. Conodonts from the Brown-Bassett #1 Well. All images are 80X magnification. The 5-digit number is the sample depth in feet below sea level in the core (see Fig. 1A for sample levels).

boulders and cobbles. Hence, we favor explanation 2, which suggests a different pattern of subsidence between the continental margin of Laurentia and the Ouachita facies during the Late Ordovician.

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CONODONT BIOSTRATIGRAPHY AND STABLE ISOTOPE STRATIGRAPHY ACROSS THE ORDOVICIAN KNOX/BEEKMANTOWN UNCONFORMITY IN THE CENTRAL APPALACHIANS

S.A. Leslie¹, M.R. Saltzman², S.M. Bergström², J.E. Repetski³, A. Howard² and A.M. Seward¹

¹ Department of Geology and Environmental Sciences, James Madison University, MSC 6903, Harrisonburg, VA 22807, USA.
lesliesa@jmu.edu, sewardam@dukes.jmu.edu

² School of Earth Sciences, The Ohio State University, Columbus, OH 43210, USA,
saltzman.11@osu.edu, stig@geology.ohio-state.edu

³ U.S. Geol. Survey, Reston, VA 20192, USA. jrepetski@usgs.gov

Keywords: Conodont, $\delta^{13}\text{C}$ excursion, Darriwilian, Ordovician, central Appalachians.

INTRODUCTION

Throughout much of eastern and central North America there is a well-documented hiatus in the Middle Ordovician that is of varying magnitude (e.g. Mussman and Read, 1986). In most areas the middle Darriwilian (uppermost *holodontata*, *polonicus*, lowermost *friendsvillensis* conodont zones) is absent. This missing interval is of particular interest because in areas of the world with a more complete succession, this interval has a middle Darriwilian $\delta^{13}\text{C}$ excursion (MDICE), which is one of the least known of the Ordovician $\delta^{13}\text{C}$ excursions (Ainsaar et al., 2004; Meidla et al., 2004; Martma, 2005; Kaljo et al., 2007, and Bergström et al., 2009). Schmitz et al. (2010) demonstrated that the MDICE is present not only in Baltoscandia but also in the Yangtze Platform of China beginning in the *pseudoplanus* Zone and continuing through the *suecicus* and lower part of the *serra* (*foliaceus*) zones. These geographically widely spaced occurrences suggest that the MDICE, which is the stratigraphically oldest of the named Ordovician $\delta^{13}\text{C}$ excursions, is likely to have a world-wide distribution and to have great potential for local and long-range chemostratigraphic correlations. In this short contribution we continue the process of systematically documenting the Middle Ordovician hiatus throughout North America in terms of conodont biostratigraphy and to add to the conodont biostratigraphic framework an isotope chemostratigraphy to test for the presence of the MDICE event in North America, where this excursion has not been recognized previously. We also recognize the occurrence of a $\delta^{13}\text{C}$ excursion in eastern North America in the same interval as the Turinian (lower Mohawkian, Sandbian) excursion in Nevada shown in Saltzman and Young (2005) and one of the Turinian Mifflin-Grand Detour excursions in Iowa illustrated by Ludvigson et al. (2004) and Bergström et al. (2010). We propose that this $\delta^{13}\text{C}$ excursion be called the Sandbian Isotopic Carbon Excursion (SAICE).

CONODONT BIOSTRATIGRAPHY IN NORTHERN VIRGINIA

The disconformity that cuts out Middle Ordovician rocks in the Appalachian outcrop belt between Tennessee and Pennsylvania represents a hiatus that becomes progressively shorter in duration from south to north (Harris and Repetski, 1982a, b; Repetski and Harris, 1982, 1986; Mussman and Read, 1986; Read and Repetski, in press). The expression of the disconformity fades in northern Virginia near Strasburg (Fig. 1) where conodont biostratigraphy does not resolve any well-defined gap in time; however, there is lithologic evidence of stratigraphic omission in the paleokarst surface preserved at the Beekmantown-New Market contact.

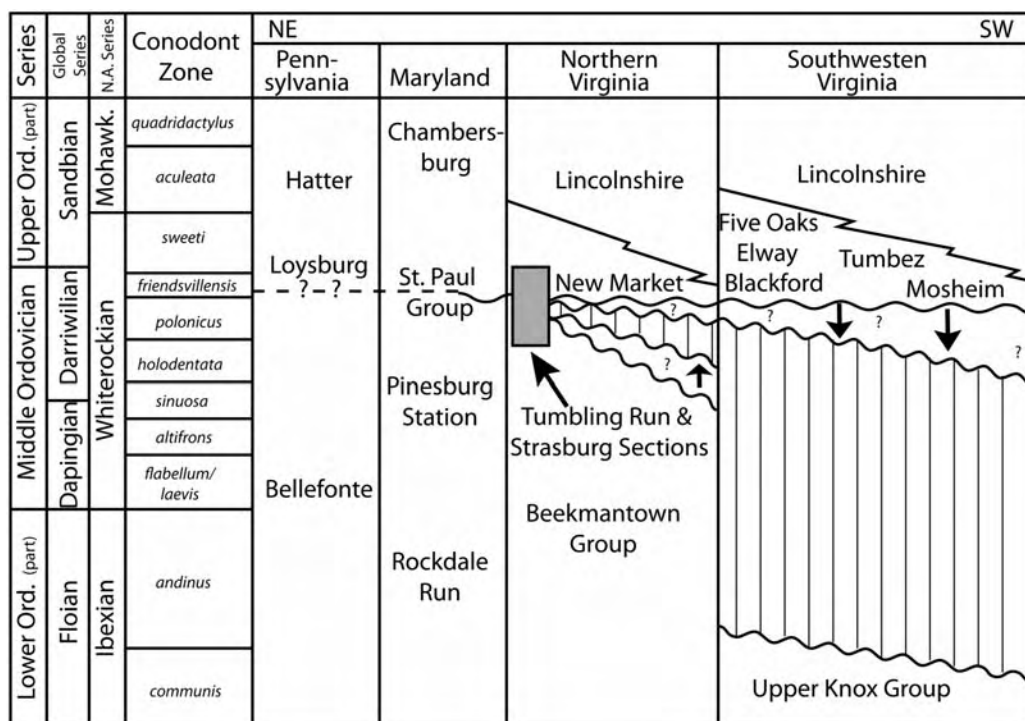


Figure 1. Correlation chart of selected Ordovician units from the central Appalachians. The proposed positions of the Tumbling Run and Strasburg sections are noted by the grey bar in the Northern Virginia column. The duration and relative position of the unconformity shown is modified from Mussman and Read (1986) and Read and Repetski (in press). It is unclear whether the top of the unconformity is nearly isochronous everywhere, as Mussman and Read's (1986) figure implies. Note that evidence for a gap in the rock succession is questioned in Pennsylvania and Maryland.

The conodont faunas from an exposure of this contact on the southbound lane of Interstate Highway 81 (I-81) near Strasburg, Virginia (Fig. 2) contains *Curtognathus* sp. and *Erismodus* sp. from the top of the Beekmantown suggesting a maximum age of the Midcontinent *holodentata* Zone (lower-middle Darrivillian). The lowermost New Market contains a conodont fauna of *Paraprioniodus* sp., *Drepanoistodus suberectus*, *Phragmodus flexuosus*?, *Curtognathus* sp., *Erismodus* sp., and *Panderodus* sp. This fauna corresponds to the Midcontinent *friendsvillensis* Zone based on the presence of this type of conodont fauna containing *C. friendsvillensis* in Maryland (Boger, 1976). The presence of *Appalachignathus*

delicatulus in the upper Beekmantown is interesting, as this suggests that the upper Beekmantown is within the *friendsvillensis* Zone. It is possible that there is stratigraphic leaking associated with the paleokarst and that the *C. friendsvillensis* fauna recovered from the Beekmantown is allochthonous. However, the sample that was collected showed no evidence of this and we think that the possibility of this being a stratigraphic leak is remote.

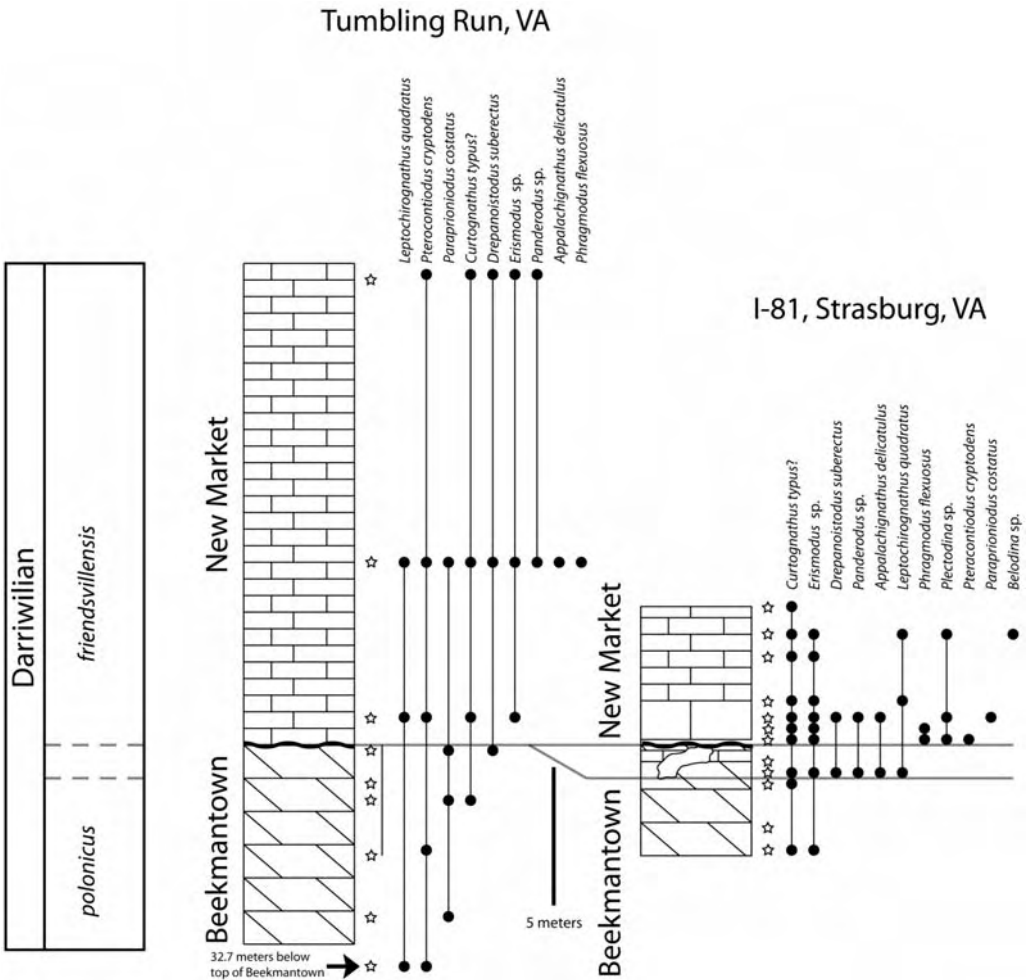


Figure 2. Conodont biostratigraphy of the Tumbling Run section (after DeMoss, 1978) and the exposures along Interstate-81 near Strasburg, VA. The level of the disconformity in the I-81 section is shown in two places in the figure. If it is above the highest dolomite in the I-81 section, then there is a *C. friendsvillensis* fauna present in the top of the Beekmantown suggesting that there is little to no time missing in the section.

The Tumbling Run section conodont biostratigraphy (DeMoss, 1978) is similar to that just described (Fig. 2). The presence of *Leptochirognathus quadratus* 32.7 meters below the top of the Beekmantown at Tumbling Run indicates that the upper Beekmantown at Tumbling Run is within the *polonicus* Zone.

Appalachignathus delicatulus first occurs in the lower New Market at Tumbling Run indicating the *friendsvillensis* Zone. The Tumbling Run section and the I-81 section are approximately 5.5 km apart with no major structure between them. Therefore, we correlate them together lithologically with a high degree of confidence. The correlation of these sections and their conodont faunas suggest that if there is a gap in time at the contact between the Beekmantown and New Market in northern Virginia its duration represents the only a portion of the upper *polonicus* Zone and/or a portion of the lower *friendsvillensis* Zone.

Along the westbound lane of I-70 near Clear Spring, Maryland, the paleokarstic expression of the unconformity is absent but the contact between the Pinesburg Station Dolomite (top Beekmantown) and the St. Paul Group (Row Park and New Market formations) is lithologically sharp. According to et al. (1999, and in press), the *holodentata* Zone corresponds to approximately the upper third of the Pinesburg Station Dolomite, the *polonicus* Zone corresponds to the lower half of the Row Park Limestone (as well as including the uppermost few meters of the Pinesburg Station) that marks a deepening event associated with the basal Tippecanoe transgression. The *friendsvillensis* Zone corresponds to the upper half of the Row Park and possibly into the overlying New Market. Along strike to the north in the New Enterprise Quarry near Roaring Spring, Pennsylvania, the contact between the Bellefonte Dolomite and the overlying Loysburg Formation is gradational with no evidence of a disconformity.

ISOTOPE STRATIGRAPHY

A major shift in Ordovician seawater $^{87/86}\text{Sr}$ corresponds largely to the North Atlantic *serra* to *anserinus* Zones, upper Darriwilian, across the Antelope Valley-Copenhagen Fm. transition in Nevada (Young et al. 2009). A similar shift is documented in the nearly time equivalent Midcontinent *friendsvillensis* to *sweeti* Zones in Oklahoma (McLish-Bromide Formation transition) (Shields et al., 2003). A similar Darriwilian $^{87/86}\text{Sr}$ shift is present in Virginia, Maryland, and Pennsylvania. Young et al. (2009) suggested that the cause of this shift is related to the weathering of non-radiogenic volcanic rocks in the uplifted Taconic mountains. Regardless of the cause of the shift, it apparently has chronostratigraphic significance based on its widespread distribution. As this Sr isotope shift becomes better calibrated to both the North Atlantic and Midcontinent zonal schemes, it may be used to project the conodont biozonation into sections with little or no conodont biostratigraphic control. Similarly, preliminary $\delta^{13}\text{C}$ stratigraphy reveals a significant and continuous shift in the Clear Spring, Maryland section near the Pinesburg Station-St. Paul contact (Fig. 3).

This shift is in the same position as the MDICE, which in Baltoscandia begins just above the interval with *H. holodentata* and ends in a part of the *serra* Zone corresponding to the *friendsvillensis* Zone. This is consistent with a relatively conformable succession, and demonstrates that the MDICE occurs in North America. In addition there is what appears to be the initiation of a second isotope excursion in the Chambersburg. It is older than the position of the GICE. This excursion is in the Sandbian and appears to be in the same interval as the Turinian (lower Mohawkian, Sandbian) excursion in Nevada shown in Saltzman and Young (2005) and one of the Turinian Mifflin-Grand Detour excursions in Iowa illustrated by Ludvigson et al. (2004) and Bergström et al. (2010). In view of its apparently wide geographic distribution (Nevada, Upper Mississippi Valley, Virginia) we feel this excursion deserves a convenient name and propose it be called the Sandbian Isotopic Carbon Excursion (SAICE). Its known range appears to be restricted to the Sandbian Stage and at least in Nevada and Virginia, its peak values are in the *gerdae* Subzone of the *tvaerensis* Zone.

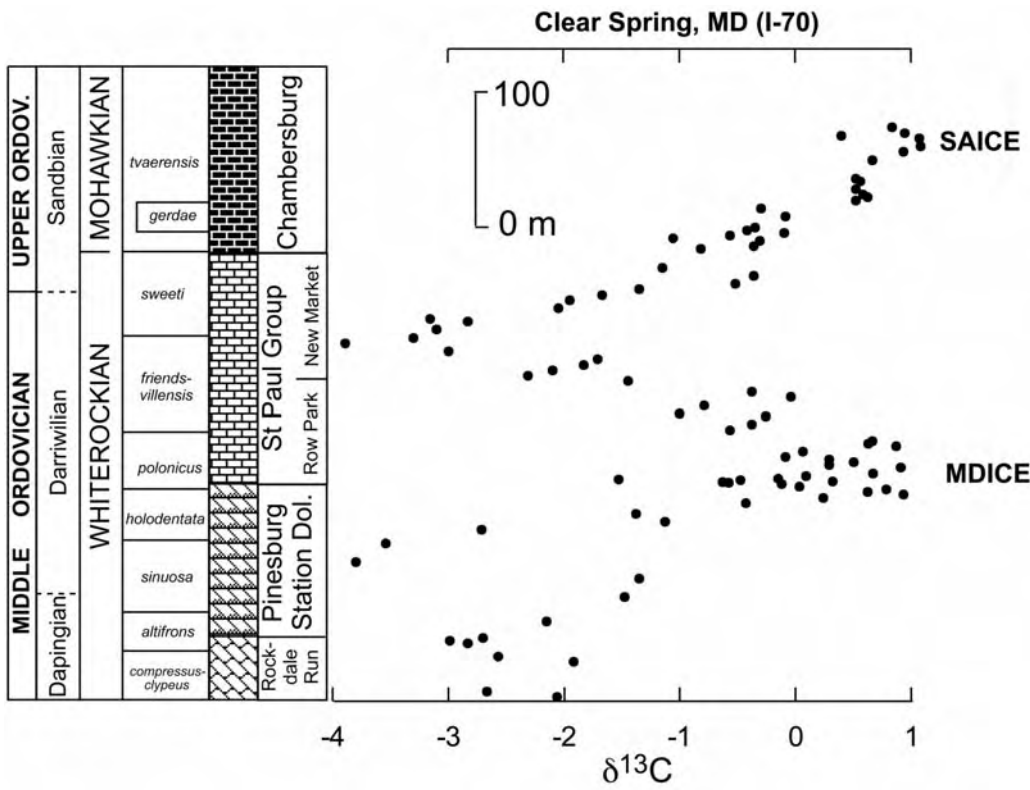


Figure 3. Carbon Isotope stratigraphy from the Clear Spring Maryland Section along Interstate-70. Note the occurrence of a major isotope excursion in the Darriwilian. This is interpreted to be the MDICE. There is another isotope excursion (SAICE) in the Chambersburg.

In the Rocky Gap section in southwestern Virginia, the MDICE may also be present. As in the Clear Spring, Maryland section, $\delta^{13}\text{C}$ values increase continuously from about -2 to -3 ‰ to +1 ‰. The shift begins in the Blackford Formation and peaks in the lower part of the Elway Formation. $^{87}/^{86}\text{Sr}$ values in the upper Blackford and lower Elway, corresponding to the $\delta^{13}\text{C}$ excursion interval, are between about 0.7087 and 0.7085. By calibration to Nevada and Oklahoma, this would be broadly consistent with the *serra* to *anserinus* Zones and upper *polonicus*, *friendsvillensis*, and lower *sweeti* Zones. The lower part of the Blackford does not yield reliable $^{87}/^{86}\text{Sr}$ values (due in part to very low Sr concentrations of the dolomites) and no conodonts have been studied from this section; therefore it is not known whether the *holodentata* and/or lower part of the *polonicus* zones are present, although no conodonts confirming the presence of these zones have been found in basal post-Knox rocks sampled at numerous sections in southwestern Virginia and northeastern Tennessee. The rest of the $\delta^{13}\text{C}$ curve in the Rocky Gap section up through the Five Oaks, Rockdell, Benbolt, Wardell, and Witten formations is generally steady at between -1 and +1‰ with only minor excursions, and $^{87}/^{86}\text{Sr}$ values stay between 0.7084 and 0.7082. The GICE is not present.

In the Union Furnace roadcut and Roaring Spring quarry composite section of central Pennsylvania, the MDICE may be present in the Loysburg and Hatter formations, although it is certainly less clear than in the Clear Spring or Rocky Gap sections. $\delta^{13}\text{C}$ values increase from about -2 to -3 ‰ to just below +1 ‰ in the

Loysburg and Hatter formations. $^{87/86}\text{Sr}$ values in the lower part of the Loysburg are between about 0.7087 and 0.7085. However, the values in the Hatter drop below 0.7084, suggesting an age of the *anserinus* Zone or younger. It is therefore possible that sediment condensation or disconformities have amalgamated $\delta^{13}\text{C}$ excursions. Conodonts in the Loysburg/Hatter suggest a lower *friendsvillensis* Zone age, but there are few identified collections available. The rest of the $\delta^{13}\text{C}$ curve in the Pennsylvania composite section up through the Linden Hall Formation is generally rising from about -1 to +1‰ with minor excursions, and $^{87/86}\text{Sr}$ values fall to as low as 0.7081.

SUMMARY

The Middle Ordovician (Darriwilian) duration of the Knox/Beekmantown unconformity in the Central Appalachians becomes progressively shorter from southwestern Virginia to central Pennsylvania. Conodont biostratigraphy demonstrates that if there is a gap in time at the contact between the Beekmantown and New Market in northern Virginia its duration represents the only a portion of the upper *polonicus* Zone and/or a portion of the lower *friendsvillensis* Zone. The Darriwilian is of particular interest because it contains significant $^{87/86}\text{Sr}$ and $\delta^{13}\text{C}$ changes that, regardless of their cause, may be useful tools in chronostratigraphic correlations.

A Darriwilian $^{87/86}\text{Sr}$ shift is present in Virginia, Maryland, and Pennsylvania that is similar to the shift reported by Young et al. (2009) across the Antelope Valley-Copenhagen Fm. transition in Nevada and by Shields et al. (2003) across the McLish-Bromide Formation transition in Oklahoma. As this Sr isotope shift becomes better calibrated to both the North Atlantic and Midcontinent zonal schemes, it may be used to project the conodont biozonation into sections with little or no conodont biostratigraphic control.

Preliminary $\delta^{13}\text{C}$ stratigraphy reveals a significant and continuous shift in the Clear Spring, Maryland section near the Pinesburg Station-St. Paul contact. This shift is in the same position as the MDICE and demonstrates that the MDICE occurs in North America. There also appears to be the initiation of a Sandbian isotope excursion in the Chambersburg that is in the same position of the Turinian (lower Mohawkian, Sandbian) isotope excursion illustrated by Saltzman and Young (2005) and the Turinian Mifflin and Grand Detour excursions illustrated by Ludvigson et al. (2005). For this excursion we introduce the name the SAICE.

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DEMISE OF EARLY ORDOVICIAN OOLITES IN SOUTH CHINA: EVIDENCE FOR PALEOCEANOGRAPHIC CHANGES BEFORE THE GOBE

J. Liu¹, R. Zhan², X. Dai¹, H. Liao¹, Y. Ezaki³ and N. Adachi¹

¹ Department of Geology, Peking University, Beijing 100871, P.R. China; Key Laboratory of Orogenic Belts and Crustal Evolution (Peking University), Ministry of Education, Beijing 100871, P.R. China. jbliu@pku.edu.cn

² State Key Laboratory of Palaeobiology and Stratigraphy, Nanjing Institute of Geology and Palaeontology, Nanjing 210008, P.R. China.

³ Department of Geosciences, Graduate School of Science, Osaka City University, Osaka 558-8585, Japan.

Keywords: Ooids, GOBE, global cooling, carbonate saturation, Ordovician.

INTRODUCTION

The "Great Ordovician Biodiversification Event" (GOBE) was one of the largest biodiversification events of marine life in Phanerozoic (Webby et al., 2004; Harper 2006). However, the primary causal mechanisms of the GOBE remain the subject of considerable debate (see Servais et al., 2010, and reference therein). Although previous studies expected dichotomically either intrinsic macroevolutionary dynamics (e.g., Sepkoski, 1979) or extrinsic physicochemical changes might have been responsible for the GOBE, more and more researches suggested that both biological and geological factors mutually controlled the onset and consequent development of the GOBE.

The Early to Mid Ordovician has been long regarded as a period of a super-greenhouse world on the basis of modeled atmospheric pCO₂ levels ranging from 14x to 18x PAL (the preindustrial atmospheric level) (Berner, 2006), high sea-level (Haq and Schutter, 2008), as well geochemical proxies (e.g., Shields et al., 2003). In contrast, a gradually global cooling through the Early Ordovician has been recently argued from the oxygen isotope data of conodonts, and consequently considered as the main trigger of the GOBE (Trotter et al., 2008).

Ooids are a kind of coated grains having spherical or ellipsoidal shapes with nuclei encompassed by calcareous cortices. They could be constructed by aragonite and/or magnesium calcite with concentric (tangential) and/or radial microfabrics (Tucker and Wright, 1990). Ooids are commonly regarded as an index of agitated, shallow-water tropical sedimentation. Moreover, the microfabrics, mineralogy and abundance of ooids appear to vary during Phanerozoic, which are well-known proxies for the changes in Phanerozoic seawater chemistry, and paleoclimatic conditions (Sandberg, 1983; Wilkinson et al., 1985; Wilkinson and Given, 1986). Wilkinson et al. (1985) distinguished marked depositional peaks of ooids in the Cambrian, Early Carboniferous, Late Jurassic. However, the temporal changes in abundance of ooids have seldom been documented from the Ordovician of South China.

This study documents the temporal distribution of ooids in the Lower Ordovician of South China. Concerning also other lines of circumstantial evidences, we propose that the decreasing and final demise

of ooid precipitation and the concurrent increasing of skeletal accumulation in Early Ordovician were probably induced by the decreasing carbonate saturation state of sea water, which was caused by a fall of atmospheric $p\text{CO}_2$ as well as the resultant global cooling. The global cooling event just opened a window for metazoan reefal constructors, and still remained the induced calcification of cyanobacteria, which were eventually ceased by the further declining carbonate saturation in early Floian of South China.

GEOLOGICAL SETTINGS

The South China paleoplate comprises mainly the Yangtze Platform, the Jiangnan Slope, and the Zhujiang Basin in most of the Early Palaeozoic (Chen and Rong, 1992). During the Early and Mid Ordovician, South China was situated in a middle latitude (Cocks, 2001), and covered by a vast epeiric sea on the Yangtze Platform. In the Tremadocian, extensive shallow-marine carbonates prevailed in the offshore area, with terrigenous clastic input in the inshore areas close to the western oldlands (Zhan and Jin, 2007). The shallow-marine carbonate deposits were shut down in the early Floian, owing to rapid sea-level rise, and the Middle and Lower Yangtze regions were overwhelmed by deeper water, carbonate-siliciclastic mixed deposits (Liu, 2006).

In this study, the Gudongkou section, located at Gudongkou village, about 2 km north of Xingshan County town, northwestern Hubei Province, South China (for the detailed locality map refer to Liu, 2009), is selected to investigate the temporal distribution of ooids in the Early Ordovician. The Early Ordovician strata at this section include the Nantsinkuan (26 m thick), Fenhsiang (21 m), Hunghuayuan (19 m), and lower Dawan formations (> 6 m) (Fig. 1), overlying conformably on the Cambrian strata, and are assigned to the Tremadocian and early Floian age, based chiefly on conodont biozones (Liao et al., in prep.) (Fig. 1).

The lower Tremadocian Nantsinkuan Formation consists of thin- to medium-bedded peloidal packstone/grainstone and oolitic grainstone, with intercalated beds of flat-pebble conglomerate and small amount of stromatolite, which were primarily deposited in a shallow subtidal environment. The upper Tremadocian Fenhsiang and basal Hunghuayuan formations are characterized by increasing deposition of medium-bedded skeletal packstone/grainstones and greenish gray shales, deposited in deep subtidal and shallow subtidal settings. Flat-pebble conglomerate is abundant in this interval. The lower Floian part of the Hunghuayuan Formation is characterized by thick-bedded skeletal packstone and skeletal peloidal packstone with patched sponge-microbial reefs as well as flat-pebble conglomerate, deposited in a shallow subtidal setting. The lower Floian part of the Dawan Formation is dominated by dark gray shales and thin-bedded nodular skeletal wackestone deposited in deep subtidal to basinal environments, due to a major rise of sea level through the Yangtze Platform (e.g. Liu, 2006).

LITHOLOGICAL CHARACTERISTICS OF OIDS

Most ooids in the Lower Ordovician have well-developed radial microfabrics, composed of calcite, and some ooids have faintly concentric laminae with microcrystal (Fig. 2B). These radial ooids range in diameter from 0.1 to 1.2 mm and have relatively thick cortices up to 0.5 mm. The nuclei of the ooids are mostly micritic peloids and subsidiary skeletal grains. The radial cortices are composed optically of radial-fibrous calcite, and exhibit poorly developed extinction crosses. Single crystals of calcite usually extend to the periphery of the ooid. Between these crystals is microcrystalline calcite exhibiting a vague banding (Fig. 2B).

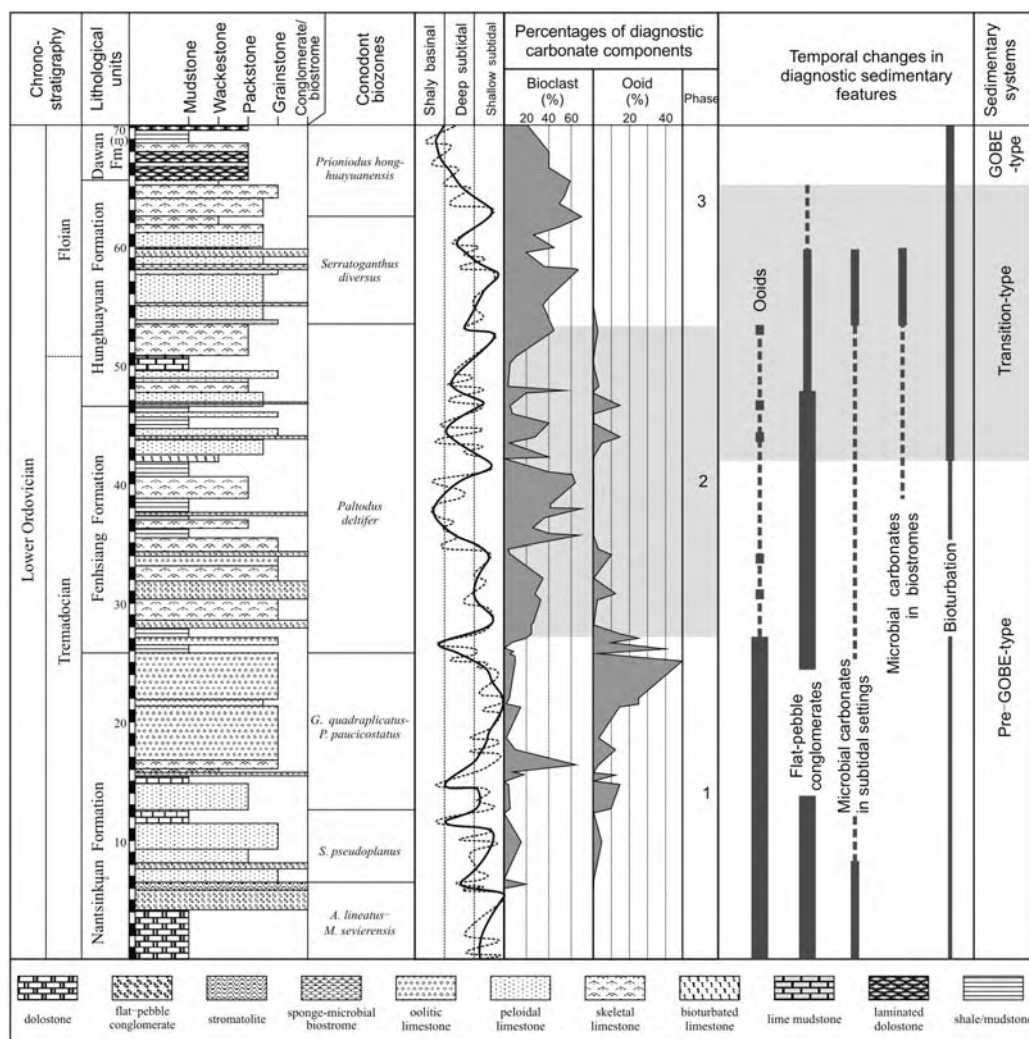


Figure 1. Lithofacies changes, relative sea-level fluctuations, temporal distributions of bioclastics, ooids, and diagnostic sedimentary fabrics of the Lower Ordovician in the Gudongkou section of Xingshan, Hubei Province.

Some other types of ooids occurred at the studied section. For example, superficial ooids with thin, radial cortices occur only in a few stratigraphic intervals, and are accompanied with radial ooids. Composite ooids are much rare, consisting of interior with two or several ooids amalgamated together and a relatively thin cortical layer (Fig. 2B). Sparry radial ooids are composed of neomorphic calcite with relics of originally radial microfabrics retained by the alignment of dark inclusions (Fig. 2C). This kind of ooids is rare, with a limited distribution in the Fenhshiang and the lowermost Hunghuayuan formations.

The ooids mainly occur in thin-bedded to massive packstone and grainstone, but also behave as a minor components of grains with round to irregular peloids and/or bioclastics in wackestone. Ooid grainstone/packstone lithofacies commonly exhibit structureless, graded-, and tabular cross-beddings.

Oolitic intraclasts are common in some oolites and flat-pebble conglomerate units. The ooids and other grains are cemented by equant calcite spar in grainstone.

Well-preserved radial microfabric of the radial ooids indicates an original calcite mineralogy, which resists disruptive structural alteration (Sandberg, 1983; Wilkinson et al., 1985). Sparry radial ooids, although consisting of equant interlocking crystals of calcite, are likely the result of aggrading neomorphic, recrystallization of calcite indicated by the alignment of inclusions. Intensive recrystallization of calcite in the cortices of sparry ooids suggests an early diagenetic stage of dissolution happened after precipitation of ooids.

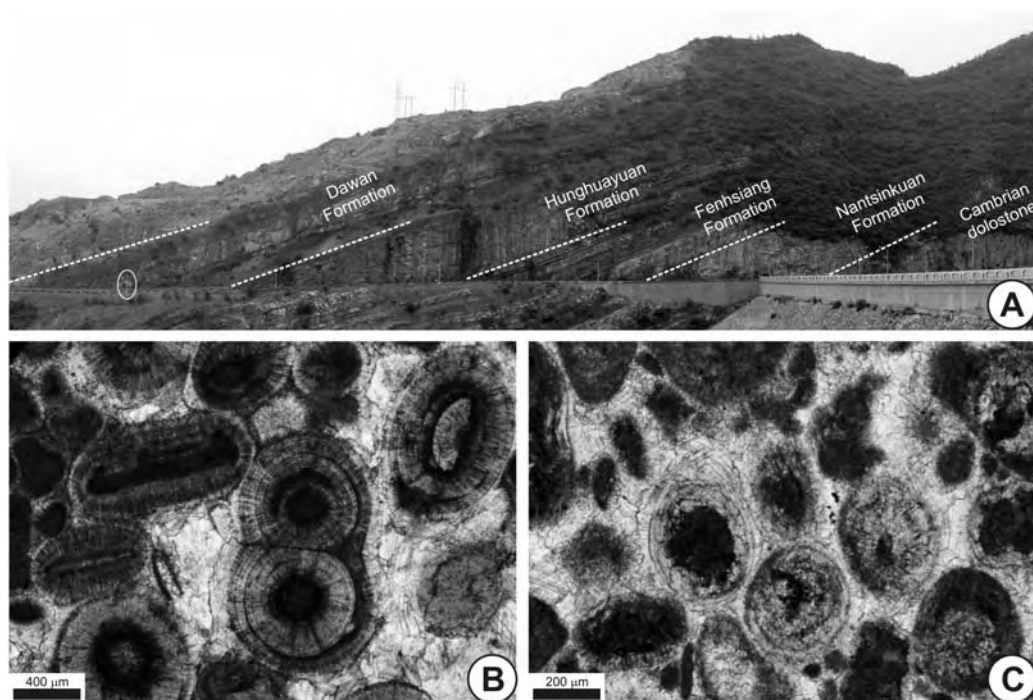


Figure 2. A, Outcrop of the Lower Ordovician in the Gudongkou section, Xingshan County. Telegraph pole (the white bar in ellipse) is about 8 m high. B, Plain light micrograph of radial ooids within oolitic grainstone of the Nantsinkuan Formation. C, Plain light micrograph of sparry radial ooids within oolitic grainstone of the Fenhsiang Formation.

TEMPORAL AND SPATIAL DISTRIBUTIONS OF OIDS

The percentages of ooids and bioclastics are established with comparison charts for visual estimates (Fig. 1). In general, the frequency of ooids increases from the lower Nantsinkuan Formation, and reaches its acme (~40%) at the top of the formation (Phase 1). A sharp decline in deposition of ooids occurs in the lower Fenhsiang Formation; only 5 ooid-containing beds with frequencies less than 25% are distinguished from the overlying Ordovician strata (Phase 2). The formation of ooids in the Lower Ordovician vanishes eventually from the lower Hunghuayuan Formation, and does not reappear in the rest of the Ordovician (Phase 3). Viewed from their microfabrics, most ooids in Phase 1 have well-preserved

radial microfabrics, whereas majority of the ooids in Phase 2 are sparry radial ooids (Figs. 2B, 2C). Composite ooids occur only in Phase 1, and superficial ooids in both Phase 1 and Phase 2.

The temporal distribution of ooids at the Gudongkou section shows a reversed trend for that of bioclastics (Fig. 1). The Nantsinkuan Formation commonly contains rare bioclastics (Phase 1). From the base of the Fenhshi Formation, the frequency of bioclastics increases gradually and shows two peaks in the middle Fenhshi Formation (Phase 2) and the middle and upper Hunghuayuan Formation (Phase 3). Additionally, the construction of sponge-microbial reefs apparently follows the disappearance of ooids formation (Fig. 1).

Such a relationship between the temporal distribution of the ooids, bioclastics, and reefs in the Early Ordovician has been observed from wide area (Liu et al., 2010; Liu, unpublished data) across the Yangtze Platform and beyond. Oolites are abundant in Early Ordovician successions (Opdyke and Wilkinson, 1990). The Tremadocian contains abundant stromatolites, oolitic grainstones, and flat-pebbled conglomerates, which become rare in the following Floian of Siberia (Kanygin et al., 2010; Zhuravlev and Wood, 2009). Accumulation of ooids was commonly associated with microbialites in the Lower Ordovician of the Appalachians (Pope and Read, 1998) and the Michigan Basin of Laurentia (Smith, 1996). James et al. (1989) documented a decline of oolite accumulation and an increase in bioclastic carbonate production in the Middle Ordovician of Laurentia, and only localized occurrences of oolites are recorded from the Katian metazoan-dominated reef on several palaeoplates (Webby, 2002). Evidently, multiple changes in carbonate factories occur successively in Early Ordovician world: (1) a decline of ooids deposition; (2) a coeval increase in skeletal mass, and (3) a subsequent inception of construction of metazoan-microbial reefs. Although these changes occurred at slightly different time according to individual regions, the overall succession of changes and their attributes are strikingly similar with each other.

DISCUSSION

A growing body of field and laboratory evidences suggest that ooids are formed by directly chemical precipitation (Davies et al., 1978; Sandberg, 1983; Morse and Mackenzie, 1990), and microbial activity does not necessarily play an essential role in the ooids formation (Schlager, 2003; Duguid et al., 2010). Modern ooids are commonly distributed in a shallow, warm, high-energy environment above a normal wave base (Hine, 1977). The formation of ooids is controlled by (1) existence of nuclei, (2) supersaturated water for carbonate minerals, (3) agitating bottom water, and (4) minimal amount of grain degradation (Flügel, 2004). Accordingly, a rapid sea-level rise or a drowning of carbonate platform may diminish ooids precipitation. However, the Early Ordovician demise of ooids on Yangtze Platform represented the change in factors apart from the facies shifts or long-term sea-level rise, since agitating setting still prevailed on the carbonate platform even after the demise of the ooids formation (Fig. 1).

Supersaturation state, as well as elevated pH, total alkalinity of sea water, is considered to be essentially controlling factors on the modern ooids production (Rankey and Reeder, 2009). Temporal changes in ooids abundance during Phanerozoic likely reflect the fluctuation of the carbonate saturation state in the ocean (Sandberg, 1983; Wilkinson and Given, 1986). During the Ordovician, a sharp decrease in $p\text{CO}_2$ as calculated from the GEOCARB and MAGic models (Berner, 2006; Guidry et al., 2007), which were approved independently by the oxygen isotope data of conodonts (Trotter et al., 2008). However, Trotter et al. (2008) further asserted that a global cooling and decrease in atmospheric $p\text{CO}_2$ possibly elevated the carbonate saturation of seawater, and may have triggered widespread carbonate

biomineralization and reef growth in the Ordovician. In fact, lower temperature tends to decrease the fluxes of calcium, DIC and total alkalinity from the continents to the ocean, and then decrease carbonate saturation over long time-scales (Riding, 2006); whereas high saturation commonly promotes ordinarily inorganic CaCO_3 precipitation (e.g., ooids, carbonate mud, etc.) (Zeebe and Westbroek, 2003). For example, when atmospheric pCO_2 declined in the Cretaceous, skeletal carbonate factory overwhelmed non-skeletal carbonate factory in neritic areas (Pomar and Hallock, 2008). Therefore, the demise of ooids deposition and increase in skeletal accumulation in the Lower Ordovician of South China and elsewhere were controlled by a probable decline of carbonate saturation induced by the decrease in pCO_2 and resultant global cooling event. The limited distribution of sparry radial ooids just after a decline of ooids precipitation in the late Tremadocian provides another line of evidence for a decline rather than a elevation of carbonate saturation during Early Ordovician.

In the Early Ordovician, cyanobacteria (e.g., *Girvanella*) are well-preserved in microbial sediments in South China (Cao et al., 2009; Adachi et al., 2011) and other palaeoplates (Webby, 2002; Riding, 2005). That is to say, the carbonate saturation of seawater, although beginning its decline to some extent from mid Tremadocian, still remained relatively high to induce the calcification of cyanobacteria and the construction of metazoan-microbial reefs in the late Tremadocian and earliest Floian in South China. From mid Floian, metazoan-microbial reefs disappeared on Yangtze Platform, and bioclastics became the major contributor to the carbonate factory, implying a further decline of carbonate saturation.

Late Tremadocian to mid Floian was a pivotal period for the biodiversification processes in South China. Brachiopod of the Paleozoic Evolutionary Fauna began its radiation from early Floian and exhibited its diversity zenith in mid Floian at generic ranks (Zhan and Harper, 2006). Bulk biodiversity trajectories of trilobites and dichograptid graptolites were also executing radiations at early Floian (Chen et al., 2006; Zhou et al., 2007), which was much earlier than the first global-scale diversification at the beginning of Darriwilian (Zhan and Harper, 2006).

Prior to the rapid biodiversification of the Paleozoic Evolutionary Fauna, the sedimentary systems had started their substantial changes (Liu, 2009). The transition-type sedimentary systems were developed in the late Tremadocian to earliest Floian, exhibiting a decrease in subtidal microbialite and flat-pebble conglomerate, and an increase in the extent of bioturbation as compared with pre-GOBE sedimentary systems (Fig. 1). In addition, a replacement of the Cambrian-type shellbeds by Paleozoic-type shellbeds occurred while the transition-type sedimentary systems were developed (Liu et al., 2010). All these changes happened prior to the rapid diversity of the Paleozoic Evolutionary Fauna in South China. According to the temporal distribution of ooids in Early Ordovician of South China, the development of the transition-type sedimentary systems were preceded by the decline of ooids precipitation in mid Tremadocian.

CONCLUSIONS

This study has documented the temporal distribution of ooids in the Lower Ordovician of South China, and recognized the decline and demise of ooids precipitation in mid and late Tremadocian respectively. Concerning also other lines of evidence, we found multiple changes in carbonate factories during the declining process of ooids: (1) a decline of ooids deposition; (2) a coeval increase in skeletal mass, and (3) a subsequent inception of the construction of metazoan-microbial reefs. We propose that the demise of ooids precipitation and the concurrent rise of skeletal accumulation were probably induced by the decrease in the carbonate saturation of sea water, chiefly due to a fall of atmospheric pCO_2 as well as the resultant

global cooling. This global cooling event just opened a window for the bloom of metazoan-reefal constructors until mid Floian in South China. The onset of the Ordovician radiation in South China might be due to the decrease in carbonate saturation of neritic seawater, subsequent turnover of carbonate factories, and mutual interactions between physical and biological processes under a long-term global cooling condition.

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NEW INSIGHTS ON THE HIRNANTIAN PALYNOSTRATIGRAPHY OF THE RIO CEIRA SECTION, BUÇACO, PORTUGAL

G. Lopes¹, N. Vaz², A.J.D. Sequeira³, J.M. Piçarra⁴, P. Fernandes¹ and Z. Pereira⁵

¹ CIMA, Algarve University, Campus de Gambelas, 8005-139 Faro, Portugal. gilda.lopes@lneg.pt, pfernandes@ualg.pt

² Trás-os-Montes e Alto Douro University, Ap. 1013, 5001-801 Vila Real, Portugal. nunovaz@utad.pt

³ LNEG Laboratório Nacional de Energia e Geologia, Coimbra, Portugal. asequeira@dct.uc.pt

⁴ LNEG Laboratório Nacional de Energia e Geologia, Rua Frei Amador Arrais, Ap.104, 7801-902 Beja, Portugal, jose.picarra@lneg.pt

⁵ LNEG Laboratório Nacional de Energia e Geologia, Rua da Amieira, Ap. 1089, 4466-901 S. Mamede Infesta, Portugal. zelia.pereira@lneg.pt

Keywords: Ordovician, cryptospores, acritarchs, Hirnantian, Rio Ceira Section, Buçaco Syncline.

INTRODUCTION

The Lower Palaeozoic successions of Portugal are well represented in the Central Iberian Zone (CIZ), one of the main tectonostratigraphic domains of the Iberian Massif.

Located in the CIZ, the Buçaco Syncline presents one of its most complete Palaeozoic sequences. With a NW-SE orientation, the Lower Palaeozoic lithostratigraphic succession, includes several Ordovician units that ranges from the Lower Ordovician (Tremadocian stage) to the Upper Ordovician (Hirnantian stage) and are unconformably overlain by the Silurian Sazes Formation, at least of Wenlock and Ludlow epochs, in the Rio Ceira Section (Fig.1).

This sequence is also well known by its rich palaeontological content in macrofauna (e.g. trilobites, brizoans, echinoderms, ostracods, brachiopods, graptolites) and microfauna (conodonts, chitinozoans, acritarchs) (Delgado, 1908; Henry and Thadeu, 1971; Mitchell, 1974; Henry et al., 1974, 1976; Elaouad-Debbaj, 1978; Henry, 1980; Paris, 1979, 1981; Romano, 1982; Romano et al., 1986; Young, 1985, 1987, 1988, 1989).

The aim of the present work is to present the new palynostratigraphic (cryptospores, acritarchs and chitinozoans) results obtained in the Ribeira do Braçal, Ribeira Cimeira and Casal Carvalhal Formations (Young, 1985, 1988) that were collected along the Rio Ceira Section, in the Buçaco Syncline. The results obtained in the Ribeira do Braçal Formation confirms the Hirnantian age attributed to this unit, based on the macrofauna content recovered at the base of this formation (Young, 1985, 1987). For the first time, Ordovician cryptospores are described in this sequence. This study will also provide information to support the undergoing surveying mapping project undertaken by the LNEG (Portuguese Geological Survey) (Sequeira, in prep.).

Future work in the north part of the syncline, to correlate the data recently obtained with the Rio Ceira Group outcropping at the SW limb, is being planned.

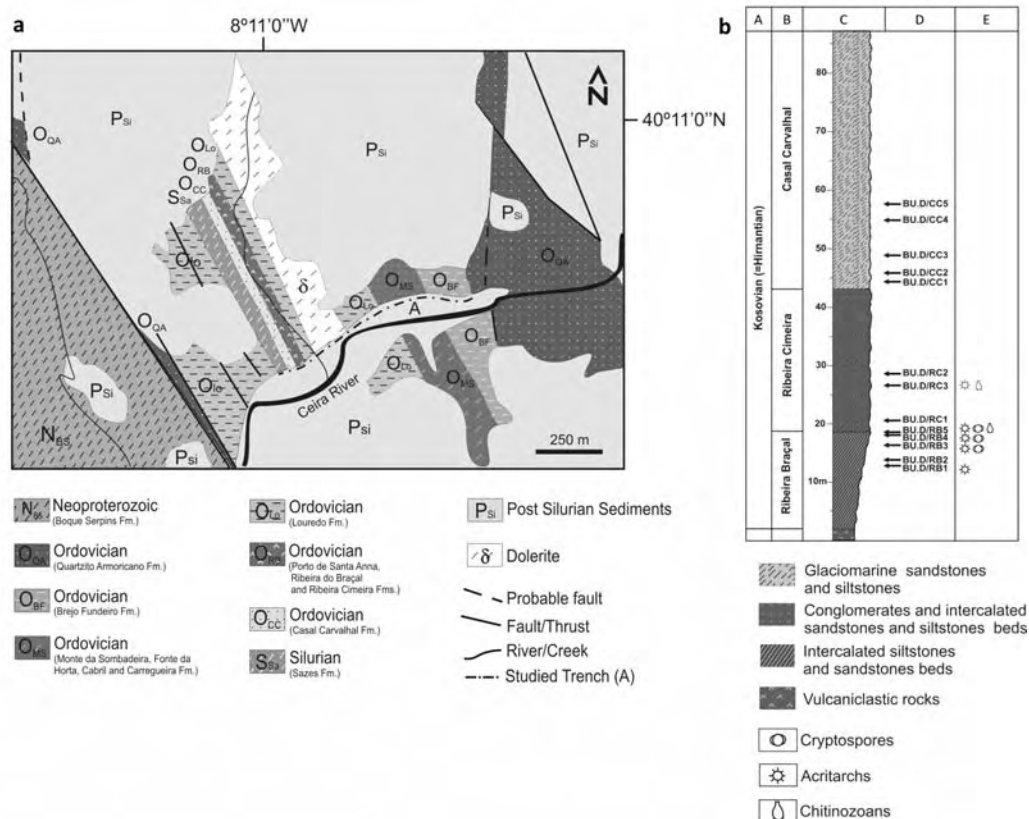


Figure 1. a, Simplified geological sketch map of the Buçaco Region, Rio Ceira Section (adapted from Soares et al., 2007), indicating the position of the studied trench (A). b, Stratigraphic log of the studied formations in the Rio Ceira Section with sample positions, modified from Young (1988) (A, Chronostratigraphy; B, Formations; C, Lithology; D, Samples; E, Fossils).

GEOLOGICAL SETTING

In the south area of the syncline, the Ordovician sedimentary sequence includes, from the base to the top (Mitchell, 1974; Young 1985, 1988; Oliveira et al., 1992; Sá, 2005):

- Armorican Quartzite Formation, with a sedimentary record that indicates a transgressive episode registered by alternations of quartzites, siltstones and pelitic beds above basal conglomerates, of Arenigian age, based in palaeontological studies (Delgado, 1908; Paris, 1981; Romano et al., 1986, Paris, 1990).
- Cákemes Group that includes the Brejo Fundero, Monte da Sombadeira, Fonte da Horta, Cabril and Carregueira Formations. The pelitic shales of the Brejo Fundero, Fonte da Horta and Carregueira Formations are intercalated with the sandstones of the Monte da Sombadeira and Cabril Formations that reflects two detrital episodes with tempestitic facies (hummocky cross-stratification) (Young, 1985, 1988; Soares et al., 2007). The age of this group ranges from the Oretanian to the early Berounian based in biostratigraphical studies (Delgado, 1908; Mitchell, 1974; Henry et al., 1976;

- Paris, 1981; Young, 1985, 1988; Brenchley et al., 1986).
- Sanguinheira Group, which starts with the Louredo Formation, composed of dominant pelitic succession that alternate with sandstones, of an early to mid Berounian age, based in several fossils groups (bivalves, ostracods, brachiopods and trilobites) (Young, 1985, 1988; Soares et al., 2007). At the base of Louredo Formation, an oolitic ironstone bed occurs, the Favaçal Bed, rich in microfossils (chitinozoans) that give an early mid Berounian age (Henry and Thadeu, 1971; Henry et al., 1976; Paris, 1979, 1981).
 - Venda Nova Group (Young, 1985, 1988; Soares et al., 2007) includes the volcano sedimentary unit of Porto de Santa Anna Formation with a thin bed of oolitic ironstone at the base. Biostratigraphic data present (Young, 1985, 1987) indicates late Berounian and Kralodvorian ages.
 - Rio Ceira Group, that includes the Ribeira do Braçal and Ribeira Cimeira Formations (Young, 1985, 1988; Soares et al., 2007). The Ribeira do Braçal Formation shows a regressive sequence composed of alternated siltstones and sandstones and is dated as Kosovian (= Hirnantian) based in biostratigraphical content (Young, 1985, 1987). The Ribeira Cimeira Formation rests unconformably over the Ribeira Braçal Formation and consists of finning-upward sequences of conglomerates, sandstones and siltstones (Young, 1985, 1988; Soares et al., 2007).
 - Casal Carvalhal Formation, characterized of sandstones and siltstones that are interpreted as glaciomarine sediments with a Kosovian (= Hirnantian) age. A Silurian age is not excluded for the top of this unit (Young, 1985, 1988; Soares et al., 2007).

PALYNOSTRATIGRAPHY

The Rio Ceira Section is located in the south area of the Buçaco Syncline, along the Ceira River. This section exposes, from northeast to southwest, a stratigraphic sequence that ranges from the Ordovician to the Silurian. The section was logged and all samples were processed for palynological studies. Standard palynological laboratory procedures using fluoridric and chloridric acids were employed in the extraction and concentration of the palynomorphs from the host sediments (Wood et al., 1996). The slides were examined with transmitted light, per BX40 Olympus microscopes equipped with a digital camera. All samples, residues, and slides are stored in the LNEG-LGM (Geological Survey of Portugal) at S. Mamede Infesta, Portugal. The acritarch biostratigraphic scheme used for the Ordovician-Silurian boundary follows Vecoli (2008). For the cryptospore biostratigraphy it is followed Burgess (1991) and Rubinstein and Vaccari (2004).

Forty two samples were collected in the Ordovician and Silurian sequence of the Rio Ceira Section and thirteen of them were collected from the Ribeira do Braçal, Ribeira Cimeira and Casal Carvalhal Formations, all of them attributed to the Hirnantian. In this section, the Ribeira do Braçal Formation contacts directly over the Porto de Santa Anna Formation and is discordant with the Ribeira Cimeira Formation at the top. The Casal Carvalhal Formation is continuous with the Ribeira Cimeira Formation and discordant with the Sazes Formation at the top.

From the five samples collected from the Ribeira do Braçal Formation, four of them were positive and yielded moderately to well preserved palynomorphs (cryptospores, acritarchs and chitinozoans) assigned to the Hirnantian age. The assemblage recovered, presents acritarchs: *Leiofusa* sp., *Leiosphaeridia* sp., *Lophosphaeridium* sp., *Multiplicisphaeridium* sp., *Veryhachium* spp., *Villosacapsula?* *setosapellicula*, *Visbysphaera* sp. (Pl. 1, figs. 7-16).

The acritarch assemblage is rather impoverished in species diversity being dominated by the veryachid forms (*Veryhachium* spp., *Villosacapsula?* *setosapellicula*), a typical feature at this age. As Vecoli (2008) refers, across the Ordovician-Silurian boundary, the presence of members of the *Veryhachium* and *Multiplicisphaeridium* complexes, as well as of netromorph acritarchs (*Leiofusa* spp.), is very common. The present assemblage includes large stratigraphic range species, Ordovician to the lower Silurian in age, *Veryachium* spp. and *Leiofusa* sp. As well, includes a latest Hirnantian specie, *Villosacapsula?* *setosapellicula*, that disappear in the Hirnantian/Rhuddanian boundary. The first occurrence of *Visbysphaera* sp., at the mid upper Hirnantian level (*Normalograptus persculptus* Graptolite Biozone; *Spinachitina oulebsiri* Chitinozoan Zone), allows constrain the age.

It was also recovered from the samples, specimens of chitinozoans moderately preserved: *?Conochitina* sp.. Completes the assemblage and presented here for the first time, are the cryptospores *?Rugosphaera* sp., *Tetrahedraletes medinensis*, *Velatitetras retimembrana*, and *Dyadospora murusattenuata*. (Pl.1, figs. 1-6) These species have a very large stratigraphic range from the Upper Ordovician to the lowermost Devonian (Burgess, 1991; Rubinstein and Vaccari, 2004) with a limited use at the Ordovician-Silurian boundary, as it has been also described by Rubinstein and Vaccari (2004).

From the Ribeira Cimeira Formation one of the three samples collected was positive, yielding moderately preserved acritarchs, that includes *Veryhachium ?trispinosum*, and a single chitinozoan specimen (*?Conochitina* sp.). No age determination was possible.

From the Casal Carvalhal Formation five samples were collected but they were barren in palynomorphs.

These data presented confirms previous determinations based on the Hirnantian brachiopod fauna from Ribeira do Braçal Formation (*Orderleyella?* cf. *bouceki*, *Plectothyrella* sp. and *Bracteoleptaena* cf. *polonica*: Young, 1985, 1987)).

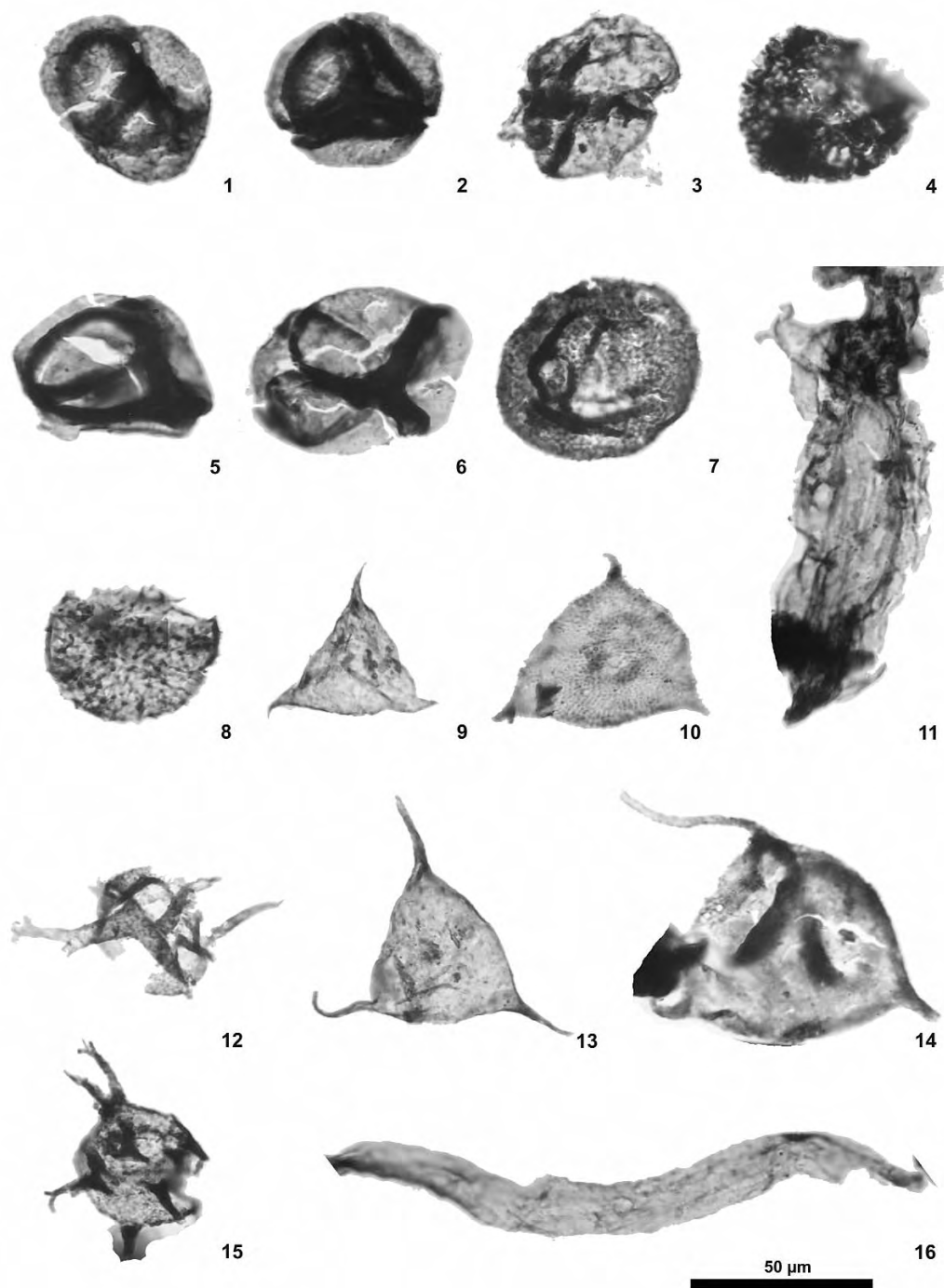
The cryptospores and acritarchs identified from these formations are the first insights to characterize the a Hirnantian microfauna in Portugal.

CONCLUSIONS

The following conclusions were reached from this study:

- This preliminary results obtained in the Ribeira do Braçal Formation indicates an acritarch assemblage assigned to the mid-late Hirnantian. These results confirm previous age determination based on macrofauna.
- The recovered assemblage could provide information to better understand and establish a more detailed acritarch biozonation for the Upper Ordovician-Lower Silurian interval.

Plate 1. Hirnantian cryptospores (1-6) and acritarchs (7-16) from the Ribeira do Braçal Formation, Buçaco syncline, central Portugal. Each specimen is referenced by sample number, slide number and microscopic coordinates. 1-2, *Velatitetras retimembrana* (Miller and Eames) Steemans, Le Hérisse and Bozdogan, 1996. Sample BU.D/RB5, slide 1(1), 1389–155; slide 1(2), 1164–109; 3, *Morphon Dyadospora murusattenuata* Strother and Traverse, 1979. Sample BU.D/RB5, slide 1(1), 1345–107; 4, *?Rugosphaera* sp. Sample BU.D/RB3, slide 1(1), 1334–205; 5-6, *Tetrahedraletes medinensis* (Strother and Traverse) emend. Wellman and Richardson, 1993. Sample BU.D/RB5, slide 1(1), 1087–149; sample BU.D/RB3, slide 1(1), 1179–226; 7, *Lophosphaeridium* sp. Sample BU.D/RB1, slide 2(1), 1233–181; 8, *Visbysphaera* sp. Sample BU.D/RB4, slide 1(2), 1350–132; 9, *Veryachium?* *reductum* (Deunff) Jekhowsky, 1961. Sample BU.D/RB4, slide 1(2), 1379–135; 10, *Villosacapsula?* *setosapellicula* (Loeblich) Loeblich and Tappan 1976. Sample BU.D/RB5, slide 1(2), 1409–116; 11, 16, *Leiofusa* sp.; samples BU.D/RB5, slide 1(1), 1373–116 and BU.D/RB5, slide 1(1), 1145–202; 12, 15, *Multiplicisphaeridium* sp., samples BU.D/RB3, slide 1(1), 1393–119 and BU.D/RB5, slide 1(2), 1354–162; 13, *Veryhachium trispinosum* (Eisenack) Stockmans and Williére, 1962. Sample BU.D/RB5, slide 1(1), 1426–93; 14, *Veryhachium subglobosum* Jardiné, Combaz, Magloire, Peniguel and Vachey, 1974. Sample BU.D/RB5, slide 1(2), 1246–71.



- For the first time a cryptospore assemblage was recovered in the Ribeira do Braçal Formation.
- More detailed sampling of these three formations should be done, in order to better characterize the Hirnantian microfauna of the Buçaco Syncline.

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DARRIWILIAN (ORDOVICIAN) GRAPTOLITE FAUNAS AND PROVINCIALISM IN THE TØYEN SHALE OF THE KRAPPERUP DRILL CORE (SCANIA, SOUTHERN SWEDEN)

J. Maletz¹ and P. Ahlberg²

¹ Department of Geosciences, Colorado State University, 322 Natural Sciences Building, Fort Collins, CO 80523-1482, USA.
yorge.maletz@colostate.edu

² Division of Geology, Department of Earth and Ecosystem Sciences, Lund University, Sölvegatan 12, SE-223 62 Lund, Sweden.
per.ahlberg@geol.lu.se

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INTRODUCTION

In Scania, southern Sweden, Lower Palaeozoic strata are preserved mainly in the Colonus Shale Trough, an elongated, fault-bounded and NW-SE-trending structure within the Sorgenfrei-Tornquist Zone. The relatively condensed Ordovician succession consists predominantly of graptolitic shales deposited in a foreland basin on a marginal portion of the Baltic plate. Outcrops are generally small and restricted to uplifted fault-blocks. Hence, our knowledge of the stratigraphy and spatial and temporal distribution of the succession is to a large extent based on drillings.

A core drilling at Krapperup, northwestern Scania, in 1946 reached a depth of 155.06 m and penetrated a significant portion of the Lower–Middle Ordovician succession. The drilling was carried out by Wargön AB at a site 1.0 km west of the Krapperup castle. The core has diameter of 63 mm, shows no evidence of significant core loss, and is housed at the Division of Geology, Lund University. Graptolites from the lower part of the core, spanning the upper Tremadocian *Hunnegraptus copiosus* Biozone through the lower Dapingian *Pseudophyllograptus angustifolius elongatus* Biozone, have been studied by Lindholm (1981, 1991a, 1991b). The succession in the Krapperup core is the only one representing an unbroken shaly sequence across the boundary between the Tøyen Shale and the Almelund Shale, two units that in Scania are usually separated by the early Middle Ordovician (Darriwilian) Komstad Limestone.

BIOSTRATIGRAPHY

The graptolite succession in the Krapperup drill core is only explored in parts, but has already provided important insights into the biostratigraphy and biogeography of the Lower to Middle Ordovician graptolite faunas of southern Scandinavia and beyond. Lindholm (1981) first recognized the base of the *Kiaerograptus supremus* [*Kiaerograptus* sp. A] Biozone at 151.56 m, followed by the *Araneograptus murrayi* [*Dictyonema* ex. gr. *murrayi*] Biozone at 147.66 m. It is followed by a considerable fault zone

(132.20–113.40 m) and overlain by the *Tetragraptus phyllograptoides* Biozone starting at 112.57 m. The bases of the *Didymograptus balticus* Biozone (88.15 m), the *Pseudophyllograptus densus* Biozone (80.78 m) and the *Pseudophyllograptus angustifolius elongatus* Biozone (75.30 m) have also been determined, but the higher part of the succession was not investigated. Lindholm (1991a) described the *Kiaerograptus supremus* and *Araneograptus murrayi* biozones for the first time from Scandinavia based on data from this drill core. The *Hunnegraptus copiosus* Biozone was not recognized in the core, but is known from surface outcrops (Lindholm, 1991a).

The Upper Dapingian (Yapeenian) may be recognized by the presence of *Arienigraptus jianxiensis* sensu Cooper and Ni (1986) at 62.95–62.98 m (Fig. 1J), as the species is neither known from Castlemainian nor from Darriwilian strata. The species is very robust and large, reaching dimensions usually only attained by the genus *Pseudisograptus*. It bears an isograptid development and possesses strong prothecal folds in the manubrium.

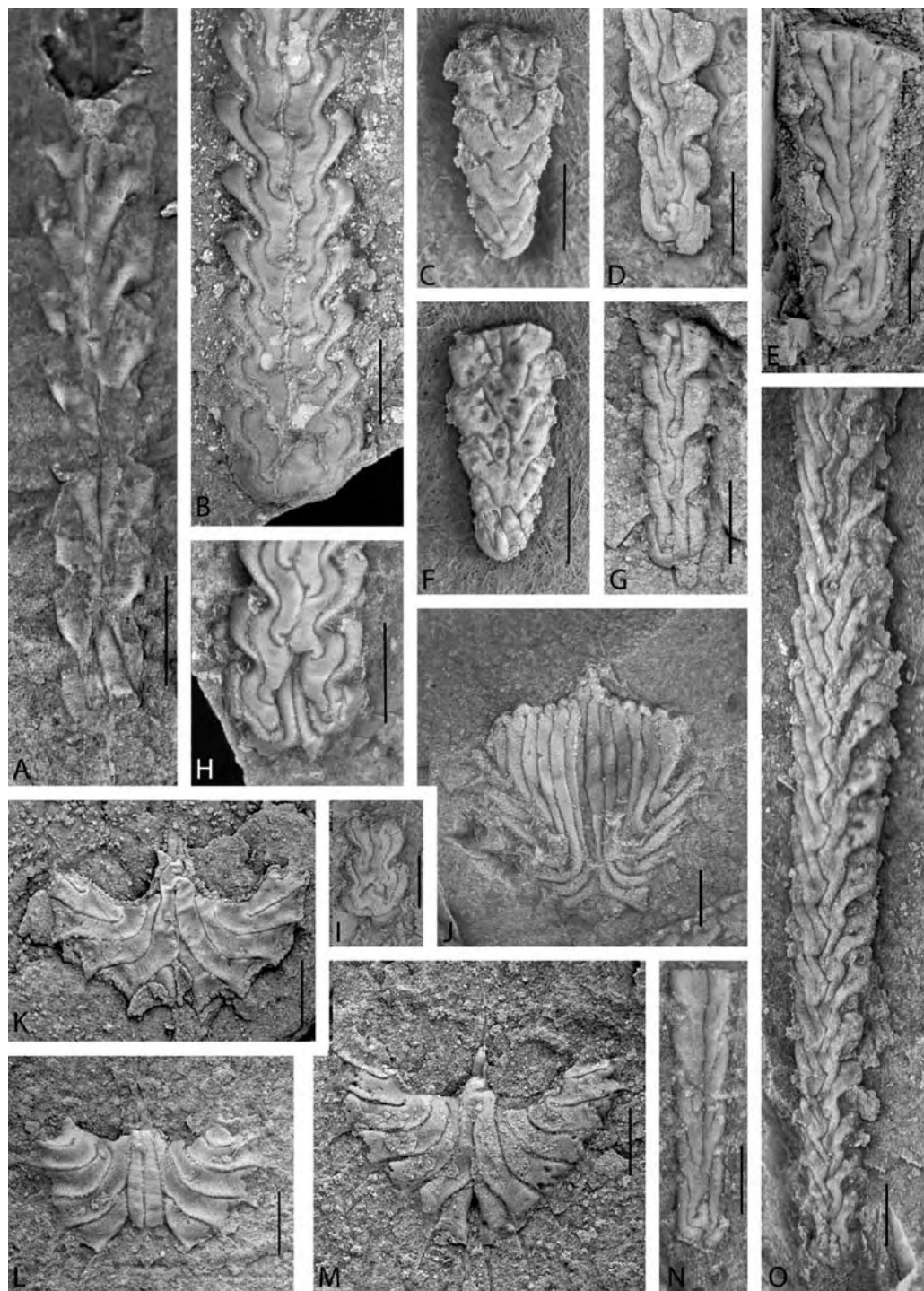
The base of the Darriwilian is here recognized by the presence of *Arienigraptus zhejiangensis* Yu and Fang at 60.67–60.68 m, where the genus is associated with *Pseudisograptus manubriatus* spp. Biserial graptolites of the genus *Levisograptus* (*L. austrodentatus* in particular) are not present and the oldest known biserial, *Levisograptus mui* (Fig. 1B, H) was found only at 54.10–54.20 m. Mitchell (1992, 1994) illustrated specimens of *Levisograptus sinicus* from 48.88–48.53 m and 50.5 m. Maletz (2005) already recognized the late appearance of biserials in the Albjära and Lovisefred drill cores of Scania. The differentiation of the early Darriwilian is difficult, even though numerous biserials of the genus *Undulograptus* are present and the next definitively identifiable level is the base of the *Holmograptus lentus* Biozone in the 24.85–25.15 m interval. The *Holmograptus lentus* Biozone includes a number of different *Holmograptus* species, some of which appear to be new. The excellent relief preservation (Maletz, 2011: figs. a, b) of a number of specimens allows to recognize the specific differences, the presence/absence of prothecal folds, and apertural differentiations.

The *Nicholsonograptus fasciculatus* Biozone is defined by the FAD of its index species at 18.88 m. All specimens are completely flattened. It is interesting to note, that in the Krapperup drill core there is a number of *Holmograptus* specimens in the *Nicholsonograptus fasciculatus* Biozone, and such a biostratigraphic overlap of both genera has not been noted before.

DARRIWILIAN FAUNAS AND BIOGEOGRAPHY

The graptolitic succession of the Krapperup drill core provides some interesting insights into the faunal diversity and composition of early to mid-Darriwilian graptolite faunas of the Atlantic Faunal Realm (Fig. 1). The faunal composition of the Floian to early Dapingian time interval is well known from the Lerhamn drill core (Maletz and Ahlberg, 2011). The interval includes a variety of characteristic *Baltograptus* species

Figure 1. A. *Normalograptus*(?) sp. nov., 22.73–22.74 m. B, H. *Levisograptus mui* (Rickards), 54.00–54.10 m. C, F. *Skaneograptus janus* Maletz, 20.95–21.00 m. D, G. *Proclimacograptus* sp. 20.15–20.19 m. E. *Undulograptus* sp. nov., robust species with straight median septum, 45.65–45.66 m. I. *Undulograptus* sp., small exposed patch of crossing canal and incomplete median septum, 46.01–46.04 m. J. *Arienigraptus jianxiensis* sensu Cooper (1973), 62.95–62.98 m. K. *Arienigraptus* sp. with diminished manubrium and shortened arienigraptid suture, 58.86 m. L, M. *Arienigraptus zhejiangensis* Yu and Fang, 59.30–59.35 m. N. *Eoglyptograptus* sp.?, delayed median septum, short interthecal septae, LO 6435t, 28.37–28.39 m. O. *Undulograptus* sp., zig-zag median septum, high thecal overlap, 20.40–20.44 m. All specimens are shown in reverse view, except for A, F–H, L (obverse views). All specimens are originals, coated with ammonium chloride, except for (A) which is a latex cast of a low relief mould. The precise magnification is provided by a 1 mm long bar in each photo.



as the most important biostratigraphic and biogeographic marker species, restricted to the Atlantic Faunal Realm and providing important biostratigraphic marker species (Toro and Maletz, 2007; Maletz and Ahlberg, 2011).

The base of the Darriwilian interval is not identified by the presence of the earliest biserials of the *Levisograptus austrodentatus* group, but the species *Arienigraptus zhejiangensis* (Fig. 1 L, M) and related forms are extremely common and often occur in nearly monospecific assemblages. A similar *Arienigraptus* species with a shorter arienigraptid suture can be differentiated (Fig. 1K). It can easily be mistaken as an isograptid in flattened specimens in which the manubrium is unrecognizable. Specimens of *Pseudisograptus* are also common at a number of levels in the basal Darriwilian of Baltoscandia (Maletz, 2005) and have been found in the Krapperup drill core.

The axonophoran (biserial) faunas are dominated by members of the genus *Undulograptus* with a rounded proximal end and lacking the typical apertural spines on $th1^1$ and $th1^2$ of the genus *Levisograptus*. A number of species can be differentiated in the Krapperup drill core, some of which are preserved in full relief, showing the proximal development in reverse and obverse views. Due to the poor taxonomic documentation of basal Darriwilian graptolite faunas, a specific identification is impossible to provide at the moment for most of the species. The earlier members often show indications of a $th1^1$ spine and the species *Undulograptus cumbrensis* has been identified in the 41.88–46.42 m interval. Species of *Undulograptus* possess a simplified proximal end development with a possible dicalycal theca at $th2^1$ and a connecting arch between $th2^1$ and $th2^2$ (Fig. 1E). The thecal shapes vary between a strongly geniculate type and a straight to curved, outward inclined, ventral thecal side without evidence of a geniculum. The thecal apertures are outwards inclined to horizontal. The thecae possess a double-sigmoid shape. The median septum is strongly zigzag (Fig. 1O) to straight (Fig. 1E). The genus *Proclimacograptus* with a modified pattern C astogeny (Mitchell, 1987) and short interthecal septae appears first in the upper part of the *Holmograptus lentus* Biozone (Fig. 1D, G), much earlier than the record from the Oslo Region of Norway (Maletz, 1997) suggested.

The evolution of a derived simple proximal end development, resembling Mitchell's (1987) pattern G and pattern H astogenies, can be seen in the genus *Skaneograptus* (Fig. 1C, F) and in a single obverse view of a *Normalograptus* specimen (Fig. 1A) from the 22.73–22.74 m level. This material may provide early evidence of a transition from complex proximal development types to simple types in the early Darriwilian. As comparable faunal elements are not found in the Pacific Faunal Realm, it may be assumed that the transition and early evolution of the Normalograptidae (sensu Mitchell et al., 2007) may have taken place in the cold water Atlantic Faunal Realm and the normalograptids invaded the Pacific Faunal Realm much later during their evolutionary history.

CONCLUSIONS

The Krapperup drill core in Scania (southern Sweden) represents one of the longest and stratigraphically most complete successions of the Scandinavian Tøyen Shale Formation and its direct transition into the Middle Ordovician Almélund Shale. A preliminary investigation indicates the presence of a number of graptolite biozones that range from the late Tremadocian *Kiaerograptus supremus* Biozone to the mid-Darriwilian *Nicholsonograptus fasciculatus* Biozone. The typical southern Swedish Komstad (Orthoceras) Limestone is not present in the succession and the Tøyen Shale Formation grades into the overlying Almélund Shale. This unusual development has not been recognized in any outcrop in

Scandinavia, where the *Orthoceras* limestones in general attains a thickness of at least a few meters. The Darriwilian graptolite fauna includes largely endemic biserial elements with a number of *Undulograptus* and *Proclimacograptus* species. The *Levisograptus austrodentatus* group of early Darriwilian biserials makes a late and only sporadic appearance in the succession, while species of the genus *Arienigraptus* are common and indicative for the basal Darriwilian strata.

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GRAPTOLITE BIOSTRATIGRAPHY AND BIOGEOGRAPHY OF THE TABLE HEAD AND GOOSE TICKLE GROUPS (DARRIWILIAN, ORDOVICIAN) OF WESTERN NEWFOUNDLAND

J. Maletz and S. Egenhoff

Department of Geosciences, 322 Natural Resources Building, Colorado State University, Ft. Collins, CO 80523-1482, USA.
yorge.maletz@colostate.edu, sven@warnercnr.colostate.edu

Keywords: Ordovician, Darriwilian, biostratigraphy, palaeogeography, western Newfoundland, Table Head Group, Goose Tickle Group.

INTRODUCTION

The Table Head and Goose Tickle groups of western Newfoundland represent the sediments of a foreland basin at the eastern edge of the North American craton. The deposits were produced during the early phases of the Taconic orogeny and document the fragmentation of the Lower to Middle Ordovician carbonate platform, subsequently overlain by deeper water carbonates and black shales and covered by a thick clastic succession of conglomerate rich Cape Cormorant Formation. Stenzel et al. (1990) established the currently used lithostratigraphical differentiation of the two lithostratigraphic groups and provided some means of correlation of the units.

The succession can be regarded as one of the best Upper Darriwilian or Upper Middle Ordovician graptolitic successions of eastern North America, dated most precisely by abundantly represented and well preserved graptolite faunas. The graptolite biostratigraphy of the successions has never been evaluated in any detail, however, and mostly faunas without precise biostratigraphical and chronostratigraphical context were collected (e.g. Morris and Kay, 1966; Finney and Skevington, 1979). The graptolite faunas were included in a broadly defined *Paraglossograptus tentaculatus* Biozone or generally referred to a "post-*U. austrodentatus* zone" interval (Williams et al., 1987). Albani et al. (2001) provided more precise information on the graptolite succession of the Mainland section of the Port au Port Peninsula and, for the first time, recognized the *Nicholsonograptus fasciculatus* and *Pterograptus elegans* biozones in North America.

UPPER DARRIWILIAN GRAPTOLITE BIOSTRATIGRAPHY

The graptolite biostratigraphy is based on a detailed investigation of a number of sections in western Newfoundland (Fig. 1). Not all intervals are equally well known due to lack of exposure and variable quality

of fossil content. The graptolite biozones used herein are defined by the first appearance (FAD) of their index species. In any case, the first occurrence of species in the sections has to be considered as a local first occurrence and the correlation is approximate at best. The discussed biozones have a considerable advantage in that they can be used on an inter-continental basis. Their particular index species are distributed worldwide and are easily recognizable even in poor preservation. The graptolite faunas of the Table Head and Goose Tickle groups are reasonably well preserved as flattened films of organic periderm in shales and limestones, and in many cases, can be chemically isolated from their host rocks for a more precise biostratigraphic and taxonomic investigation.

The *Holmograptus spinosus* Biozone

The oldest graptolite faunas of the Table Head Group are found in the West Bay Centre Quarry, where especially biserial graptolites are common and diverse in the Table Cove Formation (Finney and Skevington, 1979). The fauna is rich in specimens, but age diagnostic forms are largely lacking. This lowermost interval is here tentatively referred to the *Holmograptus spinosus* Biozone, even though the index species of the zone is not represented. Maletz (2009) discussed the *Holmograptus spinosus* Biozone from eastern North America in detail and quoted *Parisograptus forcipiformis* and *Bergstroemograptus crawfordi* as typical members of the fauna. Both species are common in the Table Cove Formation of the West Bay Centre Quarry and range into the lower part of the *Nicholsonograptus fasciculatus* Biozone in the Back Cove Formation at Black Cove.

The *Nicholsonograptus fasciculatus* Biozone

The base of the *Nicholsonograptus fasciculatus* Biozone is taken at the FAD of this easily identifiable species. The species is common in the Black Cove Formation at Black Cove and in the West Bay Centre Quarry. The fauna is dominated by a number of *Archiclimacograptus* species, associated with specimens of *Glossograptus* and *Paraglossograptus* mainly. At certain levels, *Xiphograptus robustus* is common. The shales of the Black Cove Formation grade into the green shales of the American Tickle Formation, in which graptolites are extremely rare. The limestone conglomerate in the upper part of the succession at West Bay Centre has been correlated with the Daniel's Harbour conglomerate at Daniel's Harbour (Stenzel et al., 1990), where it yielded three-dimensionally preserved graptolites (Whittington and Rickards, 1969). Albani et al. (2001) reported the presence of *Nicholsonograptus fasciculatus* from the lower part of the Mainland section. Specimens can be found at about 20 m above the base of the Cape Cormorant Formation, but the exact range of this species is uncertain. *Nicholsonograptus fasciculatus* is a widely distributed and common species that can be found in most Middle Ordovician successions worldwide and, thus, is very useful as an index species for biostratigraphic correlation.

The *Pterograptus elegans* Biozone

Pterograptus elegans is a common multiramous, pendent xiphograptid with cladal branching that is easily recognized in the Darriwilian successions worldwide. The first occurrence of *Pterograptus elegans* defines the base of this biozone. *Pterograptus elegans* is found at many levels of the Cape Cormorant Formation of the Mainland section (Albani et al., 2001), from where it was first described by Maletz (1994) from isolated specimens. The earliest specimens are present approximately at the 95 m level in the section

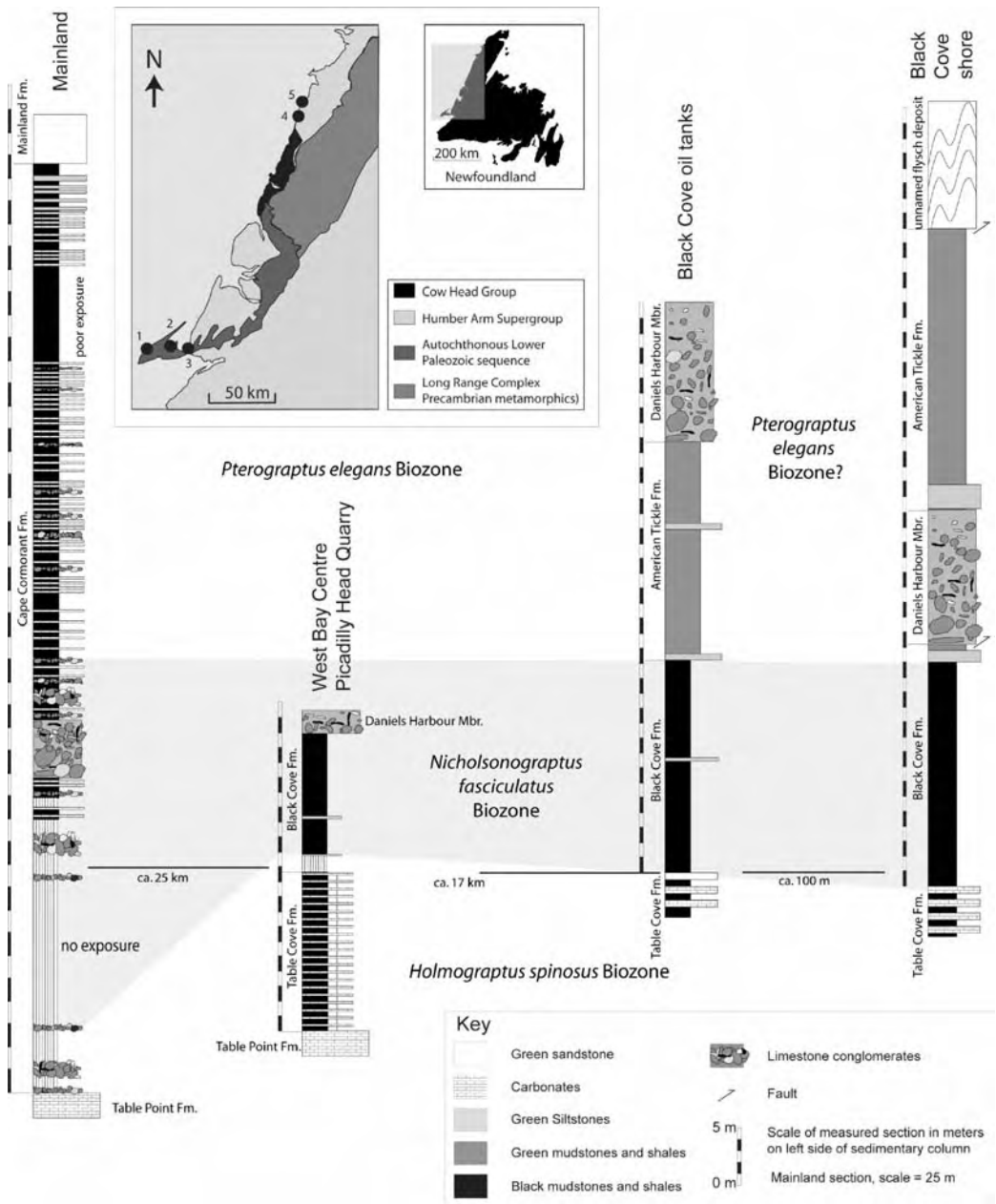


Figure 1. The lithological and biostratigraphical correlation of the Table Head and Goose Tickle group sections in western Newfoundland. The localities on the map are Mainland (1), West Bay Centre Quarry (2), Black Cove (3), Daniel's Harbour (4) and Table Point (5).

(see James and Stevens, 1986 for section). The species has not been discovered unequivocally in the American Tickle Formation at Black Cove so far, but a few stipe fragments possibly belonging to this species are present. Ruedemann (1947) misidentified specimens of this species from the Ledbetter Slate of Washington State as *Syndyograptus bridgei* and did not mention the presence of *Pterograptus* in North America.

GRAPTOLITE BIOGEOGRAPHY

Graptolites are planktic marine organisms with a wide distribution, restricted mainly by climatic conditions expressed as latitudinal temperature gradients, water depth, and ocean currents. The biogeographical differentiation of Early to Middle Ordovician graptolite faunas into the Atlantic and Pacific Faunal Realms is well established, as is a depth differentiation from shallow to deep-water faunas (Cooper and Sadler, 2010). Shallow water shelf faunas are rarely preserved in the carbonate successions of the platform regions of eastern North America, but are more commonly associated with deep water faunas and even rich benthic, dendroid faunas carried down-slope to even toe-of-slope environments as found in the Cow Head Group of western Newfoundland (Williams and Stevens, 1988) or in the Lévis Formation of the Québec Appalachians (Maletz, 1997). The associations clearly indicate transport from a shelf region into the basin, especially as the dendroids are invariably fragmented. In this case, a mixing of the shelf faunas with oceanic faunas occurs through basinward transport into deeper water regions and a precise differentiation of origin of the individual faunal elements is difficult. The planktic shelf faunas are less likely to be fragmented through the transport and appear considerably less damaged than the associated benthic dendroid elements.

The Laurentian graptolite biofacies of the Darriwilian has never been explored in detail. Western Newfoundland was part of Laurentia, positioned in a tropical region during the Paleozoic, with a carbonate platform fringing the edge of the continent. The faunas of the Table Head and Goose Tickle groups represent an open ocean environment with a considerable amount of endemic faunal elements. Thus, a mixture of endemic, shallow water and cosmopolitan elements had to be expected.

Paraglossograptus tentaculatus often dominates the faunal assemblages in the Lower Darriwilian of North America. This species is replaced in the Da 3 by *Paraglossograptus proteus* and *Paraglossograptus holmi* and cannot be found in the Table Head Group, even though it has been quoted to be present. The "dichograptid" faunal elements (*Nicholsonograptus*, *Pterograptus*, *Xiphograptus*) are the ones that represent the pandemic graptolite faunas, while the biserial faunal elements are largely endemic. The biserials *Archiclimacograptus decoratus* and *Archiclimacograptus confertus* range through a considerable time interval, probably originating in the Da 2 and reaching into the Upper Darriwilian (Da 4a/b). The biserials in the Table Head and Goose Tickle groups include the typical *Archiclimacograptus decoratus* as one of the most common species, easily recognized only through its heart-shaped nematolium. The species is widely distributed in North America and can be found in the Australasian successions (VandenBerg and Cooper, 1992). In western Newfoundland the youngest graptolites from the Middle Ordovician foreland basin belong to the *Pterograptus elegans* Biozone. Graptolites from the overlying flysch of the American Tickle Formation show a low diversity in which *Cryptograptus schaeferi* and *Archiclimacograptus* specimens dominate. The fauna does not include time indicative forms, but is unlikely to be much younger, as specimens of *Dicellograptus*, *Dicranograptus* and *Nemagraptus* are lacking altogether.

A considerable faunal endemism can be seen when the faunas are compared with faunas of the same age in Scandinavia (see Maletz, 1997; Maletz et al., 2007). The Scandinavian faunas of the *Nicholsonograptus fasciculatus* and *Pterograptus elegans* biozones include numerous members of the genera *Proclimacograptus*, *Undulograptus* and *Eoglyptograptus*, which are not represented at all in the eastern North American successions. Specimens of *Haddingograptus* and *Hustedograptus* are rare in Table Head and Goose Tickle groups, but occur in high numbers in the Scandinavian successions. The taxonomy of the frequent *Archiclimacograptus* species is difficult to verify at the moment due to a problematic systematic interpretation of the genus and its species. The species differentiation and recognition is only possible with relief specimens and most specimens cannot be identified to species level. It appears, however, that *Archiclimacograptus* appears much later, in the upper part of the *Nicholsonograptus fasciculatus* Biozone in Scandinavia, while there is a continuous representation of these in the eastern North American successions from the *Undulograptus austrodentatus* Biozone onwards, starting with the closely related "*Undulograptus*" *primus*. Thus, a distinct faunal endemism can be seen in the Darriwilian biserial graptolite faunas, matching the one obvious from other faunal, non-biserial faunal elements.

The lack of presence of any elements of the Sinograptidae (e.g. *Holmograptus* spp.) in the lower part of the Table Head Group is notable, as the genus is common in many eastern North American succession of this age (Maletz, 2009) and is associated with faunal elements common in the Table Head and Goose Tickle groups. The Sinograptidae only appear in the form of *Nicholsonograptus fasciculatus* in the eponymous biozone.

CONCLUSIONS

The Table Head and Goose Tickle groups of western Newfoundland provide a number of biostratigraphically useful graptolite species, namely *Nicholsonograptus fasciculatus* and *Pterograptus elegans*, valuable of inter-continental correlations.

A considerable faunal endemism is developed in the interval. The endemism is based on the frequency of endemic elements, often strongly outnumbering the cosmopolitan and biostratigraphically more indicative faunal elements. Endemic faunal elements make up more than 90% in many faunas from western Newfoundland, but belong to only 2-3 species.

The faunal endemism does include the biserial faunal elements, previously considered to show a less strong faunal provincialism than other (dichograptid) faunal elements.

Acknowledgements

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CORRELATION OF LOWER ORDOVICIAN (IBEXIAN) FAUNAS IN NORTH-EASTERN GREENLAND AND WESTERN NEWFOUNDLAND – NEW TRILOBITE AND LITHOSTRATIGRAPHIC DATA

L.M.E. McCobb¹, W.D. Boyce² and I. Knight²

¹ Department of Geology, National Museum of Wales, Cathays Park, Cardiff, CF10 3NP, UK. lucy.mccobb@museumwales.ac.uk

² Geological Survey, Newfoundland and Labrador Department of Natural Resources, P.O. Box 8700,
St. John's, NL, Canada A1B 4J6. dougboyce@gov.nl.ca, ianknight@gov.nl.ca

Keywords: Trilobites, Ordovician, Greenland, Newfoundland, Laurentia.

INTRODUCTION

Studies have been ongoing into Early Ordovician carbonate lithologies and faunas of North-East Greenland. GEUS-led expeditions in 2000/2001 logged detailed sections through the carbonates there, redefining the lithostratigraphy and placing it in the Fimbulfjeld Group (Stouge et al., 2001, 2002, in press; Fig. 1). Trilobites, now housed at the National Museum of Wales, were gathered by J. W. Cowie and P. J. Adams (1957) during mapping of the Cambrian—Ordovician rocks of the region. They never published descriptions but ongoing studies led by LMEM incorporate their fossils, those collected in 2000/2001, and conodont biostratigraphy within the context of the new lithostratigraphic data (McCobb et al., 2009, 2010a,b).

It has long been recognised that Lower Ordovician carbonate rocks of western Newfoundland are coeval with those of North-East Greenland and were deposited in similar settings, along the southern margin of Laurentia. Various studies (e.g., Fortey, 1979, 1983; Boyce, 1989) have done much to document the trilobite faunas of western Newfoundland, and Boyce (1989, 1997) and Boyce and Stouge (1997) erected a trilobite zonation scheme for the area. New fossil collections from the Watts Bight and Catoche (Costa Bay Member) formations on the Port au Port Peninsula (Boyce et al., 2000, 2011; Boyce and Knight, 2010) provide new data for correlation of the western Newfoundland faunas with those in North-East Greenland and elsewhere.

LITHOSTRATIGRAPHY - NORTH-EAST GREENLAND

In North-East Greenland, Ibexian rocks are represented by around 1.5km of peritidal to subtidal carbonates, best known at Albert Heim Bjerge and Ella Ø. The redefined Antiklinalbugt Formation (Peel and Cowie, 1979) includes the upper 10 m of the underlying Dolomite Point Formation, and at least 25 m of the overlying Cape Weber Formation of Cowie and Adams (1957) (Fig. 1). Five informal lithostratigraphical

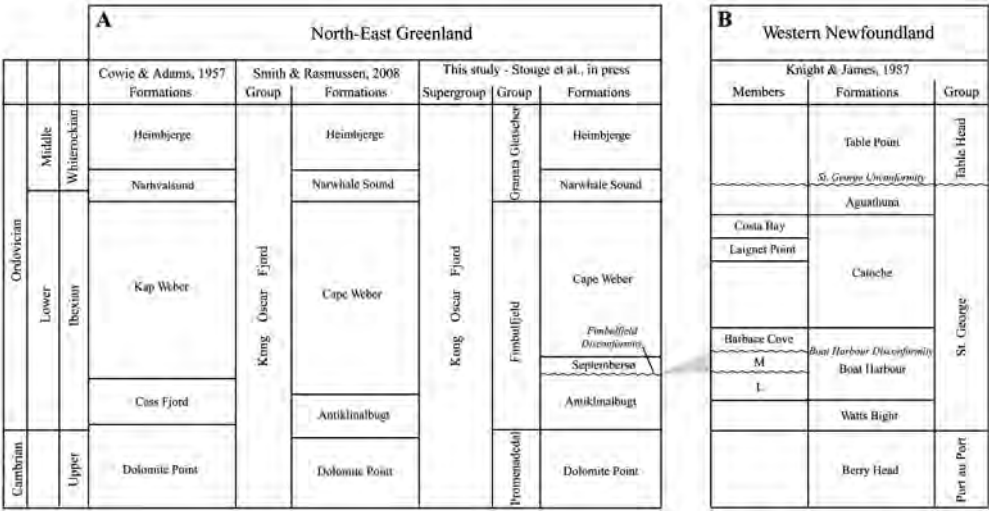


Figure 1. Early Ordovician stratigraphic nomenclature of A, North-East Greenland used in this study, versus that of Cowie and Adams (1957) and Smith and Rasmussen (2008); and B, western Newfoundland (Knight and James, 1987).

units are recognised in place of the three of Cowie and Adams (1957). The Cape Weber Formation is also redefined, with a new formation, the Septembersø Formation, erected at its base. The Fimbulfjeld Disconformity, perhaps representing the entire Stairsian Stage, separates the Tulean Septembersø formation from the underlying Skullrockian Antiklinalbugt Formation (Stouge et al., in press). The Tulean-Blackhillsian Cape Weber Formation rests conformably on the Septembersø formation, and is conformably overlain by the Narwhale Sound Formation. Five members are recognised (Stouge et al., in press), instead of three.

LITHOSTRATIGRAPHY - WESTERN NEWFOUNDLAND

In western Newfoundland, coeval platform carbonates occur in the St. George Group, which is divided into four formations (Knight and James, 1987; Fig. 1). The group hosts three depositional sequences (Knight et al., 2007, 2008). The first, a deepening-shallowing Skullrockian sequence, includes the Watts Bight Formation and lower Boat Harbour Formation and terminates at a disconformity. The second, consisting of Stairsian peritidal carbonates of the middle Boat Harbour Formation, terminates at the Boat Harbour Disconformity. Above this disconformity, the third sequence of Tulean to Blackhillsian age comprises the Barbace Cove Member (Boat Harbour Formation), the overlying Catoche Formation and the Agathuna Formation. It is terminated at the St. George Unconformity (Knight and James, 1987).

NORTH-EAST GREENLAND - TRILOBITE FAUNA AND BIOSTRATIGRAPHY

Ibexian trilobites from North-East Greenland range in age from Skullrockian to Blackhillsian; Stairsian macrofossils are absent (see also Poulsen, 1930; 1937). The Antiklinalbugt Formation yielded several

species of *Symphysurina*, along with two new species of *Tulepyge*, and the hystricurids *Millardicurus* and *Hystricurus* (Fig. 2). *Micragnostus chiushuensis* is a rare element of the fauna, as are *Bellefontia*?, *Clelandia*, *Lunacrania* and cf. "*Hystricurus*" *missouriensis*. The *Symphysurina* species place the lower Antiklinalbugt Formation within the *S. brevispicata* to *S. bulbosa* subzones of the *Symphysurina* Zone. *S. woosteri*, *Bellefontia*? sp. and *Clelandia* sp. indicate a *S. woosteri* Subzone to *Bellefontia*-*Xenostegium* Zone age for the upper Antiklinalbugt Formation. Conodonts place the lower part of the formation in the *Cordylodus intermedius* Zone and the upper part in the *Rossodus manitouensis* Zone. This combined fauna is characteristic of the late Skullrockian Stage (McCobb et al., 2009).

Trilobites from the Septembersø formation are exclusively Bathyruridae, namely *Bolbocephalus*, *Chapmania*, *Peltabellia* and *Punka* (McCobb et al., 2010a; Fig. 2). The late Tulean age suggested by the trilobites is supported by a sparse conodont fauna (Smith, 1991). The absence of definitive Stairsian trilobites suggests a major hiatus marked by the Fimbulfeld Disconformity.

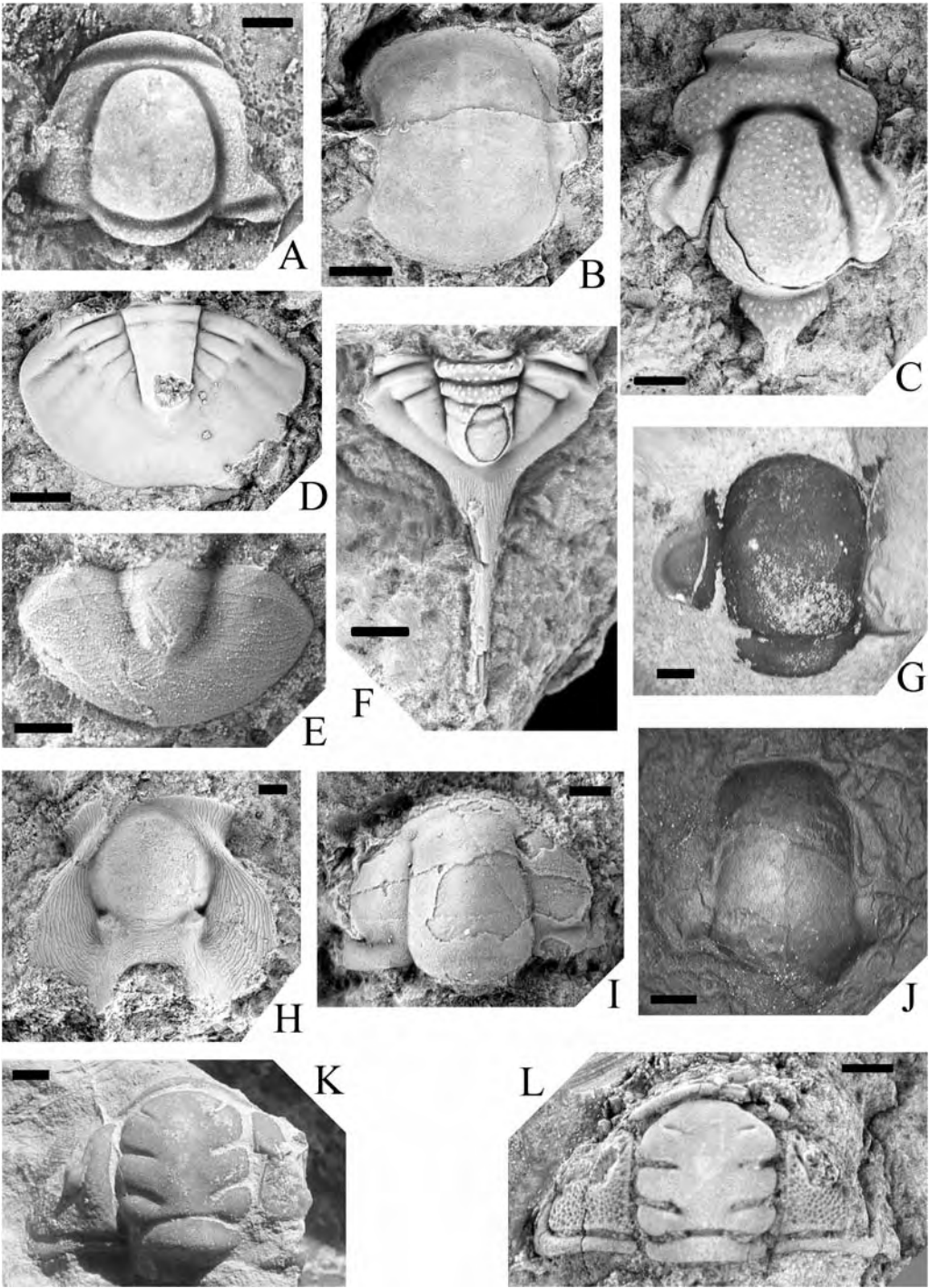
Bathyrurids, including *Acidiphorus*/*Goniotelina*, *Bathyurellus*, *Benthamaspis*, *Bolbocephalus*, *Jeffersonia*, *Petigurus*, *Punka*, *Uromystrum* and *Stigigenalis* also dominate the Cape Weber Formation (McCobb et al., 2010b; Fig. 2). Also represented are asaphids (*Isoteloides*, *Niobe*, *Paraptychopyge* and *Presbynileus* (*Protopresbynileus*)), dimeropygids (*Ischyrota*), illaenids (*Illaenus*), pliomerids (*Cybelopsis*), remopleurids (?*Eorobergia*), styginids (?*Eobronteus* and ?*Raymondaspis*) and telephinids (*Carolinites*). The faunas span the *Stigigenalis brevicaudata*, *S. caudata* and *Benthamaspis gibberula* zones of western Newfoundland (Boyce and Stouge, 1997), equivalent to the *Protopliomerella contracta* (G₂) to *Presbynileus ibexensis* (I) zones (Ross et al., 1997) of Utah-Nevada. Some trilobites in the upper two members indicate the *Pseudocybele nasuta* Zone (J) (= *Cybelopsis speciosa* Zone of Boyce et al., 2000). The trilobite-based age range, supported by three conodont faunas: Fauna D, *Oepikodus communis* biozone (Smith, 1991), and *O. intermedius* fauna (Stouge et al., in press), indicates a Tulean to Blackhillsian age.

WESTERN NEWFOUNDLAND - NEW TRILOBITE DATA AND BIOSTRATIGRAPHY

St. George Group trilobites, documented by Fortey (1979), Boyce (1989) and Boyce et al. (2000) range through several upper Skullrockian to lowermost Whiterockian zones (Boyce, 1997; Boyce and Stouge, 1997; Fig. 3). Stairsian trilobites occur in the disconformity-bounded, middle Boat Harbour Formation.

Recent fieldwork on the Port au Port Peninsula focused on the Watts Bight Formation, and Costa Bay Member of the Catoche Formation (Boyce et al., 2000, 2011). Trilobites newly collected from the lower Watts Bight Formation, include new records of *Millardicurus* sp. cf. *M. armatus* and *Symphysurina myopia* and *Bellefontia gyracantha* (Fig. 2); also present is "*Hystricurus*" *ellipticus*. These trilobites are assigned to the "*Millardicurus millardensis*" and "*Hystricurus*" *ellipticus* assemblage zones of western Newfoundland (Boyce, 1997; Boyce et al., 2011). The conodont-rich Watts Bight Formation spans the deeper-water (DW) *Cordylodus lindstromi* to *C. angulatus* Lineage Zones (Ji and Barnes, 1994), indicating correlation with the Skullrockian *S. brevispicata* to *S. bulbosa* trilobite subzones of the type Ibex. The presence of *M.* sp. cf. *M. armatus*, *Symphysurina* and *Bellefontia*, suggests a correlation with the Antiklinalbugt Formation.

Trilobites from the Costa Bay Member, Catoche Formation, primarily comprised *Acidiphorus*/*Goniotelina*, *Benthamaspis* and *Cybelopsis* (Fig. 2); the known range of *Cybelopsis speciosa* was extended. The fauna correlates with that of the *Pseudocybele nasuta* Zone (J) in western USA, and is similar to a fauna from the 'Black Limestones', Cape Weber Formation, Albert Heim Bjerge, North-East Greenland.



CORRELATIONS

Based on trilobites and conodonts, Ordovician rocks of North-East Greenland and western Newfoundland are correlated with the type Ibex area and other parts of Laurentia. Trilobites of the Antiklinalbugt Formation suggest it is equivalent to the House Formation in the type Ibexian of western Utah and eastern Nevada (Hintze, 1953), and Garden City Formation of northeastern Utah and southeastern Idaho (Ross, 1951). Common species of trilobite also allow the Antiklinalbugt Formation to be correlated in part with: lower Watts Bight Formation and uppermost part of underlying Berry Head Formation, western Newfoundland; Cape Clay Formation of western North Greenland; parts of Shallow Bay and Green Point Formations (Cow Head Group), and Cooks Brook Formation (Northern Arm group), western Newfoundland; Survey Peak Formation, Alberta; Rabbitkettle Formation, Mackenzie Mountains of Northwest Territories; Wilberns Formation of Texas; McKenzie Hill Limestone of Oklahoma; and Tribes Hill

Series	Stage	Conodont Zones (Ross <i>et al.</i> , 1997)	Shelly Fossil Zones (Ross <i>et al.</i> , 1997)	Trilobite Zones - Western Newfoundland (Boyce & Stouge, 1997; Boyce, 1997; Boyce <i>et al.</i> , 2000)
Ibexian Series	Blackhillian	Ra <i>Reutterodus andinus</i>	Hm <i>Hesperonomiella minor</i>	HIATUS
			Pn <i>Pseudocybele nasuta</i>	<i>Cybelopsis speciosa</i> <i>Gignopeltis rarus</i>
		Oc <i>Oepikodus communis</i>	Pi <i>Presbynileus ibexensis</i>	<i>Benthamaaspis gibberula</i>
			Tt <i>Trigonocerca typica</i>	<i>Strigigenalis caudata</i>
	Tulean	AO <i>Acodus deltatus</i> - <i>Oneotodus costatus</i>	Pc <i>Protopliomerella contracta</i>	<i>Strigigenalis brevicaudata</i> <i>Peltabellia knighti</i>
			Hc <i>Hintzeia celsaora</i>	HIATUS
	Slarsian	Md <i>Macerodus dianaee</i>	Rs <i>Rossaspis superciliosa</i>	<i>Randaynia saundersi</i> <i>Lelostegium proprium</i> <i>Hystricurus oculilunatus</i>
		LD "Low diversity interval"	Te <i>Tesselacauda</i>	
		Rm <i>Rossodus manitouensis</i>	LK <i>Lelostegium-Kainella</i>	HIATUS
		Skullrockian	Ca <i>Cordylodus angulatus</i> la <i>lapetognathus</i> n. sp. Cl <i>Cordylodus lindstromi</i> Ci <i>Cordylodus intermedius</i>	Pa <i>Paraplethopeltis</i>
	BX <i>Bellefontia-Xenostegium</i>			<i>Parahystricurus</i> sp. I "Hystricurus" ellipticus "Millardicurus millardensis"
	Sy <i>Symphysurina</i>			
	Mi <i>Missisquoia</i>			<i>Missisquoia typicalis</i>
	Cp <i>Cordylodus proavus</i>		Ea <i>Eurekia apopsis</i>	

Figure 3. Correlation of trilobite zones of western Newfoundland (Boyce and Stouge, 1997; Boyce, 1997; Boyce *et al.*, 1992, 2000) with Ibexian conodont and shelly fossil zones (Ross *et al.*, 1997).

Figure 2. Trilobites from North-East Greenland (A-F, H-J, L) and Port au Port Peninsula, western Newfoundland (G, I, K). A. *Tulepyge cowie* McCobb *et al.*, 2009, holotype cranium, NMW 97.56G.154.1, Antiklinalbugt Formation. B. *Symphysurina elegans* Poulsen, 1937, cranium, NMW 97.56G.190, Antiklinalbugt Formation. C. *Millardicurus armatus* (Poulsen, 1937), cranium, latex cast of NMW 97.56G.159b, Antiklinalbugt Formation. D. *Punka* sp., pygidium, NMW 97.56G.92, Septembersø Formation. E. *Benthamaaspis conica* Fortey, 1979, pygidium, NMW 97.56G.16, Cape Weber Formation. F. *Acidiphorus* cf. *brighti* Hintze, 1953, pygidium, NMW 97.56G.113, Cape Weber Formation. G. *Acidiphorus/Goniotelina* sp., cranium, NFM F-805, Costa Bay Member, Catoche Formation. H. *Isoteloides* sp., hypostoma, NMW 97.56G.234, Cape Weber Formation. I. *Benthamaaspis gibberula* (Billings, 1865), cranium, NMW 97.56G.256, Cape Weber Formation. J. *Bellefontia gyracantha* (Raymond, 1910)?, cranium, NFM F-804, Watts Bight Formation. K. *Cybelopsis speciosa* Poulsen, 1927, cranium, NFM F-806, Costa Bay Member, Catoche Formation. L. *Cybelopsis speciosa* Poulsen, 1927, cranium, NMW 97.56G.18, Cape Weber Formation. Scale bars: A, E, G, H, I, L = 1 mm; B, C, F, K = 2 mm; D, J = 3 mm.

Formation of New York State (McCobb et al., 2009). The latter also shows direct faunal links with the Watts Bight Formation (Boyce et al., 2011).

The presence of *Cybelopsis speciosa* in the Cape Weber Formation provides a faunal link with the Costa Bay Member, Catoche Formation, and a correlation with the *C. speciosa* zone of western Newfoundland (Boyce et al., 2000). It also supports a correlation with the Nunatami Formation of western North Greenland (Poulsen, 1927). Other trilobites common to the Cape Weber and Catoche Formations include *Benthamaspis conica*, *B. gibberula*, *Jeffersonia angustimarginata*, *Petigurus groenlandicus* and *Uromystrum affine* (Fortey, 1979; Boyce, 1989; Boyce and Stouge, 1997). Overall, trilobites from the Septembersø and Cape Weber formations indicate a biostratigraphic range equivalent to the *Strigigenalis breviaudata* to *Cybelopsis speciosa* zones of western Newfoundland, corresponding to *Protopliomerella contracta* to *Pseudocybele nasuta* zones of the Ibexian (Ross et al., 1997; see Fig. 3). *Jeffersonia angustimarginata* also suggests a correlation between parts of the Cape Weber and Catoche formations with the Croisaphuill Formation of the Durness Group, Scotland and the Canyon Elv Formation, Ellesmere Island; *B. conica* also occurs in the Wandel Valley Formation of eastern North Greenland. *Acidiphorus* cf. *whittingtoni* provides a faunal link with the Fort Cassin Formation of New York State and Vermont. *Parapytychopyge* cf. *disputa* indicates correlation between the Cape Weber Formation and the Valhallfonna Formation of Spitsbergen.

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ICE IN THE SAHARA: THE UPPER ORDOVICIAN GLACIATION IN SW LIBYA – A SUBSURFACE PERSPECTIVE

N.D. McDougall¹ and R. Gruenwald²

¹ Repsol Exploración, Paseo de la Castellana 280, 28046 Madrid, Spain. neilmcdougall@repsol.com

² REMSA, Dhat El-Imad Complex, Tower 3, Floor 9, Tripoli, Libya.

Keywords: Ordovician, Libya, glaciation, Mamuniyat, Melaz Shugran, Hirnantian.

INTRODUCTION

An Upper Ordovician glacial episode is widely recognized as a significant event in the geological history of the Lower Paleozoic. This is especially so in the case of the Saharan Platform where Upper Ordovician sediments are well developed and represent a major target for hydrocarbon exploration. This paper is a brief summary of the results of fieldwork, in outcrops across SW Libya, together with the analysis of cores, hundreds of well logs (including many high quality image logs) and seismic lines focused on the uppermost Ordovician of the Murzuq Basin.

STRATIGRAPHIC FRAMEWORK

The uppermost Ordovician section is the youngest of three major sequences recognized widely across the entire Saharan Platform:

Sequence C01: Unconformably overlies the Precambrian or Infracambrian basement. It comprises the possible Upper Cambrian to Lowermost Ordovician Hassaouna Formation.

Sequence C02: Truncates C01 along a low angle, Type II unconformity. It comprises the laterally extensive and distinctive Lower Ordovician (Tremadocian-Floian?) Achebayat Formation overlain, along a probable transgressive surface of erosion, by interbedded burrowed sandstones, cross-bedded channel-fill sandstones and mudstones of Middle Ordovician age (Dapingian-Sandbian), known as the Hawaz Formation, and interpreted as shallow-marine sediments deposited within a megaestuary or gulf.

Sequence C03: Is the uppermost Ordovician section assigned in Libya to the Melaz Shugran, Mamuniyat and Bir Tlacin formations. In most cases subsurface biostratigraphy confirms a possible late Katian to Hirnantian age for this sequence. As a whole, the uppermost Ordovician package is bounded at the base by a major unconformity defining a series of paleovalleys; ranging in width from ca. 1km to >20km, and remnant Mid-Ordovician palaeohighs. Given the glacial context, it is generally assumed that erosion was associated both with tunnel valleys and ice streams.

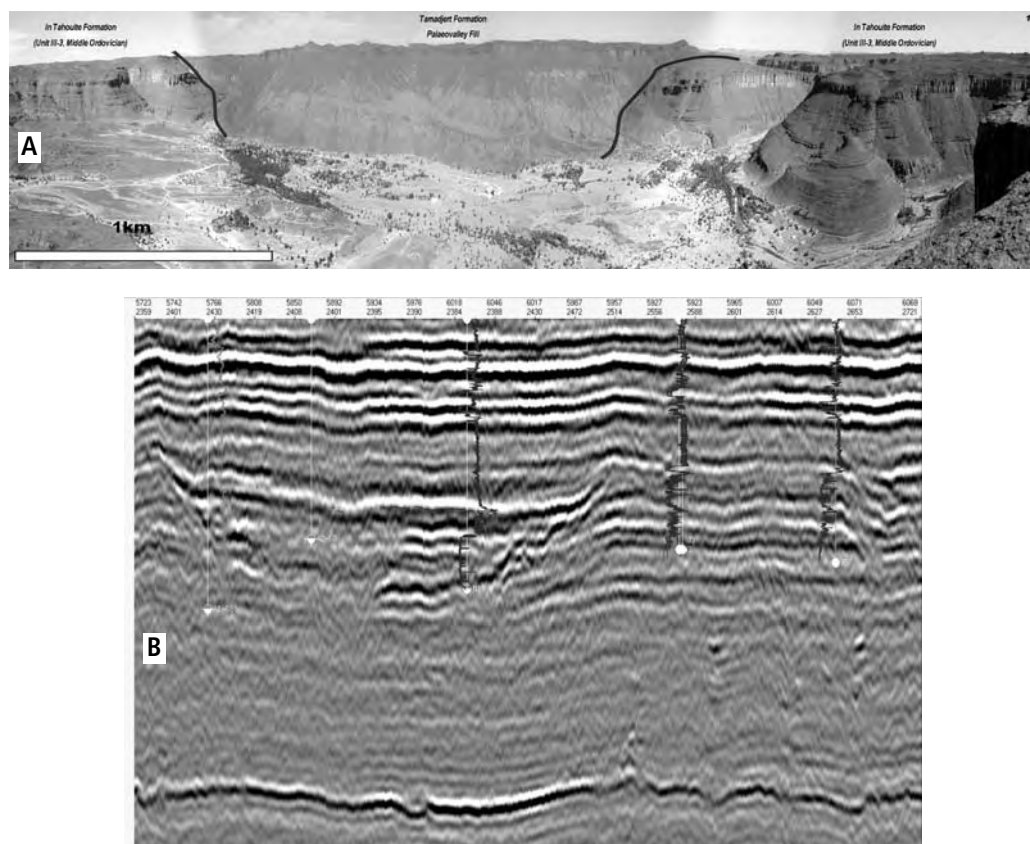


Figure 1. Upper Ordovician palaeovalleys in (A) outcrop for comparison; Iherir, Tassili N Ajers, SE Algeria and (B) in subsurface (3d seismic data) from the central Murzuq Basin of SW Libya.

Both vertical and lateral (along strike and downdip) facies changes are typically rapid resulting in a complex of glacially-influenced fluvial to glaciomarine environments. The whole succession is terminated by a major post-glacial flooding event, in the earliest Silurian (Llandovery-Rhodian) during which graptolitic shales were deposited across the region forming a sealing horizon and, locally, organic-rich source rocks.

In most cases the late Ordovician glaciogenics can be subdivided into several distinctive packages (McDougall and Martin, 2000; Ghienne et al., 2003), each bounded by Type I unconformities and are effectively a higher order depositional sequence recognizable across the entire area and associated with a series of glacial advances and retreats. Figure 2 summarizes this basic stratigraphic subdivision, as derived from both outcrop and subsurface, by reference to five genetic packages; known as Melaz Shugran, Lower Mamuniyat, Middle Mamuniyat, Upper Mamuniyat and Bir Tlacin.

SEDIMENTOLOGY

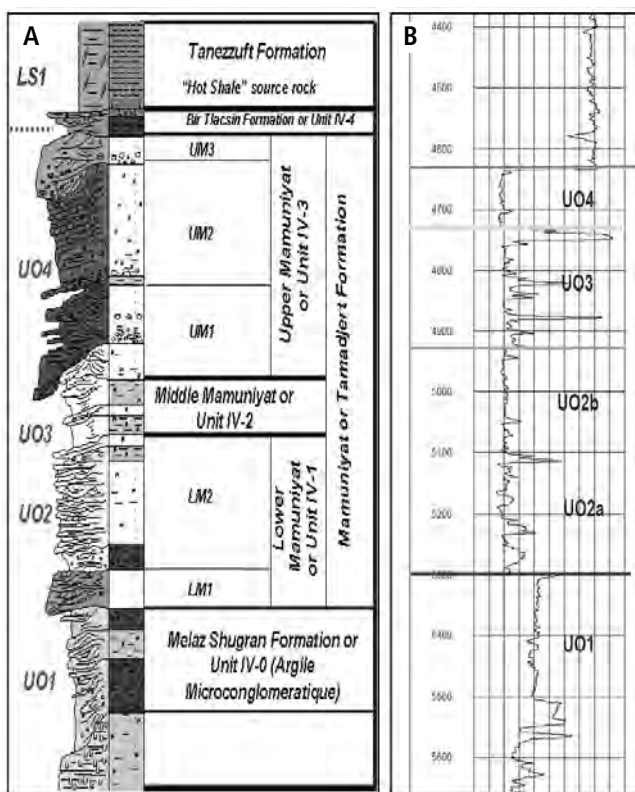
This paper is based on the study of significant volumes of seismic data, cores, image logs and conventional wireline logs interpreted within the framework of fieldwork from several key outcrops across the area of SW Libya, notably the Qarqaf Arch and Ghat-Tikiumit areas.

Melaz Shugran

This is a mud-prone, often heterolithic unit, locally >250m thick; comprising several unconformity-bound subunits, termed MS1, MS2 and MS3 possibly ranging in age from late Katian to early Hirnantian. Outcrop observations, coupled with both core and extensive image log analysis show this package to comprise several key facies types; (1) Massive, deformed sandy to pebbly mudstones, (2) Thick, generally fine-grained, relatively clean sandstones with channel geometry and (3) thinly interbedded fine-grained sandstones and mudstones all associated with pervasive soft sediment deformation. Comparison with modern analogues suggests these facies most probably represent deposition from debris flows, density underflows, turbidity currents and possibly iceberg rain-out, in morainal banks or subaqueous glaciomarine fans. In broader terms it is assumed that Melaz Shugran deposition was associated with both high relative sea levels and sediment fluxes; presumably a response to a major glacial retreat following the initial platform-wide incision event.

Lower Mamuniyat

This major sand-dominated package, up to 150m thick, sharply overlies the underlying argillaceous sediments of the Melaz Shugran, along a regional surface interpreted from seismic data and high resolution correlations to vary from a major erosional unconformity (it may even completely truncate the Melaz Shugran resting directly on the older pre-glacial sediments of Middle Ordovician age), to a minor unconformity or diastem. Outcrop observations suggest a subdivision into: (a) a lower unit (LM1) comprising fine- to coarse-grained, even locally conglomeratic, sandstones forming channel-bar complexes of probable tidal origin; and (b) Fine grained sandstones forming sheetflood complexes (LM2) which pass laterally into climbing megaripples, associated with pro-glacial outbursts (Ghienne et al., 2010). However,



this is perhaps less clear in the subsurface where many wells show abundant soft sediment deformation and locally significant downdip changes in lithology associated with deposition from density underflows and sediment flows in sand-rich glaciomarine fans.

Middle Mamuniat

This distinctive heterolithic package is separated from the sand-prone Lower Mamuniat by a major subaerial unconformity, which, in many cases, displays evidence in outcrop for glacial erosion and associated deformation (folds, step-faulting, injection structures) attributed to glaciotectonism. It records a significant episode of glacial re-incision, most probably associated with the generation of tunnel valleys, and generation of accommodation space followed by rapid post-glacial flooding and the subsequent progradation of braid-delta systems fed by the retreating ice. Initially, deposition appears to have occurred in relatively steep slope-type braid-deltas but with reduced accommodation space deposition occurred principally in lower gradient braid-delta systems.

Upper Mamuniat

This sand-prone package is areally the least extensive of the Mamuniat sequences. It tends to occur almost exclusively in the axes of relatively narrow, deep palaeovalleys. As such it is observed to erosively truncate the underlying sequences, locally resting directly on Middle Ordovician sediments, in response to a significant episode of base level fall and ice advance. The fill of the subsequent incised valleys is typically subdivided in outcrop into 3 units, bounded in proximal areas by significant unconformities:

UM1: in the most proximal outcrops, comprises coarse to very coarse, pebbly and locally conglomeratic sandstones infilling palaeorelief in the form of megachannel bodies tens to hundreds of metres in width and up to 10m thick. Internally these channel bodies are typically massive with abundant mudchips or intraclasts overlain by large-scale, low angle cross-stratification. Sequence architecture and facies combine to suggest deposition from major glacial outburst events or jokulhaups.

UM2: again in the most proximal areas this forms a single coarsening-upwards package, up to 50m thick, comprising poorly sorted, coarse-grained sandstones characterised by a dense deposit-feeding ichnofauna gradually replaced by medium to large-scale trough cross-bedding with marked bipolar palaeocurrent distribution. The whole assemblage records a significant base level fall, erosion of UM1 and the subsequent progradation of tidally-influenced braid-deltas or, in some cases Gilbert deltas.

UM3: In outcrop the least extensive of the three subunits, this package forms anastomosing channel bodies, each tens of metres in width and up to 5m in thickness. Channel bodies are notably incised into underlying Upper Ordovician sediments in response a further fall in base level and generation of small-scale tunnel valleys.

In the subsurface, correlations supported by detailed core descriptions and image logs confirm in many cases the existence of the 3-fold subdivision observed in the more proximal outcrops of the Ghat area. Grain sizes are generally finer and unit boundaries defined by the presence of mudchip-rich conglomeratic horizons. Integration of these datasets with geobody detection in 3D seismic strongly suggests that all three units were deposited within northwards or westwards-prograding braid-delta systems (sandurs) infilling probable tunnel valleys. Within this framework individual well sections record deposition in proglacial-periglacial braid-delta plain, braid-delta front and braid-delta slope environments, the latter characterised by sediment gravity flows.

The final Upper Ordovician package is poorly represented in outcrop but present in many wells across the Murzuq Basin with a maximum thickness of 81 m. Limited outcrop observations, high resolution correlations and image log data suggest that the Bir Tlacin appears to truncate, often significantly, the underlying Mamuniyat section and thus represents a final Upper Ordovician sequence although in some cases this erosional surface may pass laterally into a thin, condensed and conformable succession. Sedimentologically it is similar in many respects to the Melaz Shugran, comprising several cleaning-upwards parasequences composed of interbedded, intensely dewatered silty mudstones or muddy heterolithics and undisturbed laminated mudstones. The top of the Bir Tlacin is often marked by a condensed horizon, rich in mudchips and, locally iron-rich sandstones. These are sharply overlain by the graptolitic shales of the Lower Silurian Tanezzuft Formation.

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OSTRACODS IN BALTOSCANDIA THROUGH THE HIRNANTIAN CRISES

T. Meidla, L. Ainsaar and K. Truuver

Department of Geology, Institute of Ecology and Earth Sciences, University of Tartu, Ravila 14a, Tartu 50411, Estonia.
Tonu.Meidla@ut.ee, Leho.Ainsaar@ut.ee, Karin.Truuver@ut.ee

Keywords: Ostracods, Ordovician, Baltoscandia.

INTRODUCTION

Geological records indicative of the Hirnantian glaciation are known from many areas, in both high and low palaeolatitudes. Gradual cooling culminated in the double glacial maximum of the early Hirnantian (Brenchley, 2004) and caused dramatic changes in the composition of brachiopod faunas as well as in other invertebrate fossil groups (Sheehan, 2001 and references therein). Changes in the biodiversity of ostracod faunas through this critical interval of climate change have so far received limited attention, probably because the ostracod records from this interval are scarce, especially when compared to older parts of the Ordovician. The aim of this work is to study the changes in ostracod assemblages through the critical Hirnantian interval and their postglacial recovery in the tropical carbonate basin.

Here we present the ostracode data from the Jurmala drillcore, Latvia, and compare their distribution with $\delta^{13}\text{C}_{\text{carb}}$ isotope curve. Jurmala section is drilled in central Latvia, in the middle of the Livonian Tongue (Livonian Basin), the eastward extension of the Scandinavian Basin in between the Estonian and Lithuanian carbonate shelves of the Baltoscandian epicontinental sea (Fig. 1). Representing the deeper part of the basin in East Baltic area, the latest Ordovician deposits of the Livonian Tongue show the greatest potential to preserve a complete sedimentary succession from the interval of a glacioeustatic sea level fall. The uppermost Katian beds, the upper part of the Pirgu Regional Stage are represented by micritic limestone (Paroveja Formation) and marl (Kuili Formation). The Porkuni Stage, considered to be the Baltic equivalent of the Hirnantian Stage (Bergström et al., 2009), is characterized by argillaceous limestone (Kuldiga Formation) and overlying sandy oolitic limestone (Saldus Formation). These beds are covered by argillaceous limestone (Stačiūnai Formation), traditionally correlated with lowermost Llandovery (Juuru Regional Stage; Paškevičius, 1997). Thickness of the Porkuni Stage in the Jurmala section is 14 m (Ainsaar et al., 2010).



Figure 1. Major Ordovician facies zones of Baltoscandia (after Harris et al., 2004), distribution of Ordovician deposits (gray area) and location of the Jurmala drillcore section.

OSTRACOD FAUNA

Composition of pre-Hirnantian ostracode fauna from upper part of the Pirgu Stage in the Jurmala core is similar to that in other sections of the Scandinavian Basin (Meidla, 1996a). *Airina cornuta*, *Daleiella rotundata*, *Spinigerites spiniger*, *Sigmobolbina camarota*, *Pullvillites laevis*, *Hippula edolensis* and other taxa typical of this assemblage disappear at the lower boundary of the Kuldiga Formation with only a few species ranging into the lowermost Kuldiga Formation. This level still marks a nearly complete change in the ostracod succession.

The most diverse latest Ordovician (Hirnantian) ostracod fauna in the world is that of Baltoscandia. Two distinct ostracod assemblages are documented here, the high-diversity *beyrichioco*-dominated assemblage, termed the *Medianella aequa* association (Meidla, 1996b), and the *binodico*-dominated, low-diversity *Harpabollia* fauna (Meidla, 1996a) or *Harpabollia harparum* association (Meidla, 2007). The distribution areas of the assemblages are distinct, mainly following the general palaeodepth zonation of the Baltoscandian Palaeobasin, with *beyrichioco*-dominated assemblages within the onshore reef belt and *binodico*-dominated faunas offshore, including the Livonian Tongue. Ostracod assemblage in the coeval beds of the Jurmala core are typical, comprising the *Harpabollia harparum* association, although the nominate species has not been recorded in the Jurmala core. The assemblage is more diverse in the Kuldiga Formation (containing *Aechmina groenwalli*, *Scanipisthia rectangularis*, *Pseudoancora confragosa*,

Circulinella gailitae, etc.) and of low diversity in the Saldus Formation (only *C. gailitae* present in its middle and upper parts).

The binodicope-dominated *Harpabollia harparum* association contains a number of taxa that are also recorded in the Hirnantian of the Cellon section of the Carnic Alps (*Harpabollia harparum*, *Scanipisthia rectangularis* - (Schallreuter, 1990), whereas its generic relationships to the pre-Hirnantian ostracod assemblages of Baltoscandia are weak (Meidla, 1996a, 2007).

A completely new ostracode fauna appears in basal part of the Stačiunai Formation, right above the last occurrence of *C. gailitae*? in the topmost Saldus Formation. The first appearing species is *Longiscula smithii*. *Rectella procera*, *Microcheilinella mobile*, *M. rozhdestvenskaja* and *Bipunctoprimitia bipunctata* are appearing upward in the section. This assemblage is well known from the lowermost Silurian strata in northern and central Estonia, appearing right above the gap comprising the Late Hirnantian and likely also the earliest Llandovery in this area.

CARBON ISOTOPE CURVE AND OSTRACODS

The Porkuni interval in Baltoscandia is characterized by elevated $\delta^{13}\text{C}$ isotopic values, the HICE interval (Ainsaar et al., 2010). The Hirnantian carbon isotope excursion, the HICE event, is considered to represent the climatically triggered perturbations in carbon cycle, corresponding to the continental glaciation in southern high latitudes (e.g., Brenchley, 2004). The exact correlation of the Hirnantian isotope curves and comparison of changes in carbonate and organic material isotopic composition is matter of discussion (Melchin and Holmden, 2006; Delabroye and Vecoli, 2010). Still, there is obviously strong stratigraphic potential to correlate the isotope curves from tropical carbonate successions of different basins (Bergström et al., 2006).

Carbon isotope curve of the Jurmala section shows rapid increase of $\delta^{13}\text{C}$ values from +1 to +5‰ in the Kuldiga Formation, Porkuni Stage (Isotope zone BC16 by Ainsaar et al., 2010; Fig 2). This is followed by gradual decrease until +2‰ in the Saldus Formation (Zone BC17; Ainsaar et al., 2010). Interestingly, the $\delta^{13}\text{C}$ values continue to decrease in the Stačiunai Formation until they fall close to 0‰

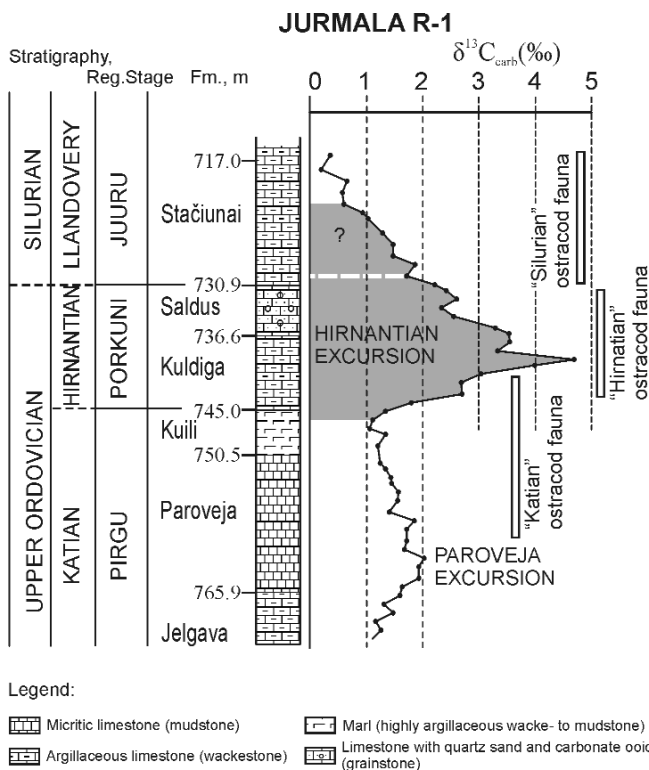


Figure 2. Stable carbon isotope curve (Ainsaar et al., 2010) and ostracode associations in the Jurmala drillcore section.

about 10 m above the traditional Ordovician/Silurian boundary, top of the Saldus Formation. This means, that lowermost beds of the Stačiunai Formation with new “Silurian” ostracod fauna comprise the upper part of the Hirnantian excursion. Carbon isotope curve of Nevada (Finney et al., 1999) shows, that Hirnantian excursion ends below the *N. persculptus* graptolite zone. According to this, there is a part of post-HICE Hirnantian interval with low $\delta^{13}\text{C}$ values (Brenchley, 2004; Melchin and Holmden, 2006). However, the *Harpabollia harparum* seems to range into the Late Hirnantian (early *persculptus* graptolite biozone) in Baltoscandia, as it is co-occurring with *N. persculptus* (identified by V. Jaanusson – see Meidla, 2007) in Sweden.

CONCLUSIONS

It is likely that Ordovician/Silurian boundary in the Jurmala section (and elsewhere in the Livonian Basin) is considerably higher than the Saldus/Stačiunai formation boundary and this should be considered in regional stratigraphic correlations. At the same time it likely means that the new ostracod fauna, considered as “Silurian” before, appeared already in the Hirnantian. This appearance may be due to global warming during the middle to late Hirnantian, in the interval characterized globally as time of “survival fauna” after the second Hirnantian extinction (Brenchley, 2004).

Acknowledgements

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FAUNAL TURNOVER NEAR THE KATIAN/HIRNANTIAN BOUNDARY IN THE PRAGUE BASIN (CZECH REPUBLIC)

M. Mergl

Department of Biology, Faculty of Education, University of West Bohemia in Plzeň, Klatovská 51, 30619 Plzeň, Czech Republic.
mmmergl@kbi.zcu.cz

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INTRODUCTION

The results of climatic changes near the Katian/Hirnantian boundary in the Barrandian area are known since mid of the 19th century. The striking lithological change at this boundary made possible the formal division of the Barrande's etage d5 into the older Königshofer Schichten (= Králův Dvůr Formation) and the younger Kosov Quartzite (=Kosov Formation). The prominent and rapid lithological change from grey-green claystones of the Králův Dvůr Formation into siltstones and sandstones of the subsequent Kosov Formation were always interpreted by the shallowing of the basin.

Chlupáč (1951) and Marek (1952) presented the first reports about the existence of the distinct horizon just below this lithological change. This horizon of calcareous claystone called "Perník Bed" at the very top of the Králův Dvůr Formation contains abundant and distinctive fauna. Havlíček and Vaněk (1965) were the firsts who evaluate this distinctive horizon. These authors referred the top of the Králův Dvůr Formation to the *Cryptolithus kosoviensis* Horizon (= *Marekolithus kosoviensis* Horizon). The source of this distinctive fauna was sought in the Baltic Province and warming of climate was suggested the main cause of immigration of the fauna into the Prague Basin.

The interpretation of the Katian/Hirnantian changes as the results of the glacioeustasy provoked new interest for this interval. Chlupáč and Kukal (1988) interpreted this so called "Basal Kosov Event" by the glacioeustatic shallowing. They interpreted the subgreywacke at the base of the Kosov Formation as regressive extremely shallow water sands with suspicion to continental fluvial deposition. New field observations of Štorch and Mergl (1989) show a remarkable uniformity of the boundary interval in the Prague Basin. They described the thin subgreywacke layer marker (level D) at the base of the Kosov Formation and distinguished the second thicker subgreywacke bed (level F) 1-3 metres above the first bed in all twelve studied localities. These localities are almost regularly distributed in the now preserved part of the Prague Basin. Foreign data and the new unique highway-cut at Levín led to rather different environmental interpretation of this boundary interval. Brenchley and Štorch (1989) interpreted the subgreywacke bed as the glaciomarine sediments (= diamictites). Additional research of Štorch (1990)

brought the new evidence about the glacial origin of mid-European subgreywackes, with proved dropstones and drifted exotic fossils.

In the topmost Králův Dvůr just below diamictites Štorch and Mergl (1989: p. 125) distinguished three distinct levels labelled A, B and C. The level B subdivided into B1, which is formed by olive-green calcareous claystone or muddy limestone. The shales rich in trilobite *Mucronaspis* forms the succeeding level B2. On the basis of the graptolites and presence of a distinctive *Mucronaspis* Fauna Brenchley and Štorch (1989) and Štorch (2006) assigned shales of B2 level and level C below the lower diamictite marker to the Hirnantian. They situated the Rawtheyan/Hirnantian boundary to the interface between B1 and B2 levels. This boundary is now accepted the Katian/Hirnantian boundary in the Prague Basin and its placement is supported by additional new data (Mergl, 2011). However, the origin of the rich fauna of the "Perník Bed" is less apparent. Up to now it was interpreted as the direct evidence of the contemporaneous warming of the Boda event.

DEPTH OF THE PRAGUE BASIN

The Prague Basin is the tectonically based linear depression of the rift origin with rapid depth gradients (Havlíček, 1981). The preserved Ordovician rocks mostly represents the deep water deposits on the basin floor. Besides the earliest Ordovician units, the autochthonous benthic associations (Havlíček, 1982) generally indicate depth corresponding to the outer shelf and upper slope environments recurrently characterised by the cyclopygid and atheloptic biofacies. Black shales with the *Paterula* association, the phyllocarid biofacies, trinucleid trilobites and cratonic diplograptid graptolite fauna are commonly preserved (see review by Fatka and Mergl, 2009). Shallow water associations are preserved only locally in the proximity of suggested shore-lines (e.g. Letná Formation), on the volcanogenic accumulations (Klabava and Bohdalec formations) or on the summits of tectonic rising zones (Zahořany Formation and "*Polyteichus*" facies of the Bohdalec Formation) (Havlíček, 1982).

Unlike the older units, the Králův Dvůr Formation is much uniform representing the monotonous succession of grey claystone. The original depth of whole now preserved basin floor was below the cyclopygid trilobites depth range. The mixture of nectonic cephalopods and cyclopygids with the remopleurid trilobite *Amphitryon*, the minute plectambonitid *Chonetoidea*, the infaunal lingulid *Rafanoglossa*, hyolithids and diverse ostracods characterize the fossil assemblage fauna. Other, slightly shallower elements of the *Foliomena* Fauna are rare in claystone with quartz silt admixture indicating the episode of higher input of shore material. There are none large rhynchoneliiform brachiopods, none bryozoans, none pelmatozoans of the typical BBP (brachiopod-bryozoan-pelmatozoan) fauna which proliferated during the Katian due the Boda event elsewhere. Unlike to the Prague Basin, the rich and diverse BBP faunas with remarkable high bryozoan diversity of Katian age are known from Spain, Sardinia, Montagne Noire, the Carnic Alps, and North Africa (Vennin et al., 1998; Jiménez-Sánchez et al., 2007) and as well as in Baltica and the Avalonia microplate (Fortey and Cocks, 2005). These south European reef-like buildups have remarkably rich fauna, among others the trilobites with distinct temperate climatic zone aspect (Hammann, 1992; Hammann and Leone, 1997, 2007).

Apart of the uppermost part, it is apparent that the Boda event has none direct influence to the composition of now fossilized benthic associations in the almost whole thickness of the Králův Dvůr Formation. The most plausible explanation of this fact is the absence of shallow water inner shelf deposits in the recent extent of the Králův Dvůr Formation. Lithology and preserved fossils of the Králův Dvůr

Formation indicate that the shoals with the BBP fauna were outside now preserved floor of the basin. In addition, the shallow-water environment could be restricted to very narrow belt along margins of the tectonically based depression. Important is low importance of tectonically based rising ridges for development of the BBP fauna. In the upper Katian, these zones likely were not active or were not enough active to compensate the depth of the basin.

SUCCESSION OF THE FAUNA NEAR KATIAN/HIRNANTIAN BOUNDARY

The fauna in the “Perník Bed” (Level B1)

Despite the local deviation in the taxonomic composition, the fauna of the “Perník Bed” is always composed from diverse invertebrate groups. Taphonomy of the “Perník Bed” indicates that all bioclasts were transported. There are no bivalved brachiopod shells, larger shells are often fragmented, deformed and collapsed after the deposition, without further motion in a sediment. Whole trilobites are extraordinary rare. Some larger bioclasts have vertical orientation. Larger bioclasts are partly nested and size-sorted. Small bioclasts of mm scale, mostly juvenile shells or fragments of large shells, entire ostracod valves, isolated plates of echinoderms and machaeridians are size-sorted forming the distinctly bedded layers. Juvenile shells and meraspid stages of trilobites are common. All features indicate the rapid transport of bioclasts by mudflows following rapidly one after another. There are rapid horizontal and vertical changes in composition of fossils assemblages and almost no breaks in sedimentation. Bioturbations of a chondritic type are uncommon and are restricted to upper part of the fossiliferous sequence. The most plausible explanation is repetitive deposition of mudflows that transported bioclasts from ecologically diverse sites. The original differences in composition of local life assemblages are rather well represented by particular laminae of bioclasts.

Almost each locality and separate beds in the particular locality has its “favourite” species and proportion of other common species gently varies. The small variation in composition shows plectambonitoid brachiopods, in which *Aegiria* and *Anoptambonites* are dominant and commonly represented by intact valves (Štorch and Mergl, 1989). This indicates that mudflows almost everywhere affected the benthic plectambonitoid assemblages. The elements of these originally deeper-water plectambonitoid assemblages were mixed with rare and more onshore fauna transported downslope towards the basin floor. New sampling in Praha-Řeporyje locality recovered in 23 cm thick sequence with a remarkable disproportion of the commoner taxa. The lower part of this claystone bed yielded mainly larger bioclasts with the trilobites *Mucronaspis*, *Duftonia*, *Actinopeltis*, *Gravicalymene*, *Stenopareia*, *Marekolithus*, the brachiopods *Dedzetina*, *Cliftonia*, *Eoanastrophia*, *Kozłowskites*, *Leptaena*, *Aegiria*, *Anoptambonites*, branched bryozoans, cystoids and other fauna. The middle part of the beds yielded mainly smaller fauna with the trilobites *Duftonia*, *Diacanthaspis*, *Stenopareia*, *Marekolithus*, the brachiopods *Jezerčia*, *Aegiria*, *Salopina*, *Pseudopholidops* and diverse bryozoans. The upper part bears mainly *Staurocephalus* and *Stenopareia* associated with the brachiopods *Epitomyonia*, *Aegiria*, *Anoptambonites*, *Proboscisambon*, few other brachiopods and almost no bryozoans. Apart of *Mimospira* and *Turbonitella*, the bivalves and gastropods are very rare. Ostracods are abundant and diverse. Bryozoans are represented by some ten species of erect bilaminar ptilodictyines, encrusting cystoporates, ramose and encrusting cyclostomes, and massive, mound-like trepostomates, but zoaria are heavily fragmented.

The fauna above the “Perník Bed” (Level B2)

Claystone and shale above the “Perník Bed” yielded much poorer and different fauna. There are also substantial differences in composition of the fauna between particular localities but the core of the fauna is the same. The trilobite *Mucronaspis grandis* is characteristic. It is associated by ostracods, nuculid bivalves, hyolithids, conularids or, in other localities, by brachiopods and other small sized fauna. Very small-sized brachiopods are strikingly different from brachiopods of the B1 level. There is the undescribed orthid similar to *Dysprosorthis* associated with minute orthotetid and the machaeridian *Lepidocoelus*. Plectambonitoids, pelmatozoans, and bryozoans are totally absent.

The fauna above the “Perník Bed” (Top of level B2)

The claystones just below the subgreywacke marker bed are distinct by a substantial admixture of sand grains and laminae of size-sorted small fossils. The last circa 5-7 cm thick bed in Praha-Řeporyje yielded a distinctive fauna, yet unknown elsewhere in this stratigraphic level of the Prague Basin. The fossil assemblage consists from two species of ostracodes, the machaeridian *Lepidocoelus*, loose plates of indetermined carpoid, very rare fragments of *Mucronaspis*, and distinct shelly fauna with *Kinnella kielanae*, *Hirnantia sagittifera* and the small *Dalmanella* (Mergl, 2011). The fauna has a distinctive character and is the taxonomically poorer example of the *Hirnantia* Fauna. Bioclasts are mostly of a millimetre scale. Larger bioclasts are rare and show traces of fragmentation, leaving only massive cardinal parts of shells intact. It is worth noting that in the eastern territory of Prague the claystone yielded fragments of *Hindella crassa*. This species suggests even a shallower environment of the Hirnantian age.

DISCUSSION

Some distinct features of the fossil association from the “Perník Bed” should be mentioned. There is very high diversity of the fauna, with more than 20 species of trilobites, 20 species of brachiopods, some 15 species of ostracodes, 10 species of bryozoans, some 5 species of hyolithids, and some other 15 species of other invertebrate groups. In total, there are almost 90 species, many of them yet undescribed. Abundance and diversity is remarkably high compared with the older succession of the Králův Dvůr Formation. Another important differences concern the total absence cyclopygidid and remopleurid trilobites, the absence of agnostids, the rarity of lingulate brachiopods and bivalves, the abundance and richness of pelmatozoans and bryozoans, the presence of odontopleurid and cheirurid trilobites, and diverse although rare elements of much shallow-water brachiopods typical for the BBP fauna elsewhere. All these features indicate that, like in the southwestern Europe, Avalonia and North Africa, also the shelves of the Prague Basin were occupied by the highly diversified BBP fauna in the late Katian. However, it is obvious, that extensive bryozoan-pelmatozoan buildups were not developed in the Prague Basin and massive limestone never originated there. The allochthonous occurrence of the BBP fauna at the uppermost Králův Dvůr Formation does not necessary indicate the first appearance of the BBP fauna of the Katian age in the Prague basin (Fig. 1A) The “Perník Bed” more likely indicates the start of the glacioeustatic sea level drop.

The sea level drop gave the impulse to channelling of the distant near-shore shelves even earlier occupied by the BBP fauna (Fig. 1B). The changes in the bathymetry of inner, probably very narrow shelves

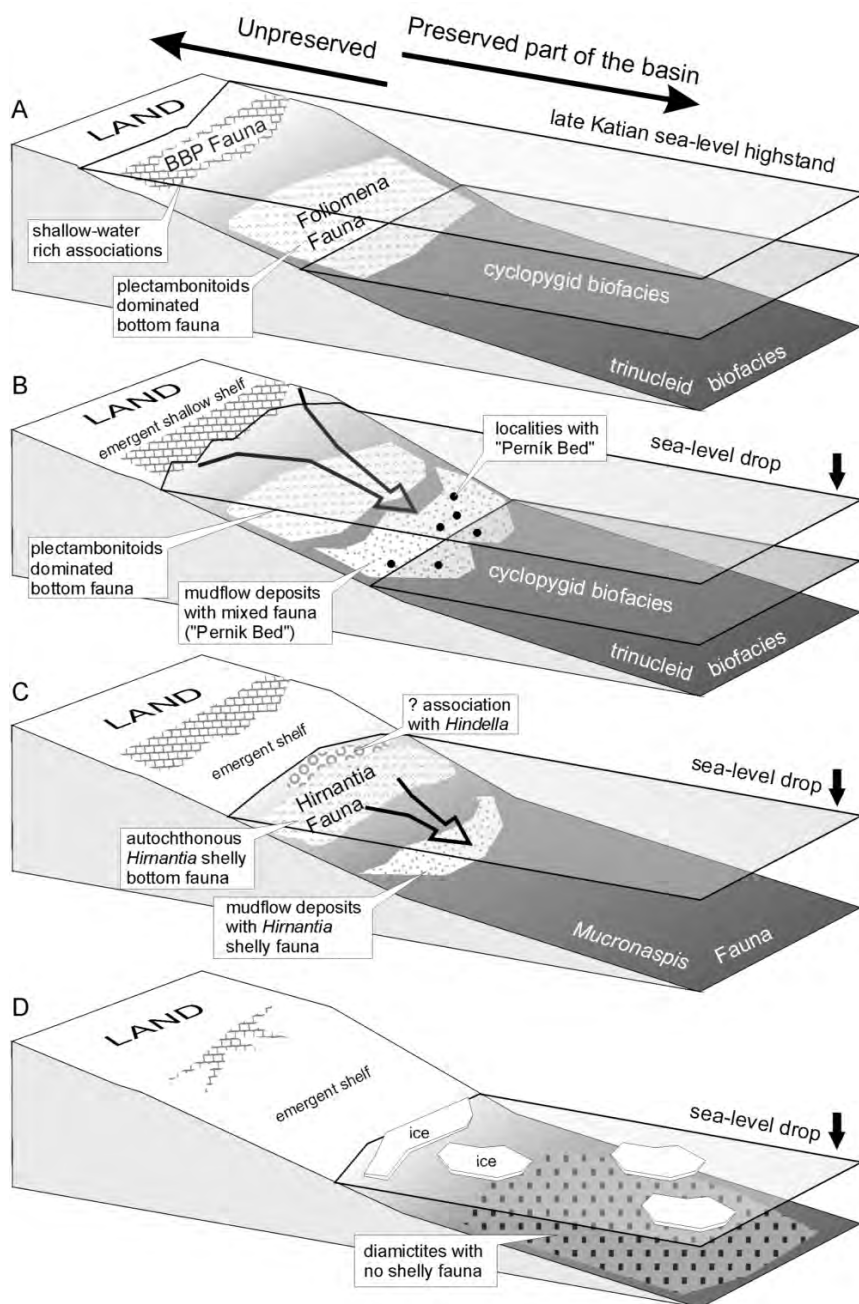


Figure 1. Diagrammatic model of environmental and sea-level changes with distributions of marine benthic associations and suggested depth-related position of significant localities with the "Pernik Bed" of the Prague Basin of in the mid-Katian (1), the latest Katian (= level B1) (2), the earliest Hirnantian (levels B2+C) (3), and in the early Hirnantian (level D) (4).

produced recurrent storm-generated local mudflows that brought a mixture of the shallow water fauna and benthic mid-shelf fauna basinwards onto the deeper basin floor. Continuing sea level drop and rapid climatic deterioration at the Katian/Hirnantian boundary led to eradication of the rich BBP fauna. It is logical to suggest that the first step was replacing of BBP fauna by the shelly *Hirnantia* fauna on now unpreserved narrow inner shelves (BA 3-4) (Fig. 1C). The deeper sea with deposited fossil-rich mudflows occupied the poor *Mucronaspis* Fauna (BA 5). The next sea-level drop and draining of shelves brought shelly *Hirnantia* Fauna basinwards, replacing, at least locally, the *Mucronaspis* Fauna. This realized shortly before the deposition of the diamictite bed marker. The occurrence of the *Hirnantia* Fauna immediately below the base of the Kosov Formation indicates the maximum sea level drop evidenced by the shelly fossils. It is the first unambiguous occurrence of the *Hirnantia* Fauna in the Prague Basin. The estimated depth of now preserved basin floor approached inner shelf (BA3-4). The local presence of *Hindella* indicates even a shallower environment.

Another important feature is moderate endemism of the Bohemian fauna. There are many cosmopolitan or widely distributed peri-Gondwanan genera: trilobites *Dindymene*, *Diacanthaspis* (Fig. 2A-E), *Phillipsinella*, *Mucronaspis*, *Staurocephalus*, *Stenopareia*, brachiopods *Aegiria*, *Anoptambonites*, *Epitomyinia*, *Jezercia*, *Leptaena*, *Ravozetina*. However, many significant brachiopod genera are absent in Bohemia (*Iberomena*, *Leangella*, *Porambonites* etc.). Many species in Bohemia are endemic and the level of endemism of some invertebrate groups (e.g. ostracods, pelmatozoans, bryozoans) has not been evaluated up to now. Preserved trilobite fauna has much deeper aspect, with abundant trinucleid *Marekolithus*, dominance of phacopids *Duftonia*, *Mucronaspis* (Fig. 2F,G) and generally rare and less diverse illaenids, proetids, cheirurids and odontopleurids. The trilobite associations described from Spain and Sardinia (Hammann, 1992; Hammann and Leone, 1997, 2007) are distinct by higher proportion of odontopleurids, illaenids, cheirurids and lichids.

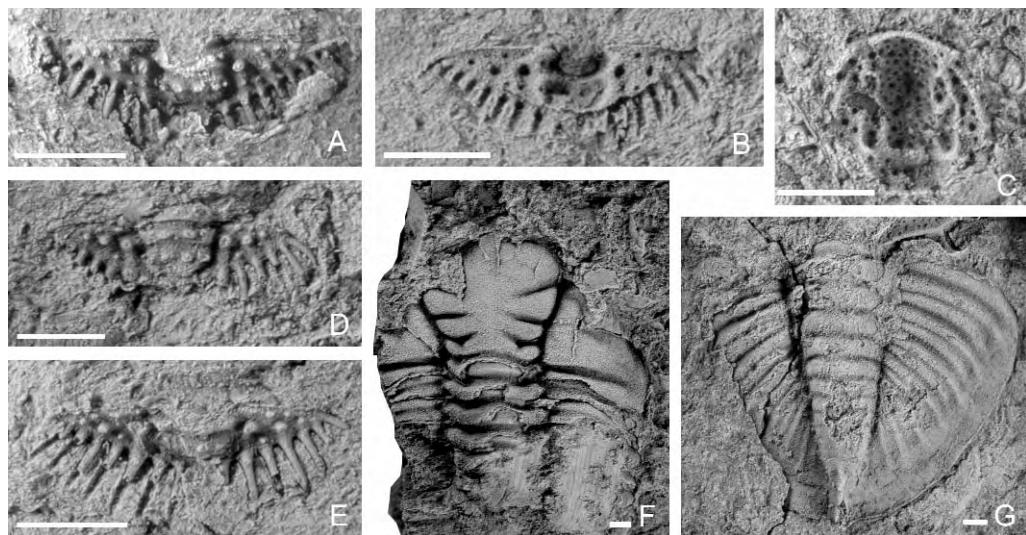


Figure 2. New and characteristic trilobites of the "Perník Bed" of the Králův Dvůr Formation: A-E: Unnamed new species of *Diacanthaspis* sp. A, B, internal and external moulds of pygidium; C, external mould of cranidium; D, E, internal moulds of pygidia. F, G: *Mucronaspis? ganabina* Šnajdr, internal mould of cranidium with incomplete thorax and pygidium. Scale bar = 1 mm.

CONCLUSIONS

Fossil associations of the uppermost Králův Dvůr Formation indicate rapid climatic changes and deterioration of the benthic fauna of the Prague Basin near the Katian/Hirnantian boundary. Although more authors formerly recorded this radical change the new detailed sampling in this interval and comparison with the published foreign data indicate, that the "Perník Bed" is the most likely the product of the initial cold pulse of the late Ordovician glaciation. This is in the striking contrast with the previous explanations (e.g. Štorch and Mergl 1989), which the presence of the rich fauna in the "Perník Bed" explained by the climatic warming of the Boda event and immigration of the "warm" water elements into the Prague Basin shortly before the first cooling. Sea level drop thrived into motion the bioclastic material accumulated onto inner shelves during the warmer period of the Boda event. The basinwards transport of the bioclastic material by mudflows from the shallower margins of the basin produced thin beds of extremely fossiliferous claystone above the unfossiliferous highly bioturbated claystones. The composition of the fauna indicates diverse sources of bioclastic material. This explains the local variation of the fossils content in the "Perník Bed". The progressive cooling with sea-level drop supported the spread of the *Mucronaspis* Fauna over the basin floor. The shallowing, evidenced by shelly fauna, culminated by appearance of the shelly *Hirnantia* Fauna just below the first diamictite bed deposition (Mergl, 2011). As whole, the succession illustrates the rapid turnover of the rich BBP fauna to the much poorer *Hirnantia* Fauna.

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EARLY ORDOVICIAN ARTHROPOD TRACE FOSSILS IN THE PRAGUE BASIN (CZECH REPUBLIC)

M. Mergl

Department of Biology, Faculty of Education, University of West Bohemia in Plzeň, Klatovská 51, 30619 Plzeň, Czech Republic.
mmmergl@kbi.zcu.cz

Keywords: Floian, *Cruziana*, Klabava Formation, Prague Basin, Czech Republic.

INTRODUCTION

Arthropod trace fossils of the Ordovician age are reviewed by Mikuláš (1999); most reports came from the *Cruziana* Ichnofacies of the Sandbian to Katian age (Zahořany and Bohdalec formations). Unlike to SW Europe, where *Cruziana* trace fossils are quite common in the early middle Ordovician in Bohemia this ichnofossil has been unknown in sediments of Tremadocian to Dapingian age. Surprisingly, the probably trilobite trace fossils has newly been observed in the succession from which arthropod fossils are unknown. Arthropod trace fossils are associated with abundant resting ?cnidarian burrows and rare other ichnofossils.

GEOLOGICAL AND GEOGRAPHICAL SETTING

The Prague Basin was been interpreted as the tectonically based linear depression of the rift origin with rapid depth gradients (Havlíček, 1981). Unlike the younger stratigraphic unit, the Tremadocian and Floian units are represented, at least partially, by the shallow-water near shore sediments. The rapid changes of bed thickness and red colour of sediments are often present. The succession of these red sediments is referred to Late Tremadocian and early Floian. However, the dating is based entirely on the macrofauna because of total absence of organic-walled fossils.

The red-coloured sediments are confined to the Milina Formation and the Olešná Member of the Klabava Formation. The main difference between these units is silicification of the sediments, which is marked by bed of cherts intercalated with thinner siltstone and greywacke beds. Several chert beds in the upper part of the Milina Formation are famous for rich although strongly fragmental eozonitids, lingulate brachiopods, echinodermates and trilobites. The trilobite fauna of this unit is the rich example of the *Ceratopyge* Fauna and yielded the lingulate *Thysanotos* which is a good index fossils of the late Tremadocian and Floian elsewhere (Mergl, 2002, 2006; Popov and Holmer, 1994). No trace fossils are associated with this trilobite occurrence. Subsequent 20 to 30 m succession of the red-coloured non-

silicified sediments belongs to the Olešná Member of the Klabava Formation. The Olešná Member is famous for poor and mostly lingulate brachiopod fauna of the *Thysanotos-Leptembolon* fauna. Sampling recovered a remarkably rich lingulate and paterinate brachiopods associated with rare eoothids, paleoscolecids, hyalosponges and demosponges (Mergl, 2002; Mergl and Durápek, 1996). Organic-walled microfauna is absent but the drepanostoid conodonts are fairly frequent. The fossils assemblage in the Olešná Member is monotonous, with ubiquitous lingulate brachiopod genera *Leptembolon*, *Orbithele*, *Celdobolus* and *Dactylotreta*.

The absence of calcareous fossils in the Olešná member is likely secondary. The remarkably high diversity of perfectly preserved phosphatic fossils contrasts with poor preservation of thick-walled shells of eoothids, which were rarely collected in lower part of the Olešná member. Trace fossils of oblique chondritic type are common in some beds. Some beds in the lower part of the Olešná Member look to be totally reworked by the bioturbation, but other beds have uninterrupted fine lamination and no ichnofossils.

Unlike to lower parts of the Olešná Member the middle and upper parts of the Olešná member are poorly exposed. However, the new exposure near village Zaječov (Komárov area, SW part of the Prague Basin, Central Bohemia) shows formerly unknown part of this unit. In three metres thick sequence, the coarse greywacke alternated with very fine brown-coloured siltstone beds. Above these red siltstones is developed a thin (3 cm) intercalation of fine tuffaceous material with graded bedding. The similar intercalations are unknown in other localities, indicating the uniqueness of this exposure and the higher stratigraphical position of the outcropping sequence within the Olešná Member. This volcanic intercalation is followed more than 50 cm thick bed of coarse unbedded greywacke that covers the bedding plane with ichnofossils. The upper bedding plane of tuffaceous intercalation yielded unique association of the trace fossils, which are well preserved in hyporelief on the lower bedding plane of the covering greywacke.

TRACE FOSSILS

There are several types of the trace fossils. The commonest trace fossils are several mm deep resting burrows of variable size (Fig. 1.3 middle left). These are almost circular in outline with a drop-like profile and are slightly inclined by single direction. These resting traces were formed after the digging of *Cruziana* traces. In Figure 1.5 is illustrated, that these circular resting traces were excavated on the bottom of the arthropod traces.

The second trace fossils are of the arthropod origin. They can be compared with the *Cruziana* trace fossils. Traces are formed up to 20 cm long traces, with the maximum width 30 mm. They have sinusoid course on the bedding plane. One track (Fig. 1.4) shows the subcircular trace of digging inside the sinusoid trace, at the deepest bottom with radially arranged endopodal scratches. This digging continues into trilobate trackway. The lateral lobes are fairly irregular, with posterior inclination of scratches. The axial excavation, however, has these scratches bend gently forwards. One bilobed trackway opens small and narrow subhorizontal *Planolites*-like trace (Fig. 1.2). One scratch dug below this subhorizontal trace. The originally smooth surface bedding plane rarely shows rare parallel scratches of the *Dimorphichnus* type.

DISCUSSION

The Olešná beds, despite more than a hundred years of collecting in this unit, did not yield any trilobite or other arthropod body fossils up to now. The presence of *Cruziana*-like trace fossil indicates the presence

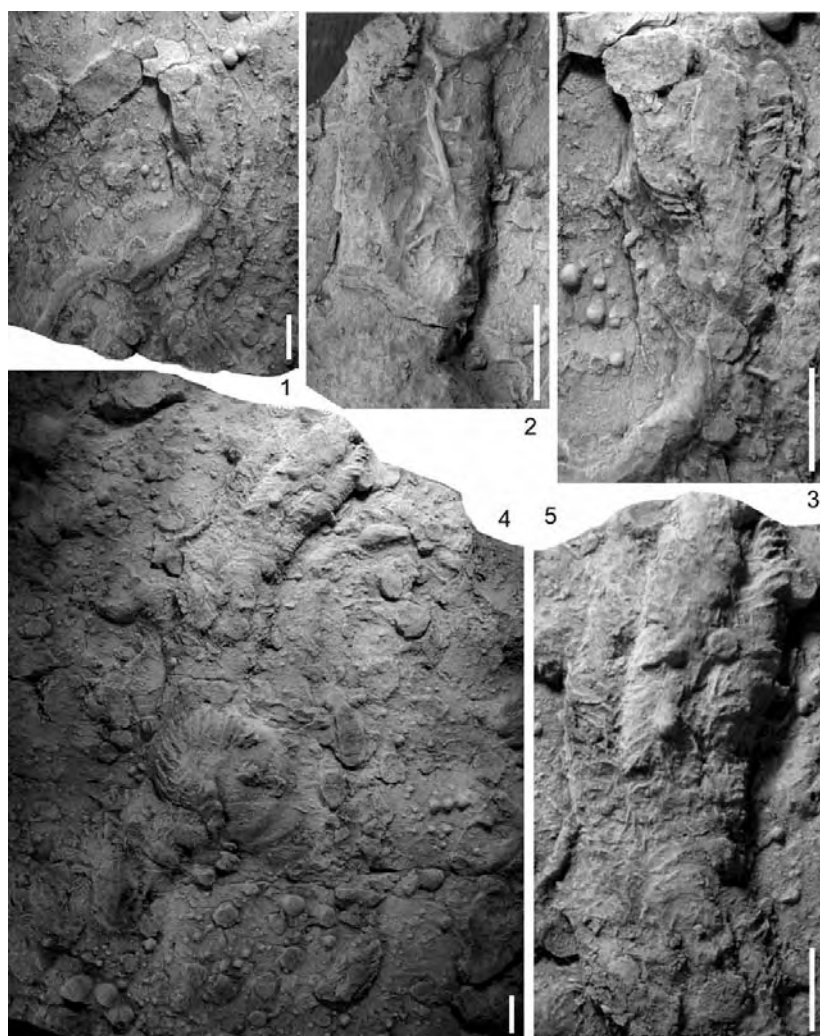


Figure 1. *Cruziana* isp. Middle part of the Olešná Member, the Klabava Formation, Floian. Locality: Zaječov near Komárov, the right bank of the Jalovýpotok Creek; Central Bohemia, the Czech Republic. 1, 3 – bedding plane with sigmoidal trackway and detail of trilobate *Cruziana* isp.; 2 – bilobate *Cruziana* isp. with small *Planolites*-like ichnofossils below. 4, 5 – bedding plane with drop-like resting traces and weakly sigmoidal trackway with the deeply trilobate *Cruziana* isp. at the upper end of the trackway; note the circular digging in the mid-way of the trackway with well preserved scratches, and (5) circular resting trace fossils below *Cruziana* isp. with posteriorly and anteriorly directed scratches.

of arthropod, most likely the trilobites, in the original benthic life associations. Only one type of *Cruziana* indicates low-diversity of the arthropod fauna which should be caused by low amount of organic material in the sediment. The presence of *Cruziana* complete changes the former suggestions about the environment of the Olešná Member. The environment of the Olešná Member was originally compared with the lagoonal environment mostly due to presence of poor lingulate brachiopod fauna (Havlíček, 1982). This model was adopted by Mergl (1986). The recovery of highly diversified lingulate fauna (Mergl, 1995, 2002)

with the taxa elsewhere characteristic for deeper and wholly marine environments, and observations about diversity of sponges and conodonts indicate, that deposition of the Olešná Member took place on moderately deep shelf with normal marine conditions. The trace fossils indicate the presence of the benthic invertebrates that are absent as the body fossils in the Olešná Member. The traces also brought evidence that absence of calcareous fossils in the Olešná Member is diagenetic, with destroy of calcareous and organic-walled shells and bioskeletons.

Drop-like resting traces preserved at the bottom of *Cruziana* indicate two successive phases of the occupation of the sea floor. After the first arthropod search followed the digging of drop-like resting traces, which can be of the cnidarian origin.

CONCLUSIONS

The first occurrence of the *Cruziana* ichnofossil in the Olešná Member of the Klabava Formation (Floian) is the only direct evidence of the arthropod presence in this lithostratigraphic unit. This agrees with otherwise high diversity of lingulate brachiopods, sponges and conodonts. Although seemingly marginal in significance, the presence of *Cruziana* radically changes the suggestions about the environment, making the original suggestions about the lagoonal conditions untenable. It is also the earliest *Cruziana* occurrence in the Ordovician of the Prague Basin.

Acknowledgements

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NEW STABLE ISOTOPE DATA AND FOSSILS FROM THE HIRNANTIAN STAGE IN BOHEMIA AND SPAIN: IMPLICATIONS FOR CORRELATION AND PALEOCLIMATE

C.E. Mitchell¹, P. Štorch², C. Holmden³, M.J. Melchin⁴ and J.C. Gutiérrez-Marco⁵

¹ Department of Geology, University at Buffalo-SUNY, Buffalo, NY, USA 14260-3050. cem@buffalo.edu

² Academy of Sciences of the Czech Republic, Institute of Geology, Rozvojová 135, 165 02 Praha 6, Czech Republic.
storch@gli.cas.cz

³ Department of Geological Sciences, Univ. of Saskatchewan, Saskatoon, SK, Canada S7N 5E2. chris.holmden@usask.ca

⁴ Department of Earth Sciences, St. Francis Xavier University, Antigonish, NS, Canada, B2G 2W5. mmelchin@stfx.ca

⁵ Instituto de Geociencias (CSIC-UCM), José Antonio Nováis 2, 28040 Madrid, Spain. jcgrapto@geo.ucm.es

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INTRODUCTION

Rocks in the Late Ordovician paleotropics commonly exhibit the now well-known, positive Hirnantian Isotopic Carbon Excursion (HICE). This event is coincident with both continental scale glaciation in Gondwana and dramatic extinction across the marine realm (Delabroye and Vecoli, 2010). Both the proximate cause of the isotopic excursion and the ultimate drivers of large scale cooling remain the subject of debate. Suggestions range from tectonic effects on weathering or changes in biological productivity, through large basaltic eruptions to gamma ray bombardment. Discussion of these alternative models is beyond the scope of this short paper, however. Our intent is to briefly report the direct association of a new high resolution $\delta^{13}\text{C}_{\text{organic}}$ record from rocks at a high latitude site that also bears biostratigraphic and sequence stratigraphic data needed to link Hirnantian oceanographic changes (especially those recorded in the paleotropics) with glacial events in the peri-Gondwanan realm (Delabroye and Vecoli, 2010; Young et al., 2010).

NEW HIGH RESOLUTION HICE IN BOHEMIA

We collected 152 samples (approx. 50-100 g each) for $\delta^{13}\text{C}_{\text{organic}}$ and total organic carbon (TOC) analyses from latest Katian to earliest Rhuddanian dark shale, siltstone, diamictite and sandy mudstone. Mid Hirnantian rocks are not readily accessible in this region at present. Samples were analyzed using facilities managed by C. Holmden at the University of Saskatchewan (Fig. 1; data and methodological details available upon request).

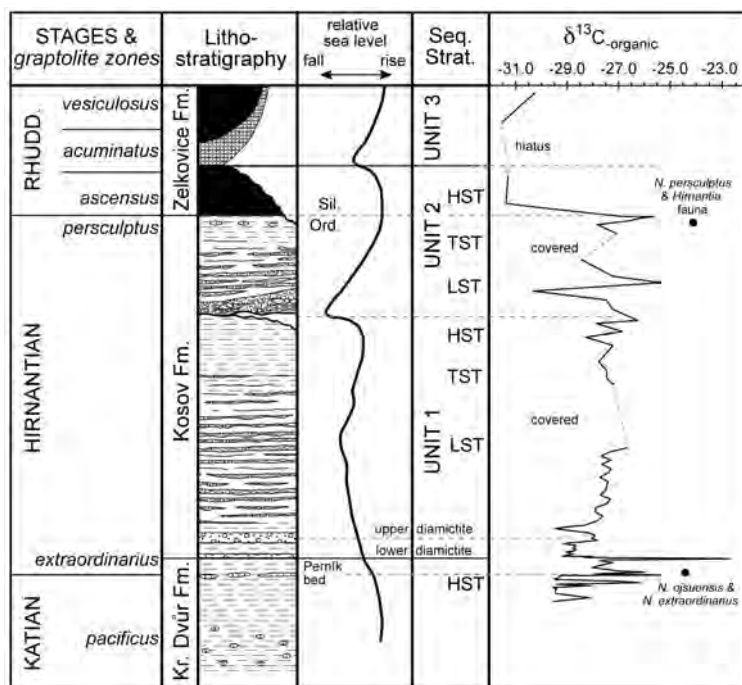


Figure 1. Summary of $\delta^{13}\text{C}_{\text{org}}$ results from Levin (lower part) and Hlásná Třebaň (upper part) plotted relative to regional stratigraphic composite through the Hirnantian based on sections in the vicinity of Praha, Czech Republic (modified from Štorch, 2006).

Early Hirnantian sections

We sampled the uppermost 4 m of the Králův Dvůr Formation (including the calcareous mudstones of the Perník Bed; Brenchley and Štorch, 1989) and 25 m of the overlying Kosov Formation in a fresh road cut (N49° 55.677', E14° 00.770') near the village of Levin (see Brenchley and Štorch, 1989, fig. 6). Samples through the lower part of this succession (including the two prominent glaciomarine "diamictite" beds at 0 and 3.5 m above the base of this unit) generally were taken at 20 cm intervals and those in the more expanded, storm-dominated succession above the diamictites at one-meter intervals (Fig. 2). The lower part of this succession was also sampled at a somewhat more weathered railway cutting at the village of Zadní Třebaň (N49° 54.865', E14° 12.299') from which Štorch (1989) reported the occurrence of *Normalograptus* cf. *ojsuensis* together with *Mucronaspis grandis* immediately above the Perník Bed (discussed further below). On the basis of that and other evidence, Štorch (2006) placed the base of the Hirnantian Stage at this graptolite-bearing level in the uppermost Králův Dvůr Formation.

TOC in the Králův Dvůr and Kosov formation samples is generally quite low (0.1 to 0.5 weight percent), with the highest TOC occurring in the lowermost Kosov Fm. diamictites. The $\delta^{13}\text{C}_{\text{org}}$ values are uncorrelated with TOC. In both of the basal Hirnantian sections, $\delta^{13}\text{C}_{\text{org}}$ values obtained in the upper Králův Dvůr Formation rise rapidly from a baseline value of about -29.3‰ (defined by the lowest 7 samples) and display initially fluctuating values in the lower part of the record before transitioning to a strong positive excursion

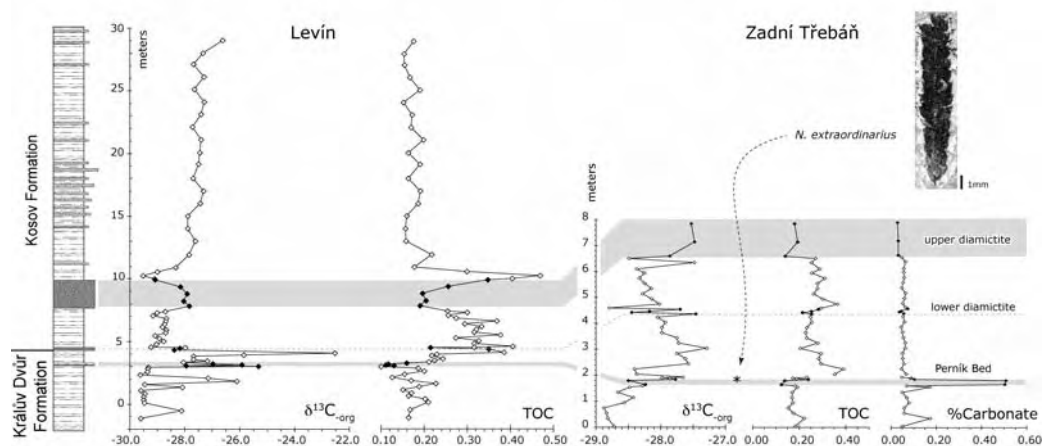


Figure 2. Organic carbon isotope data, total organic carbon (TOC) and carbonate concentration through the upper Králův Dvůr and lower Kosov formations at Levín and Zadní Třebáň.

in the lowermost Kosov Formation with peak values of about -27.5 to -26.5‰. These values are comparable to the 2-3‰ shift commonly seen in oceanic sites in the basal Hirnantian (Melchin and Holmden, 2006; Fan et al., 2009). A dark gray shale sample from immediately below the basal Kosov "lower diamictite" yields a much higher value of -22.53‰. The resulting 7‰ shift from baseline calculated using that sample is of the same scale as shifts observed in more on-shore sections in the paleotropics during the early Hirnantian (Melchin and Holmden, 2006; LaPorte et al., 2009). It is unclear whether this particularly heavy value has paleoenvironmental significance or is merely an artifact of local mixing or selective preservation of organic matter (or both) at this level in the stratigraphy.

The Perník Bed and the first appearance of Hirnantian graptolites both lie within the rising lower limb of the early HICE isotopic peak in these sections. From the lowermost Kosov Formation peak, $\delta^{13}\text{C}_{\text{org}}$ values decline somewhat toward the upper diamictite in both sections and then, at Levín where our samples continue for another 15 m, $\delta^{13}\text{C}_{\text{org}}$ values climb slowly to a maximum of about -27.0‰.

Late Hirnantian section

We collected 27 samples from an approximately 42 m-thick interval through the upper part of the Kosov Fm. and the base of the Želkovice Fm. in a natural bluff exposure above the village of Hlásná Třebáň (N 49° 55.359', E 14° 12.761'; see Štorch 2006, fig. 4). This section contains the upper part of the Kosov Fm. Sequence 1 of Štorch (2006) and the entire succession of Sequence 2 (the base of which lies at 18.5 m in our measured section; Fig. 3) as well as the lower part of the Rhuddanian-Aeronian Želkovice Formation (base at 41.5 m). Once again, TOC in the Kosov Formation samples are generally quite low (0.1-0.2 weight percent) with the exception of the lowermost Želkovice Formation strata where TOC rises abruptly to 2 to 7 weight percent. Samples through most of the upper Kosov at Hlásná Třebáň yielded $\delta^{13}\text{C}_{\text{org}}$ values that hover around -27.4‰, corresponding closely to those from the top of the lower Kosov Fm. at Levín; however, the uppermost samples show a slight positive shift to a peak value of -25.61‰. Elsewhere in the region these uppermost Kosov beds produce a shelly *Hirnantia* fauna as well as specimens

of *Normalograptus persculptus* (Štorch, 2006 and references cited therein). The black shale succession in the overlying Želkovice Formation begins with an c. 15 cm thick *Akidograptus ascensus* Zone and, after a disconformity (Štorch, 2006, fig. 4), sedimentation continues from *C. vesiculosus* Zone through middle Aeronian in this outcrop. Two samples each from the *Akidograptus ascensus* and *Cystograptus vesiculosus* zones exhibit $\delta^{13}\text{C}_{\text{org}}$ values of about -31‰, as is common in the early Rhuddanian.

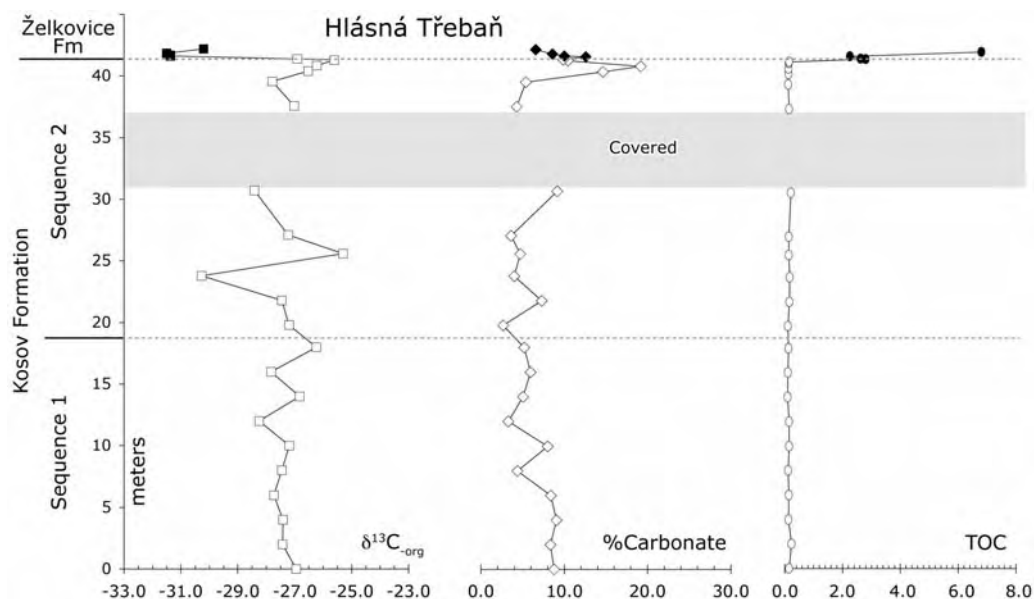


Figure 3. Organic carbon isotope data, total organic carbon (TOC) and carbonate concentration through the upper Kosov and lower Želkovice formations at Hlásná Třebaň.

NEW EARLY HIRNANTIAN GRAPTOLITES IN BOHEMIA AND SPAIN

Štorch (1989, 2006) reported specimens of *Normalograptus ojsuensis* from a thin band of shales immediately above the Perník Bed in the uppermost Králův Dvůr Formation at four sites in the study region, including Zadní Třebáň. We have now recovered *N. extraordinarius* at this site in addition to the more abundant *N. ojsuensis*. We have also recovered these two species from the basal part of the tempestite and diamictite dominated, Hirnantian Rio San Marco Fm. in Sardinia (Štorch and Leone, 2003; Leone et al., 2009). *N. ojsuensis* is also present in Niger (Legrand, 1993), but occurs there in transgressive strata deposited *between* the two major Hirnantian glacial advances (Ghienne et al., 2007). Accordingly, it may be a somewhat younger occurrence than that in the Králův Dvůr Formation.

Restudy of collections made in Spain by JCGM from the northern flank of the Guadalmez Syncline, Central Iberian Zone (N38° 45.115', E04° 58.565') reveals yet another occurrence of pre-glacial, Hirnantian graptolites (Pl. 1), in this case abundant specimens that we refer to *Neodiplograptus charis* (Mu and Ni, 1983). They occur in a laminated shale that lies stratigraphically between the top of the Bancos Mixtos and the glaciomarine Chavera Shales. The Urbana Limestone, which in many places occurs between these units, is not present at this locality. At present we are uncertain whether the graptolite-bearing beds

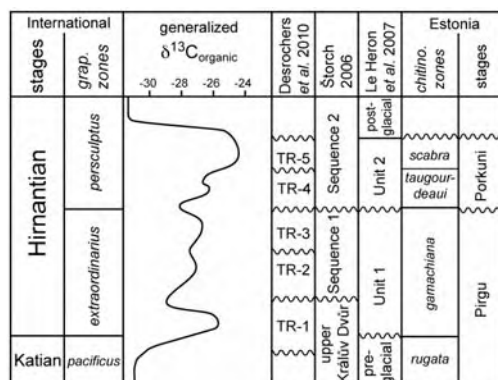


Figure 4. Summary correlation of cycles from Desrochers et al. (2010), Štorch (2006) and Le Heron et al. (2007).

are part of the depositional sequence that contains the Chavera Shale or not. In Tibet and SE China, however, *N. charis* appears to be restricted to the uppermost part of the *Paraorthograptus pacificus* Zone (upper *Diceratograptus mirus* Subzone) and the *N. extraordinarius* Zone, where it is commonly associated with *N. ojsuensis* (Chen et al., 2005). A calcareous coquina present immediately above the Bancos Mixtos at this locality (Gutiérrez-Marco, 1995) yields conodonts indicative of the *Amorphognathus ordovicicus* Conodont Biozone, which spans the Katian-Hirnantian boundary everywhere (Del Moral and Sarmiento, 2008).

REINTERPRETATION OF THE PERNÍK BED

As noted above, the Perník Bed of the uppermost Králův Dvůr Formation lies within the rising limb of the large $\delta^{13}\text{C}_{\text{org}}$ positive excursion that is present in the Bohemian succession. This unique bed is more calcareous than the other units in the predominantly clastic Late Ordovician succession in this region. It contains a moderately diverse, shelly fauna dominated by brachiopods of the to *Proboscisambon* Community of the *Foliomena* Fauna (Havlíček, 1982). The main Perník Bed fauna has been interpreted as containing warm water immigrants suggestive of a connection to the Boda Event (Fortey and Cocks, 2005). On the other hand, the Perník Bed grades both laterally and vertically into shales. Towards the top of the Perník Bed the shelly faunal assemblage becomes less diverse and is dominated by mucronaspid trilobites and varied ostracods. It is *within* these transitional layers at Zadní Třebáň that *N. ojsuensis* and *N. extraordinarius* first appear (Fig. 2; Pl. 1, fig. 6). Finally, some brachiopod index taxa of the *Hirnantia* fauna (*Hirnantia sagittifera* and *Kinella kielanae kielanae*) have been found immediately above the Perník Bed in a temporary outcrop at Praha-Řeporyje (Mergl, in press). The Hirnantian age of the shale overlying Perník Bed is, thus, well dated by both graptolites and shelly fauna.

Vertical change in the faunal composition of the Perník Bed and lateral changes in its lithology suggest condensed sedimentation in a siliciclastic-starved setting. Considering the association of the Perník Bed with the onset of the HICE, we suggest that this bed may reflect cooling and increased aridity associated with ice cap growth, which in turn reduced clastic input and permitted a temporary development of conditions conducive to deposition of cool water carbonate. If that is so, then it may also be the case that the Urbana Limestone and its equivalents in Spain as well as the marly limestones of the uppermost Domusnovas Fm. in Sardinia may reflect a similar genesis.

IMPLICATIONS FOR CORRELATION OF THE BASE OF THE HIRNANTIAN STAGE

The brief 3-5‰ $\delta^{13}\text{C}_{\text{org}}$ positive excursion documented here in association with the local occurrence of *N. ojsuensis* and *N. extraordinarius* confirms the location of the base of the Hirnantian Stage previously identified by Štorch (2006). This is followed by a sharp return to lower $\delta^{13}\text{C}_{\text{org}}$ values and then by a long

steady climb to a second positive excursion of between 4 and 5‰ above the late Katian and early Rhuddanian baseline (Fig. 1). That the HICE is a multiple isotope carbon excursion has now been documented in several sites around Laurentia (Melchin et al., 2003; Melchin and Holmden, 2006; LaPorte et al., 2009), including Anticosti Island (Desrochers et al., 2010), as well as in SE China (Fan et al., 2009). Our new data support the correlation between the paleotropics and the Gondwanan margin proposed by Desrochers et al. (2010), which they derived from their recent sequence stratigraphic interpretation of Hirnantian rocks at Anticosti Island and contradict the correlation advocated by Young et al. (2010). Our data suggest that the first HICE peak occurred during the early part of the *N. extraordinarius* Zone (Fig. 4), associated with a modest eustatic sea level rise (TR-1 of Desrochers et al., 2010), which brought early Hirnantian graptolites onto the peri-Gondwanan massifs in Spain, Sardinia, and Bohemia, and culminated in declining $\delta^{13}\text{C}_{\text{org}}$ values across the following sequence boundary. The first major Hirnantian ice sheet advance is recorded in the basal Kosov diamictites and overlying storm bed succession of Sequence 1 of Štorch (2006) and the Unit 1 synglacial and interglacial sediments of Ghienne et al. (2007) and Le Heron et al. (2007). This advance is reflected by a very modest tropical sea surface temperature decline as recorded in the Anticosti Island succession (Finnegan et al., 2011) and was evidently considerably smaller than the main later Hirnantian advance (Desrochers et al., 2010; Moreau, 2011). We suggest that this whole interval is represented by the *N. extraordinarius* Graptolite Zone and the *Belonechitina gamachiana* Chitinozoan Zone of Anticosti Island and Estonia. In most of the Estonian sections studied for carbon isotopes, however, the *B. gamachiana* Zone is absent or extremely condensed – at its thickest (at Kaugatuma – Brenchley et al.,

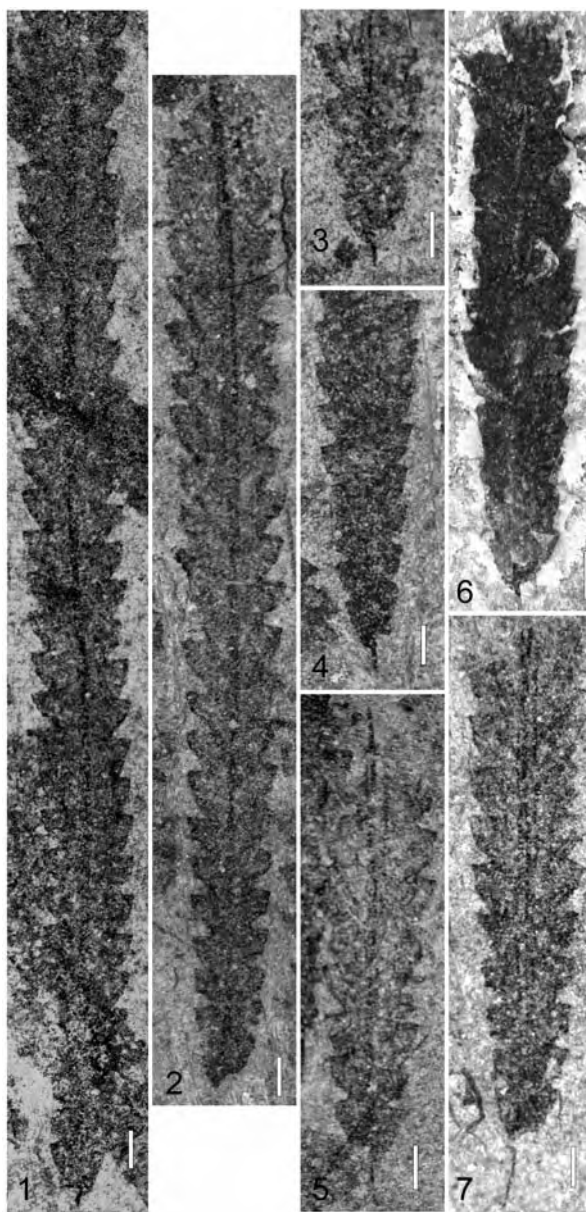


Plate 1. 1-5, 7: *Neodiplograptus charis* (Mu and Ni, 1983) from 3 km NW of Guadalmaz, central Spain; 6: *Normalograptus extraordinarius* from uppermost layers of the Perník Bed at Zadní Třebáň. Scale bar is 1 mm in length in all figures.

2003; Kaljo et al., 2008) that zone is less than 5 m thick and is bounded both below and above by discontinuity surfaces. Therefore, it is not surprising that the lower HICE peak has not been sampled in Estonia.

The main late Hirnantian ice advance appears to correspond to the interval of the hiatus and overlying low-stand in Sequence 2 within the upper Kosov Formation. The upper part of Sequence 2 succession, still within the interval of high $\delta^{13}\text{C}$ values, clearly deepens upward. If this deepening reflects eustatic sea level change rather than local subsidence, then this association suggests that high $\delta^{13}\text{C}_{\text{org}}$ values persisted into the early part of the post-glacial interval and that this pattern is obscured on Anticosti Island and in the Estonian succession by condensation or omission as result of rapid sea level rise during deglaciation. Conversely, it may be that the end of the HICE is slightly diachronous - younger in Bohemia than in the paleotropics in relation to the post-glacial transgression as a result of regional differences in carbon cycling processes. Testing of these competing hypotheses will require additional, high-resolution data sets, especially in a high-latitude site, with precise biostratigraphic and sedimentological control.

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THE TREMADOCIAN DEPOSITS OF THE ARGENTINIAN EASTERN CORDILLERA: A SCANDINAVIAN SIGNAL IN THE CENTRAL ANDES

M.C. Moya^{1,2} and J.A. Monteros¹

¹ CI-UNAS.

² CONICET. Geología, Facultad de Ciencias Naturales, Universidad Nacional de Salta, Avda. Bolivia 5150, 4400 Salta, Argentina.
moyacris@arnet.com.ar; julpaleo@unsa.edu.ar

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INTRODUCTION

One of the objectives of this paper is to present an updated biostratigraphic scheme of the Tremadocian successions of northwestern Argentina. Another objective is to show the close affinity between graptolite faunas contained in Tremadocian deposits of the Argentinian Eastern Cordillera (with Mojotoro Range as the main reference) and those referred to Tremadocian sequences in Scandinavia, particularly those exposed in the Oslo-Scania Region. The idea is to discuss an intercontinental correlation, in terms of depositional sequences and their fossil content, based on global events, mainly eustatic, which would have influenced the stratigraphic arrangement in both regions.

THE TREMADOCIAN OF THE ARGENTINIAN EASTERN CORDILLERA

Tremadocian successions cropping out in the Eastern Cordillera of Argentina (Fig. 1a, b) are the most complete in South America. Tremadocian deposits are part of the Santa Victoria Group (Furongian-Sandbian), consisting of a succession of alternating bodies of sandstone and shale accumulated in coastal and marine environments from inner and outer shelf. The Santa Victoria Group is a clastic supersequence, very fossiliferous, bound at the base and the top by Type 1 or subaerial unconformities (Vail et al., 1984; Catuneanu, 2006). This supersequence represents the records of a second order tectono-eustatic cycle or megacycle (Einsele et al., 1991). This megacycle (the *Victorian Cycle* of Moya, 2002, 2008) includes nine minor cycles, corresponding to transgression-regression episodes, documented in depositional sequences and bounded by sedimentary unconformities whose hiatuses coincide with relative sea level falls (Moya, 2008).

The most complete records of the Santa Victoria Supersequence outcrop along the eastern flank of the Argentinian Eastern Cordillera (Fig. 1b, nos. 1-6). Among the areas identified in the figure, the Mojotoro Range provides one of the best reference sections for study the Tremadocian deposits. In contrast,

Tremadocian successions of the western flank of the Eastern Cordillera (Fig. 1b, nos. 7-10) are incomplete because these areas were exposed (with non-deposition or even erosion) during the *Ceratopyge* Regressive Event (CRE, Late Tremadocian). The transgression that followed the CRE was accompanied by a tectonic event that caused the collapse and sudden subsidence of the eroded areas (Moya, 1997), so that the Tumbaya unconformity separates the coastal and inner shelf deposits (*Anisograptus matanensis* Zone) from other outer shelf deposits (*Tetragraptus phyllograptoides* Zone) (Moya et al., 1998).

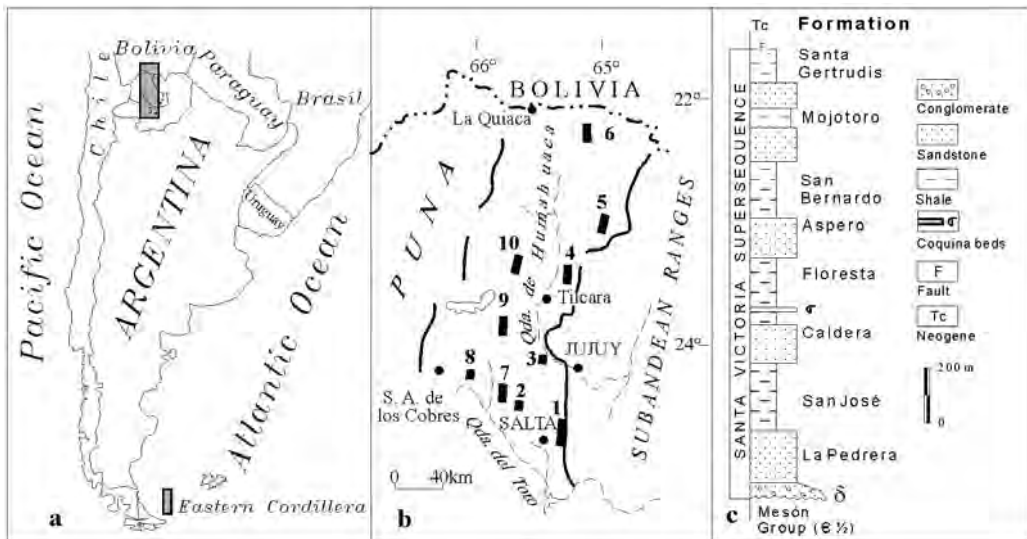


Figure 1. Location maps: a, Location of Eastern Cordillera Region in South America. b, Location map of classical sections with Tremadocian deposits: 1, Mojotoro Range; 2, Lesser Hills; 3, Yala-Reyes road; 4, Tilcara Range; 5, Zenta Range; 6, Santa Victoria Range; 7, Pascha Hills; 8, Angosto La Quesera; 9, Angosto del Moreno; 10, Aguilar and Cajas Ranges. c, Stratigraphy of the Santa Victoria Supersequence at the Mojotoro Range.

Figure 1c shows the Santa Victoria Supersequence in the Mojotoro Range. The Iruya Unconformity at the base, separates it from the Mesón Group (middle Cambrian). At the top, the supersequence is truncated by faults affecting the Santa Gertrudis Formation (Sandbian, Carlorosi et al., 2011). Furongian records are represented by conglomerates and fluvial sandstones which form the lower part of the La Pedrera Formation (Figs. 1c and 2). The hiatus of the discontinuities that limit these deposits coincide with the Lange Ranch Eustatic Event (LREE) and the *Acerocare* Regressive Event (ARE), respectively (Moya, 2008). In other parts of the Eastern Cordillera, the Furongian sequence is complete and contains, among others, *Hirsutodontus hirsutus* and *Parabolina* (*Neoparabolina*) *frequens argentina* (Moya and Albanesi, 2000). Tremadocian deposits in the Mojotoro Range include the mid-upper part of the La Pedrera Formation to the lower section of the San Bernardo Formation, where the Tremadocian-Floian transition is recorded (Fig. 2).

TREMADOCIAN GRAPTOLITE ZONES

Figure 2 integrates the information obtained in 11 stratigraphic sections located in the Mojotoro Range; the graptolite and trilobite zones that are presented here were defined based on fossil material that comes almost totally from those sections. The figure shows the development of the Tremadocian succession

through four transgression-regression cycles, represented by an equal number of sequences. The hiatus of the discontinuities that separate them coincide with relative sea level falls which were documented and paleontologically controlled in the Tremadocian of Scandinavia (Erdtmann, 1995; Nielsen, 2004). The biostratigraphic scheme of Fig. 2 is almost identical to the one that Erdtmann (1995) and Maletz et al. (2009) suggest for the Tremadocian of the Oslo-Scania region. However, there are substantial differences in the thickness and lithology of the sequences: i) In Scandinavia, the whole Tremadocian is documented in a condensed succession less than 100 m thick (Nicoll et al., 1992). In northern Argentina this interval is represented by extended successions, whose thicknesses range from 1,000 m (Fig. 2) to more than 2,000 m (Santa Victoria Range, Harrington, 1957). ii) Scandinavian sequences are composed by shale/limestone, while northern Argentinian sequences are shale/sandstone.

It is clear that the relationship between subsidence and sediment supply on each basin was the determining factor that made the difference in thickness of the Tremadocian deposits in both regions, which probably joined substrate topographies that were also different. If one accepts that both regions were in similar mid-paleolatitudes, the climate would not have been the factor that led to the different facies recorded: clastic in northern Argentina and clastic-carbonate in Scandinavia. Dronov and Rozhnov (2007 and references cited therein) indicate that the limestones intercalated in the Scandinavian sequence have characteristics of cold-water carbonate sedimentation. In addition, biostratigraphic controls in the two basins mark a synchrony in the erosion and deposition events, indicating that in both the sedimentation was influenced by eustatic factors (Catuneanu, 2006). Thus the interest for an intercontinental correlation based on eustatic controls and graptolite biozones presented here.

Rhabdinopora rustica Zone

It is the oldest biozone reported so far for the Tremadocian of the Eastern Cordillera (Moya et al., 1994), and documents the earliest appearance of planktonic graptolites in the region. *R. rustica* (Bulman) (Fig. 2a) appears in the lower interval of the San José Formation that crops out in the Yala-Reyes road (Fig. 1b), where it coexists with *Jujuyaspis keideli* Kobayashi and *Parabolinella argentinensis* Kobayashi. This biozone corresponds with the *Rhabdinopora praeparabola* Zone of Scandinavia and eastern North America (Egenhoff et al., 2004; Maletz et al., 2009) (Fig. 3).

Rhabdinopora flabelliformis Zone

It was recognized in the Mojotoro Range, Lesser Hills and Yala-Reyes area (Figs. 1b and 2b). It is developed in the middle-upper interval of the San José Formation, where undetermined subspecies of *R. flabelliformis* (Eichwald) coexist with, among others, *J. keideli*, *P. argentinensis*, *Apatokephalus exiguus* Harrington and Leanza and many agnostids. Based on this assemblage and in the fact that the fossiliferous levels definitely are located below the first appearance of *Anisograptus*, it is considered that the *A. flabelliformis* Zone in the Eastern Cordillera corresponds to the *A. f. parabola*/*R. f. socialis* Zone of Scandinavia, Newfoundland and other parts of the world (see Cooper, 1999). The *A. f. parabola* Zone from the Famatina System (Gutiérrez-Marco and Esteban, 2005), Graptolite Assemblage II (Moya et al., 1994) and Interval II (Moya et al., 2003a) recognized in various areas of the Eastern Cordillera, would correspond to the biozone discussed here.

The *R. rustica* and *R. flabelliformis* zones are developed within a single transgression-regression cycle, which starts before the appearance of *J. keideli* and the earliest planktonic graptolite, and ends before the

FAD of *Anisograptus*. Following Erdtmann (1995), the hiatuses associated to the discontinuities that define this interval, would correspond to the *Acerocare* Regressive Event (ARE) and the Black Mountain Eustatic Event (BMEE) respectively (Figs. 2 and 3). Deposits included in this cycle (San José Formation and probably part of the La Caldera Formation), would correspond to the depositional sequence that includes the oldest Tremadocian levels of the Lower Alum Shale (Fig. 3) in Scandinavia; that is, the equivalent to the *Boeckaspis* Zone (Bruton et al., 1988), in which *R. praeparabola* (Bruton, Erdtmann and Koch), *R. f. parabola* (Bulman), *R. f. flabelliformis* (Eichwald) and *J. keideli norvergica* Henningsmoen were recorded. In Scandinavia, the *Boeckaspis* Zone precedes the FAD of *Anisograptus* and follows the *Acerocare* Zone (latest Cambrian), which records *Parabolina* spp. (Bruton et al., 1988). This means that these Cambrian deposits would correspond with the beginning of the first transgression event at the base of the Victorian Cycle, developed during the Furongian, and are paleontologically documented at the Angosto del Moreno (Fig. 1b): Moya and Albanesi (2000), Moya et al. (2003b).

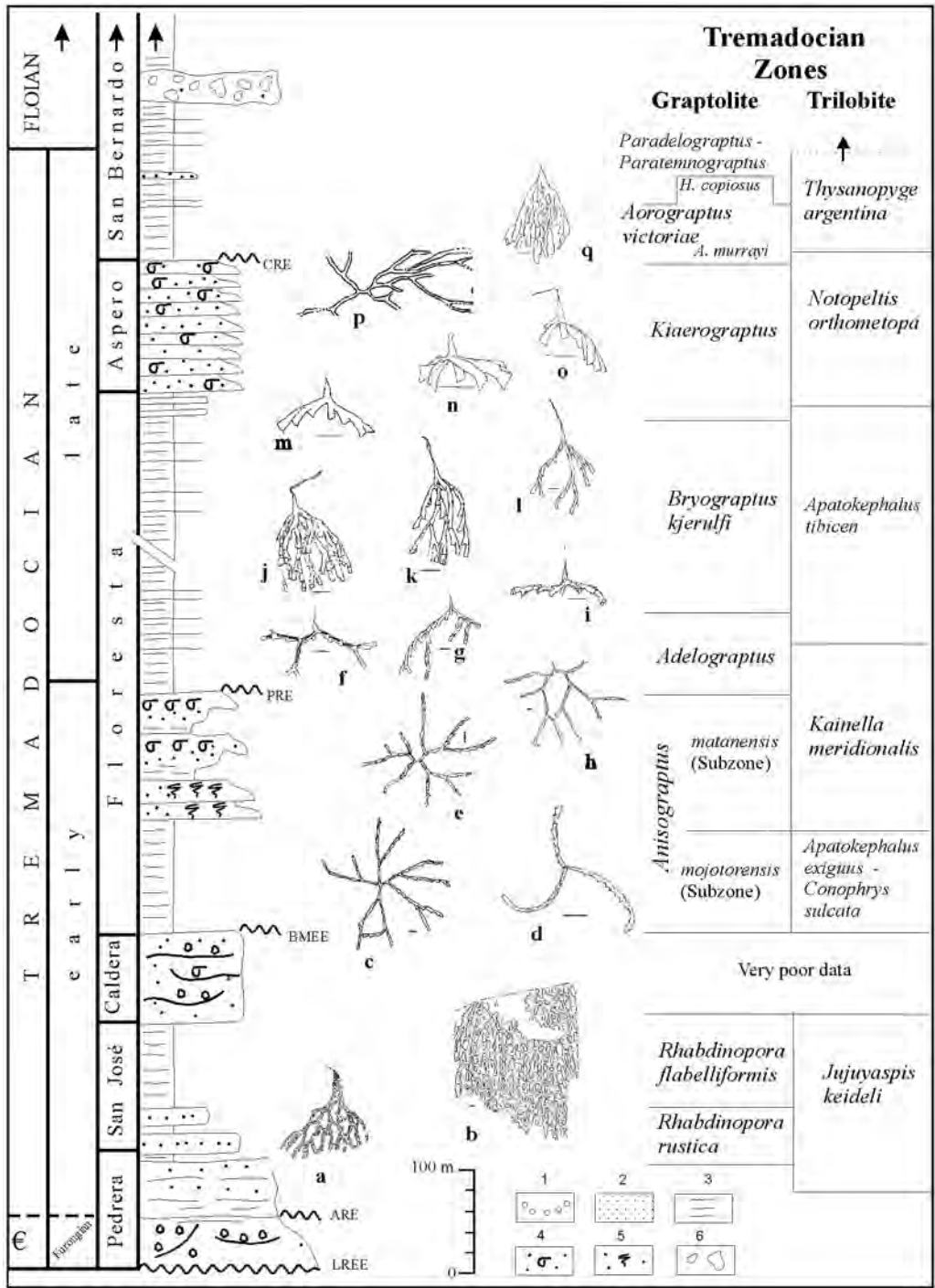
Anisograptus Zone

It develops from the base of the Floresta Formation to the top of the Coquina Beds of the same unit (Figs. 1c and 2). This biozone comprises the *A. mojotorensis* and *A. matanensis* Subzones. *Anisograptus mojotorensis* Monteros and Moya (2010a) is a small anisograptid of very delicate rhabdosome, which shows a pseudo-quadriradiate symmetry in horizontal view (Fig. 2c). The species was only recorded in the Mojotore Range, where its FAD precedes the appearance of *A. matanensis* Ruedemann. *Triograptus osloensis* Monsen (Fig. 2d) is recorded at the top of the *A. mojotorensis* Subzone (Monteros et al., 2010) predating its appearance in Oslo, where the species first occurs in the *Bryograptus ramosus* Zone (Maletz et al., 2009). The *A. mojotorensis* Subzone correlates with the *Conophrys sulcata*-*A. exiguus* trilobite Zone, this latter species persists from the previous cycle and it was not recorded from the next *Kainella meridionalis* trilobite Zone. The base of the *A. matanensis* Subzone coincides with that of the *K. meridionalis* Zone, the lapse in which both index taxa overlap is characterized by the development of abundant levels of coquina limestone (Fig. 2). These Coquina Beds display regional geographic distribution and contain *Cordylodus angulatus* (Moya et al., 2003c; Albanesi et al., 2008).

The *Anisograptus* Zone, with *A. matanensis* as indicator (Fig. 2e), was recognized in the Pascha, Cajas Range, Angosto del Moreno and Angosto La Quesera areas (Fig. 1b; Monteros, 2005; Ortega and Albanesi, 2005). The zone is developed through a depositional sequence bound by two unconformities related with two bioevents: i) The discontinuity of the base (BMEE), which predates the appearance of the *Anisograptus* triradiate pattern; ii) the discontinuity at the top, which predates the appearance of the biradiate pattern in *Adelograptus* and that, according to the scheme from Erdtmann (1995), would correspond to the *Peltocare* Regressive Event (PRE). Consequently, the deposits involved (lower-middle part of the Floresta Formation, Fig. 2) would be correlated with the upper interval of the Lower Alum Shale (Fig. 3).

Finally, Albanesi et al. (2008) defined the *Rhadinopora flabelliformis anglica* Zone, based on a punctual find of *R. f.* sp. cf. *R. f. anglica* (Bulman), together with *A. matanensis*. Apparently, this fossiliferous level

Figure 2. Integration of Tremadocian stratigraphy from the Mojotore Range with graptolite and trilobite biozones. Lithologies: 1, conglomerate; 2, sandstone; 3, shale; 4, coquina beds; 5, bioturbated layers; 6, debris flows. Fossils: a, *Rhadinopora rustica*; b, *Rhadinopora flabelliformis* ssp.; c, *Anisograptus mojotorensis*; d, *Triograptus osloensis*; e, *Anisograptus matanensis*; f, *Adelograptus tenellus*; g, *Adelograptus cuerda*; h, *Kiaerograptus altus*; i, *Kiaerograptus* sp. aff. *K. kiaeri*; j, *Bryograptus kjerulfii*; k, *Bryograptus ramosus*; l, *Bryograptus bröggeri*; m, *Kiaerograptus* sp. aff. *K. kiaeri*; n, *Kiaerograptus* sp. aff. *K. stoermeri*; o, *Kiaerograptus* sp. cf. *K. stoermeri*; p, *Clonograptus* sp. cf. *C. sarmentosus*; q, *Aorograptus victoriae*. LREE, Lange Ranch Eustatic Event; ARE, *Acerocare* Regressive Event; PRE, *Peltocare* Regressive Event; CRE, *Ceratopyge* Regressive Event (Erdtmann, 1995; Nielsen, 2004).



has no other biostratigraphic controls, making it difficult to regard it as a biozone, especially considering that in some regions such as Oslo, *A. matanensis* precedes and accompanies *R. f. anglica* and even persists after the LAD of the latter (Cooper, 1999).

Adelograptus Zone

The zone was recognized by Monteros and Moya (2005a), starting at the FAD of *Adelograptus tenellus* (Linnarsson), at levels located just above the top of the Coquina Beds of the Floresta Formation (Fig. 2f). The upper limit of this zone is marked by the FAD of *Bryograptus kjerulfi* Lapworth. Besides *A. tenellus* (Linnarsson) (Fig. 2f), this biozone includes *Adelograptus cuerdoi* Monteros and Moya (2005b) (Fig. 2g), *Adelograptus* sp., *Kiaerograptus altus* (Williams and Stevens) (Fig. 2h) and *Kiaerograptus* sp. aff. *K. kiaeri* (Monsen).

A. tenellus is represented at the lower half of the zone, while *A. cuerdoi* persists throughout it and even surpassing the upper boundary, coexisting with the first record of *B. kjerulfi*. Because of this, Moya et al. (1994) erroneously considered that the *B. kjerulfi* Zone preceded that of *Adelograptus*. In the lower part of this zone *A. matanensis* persists, and in the middle and upper intervals, the first true kiaerograptids appear. The material assigned to *Kiaerograptus altus* is congeneric with *Adelograptus altus* Williams and Stevens, which are recognizable by the free and isolated apertural part of the metascula. This feature was precisely the one that Williams and Stevens (1991) used to nominate the species, but is also one of the defining features of *Kiaerograptus* (Maletz, 1999).

The *Adelograptus* Zone documents the transitional beds between the lower and upper Tremadocian in northern Argentina, rather than the base of the upper Tremadocian (Figs. 2, 3), owing to the presence of *A. matanensis* at the base and to the FAD of *Kiaerograptus* in the middle to upper intervals. Maletz and Erdtmann (1987) noticed this and made separate mentions of the assemblages containing *A. tenellus* with *R. f. anglica* and *R. f. flabelliformis* on one side and upper Tremadocian graptolites on the other. The *Adelograptus* Zone of northern Argentina may be directly correlated with the *Adelograptus* Zone of southern Bolivia, which was equated with *Adelograptus tenellus* Zone from Scandinavia (Egenhoff et al., 2004; Maletz et al., 2009); the latter being developed in the lower interval of the Upper Alum Shale (Erdtmann, 1995), in a similar stratigraphic position to the *Adelograptus* Zone of northern Argentina (Fig. 3).

Bryograptus kjerulfi Zone

González Barry and Alonso (1984) recognized *Bryograptus* sp. aff. *B. kjerulfi* for the first time in the Mojotoro Range, and assigned the fossiliferous levels to the upper Tremadocian. Moya et al. (1994) ratified the find of *Bryograptus* in the Mojotoro Range, identified the material with *B. kjerulfi* Lapworth, and recognized the interval defined by this species in Pascha, Angosto La Quesera, Angosto del Moreno and other localities in the Mojotoro Range. Ortega and Albanesi (2003) recognized the *Bryograptus* Zone in Pascha.

The *B. kjerulfi* Zone is defined by the LAD and FAD of *B. kjerulfi* Lapworth (Fig. 2j), occurring through the upper third of the Floresta Formation (Monteros and Moya, 2006a). Besides *B. kjerulfi*, in this zone *B. ramosus* Brögger (Fig. 2k) and *B. bröggeri* Monsen (Fig. 2l) were also recorded, as well as some relict forms such as *Kiaerograptus* sp. aff. *K. kiaeri* (Fig. 2i), *Adelograptus cuerdoi*, *Adelograptus* sp. and a very late record of *Staurograptus* sp. aff. *S. dichotomus* Emmons. Recently, Maletz et al. (2009) carried out a

Age		Scandinavia Maletz et al. 2009	Oslo - Scania Erdtmann 1995	Mojotoro Range Moya 1998, 2002	Northwest Argentina Monteros and Moya 2006a This paper
TREMADOCIAN	late	<i>Hunnegrapthus copiosus</i>	Tøyen Shale	San Bernardo	<i>Hunnegrapthus copiosus</i> <i>Aorograptus victoriae</i>
		<i>Araneograptus murrayi</i>			<i>Araneograptus murrayi</i>
		<i>Kiaerograptus kiaeri</i>	Björkåsholmen	Áspero	<i>Kiaerograptus</i>
		<i>Bryograptus ramosus</i>	Upper Alum Shale	Floresta (Upper)	<i>Bryograptus kjerulfi</i>
		<i>Adelograptus tenellus</i>			<i>Adelograptus</i>
	early	<i>Rhabdinopora f. anglica</i>		Coquina beds	<i>Anisograptus matanensis</i>
		<i>Anisograptus matanensis</i>	Lower	Floresta (Lower)	<i>Anisograptus</i>
		<i>Rhabdinopora f. parabola</i>		Caldera	<i>Anisograptus mojotorensis</i>
		<i>Rhabdinopora praeparabola</i>	Alum Shale	San José	<i>Rhabdinopora flabelliformis</i> ssp. ?
					<i>Rhabdinopora rustica</i>

Figure 3. Correlation scheme of the Tremadocian Units from Scandinavia and Argentinian Eastern Cordillera.

detailed taxonomic, biostratigraphic and biogeographic analysis of *Bryograptus*, and established a correlation between the *B. ramosus* Zone in Scandinavia and the *B. kjerulfi* Zone discussed here.

The record of *Bryograptus* representatives seems restricted to the zone involved here. The specimen that Monteros and Moya (2003) carefully assigned to *Bryograptus*? nov. sp. in the *Aorograptus victoriae* Zone, corresponds to the new species *Aorograptus andinus* (Monteros and Moya, 2010c).

The *Adelograptus* and *Bryograptus kjerulfi* zones, together with the *Apatokephalus tibicen* trilobite Zone date the most important Tremadocian transgression occurred in northern Argentina. In Scandinavia, the *A. tenellus* and *B. ramosus* zones correspond to a similar event. The deposits accumulated during this transgression would be equivalent to the middle and upper intervals of the Floresta Formation, and correlate with the Upper Alum Shale (Fig. 3). The regression that follows this important transgressive event is equally intense and recorded within the *Kiaerograptus* Zone.

Kiaerograptus Zone

It was defined by Monteros (2005) and it is developed in the interval between the LAD of *B. kjerulfi* and the FAD of *Aorograptus victoriae* (T.S. Hall) (Fig. 2q). This range includes the terminal beds of the Floresta Formation and the entire Áspero Formation. The record of graptolites in the Áspero Formation is very poor both in quantity and diversity due to unfavorable depositional conditions, linked with a shallow, high energy environment. The relict *Kiaerograptus* sp. aff. *K. kiaeri* is the only form found in the lower part of the zone, co-occurring in the upper part with some new forms, represented by *Kiaerograptus* sp. aff. *K. stoermeri* (Erdtmann) (Fig. 2n), *Kiaerograptus* sp. cf. *K. stoermeri* (Fig. 2o), *Clonograptus* cf. *C. sarmentosus* (Moberg) (Fig. 2p) and *Clonograptus* sp. (Monteros and Moya, 2006b).

The *Kiaerograptus* Zone has close affinities with the graptolite assemblage indicated by Erdtmann

(1965) in the *Ceratopyge* Limestone (3a interval), which includes, among others, *Clonograptus callavei* (Lapworth), *K. kiaeri* and *K. stoermeri*. In turn, the 3a interval would correspond to the *Kiaerograptus* Zone of Erdtmann and Paalits (1994) or to the *K. kiaeri* Zone of Maletz et al. (2009), both recognized in the Oslo-Scania region.

On the other hand, in the lower part of the *Nothopeltis orthometopa* trilobite Zone (Fig. 2), the index fossil coexists, among others, with *Apatokephalus tibicen* Přibyl and Vanek and with *Ceratopyge forficuloides* Harrington and Leanza. Similarly, in Scandinavia *A. tibicen* (= *A. "serratus"*) coexists with *Ceratopyge forficula* (Sars) in the Bjökåsholmen Formation and in the *Ceratopyge* Limestone (3a interval), the units that referenced the *Ceratopyge* Regressive Event (CRE; Erdtmann, 1986). Globally, the CRE was followed by an important transgressive event developed at the Tremadocian-Floian boundary (Erdtmann and Paalits, 1994). In northern Argentina, this transgression is documented in the San Bernardo, Parcha, Cieneguillas and Acoite formations (Shale 3, Moya, 2002).

Aorograptus victoriae Zone

It is developed in the lower portion of the San Bernardo Formation and its range is extended from the FAD to the LAD of *Aorograptus victoriae* (T.S. Hall). Monteros and Moya (2008) consider this as the latest Tremadocian zone, and indicated that the basal Floian boundary is recorded in the San Bernardo Formation by the *Paradelograptus-Paratemnograptus* Zone, in absence of *Tetragraptus approximatus* (Nicholson) that internationally defined the base of the Floian stage.

In Fig. 2 the index species of *A. victoriae* Zone were only plotted, because the majority of the accompanying species were described and illustrated by Monteros and Moya (2003). In this biozone, the true kiaerograptids disappear, together with the maximum diversification of the anisograptids and the appearance of the first kinnegratids and dichograptids. The assemblage includes *Aorograptus victoriae* (T.S. Hall), *A. andinus* Monteros and Moya, *Adelograptus?* sp., *Kiaerograptus?* *supremus* Lindholm, *Clonograptus* sp. cf. *C. flexilis* (J. Hall), *Dichograptus?* sp. *D. octobrachiatus* (J. Hall), *Paradelograptus* sp. cf. *P. pritchardi* Erdtmann, Maletz and Gutiérrez-Marco, *P. onubensis* Erdtmann, Maletz and Gutiérrez-Marco, *P. mosseboensis* Erdtmann, Maletz and Gutiérrez-Marco, *Paratemnograptus isolatus* Williams and Stevens, *Tetragraptus?* *bulmani* (Thomas) and *Tetragraptus* sp. cf. *T. sanbernardicus* Loss.

The *A. victoriae* Zone in the Mojotoro Range correlates with the eponymous Zone in western Newfoundland, Canada (Williams and Stevens, 1991) and with the upper interval of the *A. victoriae* Zone in Australia (Cooper, 1999). According to Cooper et al. (2004), the latter is correlated with the Scandinavian *Araneograptus murrayi* and *Hunnegraptus copiosus* zones. In the Oslo-Scania region, these biozones are developed in the lower part of the Tøyen Shale where they precede the *Tetragraptus phyllograptoides* Zone in the base of the Floian. If one accepts the correlation proposed by Cooper et al. (2004), the *A. victoriae* Zone of Argentina would be equivalent to the *A. murrayi* plus the *H. copiosus* zones of Scandinavia; therefore, the lower part of the San Bernardo Formation could be correlated with the lower part of the Tøyen Shale (Fig. 3), which contains conodonts from *Paroistodus proteus* Zone (Erdtmann, 1995). Precisely Zeballo et al. (2008) recognize conodonts from the *Paroistodus proteus-Acodus deltatus* Zone (= *A. deltatus* Lindström *sensu lato*) in beds with *A. victoriae* in the Chalala-Coquena area (Quebrada de Humahuaca), reinforcing the proposed correlation.

On the other hand, the *A. murrayi* and the *H. copiosus* zones were recognized by Ortega and Albanesi (2003) in the Parcha Formation, a unit equivalent to a portion of the lower half of the San Bernardo Formation. The Parcha Formation has not produced *A. victoriae* nor *T. phyllograptoides* or *T. approximatus*.

and, as a consequence, Ortega and Albanesi (2003) assigned the entire Parcha Formation to the Tremadocian. However, the information provided by conodonts is somewhat different: Albanesi et al. (2008) point out that *A. murrayi* (J. Hall) coexists there with *A. deltatus deltatus* Lindström and with *Didymograptus* spp. and *Paradelograptus* spp.

According to Maletz et al. (1996) and Bergström et al. (2004), the base of the *Oelandodus elongatus-Acodus deltatus deltatus* Subzone in the Diabasbrottet GSSP, is located immediately below the Tremadocian-Floian limit and almost the entire range of *A. deltatus deltatus* occurs within the *T. approximatus* Zone.

On the other hand, in the *H. copiosus* Zone Albanesi et al. (2008) listed *Paradelograptus* spp., *Tetragraptus* sp. and conodonts from the *A. deltatus*-*P. proteus* Zone. This biozone is developed above the *A. murrayi* Zone and of course, its age should be younger.

Previous concepts and the very wide range of *A. murrayi* in northern Argentina (Monteros, 2005), induce us to consider that the *A. victoriae* Zone is the last of the Tremadocian zones in the region (Fig. 2). Intervals of *A. murrayi* and *H. copiosus* were plotted as subzones, and must be considered with caution. In none of the Hunneberg sections analyzed by Maletz et al. (1996) and Bergström et al. (2004), the *A. murrayi* and/or *H. copiosus* zones were recorded. The biostratigraphical control of the Tremadocian-Floian boundary as conventionally accepted was established by the FAD of *T. approximatus* and characteristic conodonts.

CONCLUSIONS

Seven graptolite zones were recognized in the Tremadocian succession of the Argentinian Eastern Cordillera: *R. rustica*, *R. flabelliformis*, *Anisograptus*, *Adelograptus*, *B. kjerulfi*, *Kiaerograptus* and *Aorograptus victoriae*. In our scheme, a gradual diversification in the first five anisograptid biozones is evident.

During the *Kiaerograptus* Zone, coinciding with the CRE, a sudden impoverishment of the fauna occurs, where the delicate multiramous forms that characterized much of the Tremadocian were replaced in the *Kiaerograptus* Zone by other more robust, also multiramous species. During the transgression that followed the CRE, a maximum anisograptid diversification is reached, together with the FAD of the dichograptids in the *A. victoriae* Zone.

Based on the seven recognized biozones, a fit correlation between eustatic and sedimentary events documented in the Tremadocian successions of the Oslo-Scania region and the Eastern Cordillera of Argentina is produced.

The very close affinity of Scandinavian and Argentinian faunas described from the Furongian (*Acerocare* and *Neoparabolina frequens* trilobite Zones) to the late Tremadocian (*Kiaerograptus* Zone), suggests not only that the respective basins were situated in similar intermediate paleolatitudes, but they were probably interconnected. From the latest Tremadocian onwards (*A. murrayi*-*H. copiosus* and *A. victoriae* zones) the faunal affinity decreases, and the clear predominance of shared Atlantic forms that characterized Argentinian and Scandinavian faunal assemblages, gives way to the incorporation of typical elements from the Pacific Realm to the north Argentinian basin. This time coincides with a tectonic event that causes paleogeographic changes in the Ordovician basin of northern Argentina.

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EARLY ORDOVICIAN MAGMATISM IN THE NORTHERN CENTRAL IBERIAN ZONE (IBERIAN MASSIF): NEW U-Pb (SHRIMP) AGES AND ISOTOPIC Sr-Nd DATA

M. Navidad and P. Castiñeiras

Departamento de Petrología y Geoquímica e Instituto de Geología Económica (centro mixto UCM-CSIC), Facultad de Ciencias Geológicas, Universidad Complutense de Madrid, José Antonio Novais 2, 28040 Madrid, Spain. navidad@geo.ucm.es

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INTRODUCTION

In the Central Iberian Zone (CIZ), an important volume of terrigenous sediments was deposited during the Early Ordovician, among them, the widespread Armorican Quartzite. This formation outlines the Variscan orocline from the northwest to the central outcrops of the Iberian Massif (Fig. 1). Late Cambrian-Early Ordovician magmatism produced a great amount of crustal melts, giving rise to large bodies of aluminous granites and explosive volcanic rocks, the so-called Ollo de Sapo volcano-sedimentary formation (Hernández Sampelayo, 1922; Parga Pondal et al., 1964). Both lithologies formed during an acid magmatic episode derived from crustal melts, which is related to the birth of the Rheic Ocean (Navidad et al., 1992; Gebauer, 1993; Valverde Vaquero and Dunning, 2005; Díez Montes, 2007; Montero et al., 2007). There is evidence of intra-ordovician tectonic movements (Sardic event, Julivert et al., 1972) that originated unconformities at the base of the Early Ordovician series (Toledanic unconformity, Díez Balda et al., 1990; Pérez Estaún et al., 1991), and a passive margin sequence from Early Ordovician to Early Devonian. So far, there is no clear evidence for thermal metamorphism coeval with this important Ordovician magmatism in the CIZ.

In this paper, we report two new SHRIMP ages of augen orthogneisses from the Guadarrama Sierra (Spanish Central System) and new isotopic Sr-Nd data from orthogneisses and Ollo de Sapo volcanic rocks from the northern CIZ. We complete our dataset with data from other authors to obtain a geodynamic model for the early Ordovician crust in the CIZ.

GEOLOGICAL SETTING

The Ollo de Sapo Formation crops out in the core of a Variscan antiform in the limit between the West Asturian-Leonese and the Central Iberian zones (Julivert et al., 1972). It delineates an arc from northern

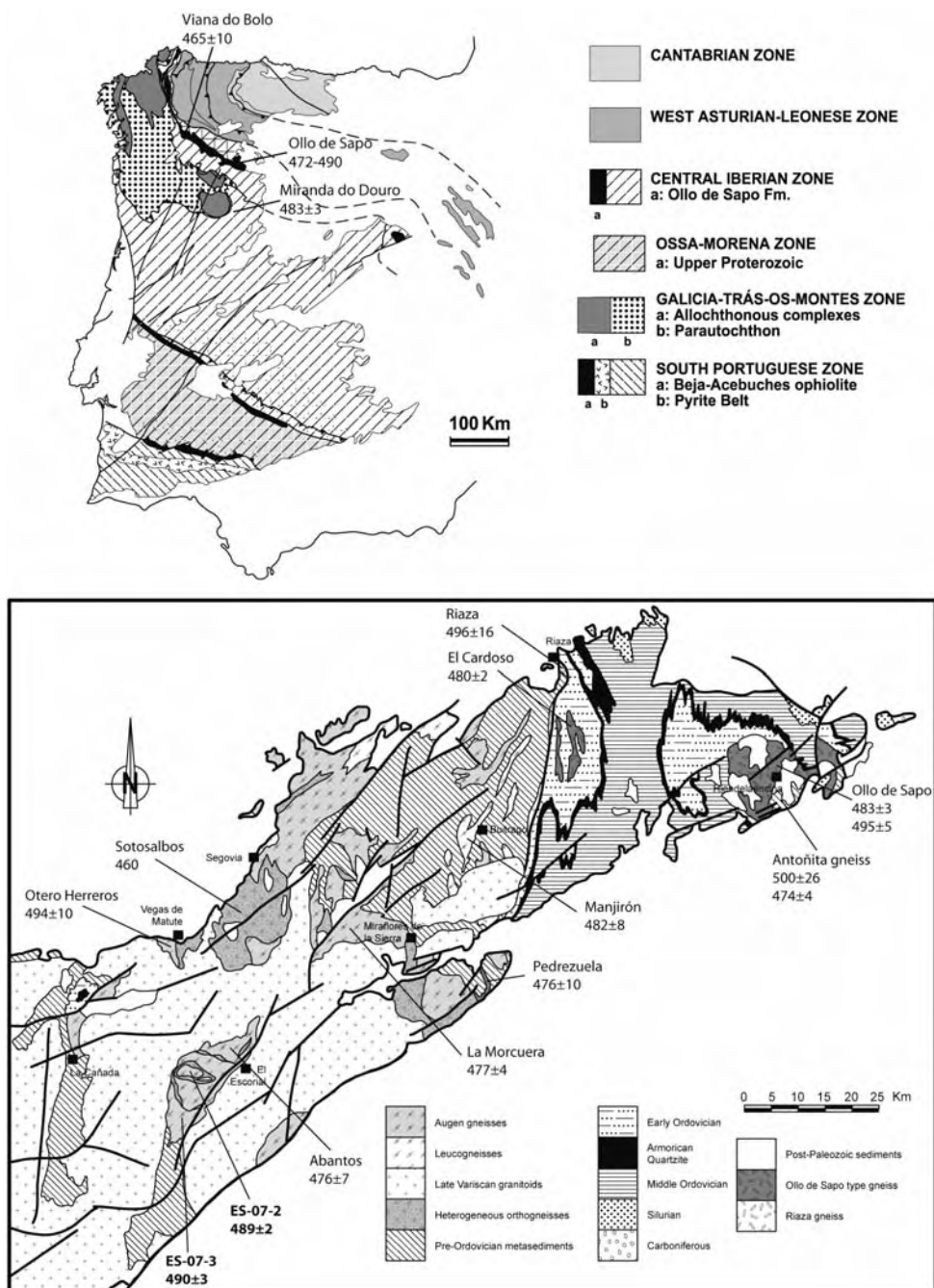


Figure 1. Geological sketch of the Ollo de Sapo domain in the northern Central Iberian Zone. Ollo de Sapo antiformal and Spanish Central System. Ages of metagranitic orthogneisses and Ollo de Sapo volcanic rocks (in Ma) are distributed in different metamorphic Variscan massifs.

Galicia to the easternmost outcrops of the Spanish Central System (Hiendelaencina gneissic dome: Fig. 1). The Armorican Quartzite also outlines this structure and below it, the Ollo de Sapo Formation crops out at the lower part of the Tremadocian metasediments or interbedded within them. The volcano-sedimentary rocks were deformed and metamorphosed during the Variscan orogeny, varying in grade from the chlorite and biotite zones, where they show the characteristic blue quartz, to the Ms out-Kfs in zone, where they are migmatitic augen gneisses. It is formed by meta-dacitic and meta-rhyolitic explosive rocks and volcanoclastic sediments of greywacke composition; sub-volcanic meta-granite bodies are enclosed in the former (Navidad, 1978; Navidad et al., 1992; Díez Montes et al., 2004). Their characteristic aspect is a porphyroid rock with feldspar megacrysts of different size and blue subvolcanic quartz crystals embedded into a quartz-feldspathic matrix that show occasional relict of ignimbrite structures.

This Formation has been repeatedly dated by different methods, with results ranging from 618 Ma to 465 Ma (Lancelot et al., 1985; Wildberg et al., 1989; Gebauer, 1993). Recent U-Pb ages on zircon crystals offer more precise ages: ~480 Ma for Valverde Vaquero and Dunning, 2000; whereas Montero et al. (2007), dated the explosive rocks between 495 and 483 Ma, similar to the ages obtained by Díez Montes et al. (2010) for the intrusive metagranites (between 483 to 474 Ma). Based on inherited zircons some authors place Central Iberia to the east of the African Craton in the Early Ordovician (Díez Montes et al., 2010; Bea et al., 2010). This location is in agreement with the faunal data and the paleogeographic situation proposed by Gutiérrez Marco et al. (2002). This proposal contrasts with those where the location of the CIZ next to the West African Craton, based on TDM isotopic model ages and inherited zircon populations from sediments of the CIZ, put forward by Martínez Catalán et al. (2004), Fernández Suárez et al. (2000) and Ugidos et al. (2003).

The metagranitic augen gneisses represent an important volume of aluminous granitoids placed into a pre-Tremadocian sequence that crops out underlying the Ollo de Sapo Formation. They form massifs of different extension along the northern CIZ, small bodies in the northwestern Iberian Peninsula (Porto, Sanabria and Viana do Bolo), and massifs of kilometric extension in the Guadarrama Sierra (Spanish Central System, Fig. 1). They crop out from the biotite to moscovite out-Kfs in metamorphic zones, where they are migmatized. The massifs are constituted by different compositional types, being the most frequent mesocratic and feldspar-rich augen gneisses, leucogneisses rimming the augen gneisses or forming independent massifs, and the banded gneisses are more frequent to the east of the Somosierra domain in the medium grade metamorphic zones. There is no contact metamorphism related to the intrusion of these rocks, but in several cases there are tourmaline rims in the contact with the medium grade metasediments.

Navidad (1979) and Navidad et al. (1992) point out, in the first whole-rock geochemical studies, the genetic similarity between the augen orthogneisses and the Ollo de Sapo volcanic rocks and suggest a crustal origin for them. Later, Valverde Vaquero and Dunning (2000) link both protoliths to a felsic magmatic belt active during the break-up of the Gondwanan margin and the birth of Iapetus Ocean.

Based on Rb-Sr isochrons, Viallet et al. (1987) ascribe ages of emplacement from 494 to 471 Ma, related to the Armorican Quartzite sedimentation. U-Pb ages show emplacement ages spanning from 482 to 468 Ma related to the Sardic *sensu lato* event (Lancelot et al., 1985; Gebauer et al., 1993; Valverde Vaquero and Dunning, 2000; Montero et al., 2007; Díez Montes et al., 2010). Regarding the ages present in the crust hosting these igneous rocks, Wildberg et al. (1989), based in inherited zircon populations, point out a protracted evolution during the Paleozoic from 540 Ma. Lancelot et al. (1985) cited Precambrian crustal components (2.4 and 2.0 Ga) that could be derived from the West Africa Craton, whereas Montero et al. (2007) found inherited zircon populations with ages between 0.85-0.90 and 0.7-0.65 Ga.

GEOCHRONOLOGY

Two orthogneiss samples from the El Escorial metamorphic massif (Spanish Central System: Fig. 1) were selected for SHRIMP analysis, including U-Pb geochronology and REE (rare earth elements) and elemental Hf determinations in zircon. Sample ES-07-2 corresponds to the so-called Santa María de la Alameda orthogneiss and its zircon grains are colorless to yellow, forming simple bipyramidal prisms with aspect ratios between 2:1 and 3:1. Under cathodoluminescence (CL) they usually display a moderately luminescent oscillatory zoning, with scarce xenocrystic cores. Sample ES-07-3 is a melanocratic augengneiss, locally known as the Robledo orthogneiss. Zircon from this sample is generally less colored and forms prisms with higher aspect ratios (4:1). There are also subrounded grains which are probably inherited. Under CL, the hand-picked zircons show oscillatory zoning.

Forty-seven analyses were carried out in sample ES-07-2, four of them are discarded because of their high common Pb or their inherited character. The remaining forty-three are continuously distributed between 506 and 416 Ma. There are at least two possible explanations for this dispersion: lead loss or inheritance. Owing to the absence of inherited xenocrysts, we consider that the magmatic age of this sample is closer to the older end. Considering the twelve oldest analyses, the best estimate for the age is obtained from a group of eight analyses which yield a mean age of 489 ± 3 Ma, with a mean square of weighted deviation (MSWD) of 1.6.

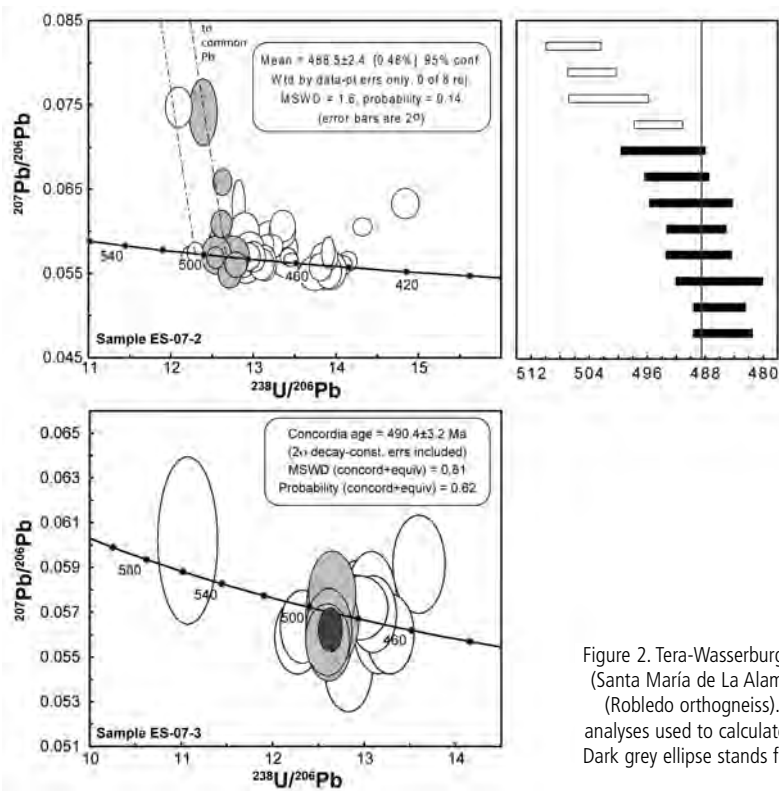


Figure 2. Tera-Wasserburg diagrams for samples ES-07-2 (Santa María de La Alameda orthogneiss) and ES-07-3 (Robledo orthogneiss). Light grey ellipses represent analyses used to calculate the age (mean or Concordia). Dark grey ellipse stands for Concordia age in sample ES-07-3.

In sample ES-07-3, eighteen analyses spread out between 506 and 456 Ma. The best statistical estimate for the crystallization age of this sample is obtained from six analyses that yield a concordia age of 490 ± 3 Ma (MSWD=0.81).

GEOCHEMISTRY

The Ollo de Sapo volcanic rocks and the metagranitic orthogneisses have been repeatedly analyzed in the last thirty years (Navidad, 1978, 1979; Navidad et al., 1992; Ortega et al., 1996; Díez Montes et al., 2004). Whole-rock geochemistry allows us to classify the volcanic rocks as dacite and rhyodacite tuffs of calc-alkaline affinity and the orthogneisses as aluminous granites. Both derived from melts enriched in potassium with trace elements normalized profiles showing a subduction geochemical signature marked by a Ta-Nb negative anomaly. This anomaly, in the case of volcanic rocks, has been interpreted as representative of a continental arc. Rare earth normalized profiles are fractionated with a negative anomaly in europium owing to the fractionation of plagioclase, which is characteristic of calc-alkaline melts.

New Sr-Nd isotopic data from the Ollo de Sapo Formation and the metagranitic orthogneisses are presented here. The Sr-Nd analyses were carried out at the Centro de Geocronología y Geoquímica Isotópica of the Complutense University of Madrid using the ID-TIMS method.

Metavolcanic and metagranitic orthogneiss samples show an important variation of $(^{87}\text{Sr}/^{86}\text{Sr})_{490-470}$, spanning from 0.7068 to 0.7126 and the ϵSr between 8 and 123; contrastingly, the $(^{143}\text{Nd}/^{144}\text{Nd})_{490-470}$ is more homogenous and the samples are concentrated in a little range, between 0.5117 and 0.5119, with high ϵNd values varying between -3 and -6. In a ϵSr , ϵNd diagram (Fig. 3a), the samples plot in the crustal provenance quadrant and they define a straight line suggesting mixing petrogenetic processes between juvenile melts and primitive arc-type isotopic signature related with their high ϵNd and crustal origin (Farmer and DePaolo, 1983). The plot $f_{\text{Sm/Nd}}$ versus ϵNd (Fig. 3b from DePaolo and Wasserburg, 1976), show a distribution from active to passive margins, but preferably in the latter geodynamical context.

With regard to the model ages, the ϵNd signature respect to the early Ordovician ages (Fig. 3c), shows a coincidence between the Ollo de Sapo and the metagranite values and their intersection with the CHUR evolution line point to a time span of the TCHUR ages between 0.7 and 1.6 Ga for the metagranites and from 1.0 to 1.5 for the Ollo de Sapo volcanic rocks suggesting a more protracted extraction period for the metagranitic than for the volcanic melts. TCHUR and provenance zircon ages are equivalent (Late Neoproterozoic and Mesoproterozoic), suggesting that the most probable sources for the Late Cambrian-Early Ordovician magmatism are melts from a Late Neoproterozoic crust essentially formed by a mixture of juvenile and Mesoproterozoic components.

INTERPRETATION AND CONCLUSIONS

Late Cambrian-Early Ordovician magmatism is constituted by aluminous metagranites intruded into pre-Early Ordovician sequences, and calc-alkaline explosive volcanic rocks are syngedimentary with Early Ordovician metasediments. The age of this magmatism varies in the northern CIZ between 490 and 470 Ma, being slightly previous and synchronous to the Armorican Quartzite *sensu stricto*.

Whole-rock geochemistry characterizes this magmatism as aluminous crustal melts intruded in an

extensional geodynamic context. However all magmatic protoliths preserve a subduction signature from an orogenic arc and the rare earth profiles are in agreement with calc-alkaline melts.

Contrastingly, Sr-Nd isotopic data indicates inhomogeneous $^{87}\text{Sr}/^{86}\text{Sr}$ content, and the alignment of the samples in the graphs suggests mixing processes between different melts. The high and homogeneous ϵNd signature (between -2.8 and -5.8), suggest that juvenile magmas or melts from a young crust are involved in the source of this magmatism.

Finally, Neoproterozoic zircon and TCHUR ages (0.6-0.7 Ga) are always present in both protolith types along the CIZ, implying the most probable age of crust source. Additionally, mafic protoliths are scarce or absent in the Early Ordovician sequences. However, in the Eastern Iberian Massif (Navidad and Carreras, 1998; Castiñeiras et al., 2008) syn-sedimentary metabasites from E-MORB tholeiitic melts and aluminous metagranites with subduction signatures are present in the Ediacaran sequences, and in the southern CIZ there are mafic fragments with TDM of 650 Ma in Silurian breccias (López Guijarro et al., 2008). We consider that the youngest Neoproterozoic sequences formed by sediments with Mesoproterozoic and Paleoproterozoic components, with juvenile rocks emplaced in them, are the most probable crustal source for the Late Cambrian-Early Ordovician magmatism that is developed in an extensional geodynamic context during the opening of the Rheic Ocean.

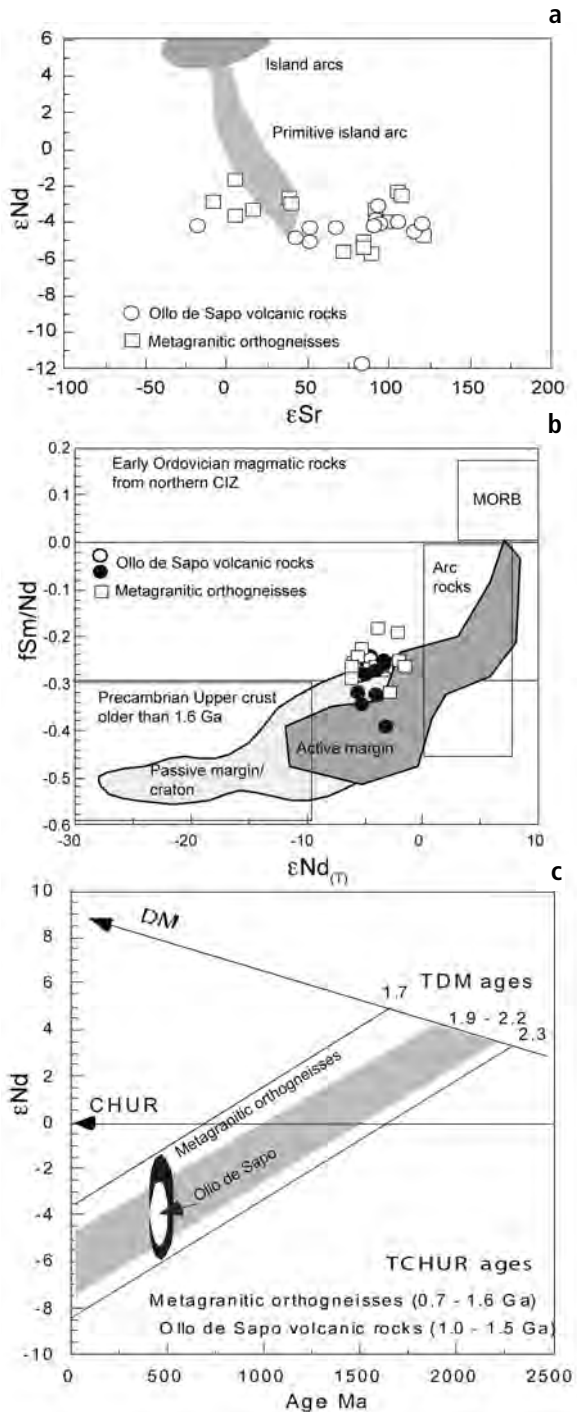


Figure 3. a, ϵSr – ϵNd binary diagram. Metagranitic orthogneisses and volcanic rocks samples draw an horizontal line of mixed processes. b, $f\text{Sm}/\text{Nd}$ – ϵNd binary diagram with orogenic context for the CIZ magmatic rocks. c, ϵNd –age with CHUR and DM lines from DePaolo and Wasserburg (1976). The plot of samples are enclosed in the areas that intersected in CHUR and DM lines and marked the model ages.

Acknowledgements

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A RE-CALIBRATED REVISED SEA-LEVEL CURVE FOR THE ORDOVICIAN OF BALTOSCANDIA

A.T. Nielsen

Natural History Museum, University of Copenhagen, Øster Voldgade 5-7, DK-1350 Kbh K, Denmark. Arnet@snm.ku.dk

A detailed assessment of Ordovician sea level changes in Baltoscandia was published by Nielsen (2004). Subsequently the Ordovician chronostratigraphic frame has been adjusted and emended (Ogg et al., 2008; Bergström et al., 2009) and minor errors and revisions of the published sea level curve have been detected. A revised and re-calibrated relative sea level curve for Baltoscandia is presented here. It mainly outlines 3rd order oscillations, but some of the indicated minor changes, notably in the Dapingian-early Darriwilian, are likely of 4th order. The black and white bar on the right hand side in Figure 1 indicates 2nd order oscillations.

Work is in progress trying to quantify the changing sea level, but the task is difficult due to the generally very high Ordovician sea level. As a result Scandinavia was continuously transgressed for most of the Ordovician with full regression only taking place associated with the most extreme lowstands in the Late Ordovician. As a result a quantification of the sea level remains a cumulative exercise without a return to "zero". The extensive flooding also makes it impossible to reconstruct an onlap curve; the relevant shallow water successions are not preserved in Baltoscandia. An additional obstacle is the changing depositional system associated with the climatic changes caused by the northwards drift of Baltica into lower latitudes during the Ordovician.

Despite these obstacles the condensed Ordovician sedimentary succession of Baltoscandia is an almost ideal basis for reconstructing a sea level curve. The area was tectonically quiescent and the depositional rates were comparatively small or very small (Lindström, 1971) and basically not influencing the local depth of deposition. Besides, most strata are well-dated biostratigraphically. The craton was located in the southern hemisphere and gradually shifted from intermediate latitudes towards equator during the period. As a result the climate changed from temperate to subtropical, and the carbonate production increased. However, this primarily affected the Estonian area, located in a relatively more nearshore setting than Scandinavia. The majority of Sweden was characterized by deposition in deep water of strongly condensed, cool water carbonates fringed further offshore by mudstones through most of the Ordovician, although the local appearance of Upper Ordovician carbonate mounds heralded the warm water carbonate deposition of the Silurian. In southern Norway the succession is thicker than in Sweden, being deposited in a distal foreland basin associated with the incipient Caledonide collision.

The Ordovician has been considered a green-house period, but major, fast sea level oscillations, nonetheless, suggest that glaciations occurred also prior to the well-known Hirnantian glaciations. Thus

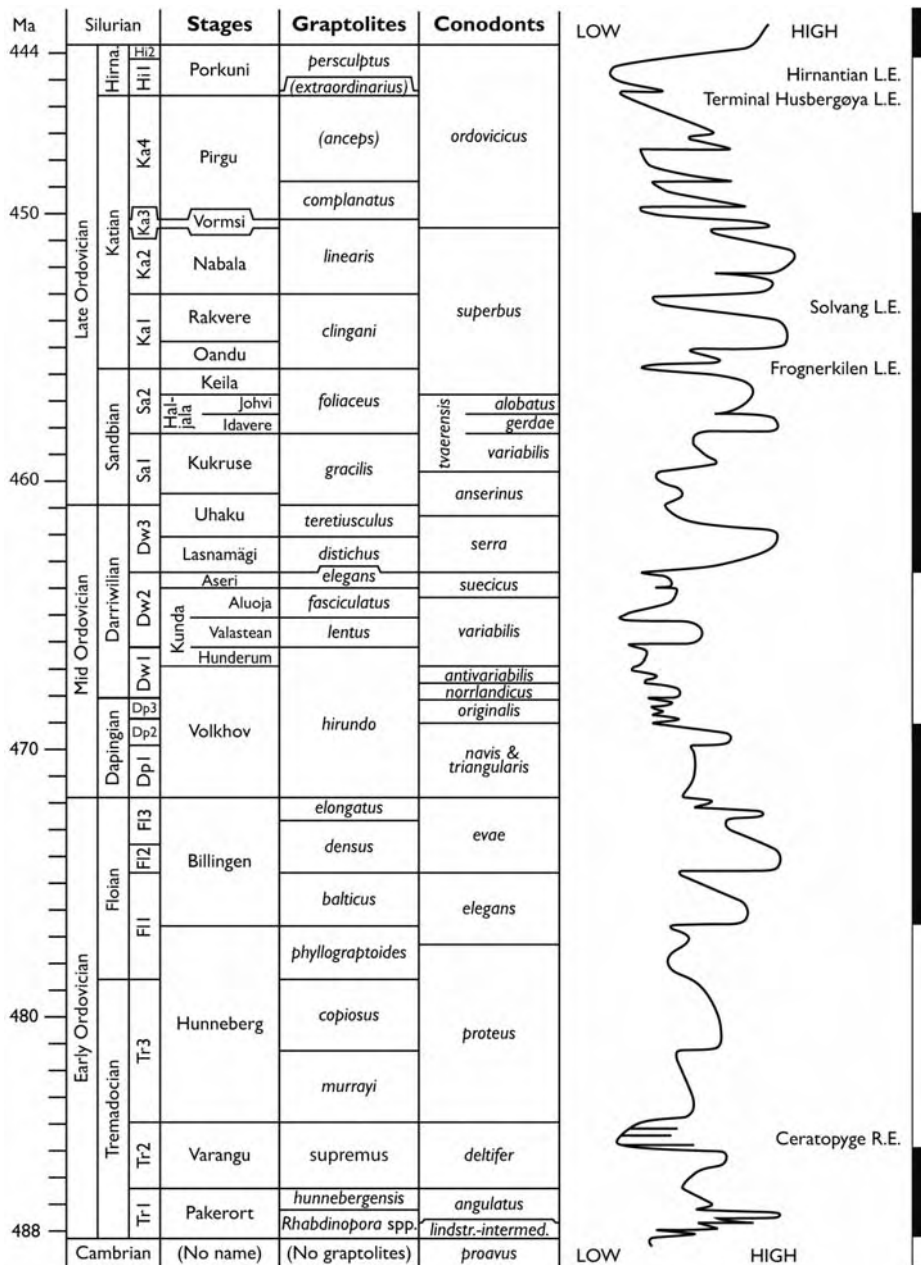


Figure 1. Ordovician sea level changes in Baltoscandia. In comparison with the Ordovician sea level curve published by Nielsen (2004) the curve has been re-calibrated according to the revised time frame for the Ordovician published by Ogg et al. (2008), including the global stages proposed by Bergström et al. (2009). In addition, several changes have been introduced, primarily in the Late Ordovician interval. The black and white bar on the right hand side indicates 2nd order oscillations (black: highstand interval, white: lowstand interval). Lowstand events referred to in the text are indicated. For naming of other sea level events, see Nielsen (2004).

the Tremadocian Ceratopyge Regressive Event and the Sandbian-Katian Frognerkilen and Solvang lowstands likely signal glaciations. Following this line of interpretation the numerous late Katian sea level changes may also be speculated to be of glacioeustatic nature. Regarding the CRE it may be noted that the "Ceratopyge Limestone" (now Bjørkåsholmen Fm) marks a poly-phased sea level rise, not the lowstand per se.

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NEW DATA ON UPPER ORDOVICIAN RADIOLARIANS FROM THE GORNY ALTAI (SW SIBERIA, RUSSIA)

O.T. Obut and A.M. Semenova

Trofimuk Institute of Petroleum Geology and Geophysics SB RAS, Acad. Koptyug av., 3, 630090, Novosibirsk, Russia.
ObutOT@ipgg.nsc.ru

Keywords: Upper Ordovician, Gorny Altai, stratigraphy, radiolarians.

INTRODUCTION

Upper Ordovician (Katian-Hirnantian) radiolarians were recovered from several localities around the world, including Nevada U.S. (Dunham and Murphy, 1976; Renz, 1990), Kazakhstan (Nazarov, 1975, 1988; Nazarov and Ormiston, 1993), western China (Li, 1995; Buckman and Aitchison, 2001), Baltic and Pomerania erratic boulders (Nazarov and Nylvak, 1983; Gorka, 1994), Scotland U.K. (Danelian and Floyd, 2001) and east and southeast Australia (Webby and Blom, 1986; Goto et al., 1992; Iwata et al., 1995; Noble and Webby, 2009). Their assemblages are moderately diverse and common in composition. Noble and Webby (2009) proposed to assign these faunas to the Pylomate-Large Concentric Sphaerellarian Zone 1, a globally recognized first occurrence zone based on the first appearance of *Secuicollacta*, that persists in geographically separate areas, including terranes of peri-Gondwanan of China, eastern Australia, Kazakhstan and continental margin regions of Laurentia. First report on Upper Ordovician (Ashgill) radiolarians from the Gorny Altai, SW Siberia was made by Obut and Iwata (2005, 2006). This faunas possessed common features with Late Ordovician assemblages known worldwide. Since then new data on taxonomic composition and stratigraphic distribution were recovered.

UPPER ORDOVICIAN RADIOLARIANS FROM OF THE GORNY ALTAI

Altai Ordovician basin is characterized by wide range of sedimentary facies: from shelf to oceanic genesis. Examination of siliceous deposits resulted in substantial progress in Paleozoic radiolarian studies in SW Siberia. It revealed presence of radiolarians in Lower and Upper Cambrian, as well as from Ordovician (Obut and Iwata, 2000, 2006). Upper Ordovician radiolarians (more than 1000 specimens) were recovered the Katian - Lower Hirnantian siliceous rocks belong to Tekhten' Formation and Siliceous-terrigenous Sequence cropped out in three sections in north-western Gorny Altai (Figs. 1, 2). They were accumulated on the on a relatively deep outer shelf and on a shallow-marine carbonate platform along the shelf edge at the foot of a continental slope (Sennikov et al., 2008).

The Tekhten' Formation is a carbonate-terrigenous succession with mainly reefal carbonates. Siliceous (radiolarites) shelf sediments are in close contact with the shallow-water carbonate rocks. Age of strata was defined by graptolites of *Climacograptus supernus* Zone, yielded in sandstones and siltstones, and by conodonts of *A. ordovicicus* Zone, recovered from limestones (Sennikov et al., 2008). Radiolarian associations were revealed from grey and greenish-grey siliceous mudstones from several localities on the left bank of Tachalov Brook, left tributary of Chagyrka river ("Tachalov" section) and on the right bank of Barany Brook, near Ust'-Chagyrka Village ("Baranyi-2" section). They are represented by 8 species belonging to 5 genera: *Secuicollacta ornata* Goto, Umeda and Ishiga, *S. silex* Goto, Umeda and Ishiga, *S. cf. sceptri* McDonald, *Kalimnosphaera maculosa* Webby and Blom, *Borisella subulata* (Webby and Blom), *Borisella* sp., *Protocera-toikiscum chinocrystallum* Goto, Umeda and Ishiga, *Inanigutta complanata* Nazarov (Fig. 3). Radiolarians are very well preserved and abundant, especially that of "Tachalov" section.



Figure 1. Location of Upper Ordovician sections, Gornyy Altai, south of West Siberia, Russia.

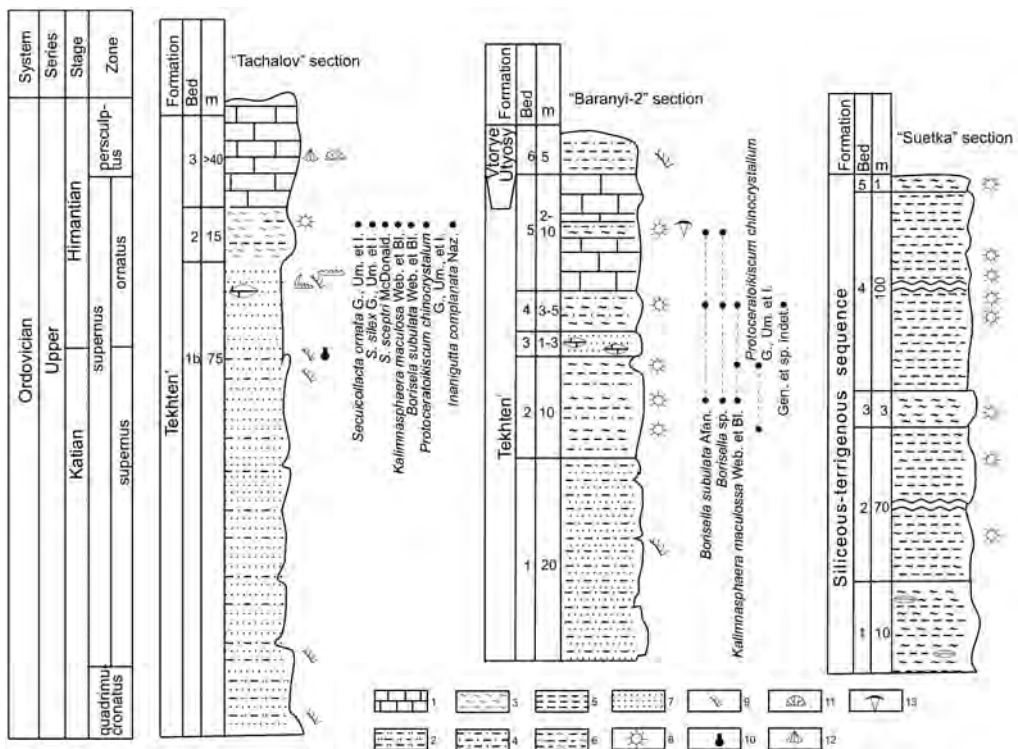


Figure 2. Ranges of radiolarians from Upper Ordovician sections of the north-western Gornyy Altai.

Legend: 1, limestones; 2, siltstones; 3, cherts; 4, silty sandstones; 5, mudstones; 6, siliceous mudstones; 7, sandstones; 8, radiolarians; 9, graptolites; 10, chitinozoans; 11, corals; 12, brachiopods; 13, trilobites.

The Siliceous-terrigenous Sequence consists of mudstones intercalated with silicilites, with few limestone lenses in the lower part of the sequence and black siliceous mudstones and laminated cherts in the middle-upper part. Rocks are believed to be deposited in the slope facies. Mudstones contain graptolites of the middle Katian *Cl. supenus* Zone. Radiolarian association contain 11 species assigned to 5 genera: *Protoceratoikiscum chinocrystallum* Goto, Umeda and Ishiga, *Pr. arachnoides* Goto, Umeda and Ishiga, *Kalimnosphaera maculosa* Webby and Blom, *K. sp.*, *Borisella subulata*, *Borisella sp.*, *Secuicollacta ornata* Goto, Umeda and Ishiga, *S. silex* Goto, Umeda and Ishiga, *S. sp.*, *Inanigutta complanata* Nazarov, *Inanigutta sp.* (Fig. 3). It was noted that radiolarians obtained from the lowermost beds of this sequence, characterized by underwater-sliding structures, are poorly preserved (probably as a result of skeletal dissolution). In the middle part the Siliceous-terrigenous Sequence – radiolarians are better preserved, and the most well preserved and abundant radiolarians were collected from the upper part of sequence where underwater-sliding structures were not observed.

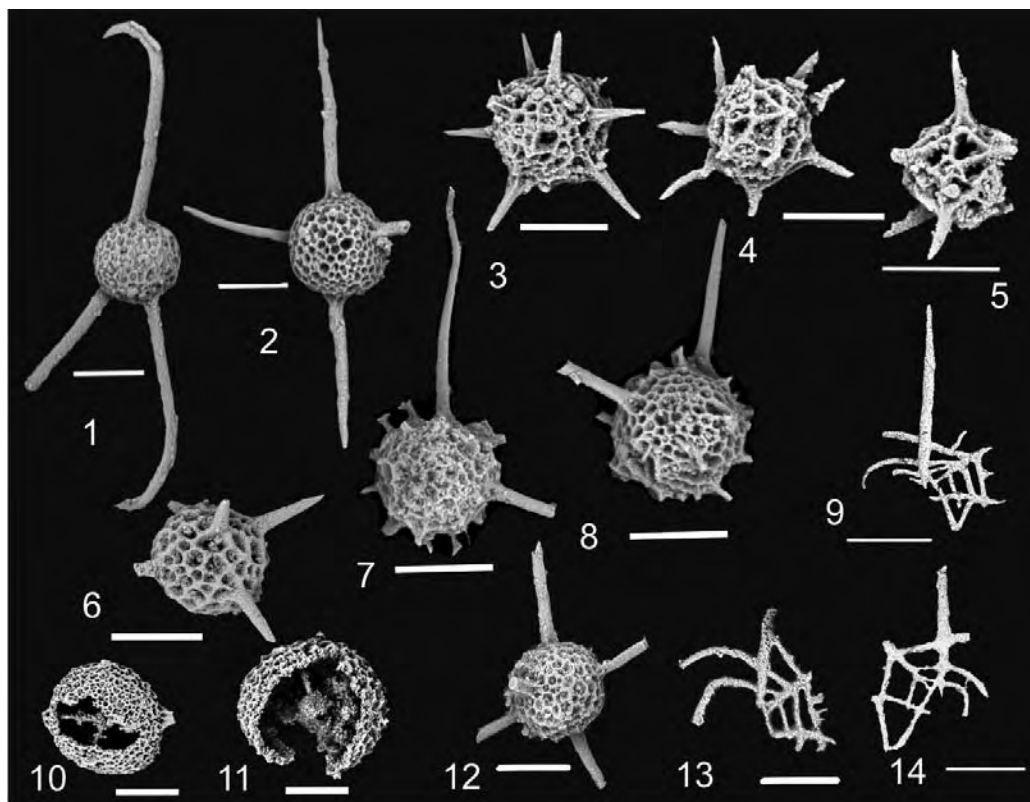


Figure 3. Scanning electron micrographs of selected radiolarians from the sections "Tachalov" (loc. 98080406: 1-5, 7, 8, 12), and "Suetka" (loc. SA 1007303: 6, 9, 10-14). 1, 2, *Borisella subulata* (Webby and Blom); 3, *Secuicollacta ornata* Goto, Umeda and Ishiga; 4, 5, *Secuicollacta sceptri* McDonald; 6, 12, *Inanigutta complanata* Nazarov; 7, 8, *Kalimnosphaera cf. maculosa* Webby and Blom; 9, 14, *Protoceratoikiscum chinocrystallum* Goto, Umeda and Ishiga; 10, 11, *Inanigutta sp.*; 13, *Protoceratoikiscum sp.* Scale bar is 100 μm .

DISCUSSION

Radiolarian fauna of the Tekhten' Formation and Siliceous-terrigenous Sequence possess many common taxa with assemblages reported from other Upper Ordovician localities worldwide. It is most comparable with fauna described from carbonates of Malongulli Formation of New South Wales, Australia and Hanson Creek Formation of Nevada (5 and 3 common species respectively) and siliceous mudstones and cherts of Lachan Orogene, SE Australia (5 common species). The Malongulli and Hanson Creek formations are dominated by spherical single shelled *Borisella subulata* and *Inanigutta complanata*, as well as pylomate spherical *Kalimnosphaera maculosa*, whereas abundant latticed spherical *Secuicollacta ornata* and spiny spider-web like *Protoceratoikiscum chinocrystallum* were recovered from Lachan Orogene siliceous sequences.

Radiolarians from Upper Ordovician siliceous sequences of Kazakhstan are dominated by spherical Inaniguttidae, *Haplentactinia*, and pylomate spherical *Kalimnasphaera*, whereas spiny Palaeoscenidiidae are rare. Radiolarians from Upper Ordovician of Gorny Altai include abundant *Secuicollacta ornata*, *S. silix* and *S. cf. sceptri*, *Inanigutta* spp. and *Borisella subulata*, with few *Protoceratoikiscum* spp. and rare *Kalimnosphaera maculosa*. Presence of common taxa as for carbonate as for siliceous facies allow use this fauna for correlation of vary-facies sedimentary strata. These faunas may be useful for inter-regional correlation.

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DARRIWILIAN GRAPTOLITES FROM THE LINA RANGE, NORTHWESTERN PUNA OF JUJUY, ARGENTINA

G. Ortega¹, G.L. Albanesi¹ and C.R. Monaldi²

¹ CONICET - Museo de Paleontología, FCEfN, Universidad Nacional de Córdoba, Casilla de Correo 1598, Córdoba X5000FCO, Argentina. gcortega@arnet.com.ar, galbanesi@arnet.com.ar,

² CONICET - Facultad de Ciencias Naturales, Universidad Nacional de Salta. crmonaldi@arnet.com.ar

Keywords: Graptolites, Biostratigraphy, Lina Formation, Ordovician, Puna, Argentina.

INTRODUCTION

The Lina range is located in the northwestern Puna of Jujuy Province, at Susques Department, northwestern Argentina. The average altitude at the study area in the Lina range is over 4000 m. At this area extensive outcrops of the Lower Paleozoic are exposed, which are in turn covered by Cenozoic volcanogenic rocks with abundant Pleistocene-Holocene ignimbrites.

Graptolites were firstly discovered in the Lina range by Ramos (1972), who referred the bearer strata to the Upper Ordovician. The author named Lina Formation to a thick succession of shales and graywackes affected by deformation and low grade metamorphism that is exposed in the El Toro village area and to the west, through the road to the Jama pass to Chile. He distinguished a lower sequence cropping out close to the village, where fossil remains were not recorded, and an upper one in the eastern flank of the Lina range. Due to the intense folding affecting these rocks, the real thickness of the formation could not be calculated. The base of the formation is not exposed and at some localities the unit is in tectonic contact with Cenozoic rocks. The graptolite fauna described by Ramos (1972) consist of *Glyptograptus euglyphus* Lapworth var. *linensis* n. var. It was recorded from the upper section of the Lina Formation, ca. 17 km west of El Toro, and was correlated tentatively with the *Nemagraptus gracilis* Zone that globally characterizes the early Sandbian.

Bahlburg et al. (1990) defined the Puna Turbidite Complex, as a thick succession reaching up to 3500 m incorporating Floian to late Darriwilian – Sandbian? deposits represented by the Coquena, Falda Ciénaga, and Lina formations (Aceñolaza and Baldi, 1987). The thickest part of the Falda Ciénaga and Lina formations correspond to the Upper Turbidite System. The graptolite species collected by these authors in the Lina range; i.e., *Eoglyptograptus* cf. *E. dentatus*, *Glossograptus hincksii fimbriatus*, and *Glyptograptus* (?*Oelandograptus* sp.), constrain the unit to the late Darriwilian. The authors proposed that the graptolite locality of Ramos (1972) would not be correlated with the upper part of the unit as previously suggested, but it would correspond to the basal part. They indicate that a homoclinal succession, ca. 800 m thick, contains the same graptolites in its uppermost part.

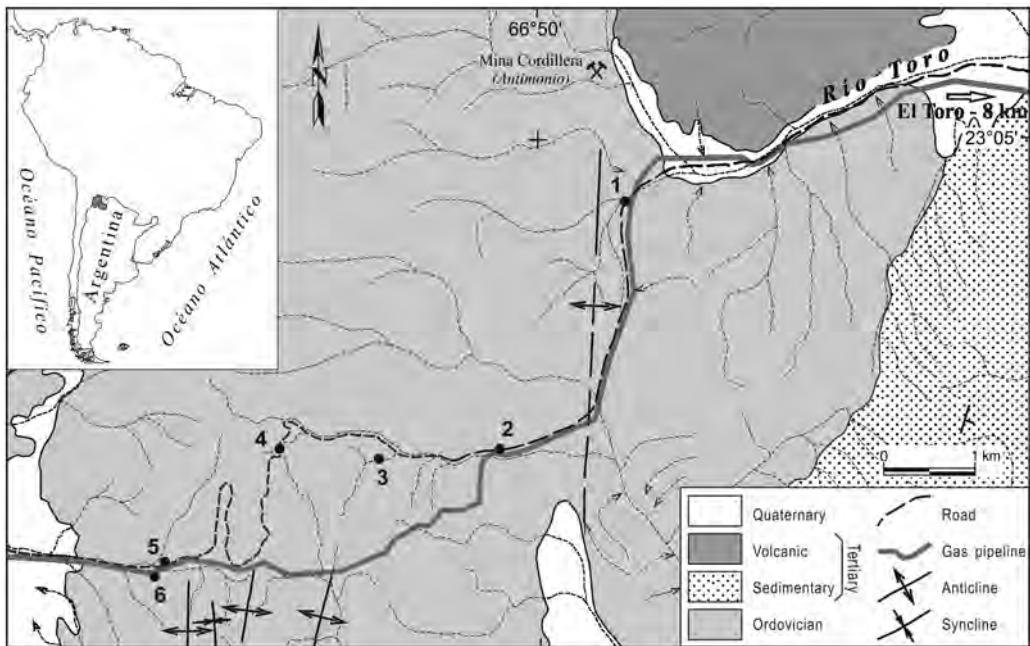


Figure 1. Location map of the study area and fossiliferous localities (numbered 1-6).

GRAPTOLITES AND BIOSTRATIGRAPHY

The present contribution deals with the discovery of late Darriwilian graptolites (Middle Ordovician) in outcrops of the Lina Formation (*sensu* Ramos, 1972), located in the eastern margin of the Lina range, on the road to Jama pass, west of the Toro village (Fig. 1). The fossils were collected from gray shales, yellowish siltstones, and grayish fine sandstones showing yellowish alterations. The rhabdosomes are affected by the low grade metamorphism that presents this unit, mainly those preserved in shales, making the identifications difficult. A preliminary study reveals the presence of a graptofauna composed by biserials that dominate the associations, being the glossograptids scarcely represented.

The rhabdosomes are replaced in pyrite, altered to hematite or limonite, and they are frequently oriented indicating transport. This effect is more evident in sandstones. Mature forms are the more frequent, although they are generally incomplete. The siculae and proximal ends were recorded in particular bedding planes, but they are usually deformed or broken.

The graptolite fauna is dominated by biserial specimens reaching more than 30 mm in length, in some cases. The rhabdosome widens from 0.9-1 mm in the first thecal pair to 2.2- 2.5 mm in medial to distal part. The median septum is slightly undulating in the first part becoming straight distally. The proximal end of these colonies is poorly preserved but it is possible to see the virgella and one apertural spine in th^1 . The strong asymmetry of the first two thecae showed by some specimens could correspond to deformation. The thecal apertures are usually straight and supragenicular walls are short. There are 11-11.5 thecae in the first 5 mm and 7-8 thecae in 5 mm to the distal part. These specimens are referred to as *Pseudamplexograptus* cf. *P. distichus* (Eichwald). Moreover, the fauna contains poorly preserved rhabdosomes that can be determined as

Archiclimacograptus sp., *Oelandograptus*? sp., *Hustedograptus*? sp., and *Glossograptus hincksii fimbriatus* Hopkinson.

The mentioned graptolites, the only fossils recovered from this range, confirm the late Darriwilian age *sensu lato*, as proposed by Bahlburg et al. (1990) for the area. If the classification of *P. distichus* was correct, it would be possible to refer the bearing strata to a level overlying the *Pterograptus elegans* Zone. The material named by Ramos (1972) as *Glyptograptus euglyphus* Lapworth var. *linensis* n. var. should be restudied on the light of new information.

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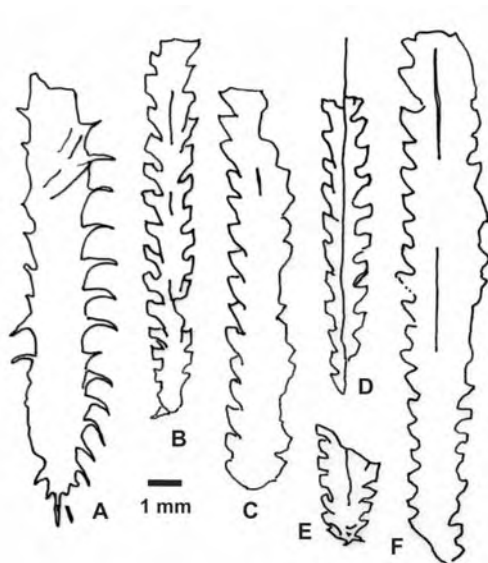


Figure 2. Darriwilian graptolite faunas from the Lina range, northwestern Puna of Jujuy, Argentina. A, *Glossograptus hincksii fimbriatus* Hopkinson (locality 6); B, *Archiclimacograptus* sp. (locality 6); C, *Hustedograptus*? sp. (locality 2); D, *Oelandograptus*? sp. (locality 2); E, F, *Pseudamplexograptus* cf. *P. distichus* (localities 2, 3).

PATTERNS OF ORIGINATION AND DISPERSAL OF MIDDLE TO LATE ORDOVICIAN BRACHIOPODS: EXAMPLES FROM SOUTH CHINA, EAST GONDWANA, AND KAZAKH TERRANES

I.G. Percival¹, L.E. Popov², R.B. Zhan³ and M. Ghobadi Pour⁴

¹ Geological Survey of New South Wales, 947-953 Londonderry Rd, Londonderry 2753, NSW, Australia.
ian.percival@industry.nsw.gov.au

² Department of Geology, National Museum of Wales, Cardiff CF10 3NP, Wales, United Kingdom.
leonid.popov@museumwales.ac.uk

³ State Key Laboratory of Palaeobiology and Stratigraphy, Nanjing Institute of Geology and Palaeontology, Chinese Academy of Sciences, Nanjing 210008, China. rbzhan@nigpas.ac.cn

⁴ Department of Geology, Faculty of Sciences, Golestan University, Gorgan 49138-15739, Iran. m.ghobadipour@gu.ac.ir

Keywords: Brachiopods, biogeography, terranes, Darriwilian, Sandbian, Katian.

INTRODUCTION

Brachiopod and other faunas (e.g. trilobites, sponges) from several of the mosaic of terranes now forming Kazakhstan have long been recognised as sharing close affinities with those of central New South Wales (NSW) during the Late Ordovician (e.g. Webby et al., 2000), as is increasingly evident from quantitative analyses (e.g. Nikitin et al., 2006; Popov et al., 2009). Reconstructions of Middle to Late Ordovician palaeogeography (Fig. 1) interpose South China, North China and Tarim between the supercontinent of Gondwana (with eastern Australia and New Zealand on its eastern margin), and the Kazakh terranes further west. Assessment of trilobite faunas suggests strong linkages between North China and Eastern Australia in the Late Ordovician, more so than between South China and Eastern Australia (Ghobadi Pour et al., 2011). Unfortunately, Ordovician brachiopod faunas from North China, Tarim and Iran are generally insufficiently known to permit accurate assessment of their biogeographic affinities. However, those from South China are well documented, enabling detailed evaluation to be made of their relationship to those of Eastern Gondwana and the Kazakh terranes.

In this paper we utilise a considerably improved and expanded database listing all brachiopod genera known from the Kazakh terranes, South China and Eastern Gondwana, constrained wherever possible by graptolite and conodont-based biostratigraphy to ensure precision in correlation, to assess their biogeographic affinities. We follow the scheme of Rong et al. (1995) to differentiate *endemic*, *regional*, and *cosmopolitan* genera. The latter were removed from the analysis, as advocated by Candela (2006), to clarify the signal of regionally-distributed and endemic genera. We were careful to compare only those brachiopod associations existing in similar benthic assemblages (BA), as faunas from shallow- and deep-water settings are generally markedly disparate and can lead to erroneous conclusions as to their biogeographic affinity. Multivariate cluster analysis, using the PAST computer program (Hammer et al., 2001), demonstrates that as the position of South China changed during the Darriwilian to Hirnantian

interval, it developed increasing faunal affinities with the Kazakh terranes and with East Gondwana (particularly NSW). By examining records of first and subsequent appearances of genera within these regions, we aim to recognise relative centres of origination and timing and direction of faunal migration amongst these regions.

PRE-DARRIWILIAN BRACHIOPOD DISTRIBUTION IN SOUTH CHINA

The South China block differs from most of the other regions included in this analysis (only the Tasmanian Delamerian margin being comparable) by virtue of its tectonic setting – a broad shallow marine shelf (the Yangtze Platform), flanked by a slope environment (Jiangnan Slope) leading into the deepwater Zhujiang Basin. A continuous record of sedimentation extends from the earliest Ordovician to the top of the Hirnantian, with brachiopods present in nearly all levels at different localities. The Ordovician brachiopod radiation in China initially was concentrated in the central part of the Upper Yangtze Platform, with subsequent gradual expansion of new communities into benthic regimes that were both offshore, deeper water and nearer shore, shallower water, as South China drifted into the tropical latitudes (Zhan et al., 2006).

Zhan et al. (in press) analysed the record of gamma-diversity of brachiopod genera throughout the Early and Middle Ordovician (up to the early Darriwilian) in South China. Their biogeographic analysis reveals that the Tremadocian brachiopod fauna of South China was initially dominated by cosmopolitan genera, with increasing similarity to Laurentian faunas developing during the later Tremadocian. From the Floian, as South China drifted away from Gondwana, affinity of its brachiopod faunas gradually shifted from Europe (peri-Gondwanan terranes e.g. Bohemia and England), North Africa (Morocco) and South America (Bolivia and Argentina) to closer relationships with the terranes of Baltica, Avalonia, Sibumasu and southern Kazakhstan (the latter first becoming apparent from the *clavus* Biozone of the latest Dapingian).



Figure 1. Palaeogeographical reconstruction for the Late Ordovician (Katian), adapted from Popov et al. (2009) and Ghobadi Pour et al. (2011). Abbreviations for Kazakhstani island arcs and microplates: A-Zh = Atasu-Zhamshi, Ak = Akbastau, Ch-T = Chingiz-Tarbagatai, K-N = Karatau-Naryn, NTS = North Tien Shan.

REMARKS ON KAZAKHSTANIAN TERRANES

Three major clusters of Kazakhstani terranes can be recognised in the Middle to Late Ordovician (Popov et al., 2009). The southern cluster includes three crustal terranes, including the Chu-Ili and Karatau-Naryn blocks that were amalgamated by the Late Silurian, with the North Tien Shan microplate sandwiched

between them. This southern cluster is separated from the Atasu-Zhamshi microplate by an oceanic suture (see Popov et al. 2009). At least three major island arc systems (some of which may be of intraoceanic origin) can be recognised, including the Akbastau, Chingiz-Tarbagatai and Boshchekul terranes.

Another group of early Palaeozoic terranes is preserved in north-central Kazakhstan, including (according to differing interpretations) the Kalmyk Kol-Kokchetav unit of Şengör and Natal'in (1996), the Shatsk and Kokchetav microplates of Dobretsov et al. (2006) and adjacent island arcs. In various plate tectonic models they are considered either as separate Early Palaeozoic microplates, or as an integral part of a larger microcontinent that also included North Tien Shan. Differing polarities of the surrounding Selety, Ishim and Stepnyak volcanic arcs, the duration of island arc volcanism, and dating of major accretionary events, imply that the various independent Early Palaeozoic units did not interact with the south Kazakhstanian cluster of terranes until at least the Late Ordovician (Dobretsov et al., 2006).

Knowledge of Middle and Late Ordovician brachiopod faunas from Kazakhstan is uneven; data is most complete for Chu-Ili, and reasonable for the Chingiz-Tarbagatai and Boshchekul terranes whereas information presently available for other terranes is inadequate. There is also a considerable problem with dating of Kazakhstanian faunas. Conodonts remain very poorly known (except from cherts) and graptolite biostratigraphy requires significant revision. Thus attribution of faunas to certain graptolite biozones must be considered as provisional.

EASTERN GONDWANA TERRANES AND CONTINENTAL MARGIN

Brachiopod faunas spanning the Darriwilian to Katian interval are represented in Eastern Australia from Tasmania and central NSW. The tectonic settings of these areas are dissimilar, with the Tasmanian succession developed as predominantly shallow water shelfal carbonates on the Delamerian margin of Gondwana, whereas in NSW contemporaneous brachiopods are preserved in limestones and deeper-water clastic rocks fringing volcanic islands of the Macquarie Arc, which is interpreted as having formed offshore to the continental margin. Despite their relatively close proximity in the later Ordovician, the brachiopod faunas of NSW (based on a database comprising 80 genera) and Tasmania (including 30 genera) share surprisingly few close biogeographic similarities. This has been related to separation of the two regions by deep ocean basins swept by strong currents (Webby et al., 2000). Ordovician strata in New Zealand, confined to the northern part of the South Island, include remnants of island-flanking limestones surrounded by deepwater graptolitic shales that also lay offshore to the Gondwanan margin. Brachiopods from this region have so far only been described from two levels, one middle Darriwilian and the other Hirnantian.

METHODOLOGY

In order to more precisely analyse biogeographic linkages amongst the terranes and regions that are the subject of this study, the stratigraphy of each has been as finely subdivided as is practicable using graptolite and conodont zones, and where possible each fossiliferous horizon has been assigned to a Benthic Assemblage (BA). The latter is especially significant in biogeographical analysis as many brachiopod taxa are restricted to generally one and occasionally two adjacent BAs. The database encompassed 66 stratotectonic units and subunits, spanning the early Darriwilian to late Hirnantian interval, with ranges recorded for a total of 342 brachiopod genera (excluding 12 cosmopolitan taxa). Data was incorporated from 9 levels in South

China, 7 in central NSW, 7 in Tasmania, 2 in NZ, 2 in the Alborz Terrane of Iran, with the remainder from various terranes in Kazakhstan including the Chu-Ili Terrane (16 levels and BAs), Karatau-Naryn Terrane (2), North Tien-Shan Terrane (2), Chingiz-Tarbagatai Terrane (7), Boshchekul Terrane (5 levels), Ishim-Selety terrane cluster (4), Atasu-Zhamshi microplate (1), and the relatively poorly-known Zerafshan-Hissar region (2).

Interpretation of faunal affinity was facilitated by multivariate cluster analysis (Raupe-Crick similarity), using the computer program PAST, of several biostratigraphic intervals, i.e. Darriwilian, Sandbian, early Katian (equivalent to Eastern Australian zones Eastonian 1 and 2), and middle to late Katian (Eastonian 3 to Bolindian 3). Two of the major intervals were selected for illustration (Fig. 2A-B). The Hirnantian was not analysed due to the predominance of cosmopolitan genera. The clusters were checked to identify any inconsistencies related to differences in Benthic Assemblages. Finally, the distribution matrix (in an Excel spreadsheet) was colour-coded to show first appearances and subsequent records of regionally-distributed genera, in order to enable recognition of potential origination centres and migration pathways.

RESULTS

Faunal affinities related to biogeography

Darriwilian – Sandbian interval (Fig. 2A): Most assemblages from South China through this interval cluster closely, with Sandbian faunas strongly linked to a contemporaneous fauna from the Chu-Ili Terrane. An early Darriwilian assemblage from Tasmania shows more distant affinities with these South China faunas. Other Tasmanian faunas (Da2-3, Da4 and Gi2 ages) are tightly clustered, and broadly linked with those from the Chu-Ili, Boshchekul and Ishim-Selety terranes. Two late Darriwilian to earliest Sandbian faunas from South China

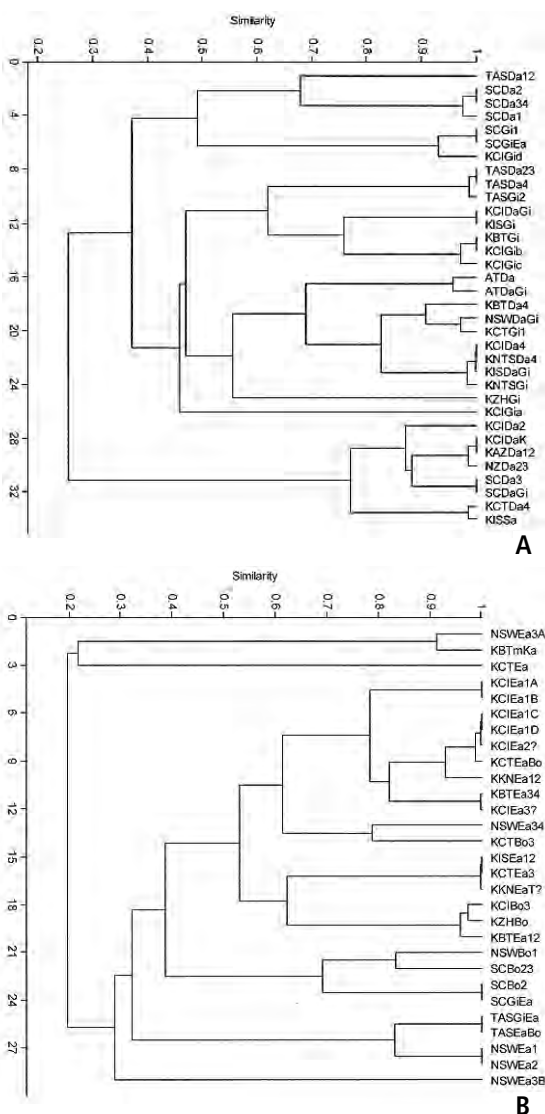


Figure 2. Multivariate cluster analysis (Raupe-Crick similarity) of brachiopod distribution through (A) Darriwilian – Sandbian and (B) Katian intervals, in South China, East Gondwana, Kazakh terranes and Alborz Terrane. For interpretation, see text. Abbreviations: TAS = Tasmania, NSW = New South Wales, NZ = New Zealand, SC = South China, AT = Alborz Terrane, Kazakh terranes include KZH = Zerafshan-Hissar region, KCI = Chu-Ili, KIS = Ishim-Selety, KNTS = North Tien Shan, KBT = Boshchekul, KCT = Chingiz-Tarbagatai, KAZ = Atasu-Zhamshi, KKN = Karatau-Naryn; age abbreviations (followed where known by zone number) are Da = Darriwilian, Sa = Sandbian, K = Katian, Gi = Gisbornian, Ea = Eastonian, Bo = Bolindian.

form part of a diverse cluster that incorporates early to middle Darriwilian assemblages from the Chu-Ili Terrane, Atasu-Zhamshi microplate, and New Zealand, and is more distantly related to late Darriwilian faunas from the Chingiz-Tarbagatai and Ishim-Selety terranes. Alborz Terrane faunas group closely together as expected, and are broadly related to one cluster encompassing late Darriwilian to early Sandbian faunas from the Chingiz-Tarbagatai and Boshchekul terranes, and to another tight cluster grouping latest Darriwilian to earliest Sandbian faunas from the Chu-Ili, North Tien-Shan and Ishim-Selety terranes.

Katian: four main groupings are evident (Fig. 2B). The first includes faunas mainly inhabiting middle shelf (BA3) environments from Tasmania and NSW. A second group includes assemblages of earliest Katian and late Katian (Bolindian 2-3) age from South China that are linked with a fauna of early Bolindian age from NSW. The third and most diverse group is dominated by faunas from slightly older (early to middle Katian) terranes from Kazakhstan, including the Chu-Ili, Chingiz-Tarbagatai, Boshchekul and possibly the Karatau-Naryn terranes and the Ishim-Selety cluster; this group is weakly associated with cluster grouping a middle Katian fauna from NSW and a latest Katian fauna from the Chingiz-Tarbagatai Terrane. A fourth, rather loosely linked group showing little in common with the others, includes a deeper-water fauna of Eastonian 3 age from NSW, together with faunas from the Boshchekul and Chingiz-Tarbagatai terranes.

Origination and migration trends in selected genera

Genera originating in or spreading from Kazakh terranes: *Dulankarella* is first noted in Sandbian and early Katian strata of the Chu-Ili Terrane; it subsequently occurs in middle Katian rocks of the Boshchekul Terrane and in NSW, and finally appears in the late Katian of the Chingiz-Tarbagatai Terrane. *Mabella* is another genus with a well-defined migration pathway, first appearing at the Darriwilian-Sandbian boundary in the Chu-Ili Terrane where it ranges into the Sandbian; it then spreads to the Ishim-Selety region in Sandbian and early Katian times, becomes established in NSW in the early and middle Katian, and is also present in middle Katian rocks of the Chingiz-Tarbagatai and Karatau-Naryn terranes. The distinctive acrotretide *Atansoria* only occurs in two regions: the Selety Terrane where it originates during the Sandbian, then in NSW in middle Katian slope-edge limestones. *Gunningblandella*, previously thought to be endemic to middle Katian strata in NSW, is also known from early Katian rocks of the Chu-Ili Terrane (and has recently been recorded by Robin Cocks from a slightly older level in Avalonia). *Metambonites* also makes its first appearance in the Chu-Ili Terrane in early Katian time, before reaching NSW in the late Katian and finally South China in the latest Katian. *Synambonites* originates in the Boshchekul Terrane in the middle Katian, before following a migration pathway to NSW and then South China identical to that of *Metambonites*. *Altaethyrella*, which is widely distributed amongst the Chu-Ili, Boshchekul and Karatau-Naryn terranes throughout the Katian, only reaches South China in the latest Katian. Trimerellides are known to have originated in the Chingiz-Tarbagatai Terrane (*Ovidiella*, *Palaeotrimerella* and *Ussunia*, of latest Darriwilian age) and Chu-Ili Terrane (*Adensu*, early Sandbian) before migrating to NSW in the earliest Katian and appearing in South China by the middle Katian (Popov et al. in press). Of several groups of atrypides first appearing in Kazakh terranes, lissatrypides are probably the oldest and occur in the latest Darriwilian of the Chingiz-Tarbagatai Terrane, represented by *Rozmanospira*; the earliest atrypides (*Sulcatospira*) follow in the Sandbian with this genus spreading to other Kazakh terranes in the Katian (but apparently not further afield).

Genera originating in or spreading from South China: Strangely, the *Saucrorthis* Fauna that is so characteristic of the Darriwilian in South China does not seem to have spread to any of the Kazakh

terrane. Many of the associated genera present at this time first appeared in South China in the Early Ordovician, e.g. *Martellia* has a long record there from middle Floian to the late Darriwilian, reaching the Chu-Ili Terrane only in the Dapingian (and not persisting past the early Darriwilian). South China also carried several stocks of Scoto-Appalachian origin with it as it moved towards the Kazakh terranes. An example is *Glyptomena*, which has a long record throughout the Darriwilian in South China, prior to its appearance during the Sandbian in the Chu-Ili and Selety terranes, persisting in the latter region into the early Katian.

Genera originating in East Gondwana: *Durranelia* provides a link between NSW (first appearing in the middle Katian) and South China (late Katian), as does the trimerellide *Belubula*, only occurring otherwise in the late Katian of South China, where it is represented by its junior synonym *Zhuzhaiia*. Three genera of Tasmanian origin in the early to middle Darriwilian have very restricted distributions: *Lepidomena* also occurs in late Darriwilian strata of the Boshchekul Terrane, *Maydenella* subsequently appears in the late Darriwilian of South China, and *Teratelasnella* is found both in Sandbian rocks of the Chu-Ili Terrane and in the late Katian of NSW. However, 13 genera found in Darriwilian to Katian rocks of NSW and another 3 from Tasmania are strictly endemic to each of those regions, despite their relative proximity.

DISCUSSION

Previous analyses of Ordovician brachiopod biogeography of the Kazakh terranes, South China and East Gondwana (e.g. Candela, 2006 and Nikitin et al., 2006) relied on considerably smaller listings of genera from a limited number of localities and horizons, compressing the age range to obtain a generalised picture. This tended to obscure the changing pattern of faunal affinities as terranes and continental blocks (particularly South China) moved through time. Only with the benefit of the expanded databases we have at our disposal, based on recent and unpublished systematic studies, and subdivided with substantially increased precision, do faunal relationships start to clarify and some broad trends become apparent. Firstly, the northward path of South China is reflected by changing faunal affinities, with increasing linkages to Kazakhstani terranes developing from the Sandbian, and strong affinities to NSW faunas becoming evident by the middle to late Katian as South China intersected migration pathways defined by surface currents. Secondly, the Chu-Ili Terrane of Kazakhstan stands out as an origination centre for a number of biogeographically significant genera, with the Chingiz-Tarbagatai, Boshchekul and Selety terranes providing secondary centres of origination. Even where the latter three terranes do not have first appearances, they are frequently the second port of call on migration pathways from the Chu-Ili Terrane.

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RECENT DISCOVERIES AND A REVIEW OF THE ORDOVICIAN FAUNAS OF NEW ZEALAND

I.G. Percival¹, R.A. Cooper², Y.Y. Zhen³, J.E. Simes² and A.J. Wright⁴

¹ Geological Survey of New South Wales, 947-953 Londonderry Rd, Londonderry 2753, NSW, Australia.

ian.percival@industry.nsw.gov.au

² Institute of Geological and Nuclear Sciences, P.O. Box 30 368, Lower Hutt, New Zealand.

r.cooper@gns.cri.nz, j.simes@gns.cri.nz

³ Palaeontology Section, Australian Museum, 6 College St, Sydney NSW 2010, Australia.

yongyi.zhen@austmus.gov.au

⁴ School of Earth and Environmental Sciences, University of Wollongong, Wollongong, NSW 2522, Australia.

tony_wright@uow.edu.au

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INTRODUCTION

Fossiliferous Ordovician rocks are of limited extent in New Zealand, being largely restricted to northwest Nelson and Westland in the northern part of the South Island, with some isolated exposures at the southern extremity of Fiordland (Fig. 1). The known stratigraphic record, though incomplete, covers much of the period, and new information from study and revision of old collections (mostly dating from the 1960s and 1970s) continues to fill in the gaps. These data are critical to a better understanding of New Zealand's place in the Ordovician world, when it occupied an isolated position facing the palaeo-Pacific Ocean offshore to East Gondwana. The closest contemporaneous strata are located in Tasmania, Victoria and New South Wales in southeastern Australia, now separated from New Zealand by the Tasman Sea but in Ordovician times probably situated even further west (though at a similar 10-20 deg. N palaeolatitude). In this paper we review all known records of Ordovician fossils described from New Zealand, placing these in their currently-accepted stratigraphic and tectonic context. New data are presented on research currently underway into faunas from the Maruia–Springs Junction–Lake Daniels region, southeast of Reefton (Fig. 1).

Ordovician geology of northwest Nelson and Westland falls into two lithologically distinct terranes (Cooper, 1989), separated by a north-south trending zone of major transcurrent and thrust faulting called the Anatoki Fault. Ordovician faunas from in situ carbonate bedded formations, and from isolated allochthonous limestone blocks, are confined to the Takaka Terrane, on the eastern side of the Anatoki Fault. The carbonate-dominated formations (Patriarch Formation, Summit Limestone, Owen Formation, and Arthur Marble) of the Takaka Terrane extend in age from late Furongian (Late Cambrian) to probably near the latest Ordovician (Fig. 2). The Wangapeka Formation, consisting of dark grey shale and strongly bioturbated quartz-sandstone, is laterally equivalent to the upper Arthur Marble and contains a Hirnantian fauna in its uppermost beds. The latest Cambrian to Ordovician succession of the Takaka Terrane

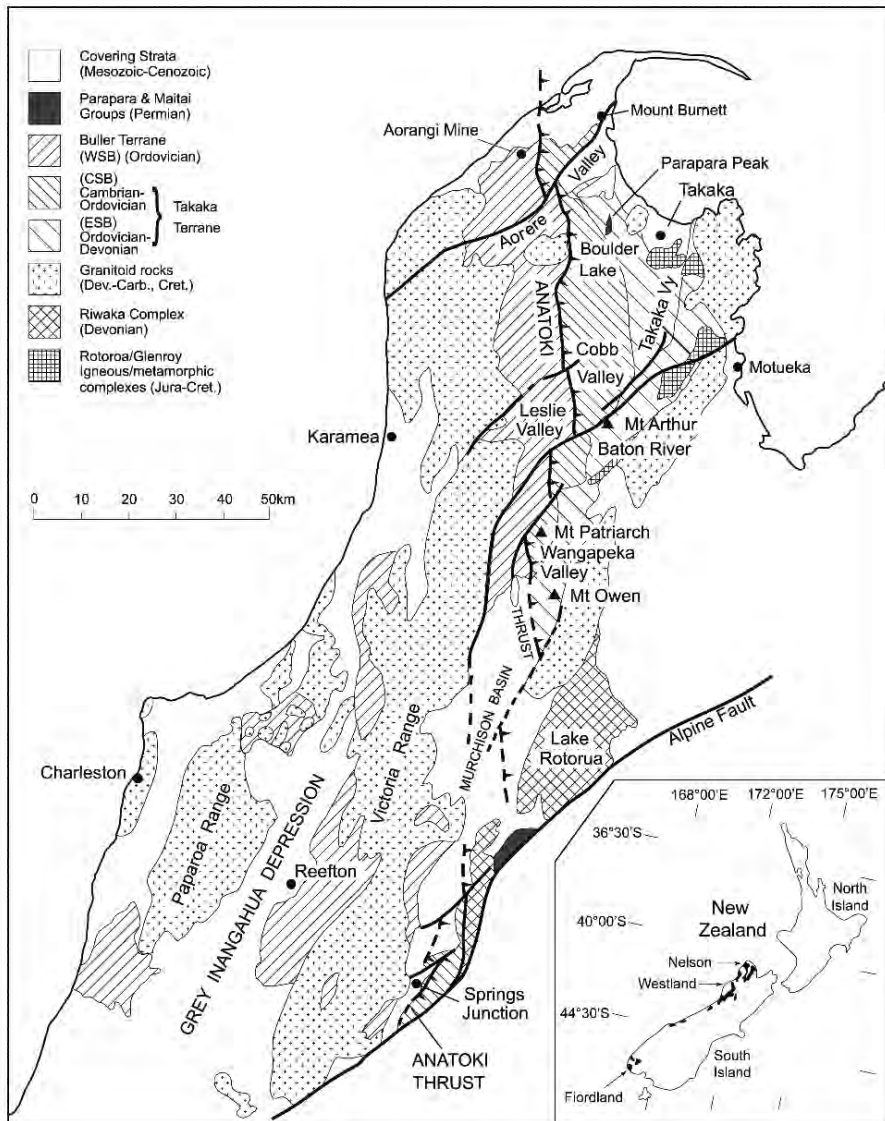


Figure 1. Inset map of New Zealand shows position of outcropping Ordovician rocks in Nelson, Westland and Fiordland in the South Island. Main map depicts simplified geology of the northwestern part of the South Island, showing localities mentioned in the text.

conformably overlies the Anatoki Formation, comprising siliceous and volcanogenic green sandstone of Late Cambrian age.

The Ordovician sequence of the Buller Terrane, west of the Anatoki Fault (Fig. 2), differs from that of the Takaka Terrane in being dominated through the Tremadocian to latest Darrivilian interval by graptolitic shales, including the uppermost Webb Formation, Aorangi Mine Formation, and Slaty Creek Formation. This succession is conformably overlain by dark shale, quartz sandstone and green laminated siltstone of the

Douglas Formation, containing earliest Sandbian fossils at the base and probably extending well into the Katian (although age-diagnostic fossils have not been found in its middle and upper parts).

REVIEW OF PREVIOUS ORDOVICIAN STUDIES

Fiordland

Poorly preserved obolid brachiopods and phyllocarids were documented from strata at Preservation Inlet and Cape Providence in Fiordland by Chapman (1934a, b). Early Ordovician graptolites were subsequently described from these rocks, now known as the Preservation Formation, by Benson and Keble (1935) and Skwarko (1958). The sequence of graptolitic shales at Cape Providence is complete over an interval ranging from late Tremadocian (Lancefieldian 2, in terms of the Victorian graptolite zonation – see VandenBerg and Cooper 1992) to latest Floian (Castlemainian 1) according to Cooper (1979a, 1981), but no new descriptive work on the fauna from Fiordland has been undertaken for over 50 years.

Takaka Terrane, Nelson and Westland

All other studies on New Zealand's Ordovician faunas have concentrated on exposures in northwest Nelson and adjacent Westland. From here, trilobites supposedly of Ordovician age were first described from two quite separate stratigraphic levels by Reed (1926). One of these trilobites, since revised by Wright et al. (1994) as *Hysterolesus hectori* (Reed), comes from the latest Cambrian Faunule 1 level near the base of the Patriarch Formation, which extends into the earliest Ordovician and is overlain by the Summit Limestone of Tremadocian to Floian age. Wright et al. (1994) described an additional 40 species of trilobites from the Tremadocian section of the Patriarch Formation, together with a small conodont fauna and two graptolites. Conodonts described by Cooper and Druce (1975) from the middle and upper parts of the Summit Limestone were reassessed by Wright et al. (1994), who revised the age of the middle Summit Limestone (basal Faunule 4) as latest Tremadocian; the age of the uppermost exposed Summit Limestone (Faunule 5) at Mount Patriarch is earliest Floian.

Simes (1980) recorded latest Darriwilian to earliest Sandbian conodonts, including *Pygodus anserinus*, from the lower part of the Arthur Marble at Mount Owen. In the Cobb Valley and Mount Patriarch areas, the Summit Limestone contains conodonts which indicate an age range of latest Cambrian to Middle Ordovician (Darriwilian: Cooper and Bradshaw, 1986; Cooper, 1989). Middle Ordovician conodonts from allochthonous limestone at Thompson Creek near the Paturau River were first reported by Wright (1968) who suggested a Llanvirnian age on the basis of a limited and undescribed fauna. From its age and lithology, the Thompson Creek lens (Zhen et al., 2009; Percival et al., 2009) is interpreted as equivalent to the uppermost part of the Summit Limestone. The presence of conodonts (Plate 1, figs 1-24) including

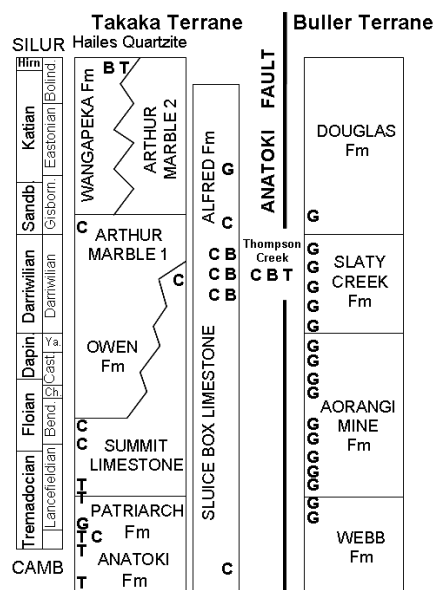


Figure 2. Correlation chart (modified and updated from Webby 1981) showing equivalence of Ordovician stratigraphic units in the Takaka and Buller terranes, and succession in the Maruia – Springs Junction area; graptolite (G), conodont (C), brachiopod (B) and trilobite (T) occurrences are shown.

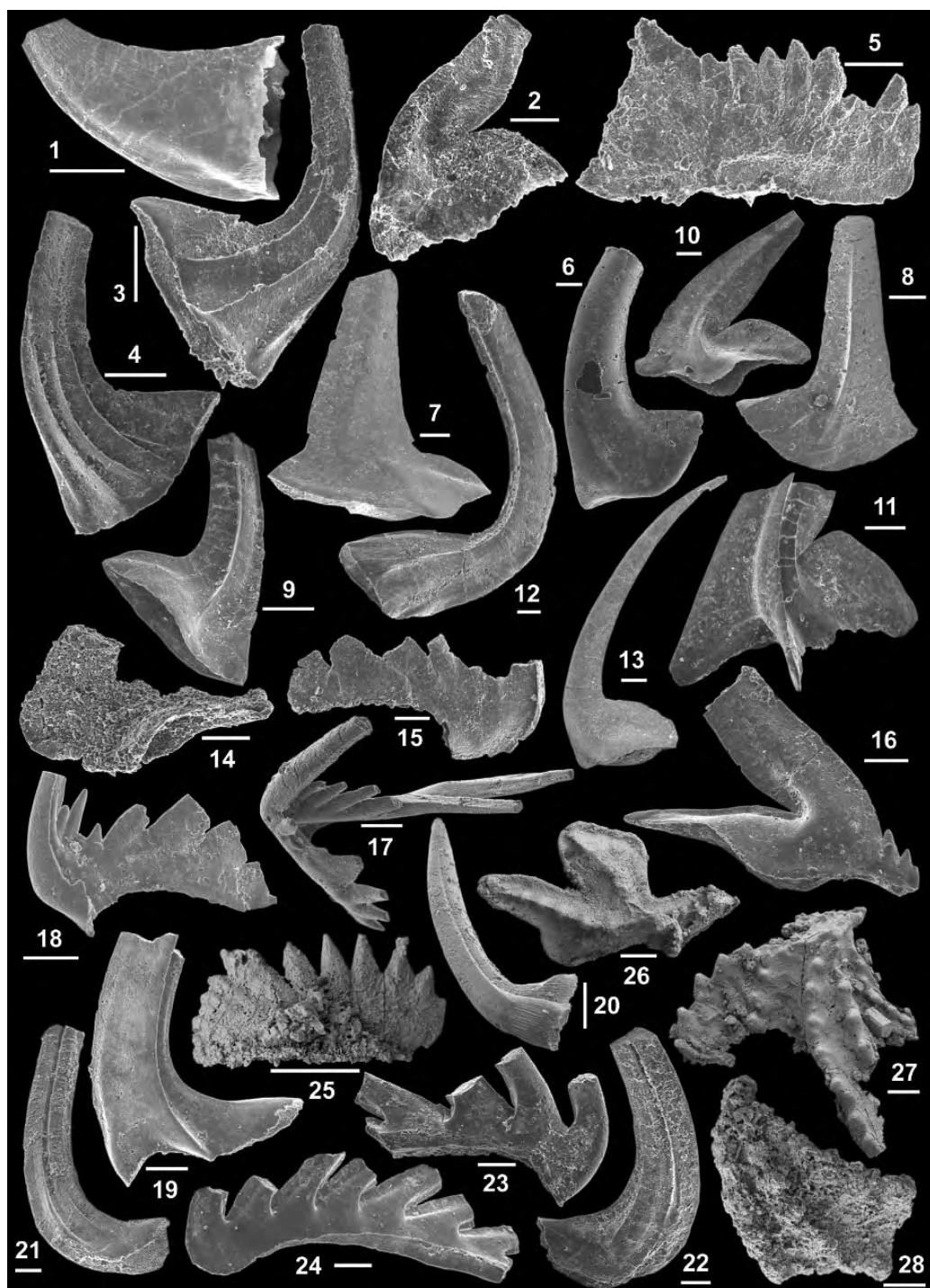
Histiodela holodentata, *Baltoniodus?* sp., *Paroistodus originalis*, *P. horridus*, *Periodon macrodentatus*, *Protopanderodus* sp. cf. *P. varicostatus*, *Costiconus ethingtoni* and *Venoistodus balticus* in the fauna indicates a Darriwilian (late Da2 to mid Da3) age. Occurrence of *Ansella jemtlandica*, *Baltoniodus?* sp., *Periodon macrodentatus*, *Spinodus* sp., *Spinodus?* sp. and *Histiodela holodentata* in this fauna suggests a relatively deeper water (outer shelf to slope) setting. Lingulate brachiopods described from Thompson Creek include the new species *Hyperbolus?* *thompsonensis*, *Cyrtontreta robusta*, *Scaphelasma paturauensis*, *Torynelasma takakaea* and *Nushbiella neozealandica*, together with representatives of *Spinilingula*, *Schizotreta*, *Trematis*, *Cyrtontreta*, *Physotreta?* and *Lurgiticoma?* *Undiferina nevadensis* has previously been described from Nevada and west Kazakhstan. The brachiopod fauna, the first of Middle Ordovician age to be described from New Zealand, shows strong affinities to Middle Ordovician faunas from Kazakhstan, Nevada, and to a slightly younger assemblage from the Pratt Ferry Formation of Alabama. A trilobite documented from Thompson Creek is tentatively assigned to *Gogoella*, a genus previously described from Western Australia and Argentina.

The youngest known Ordovician fossils described from New Zealand include representatives of the globally distributed *Hirnantia* Fauna of latest Ordovician (Hirnantian) age, described by Cocks and Cooper (2004) from the uppermost Wangapeka Formation in Wangapeka Valley. Genera recorded include the brachiopods *Eostropheodonta*, *Plectothyrella*, *Cliftonia*, *Leptaena*, together with several tentatively-identified forms, the trilobite *Mucronaspis* and remains of several other poorly preserved trilobites, bryozoans, echinoderms, molluscs, corals and ostracodes. Cooper (1968) also reported the occurrence of corals, including *Proheliolites*, *Plasmoporella* and *Favistella*, of Late Ordovician (Bolindian?) age from the upper Arthur Marble in Takaka Valley, but these have not been described.

Buller Terrane, Nelson and Westland

Cooper (1979b) described an almost complete and well-preserved graptolite faunal succession ranging in age from late Tremadocian to early Sandbian, from siltstones and shales of the Aorangi Mine, Slaty Creek and Douglas formations in the vicinity of Aorangi Mine between the Paturau River and the West Coast. The Anthill Black Shale member, overlying the basal Malone member of the Aorangi Mine Formation, contains the first identifiable graptolites in the succession, ranging in age from upper Tremadocian (Lancefieldian 2, Zone of *Adelograptus victoriae*) to early Dapingian (Castlemainian 3, Zone of *Isograptus victoriae maximus*) age. The graptolite succession in the overlying Battery Member of the Aorangi Mine Formation continues with the Castlemainian 4 Zone of *Isograptus victoriae maximodivergens* and the lower

Plate 1. Representative Darriwilian conodonts from New Zealand. Figs 1-24 are from the Thompson Creek area [figs 1, 3-4, 6-13, 16-24 from locality CN641, 2 from locality CN917, 5 from locality 1-39, 14-15 from locality CN918 – see Zhen et al. (2009) for details] and figs 25-28 are from the Maruia – Springs Junction area. 1-2, *Ansella jemtlandica* (Löfgren, 1978); 1, Pb element; 2, M element. 3-4, *Costiconus ethingtoni* (Fähraeus, 1966); 3, Sb element; 4, Sc element. 5, *Histiodela holodentata* Ethington & Clark, 1982; Pa element. 6-7, *Drepanoistodus tablepointensis* Stouge, 1984; 6, Sc element; 7, Sa element. 8-9, *Drepanoistodus costatus* (Abaimova, 1971); 8, Sa element; 9, Sd element. 10-11, *Oistodus* sp. cf. *O. lanceolatus* Pander, 1856; 10, Sc element; 11, Sd element. 12, *Drepanodus* sp. cf. *D. reclinatus* (Lindström, 1955); Sa element. 13, *Paroistodus originalis* (Sergeeva, 1963); Sd element. 14-15, *Paroistodus horridus* (Barnes and Poplawski, 1973); 14, Pb element; 15, Sc element. 16-18, *Periodon macrodentatus* (Graves and Ellison, 1941); 16, M element; 17, Sb element; 18, Sa element. 19, *Protopanderodus cooperi* (Sweet and Bergström, 1962); Sa element. 20, *Protopanderodus?* *nogamii* (Lee, 1975); Sa element. 21-22, *Protopanderodus* sp. cf. *P. varicostatus* (Sweet and Bergström, 1962); 21, Sc element; 22, Sb element. 23-24, *Spinodus* sp.; 23, Sc element; 24, M element. 25, *Histiodela kristinae* Stouge, 1984; Pa element, locality CN463. 26, *Eoplacognathus suecicus* Bergström, 1971; Pa element, locality CN579. 27, *Pygodus anitae* Bergström, 1983; Pa element, locality CN574. 28, *Pygodus serra* (Hadding, 1913); Pa element, locality CN487. Scale bars: 100 µm.



Yapeenian Zone of *Ocograptus upsilon* (mid to late Dapingian). This zone is also represented in thin black shales of the overlying Jimmy Creek Quartzite at the top of the Aorangi Mine Formation. The succeeding Slaty Creek Formation comprises 3 informal members; *Undulograptus austrodentatus* in the lowermost of these indicates an earliest Darriwilian age, the middle member contains representatives of the Darriwilian 3 Zone of *Pseudoclimacograptus? decoratus*, and this zone continues into the overlying third member. Formation A of Cooper (1979b), now known as the Douglas Formation, contains the first appearance of *Dicellograptus* in the form of *D. cf. vagus*, indicative of Subzone Da4b at the top of the Darriwilian (VandenBerg and Cooper, 1992). Slightly higher in this formation is found *Nemagraptus gracilis*, zonal indicator of the basal Sandbian. It is there associated with the second of Reed's (1926) trilobites, redescribed by Wright (2009) as *Basiliella collingwoodensis* (Reed), together with the trinucleid trilobite *Incaia bishopi* Hughes and Wright, 1970.

GEOLOGY OF THE MARUIA – SPRINGS JUNCTION AREA

The Lower Paleozoic sequence in the Maruia – Lake Daniels – Springs Junction area is interpreted as the southernmost extension of the fossiliferous Lower Palaeozoic terranes of NW Nelson (Cooper 1989). Rocks lithologically matching the Leslie and Douglas Formations of the Buller Terrane outcrop along the western side of the ridge separating Lake Daniels from the Maruia Valley. To the east is a heterogeneous mixture of fossiliferous sedimentary and volcanic rocks that match lithologies in the Takaka Terrane. Contact between the two terranes is nowhere exposed but a major fault is inferred, equivalent to the Anatoki Fault of northwest Nelson. Outcrop is generally poor and weathered, except for the limestones.

In the Takaka Terrane succession, a major fault separates an 'upper plate', composed of Sluice Box Limestone overlain by siltstone of the Alfred Formation, from a 'lower plate' composed of three units: (1) grey dolomitic thin-bedded mudstone, orange-weathering ankeritic sandstone, dark muddy diamictite and monomict conglomerate; (2) polymict pebble and granule conglomerate and coarse-grained lithic sandstone; and (3) grey laminated and thin bedded dolomitic siltstone and grey orange-weathering ankeritic mudstone with abundant interlayered acid to intermediate volcanic rocks. None of the three lower plate units contain fossils. The fault is subparallel to bedding in the upper plate and truncates the base of the Sluice Box Limestone. Based on the mapped distribution and structure of the Sluice Box Limestone, the fault, along with the whole upper plate, has been folded into a north-plunging antiform. The fault and antiform were subsequently disrupted by many NE trending faults, probably associated with the Alpine Fault. Fossils are present in both upper plate units, and indicate that age decreases away from the antiform axis, so on this basis the structure is interpreted as an anticline rather than an inverted syncline. The upper plate units are less tectonised than the lower plate units and clearly match the Summit Limestone (Late Cambrian to Darriwilian) and Wangapeka Formation (Darriwilian to Eastonian) of Northwest Nelson, in both age and lithofacies.

Sluice Box Limestone

Two informal members (SB1 and SB2) are recognized in this limestone. No complete sections are exposed through the limestone, so that its internal stratigraphy is inferred from numerous partial sections.

Member SB1 consists of dark flaggy micritic limestone and calcareous shale that varies in thickness and lithology. It is approximately 120 m thick in the eastern limb in Station Creek but in the western limb

it is thinner and is probably fault-truncated. To the north, in Gorge Creek, the member contains abundant thin interbedded chert layers and is less calcareous than to the south. Soft sediment deformation, including probable slump folds, is common. No age diagnostic fossils are known. Member SB2 is about 70 m thick in the upper reaches of Station Creek, increasing in thickness to the north in Gorge Creek. It consists of pale coloured micritic, partially recrystallised massive to flaggy limestone with rare interbedded shale. Oolite-rich lenses are common and lenses of sparry limestone are found at several localities, particularly near the upper boundary. Rich microfossil assemblages (currently under study) have been recovered from several of these sparry layers. Conodonts (Plate 1, figs 25-28) provisionally identified include *Histiodela kristinae*, *Erraticodon* sp., *Polonodus* sp., *Periodon macrodentatus*?, *Costioconus ethingtoni*, *Ansella jemtlandica*, *Paroistodus horridus*, *Pygodus serra* and *P. anitae*. Several middle to late Darriwilian zones may be represented in these faunas. Associated microbrachiopods include *Scaphelasma* sp. and *Dictyonites* sp.

Alfred Formation

Dark, strongly cleaved siltstone and sandstone that overlies the Sluice Box Limestone on both flanks of the Thompson Flat anticline is termed the Alfred Formation. Two informal members (A1, A2) are distinguished.

Member A1 forms the lower beds, that consist of dark calcareous siltstone with rare limestone nodules and bands. Soft sediment deformation (possibly due to slumping) is common. A black shale band less than 15 m above the base on the western anticline limb has yielded graptolites of indeterminate age. Conodonts from the limestone lenses include *Periodon aculeatus*, indicating a late Darriwilian age. Member A1 passes upwards into Member A2, consisting of dark coloured strongly cleaved shale, with alternating quartzose sandstone beds. Graptolites indicate a Late Ordovician (Gisbornian to possibly Eastonian) age.

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ORDOVICIAN GRAPTOLITES AND ACRITARCHS FROM THE BARRANCOS REGION (OSSA-MORENA ZONE, SOUTH PORTUGAL)

J. Piçarra¹, Z. Pereira² and J.C. Gutiérrez-Marco³

¹ Laboratório Nacional de Energia e Geologia (LNEG), Ap. 104, 7801-902 Beja, Portugal. jose.picarra@lneg.pt

² Laboratório Nacional de Energia e Geologia (LNEG), Ap. 1089, 4466-901 S. Mamede Infesta, Portugal. zelia.pereira@lneg.pt

³ Instituto de Geociencias (CSIC-UCM), Facultad de Ciencias Geológicas, José Antonio Novais 2, 28040 Madrid, Spain.
jcgrapto@geo.ucm.es

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INTRODUCTION

Ordovician fossils are relatively scarce in the Portuguese part of the Ossa-Morena Zone, one of the most distinctive Paleozoic domains of the Iberian Massif, where most of the available data come from the Barrancos region and needs an accurate review.

The first account on the existence of Ordovician rocks near the Portuguese small city of Barrancos were published by Delgado (1901, 1908, 1910), who mentioned some graptolites and ichnofossils and defined three lithostratigraphic units from his "lower Silurian" division. In ascending order, these are the Fatuquedo shales, the Barrancos shales (including the "*Phyllodocites* shales" at its upper part) and the Colorada greywackes and quartzites formations. The paleontological record was virtually restricted to the fine shales of the Barrancos Fm. and especially to their "*Phyllodocites* shales", a local facies of very micaceous, grey, green or red slates, not represented in the Spanish counterpart of the Estremoz-Barrancos-Hinojales domain of the Ossa-Morena Zone (Robardet et al., 1998). The age of the overlying Colorada Formation remains unknown, but graptolites recorded just above its top are already representative of the earliest Silurian (Piçarra et al., 1995).

Ordovician graptolites from the Mestre André quarry east of Barrancos (Fig. 1) were first identified by Delgado (1901, p. 215) as *Didymograptus geminus* His., and then as *Didymograptus sparsus* Hopk. (Delgado, 1908 p. 187; 1910, pl. 27). New graptolite findings from the same locality were illustrated by Perdigão (1967, pl. 1, figs. 3-4), who recorded a broken stipe of *D. sparsus* and a proximal fragment of *Didymograptus hirundo* Salter, being the latter species indicative of an "upper Skiddawian" (= late Arenigian) age for the "*Phyllodocites* shales".

The existence of Ordovician rocks in the Barrancos region was questioned in 1981 by Teixeira (publ. 1982), who considered the graptolites from the Mestre André quarry as of Silurian age, following the ideas of Romariz (1962). These authors, in agreement with certain data from Delgado (1908), extended the range of some Ordovician genera like *Didymograptus*, *Phyllograptus* and *Tetragraptus* into the Silurian (see also Teixeira, 1984). Romariz (1962) also cited the occurrence of *Pristiograptus* sp. immediately east of the

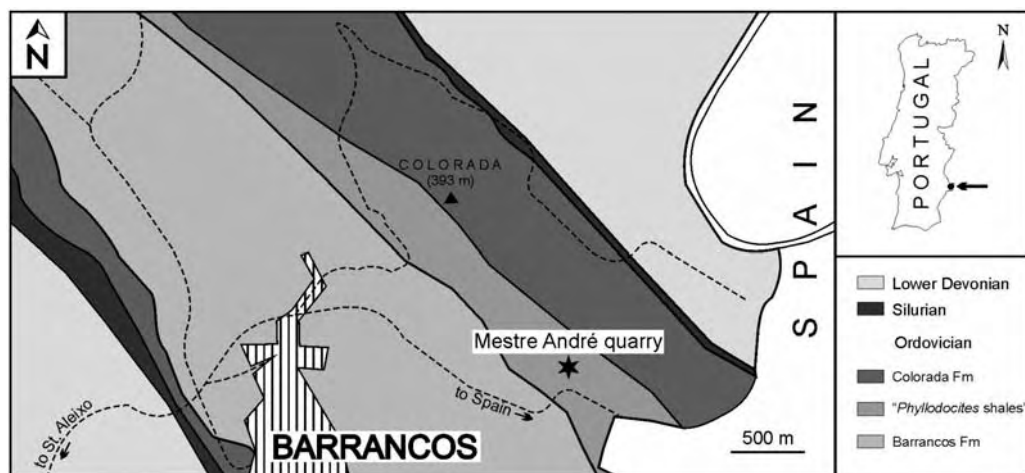


Figure 1. Simplified geological map of the Barrancos Region, with indication of the Mestre André quarry in the "Phyllococites shales", bearing ichnofossils, acritarchs and graptolites.

Mestre André quarry, a record that was reviewed by Gutiérrez-Marco (1981) and reassigned to indetermined dichograptoid stipe fragments coming from an Ordovician outcrop. In his report, Gutiérrez-Marco (1981) confirmed the Ordovician age of the graptolite specimens illustrated by Perdigão (1967), as being most probably indicative of the upper Arenigian *Expansograptus hirundo* Zone.

Palynological research in the "Phyllococites shales" led Cunha and Vanguetaine (1988) the discovery of Lower Ordovician acritarchs in two samples taken west of Barrancos, at km 94.2 of the road to Santo Aleixo, in the western flank of the Barrancos anticline (outside the frame of Fig. 1). The moderately preserved association was placed close to the Tremadocian-Arenigian boundary. A second report on acritarchs from the same unit was recently presented by Borges et al. (2008). The latter samples come from several pieces of rock labelled as "Mestre André quarry, Barrancos", originally from the Delgado collection, deposited at the LNEG Geological Museum in Lisbon. A preliminary palynological study presented by Borges et al. (2008) concludes a Floian age for the acritarch assemblage, which is briefly examined below.

Further research of the present authors in typical outcrops of the "Phyllococites shales" in the Barrancos region, including the Mestre André quarry, didn't produce any positive result for macrofossils or microfossils, despite several sampling campaigns carried out in the region by the present team on the last three decades. For this reason, we suspect that previous records of graptolites by Delgado (1901, 1908, 1910) and Perdigão (1967) may have involved lenticular intercalations of fossiliferous shales temporally exposed during the exploitation of the slate quarry. The review of the original samples preserved in the Delgado collection show that these graptolite-bearing shales are very different to the green to reddish micaceous slates, with common ichnofossils, obtained in the Mestre André quarry for building and roofing purposes. In this sense, the matrix of the fossiliferous samples is usually of grey colour, only lightly micaceous, and have a higher argillaceous content, being more affected by the regional foliation.

The purpose of this note is to present a more detailed revision of the original samples labelled as "pedreira do Mestre André, Barrancos" coming from the Delgado collection. Among them we have restudied the graptolites illustrated in his posthumous monograph (Delgado, 1910), as well as some sponges also recovered from his collection, but coming from a different site (samples no. IGM-10862). Reinvestigation of selected samples has added more detailed palynostratigraphic information to that

presented by Borges et al. (2008). In addition to this, information derived from the abundant, but confusing, record of trace fossils in the same strata is briefly examined for paleoenvironmental discussion.

GRAPTOLITES

We have examined seven specimens of Ordovician graptolites from the Delgado collection, nos. 6000-6005, plus one unnumbered. Specimens 6000, 6004 and those without number are stipe fragments, and the remaining correspond to proximal parts of the rhabdosome usually including the complete sicula. They are all labelled as "*Didymograptus sparsus* Hopk.", except specimen no. 6004 that was misidentified by Delgado's handwriting as the Silurian graptolite "*Monograptus lobiferus* McCoy" (Delgado, 1910, pl. 27, fig. 5). The latter clearly corresponds to a badly preserved stipe of an Ordovician dichograptoid, strongly affected by cleavage. Besides this material, Perdigão (1967, pl. 1, fig. 2) illustrates another specimen of *D. sparsus* from the Delgado collection, but this might not have been reintegrated to the collection and is probably lost.

All graptolites recorded from the Mestre André quarry belong to the form genus *Expansograptus* Bouček and Přibyl, which includes a variety of horizontal didymograptids lacking the dorsal virgella characteristic of *Xiphograptus* Cooper and Fortey, and that could be partly related with the genus *Didymograptellus* Cooper and Fortey *sensu lato* (its type species also possess virgella: Maletz, 2010). The Portuguese material is somewhat expanded tectonically, and the details of proximal developments are not well enough preserved for an unequivocal identification.

The most prominent element occurring in the assemblage (Fig. 2a, e) is a robust *Expansograptus*. This form has a large sicula (3-4 mm in length, with long supradorsal and ventral-apertural free parts), thecae widely spaced (2TRD 2.5-3 mm) with flared apertures, two or three declined thecae in the proximal region with relatively low stipe expansion, spreading from 2 mm at the first theca of each stipe to a maximum width of about 3 mm. All these characters are very distinctive of the species *Expansograptus sparsus* (Hopkinson), already identified in this locality by Delgado (1908, 1910) and Perdigão (1967), which is a form apparently restricted to the Fennian stage of the British Arenig, perhaps crossing its upper boundary with the Llanvirn (Fortey and Owens, 1987; Zalasiewicz et al., 2009). *Expansograptus sparsus* differs from other robust species with long prominent sicula coming from older beds, like *E. protobalticus* (Monsen), by its higher thecal inclination, lower thecal content and shorter declined part (see Maletz, 1996a). *Expansograptus praenuntius* (Törnquist) also superficially recalls the Portuguese form, especially by the high supradorsal part of the sicula exposed; but their stipes are thinner (< 3 mm in width), the thecal density higher (2TRD= 2.2 mm), the stipe expansion different, and the sicula shorter (2.7 mm): see Rushton in Zalasiewicz and Rushton (2008). The rhabdosome of *D. sparsus* Hopk. figured by Perdigão (1967, pl. 1, fig. 2: a retouched photograph) is clearly conspecific with our specimen and has a very similar aspect, but it is not its actual counterpart, since it has longer proximal stipes.

A second species with a shorter sicula also has two robust stipes that widens quickly in the first ten thecae from 2.5 mm to a maximum of 3 mm. The rhabdosome is slightly reflexed (Fig. 2c), with an origin of th ¹¹ high on the sicula, showing a certain degree of isograptid symmetry in obverse view. This form somewhat resembles *Expansograptus hirundo* (Salter) by the general aspect of rhabdosome, including the slight reclination of the proximal part (Rushton, 1985; Fortey and Owens, 1987). However, the Portuguese specimen has shorter proximal thecae, and the remaining thecae are less inclined than in the cosmopolitan Arenigian form. Several older species with a proximal part slightly reflexed, like *E. similis* (Hall), *E.*

constrictus (Hall), *E. grandis* (Monsen) or *E. suecicus* (Tullberg) have different dimensions and present a longer ventral apertural wall of the sicula (Williams and Stevens, 1988; Maletz, 1996, 1997). As the Portuguese material is limited to a single flattened specimen, we identified it provisionally as *Expansograptus* sp. A.

A third graptolite species recognized in the assemblage is a thinner horizontal didymograptid, with a relatively thin proximal end and few-inclined thecae, of very low thecal density (2TRD= 3 mm). Its general shape (Fig. 2b, f) is reminiscent of *Didymograptus* (s.l.) *nitidus* (Hall) and of some other species variously included in the genera *Expansograptus*, *Didymograptellus* or *Xiphograptus*. The poor preservation of our material, which is strongly flattened and slightly deformed, prevents an accurate identification, and the two Portuguese specimens are here regarded as *Expansograptus*? sp. B.

Finally, a fourth graptolite species from the Mestre André quarry is represented by the wide-stiped horizontal didymograptid described by Perdigão (1967, pl. 1, fig. 4) as *Didymograptus hirundo*. The retouched photograph provided by this author makes the presence of this species at the locality very probable, with the specimen showing a relatively short sicula, broad stipes (3.5 mm in width), and thecae highly inclined (aprox. 60° in the free ventral part). However, not having been possible to examine the original specimen, its provisional identification stands as *Expansograptus* cf. *hirundo* (Salter). According to Zalasiewicz et al. (2009) and previous works, *E. hirundo* is a long-ranging species, widespread (at least) from the earliest Dapingian to the early Darriwilian. It was the nominal species of a British graptolite zone, formerly representative of the late Fennian regional stage, which has recently been renamed as *Aulograptus cucullus* Zone, to take into account the fact that *E. hirundo* originates much earlier, in the *Isograptus victoriae* Zone.

PALYNOSTRATIGRAPHY

Preliminary palynostratigraphic research was established in several samples from the N. Delgado collection of the Geological Museum. Standard palynological laboratory procedures were employed in the extraction and concentration of the palynomorphs from the host sediments (Wood et al., 1996). The slides were examined with transmitted light, using a BX40 Olympus microscope equipped with an Olympus C5050 digital camera. All samples, residues and slides are stored in the LNEG-LGM (Geological Survey of Portugal) at S. Mamede Infesta, Portugal.

The acritarchs are abundant and reasonably well preserved in the studied samples. The assemblage includes the forms *Acanthodiacrodium costatum* Burmann (Pl. 1, fig. 1), *A.?* *dillatum* Molyneux (Pl. 1, fig. 5), *Acanthodiacrodium* spp., *Arbusculidium* cf. *filamentosum* (Vavrdová) (Pl. 1, fig. 4), *Coryphidium* cf. *bohemicum* Vavrdová (Pl. 1, fig. 2), *Cymatiogalea messaoudensis* Jardiné et al. (Pl. 1, fig. 14), *Cymatiogalea* sp. (Pl. 1, fig. 11), *Pachisphaeridium* sp., *Peteinosphaeridium trifurcatum* Eisenack (Pl. 1, fig. 10), *Polygonium* sp., *Michrystidium* sp. (Pl. 1, fig. 7), *Stelliferidium stelligerum* Deunff et al. (Pl. 1, fig. 6), *Stelliferidium* sp., *Striatotheca principalis parva* Burmann (Pl. 1, fig. 8), *S. rugosa* Tongiorgi et al., *S. rarirrugulata* Eisenack et al. (Pl. 1, fig. 9), *Vavrdovella areniga* Loeblich and Tappan (Pl. 1, fig. 3), *Verhachium trispinosum* (Eisenack) (Pl. 1, fig. 12) and *V. lairdii* (Deflandre) (Pl. 1, fig. 13). Also common in the studied assemblage are the cryptospores *Virgastaporites rudi* Combaz (Pl. 1, fig. 15) and *Gneudnaspora divellomedia* (Tchibrikova) (Pl. 1, fig. 16).

The palynological assemblage is mainly composed of long ranging acritarch species recorded in a wide rank of Early Ordovician ages. Some forms such as *Arbusculidium filamentosum*, *Coryphidium* and

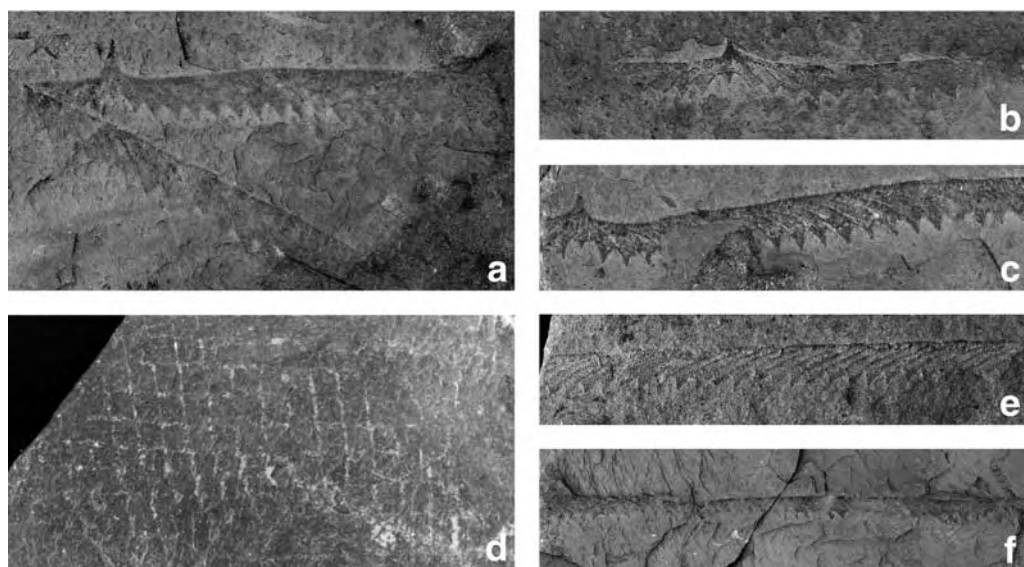


Figure 2. Some fossils from the “*Phyllodocites* shales” belonging to the Delgado collection, LNEG Geological Museum of Lisbon. a, *Expansograptus sparsus* (Hopkinson), proximal part with very large sicula (figured in Delgado, 1910, pl. 27, fig. 2), coll. no. 6001, x 2.2; b, *Expansograptus?* sp. B (figured in Delgado, 1910, pl. 27, fig. 6), coll. no. 6005, x 2.2; c, *Expansograptus* sp. A (figured in Delgado, 1910, pl. 27, fig. 4), coll. no. 6003, x 2.2; d, articulated spicules of a hexactinellid sponge showing a subquadrate pattern, coll. no. IGM-10862, x 1.3; e, *Expansograptus sparsus* (Hopkinson), fragment of stipe (figured in Delgado, 1910, pl. 27, fig. 1), coll. no. 6000, x 1.7; f, *Expansograptus?* sp. B, proximal fragment with broken sicula (figured in Delgado, 1910, pl. 27, fig. 3), coll. no. 6002, x 1.1.

Striatotheca have their FADs in the *messauoudensis-trifidum* assemblage from late Tremadocian-early Floian strata (Servais and Molyneux, 1997; Servais and Mette, 2000; Molyneux et al., 2007), but some others are more typical of “*Arenig sensu lato*” assemblages (i.e. Floian to lower Darriwilian) that may continue into younger strata, such as *Cymatiogalea messauoudensis*, *Peteinosphaeridium* sp., *Vavrdovella areniga*, *Veryachium trispinosum* and *V. lairdii*.

In consequence this acritarch assemblage, although incomplete, indicates a palynostratigraphic “*Arenig sensu lato*” (i.e., Floian to early Darriwilian) age, which agrees with the more precise biostratigraphic data provided by the graptolites.

SPONGES

Among the samples of the Delgado collection from the lower part of the Barrancos Formation, we have recognized two sponge fragments, not referred to in his work, coming from a locality placed 250 m north of Monte do Pombal (32 km northwest of Barrancos). They represent the only Ordovician sponges so far recorded in Portugal, and were also mentioned by Piçarra and Rigby (1996) and Rigby et al. (1997), but are figured here for the first time. The best preserved specimen (Fig. 2d) is a flattened mesh of spicules showing a subquadrate pattern, of very similar aspect to the reticulate pentactine clusters described from the Klabava Formation of Bohemia (Mergl, 2008). However, the skeletal structure is not preserved in the

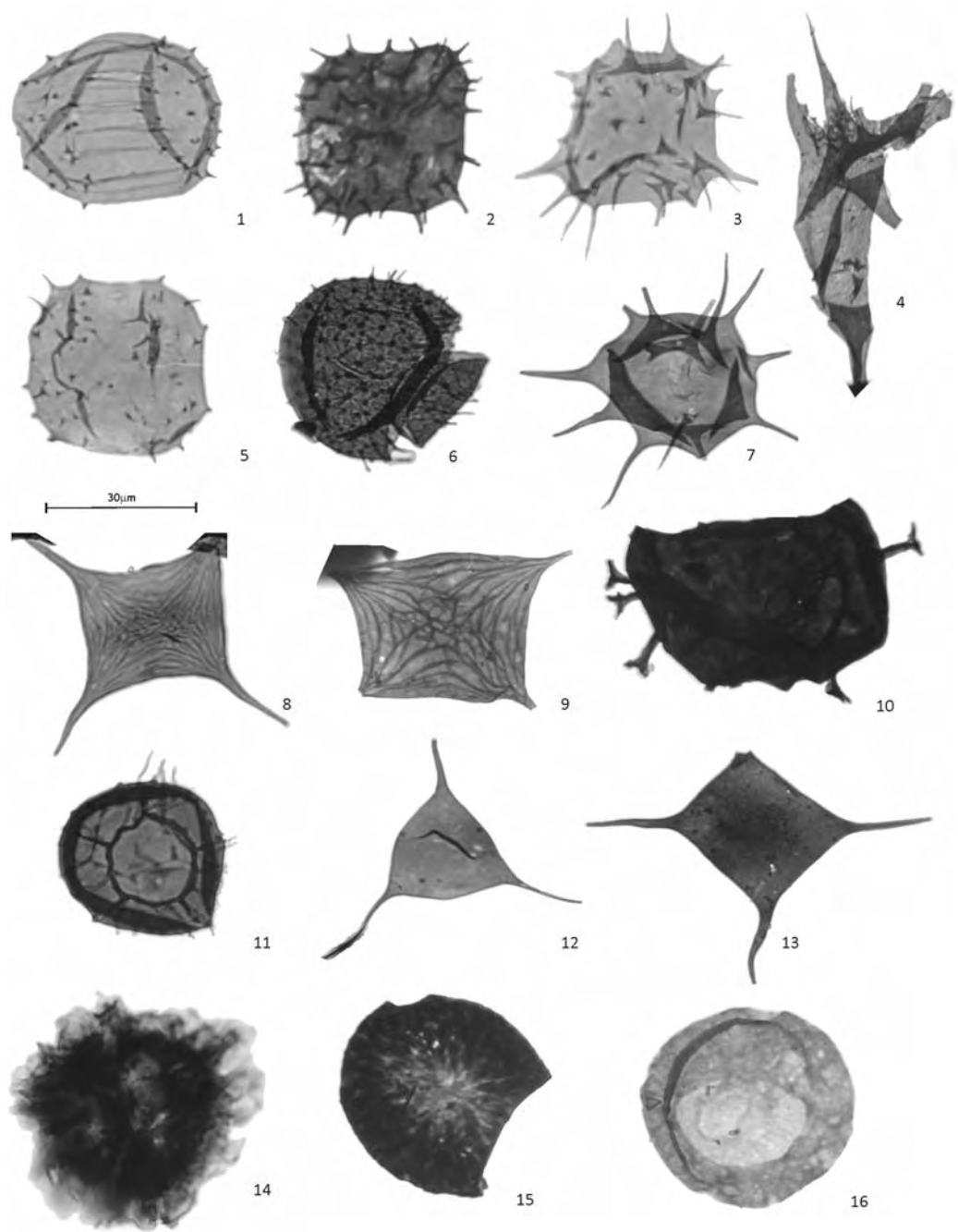
Portuguese specimens, perhaps indicating a dictyonine grade, but are taxonomically indeterminable due to the state of preservation.

ICHTHOLOGICAL RECORD

The “*Phyllodocites* shales” were originally named by Delgado (1908) according with the frequent record of trace fossils, partially interpreted by him as annelid impressions (*Phyllodocites*, *Nereites*, *Myrianites*), worm traces (*Arenicolites*), cnidarian body fossils (*Oldhamia*, *Lophoctenium*), molds of marine plants (*Palaeochorda*, *Palaeophycus*, *Chondrites*, *Bythotrephys*, *Alectorurus*, *Fraena*), and undetermined crustacean traces. In his most celebrated papers, Delgado (1908, 1910) equated the “*Phyllodocites*” assemblage of Barrancos with similar traces found in nearby outcrops of Devonian flysch (S. Domingos mine and Aljustrel area), causing both to be confused by later authors. Thus, Seilacher (1955) partially redrew the Ordovician and Devonian ichnofossils from Delgado’s (1910) photographs, as being part of a single assemblage representative of deep-marine environments in the Paleozoic of southern Portugal. The Ordovician and Devonian “*Nereites* shales” of the region were also studied by Perdigão (1961, 1967). In the complementary data to Uchmann’s chapter (2004), the database from the Ordovician of Barrancos appears restricted to *Nereites* isp., *Phyllodocites saportai*, *Myrianites tenuis*, *M. andrei* and *Crossopodia* isp. According to the actualized view of Seilacher (2007), in the “Barrancos Shales” there would be *Nereites*, *Dictyodora*, *Urohelminthoida*, *Lophoctenium*, *Paleodictyon*, *Zoophycos* and *Chondrites*. However, some of these are not documented in any previous paper by the author, nor in the illustrations from Delgado (1910).

A comprehensive review of the trace fossil association from the “*Phyllodocites* shales” was presented by Piçarra (2000), who updated the synonymies of former ichnotaxa and recognized the presence in the Mestre André quarry of the following traces: *Nereites jacksoni* Emmons, *Nereites* isp., *Phyllodocites saportai* Delgado, *Dictyodora? andrei* (Delgado), *Dictyodora? lorioli* (Delgado), *Dictyodora? bocagei* (Delgado), *Dictyodora* isp., *Lophoctenium geinitzi* Delgado, *Chondrites* isp., *Glockerichnus? isp.*, *Oldhamia* n. isp., *Zoophycos* isp., *Palaeophycus* cf. *striatus* Hall, *Palaeophycus* isp., *Gordia marina* Emmons, *Gordia* isp., *Phycodes? isp.*, *Cochlichnus? isp.*, aff. *Didymaulichnus* isp., *Megagraptos? isp.*, *Diplichnites* isp. and *Dimorphichnus? isp.* The type specimen of *Helicolithus delgadoi*, defined from the Delgado collection (Azpeitia Moros, 1933) was not revised by this author. A constructional model of *Dictyodora tenuis* (M’Coy) based in specimens from the Mestre André quarry was presented by Neto de Carvalho (2001), who synonymized with this ichnospecies some forms described by Delgado as *Myrianites andrei*, *M. lorioli* and *M. tenuis*. To the ichnotaxa list derived from the “*Phyllodocites* shales”, we must also add “*Oldhamia pinnata*”, mentioned apparently without any ichnotaxonomical formalization (Seilacher,

Plate 1. Acritarchs of the “*Phyllodocites* shales” of Barrancos, Portugal. Each specimen is referenced by a collection number, slide number and microscope coordinates (MC). 1, *Acanthodiacrodium costatum* Burmann, PMA3-1b, MC 1342-56; 2, *Coryphidium* cf. *bohemicum* Vavrdová, PMA3-3, MC 1125-115; 3, *Vavrdovella areniga* Loeblich & Tappan, PMA3-3a, MC 1245-55; 4, *Arbusculidium* cf. *filamentosum* Vavrdová, PMA3-3a, MC 1445-125; 5, *Acanthodiacrodium? dilatatum* Molyneux, PMA3-1b, MC 1224-89; 6, *Stelliferidium stelligerum* Deunff, Górka and Rauscher, PMA 3-1, MC 1228-35; 7, *Michystridium* sp., PMA3-1, MC 1265-128 (800x); 8, *Striatotheca principalis* parva Burmann, PMA3-1b, MC 1188-212; 9, *Striatotheca rarirugulata* Eisenack, Cramer and Díez, PMA3-3, MC 1425-145; 10, *Peteinosphaeridium trifurcatum* Eisenack, PMA3-3a, MC 1115-124; 11, *Cymatiogalea* sp., PMA3-1a, MC 1168-155; 12, *Veryhachium trispinosum* (Eisenack) Deunff, PMA3-1a, MC 1355-13513; 13, *Veryhachium lairdii* (Deflandre) Deunff, PMA3-1C, MC 1186-120; 14, *Cymatiogalea messaoudensis* Jardiné, Combaz, Magloire, Peniguel & Vachey, PMA3-1b, MC 1165-116; 15, *Virgastaporites rudi* Combaz, PMA3-1a, MC 1455-75; 16, *Gneudnaspora divellomedia* (Tchibríkova) Balme, PMA3-1, MC 1035-125.



1997, 2007; Seilacher et al., 2005), which was already illustrated by Delgado (1910, pl. 38, fig. 1) and is usually confused with "*Lophoctenium geinitzi*" owing to a plate mistake in his posthumous publication.

FINAL REMARKS

According with graptolite and palynological data, the age of the "*Phyllocytes* shales" of the Barrancos anticline can be mainly envisaged as early Darriwilian, equivalent to the Da1 stage slice of Bergström et al. (2009). This dating is close to the provided by another acritarch assemblage recorded from the Barrancos Formation in Spain, in a section located SSE of Cañaveral de León (Mette, 1989), belonging to the same Estremoz-Barrancos-Hinojales domain of the Ossa-Morena Zone.

Owing to the common record of trace fossils and sedimentary structures in the Mestre André quarry, the "*Phyllocytes* shales" have been linked to littoral (Delgado, 1908; Perdigão, 1967) or deep marine environments (Seilacher, 1955, 1974; Burke et al., 2005), with some authors even suggesting a distal turbiditic sequence (Neto de Carvalho, 2001). However, the dominant siltstones and shales do not show any sedimentological characteristic that can support their interpretation as turbidites. Instead, the high concentration of clastic muscovite in these rocks, and especially along bedding planes, seems more typical of shelf sediments. The rare extensiform didymograptids belong to a graptolite ecomorphotype widely distributed in the epipelagic biotope, but normally confined to shelf sediments (Cooper et al., 1991; Cooper and Sadler, 2010). The presence of cryptospores (terrestrial primitive contribution), as well as *Leiosphaeridia* spp. and common acritarchs (acanthomorphs, sphaeromorphs, netromorphs and polygonomorphs) may indicate proximity to an inner-shelf environment (Al-Ameri, 1983).

On the other hand, the trace fossil record of the "*Phyllocytes* shales" has been ascribed to the *Nereites* ichnofacies, typical of deep-sea environments (bathyal-abyssal according to Seilacher, 1967, 2007), with sporadic matgrounds representative of the *Oldhamia* ichnofacies. In our opinion, the assemblage would be alternatively explained as the opportunistic colonization of a dissiaerobic environment by marine benthos, developed in an outer-shelf setting, not necessarily in a deep environment. This oxygen-depleted bottom could be consequence of a water-mass stratification or paleotopographic differentiation of the platform by extensional tectonics, similar to the fact observed in the Middle Ordovician shales of central Portugal that bear some flysch-like trace fossils locally recorded in an inner shelf setting (Gutiérrez-Marco and Sá, 2006).

Thus, the paleoecological restriction observed in the "*Phyllocytes* shales", cannot be primarily related with its placement in a deep-turbiditic or abyssal zone; their lateral continuity over a relatively reduced area with regard to the remaining outcrops of the Barrancos Formation both in Portugal and Spain also conflicts with the interpretation of an oceanic-related paleoenvironment.

In any case, the "*Phyllocytes* shales" are not representative of typical inshore settings and its sedimentary circumstances fully concur with the general paleogeographic trend of the Ossa-Morena Zone. The Ordovician and Silurian sequences of this zone are characterized by more distal and deeper environments and faunas (cyclopygid trilobite biofacies) than the remaining successions coevally known for the same periods in the Iberian Massif, to which the Central Iberian Zone (the proximal part of the same platform) was tectonically juxtaposed as the result of Variscan transcurrent displacement along the Badajoz-Córdoba Shear Zone (Robardet et al., 1998; Robardet and Gutiérrez-Marco, 2004 and references therein).

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NEW INSIGHTS INTO THE STRATIGRAPHY AND STRUCTURE OF THE UPPER ORDOVICIAN ROCKS OF THE LA Cerdanya AREA (PYRENEES)

C. Puddu¹ and J.M. Casas²

¹ Department of Earth Sciences, University of Cagliari, Via Trentino 51, 09127 Cagliari, Italy. cpuddu@unica.it

² Departament de Geodinàmica i Geofísica-Institut de recerca GEOMODELS, Universitat de Barcelona, Martí i Franquès s/n, 08028 Barcelona, Spain. casas@ub.edu

Keywords: Middle Ordovician folding event, brachiopods, Upper Ordovician fractures, Upper Ordovician stratigraphy.

INTRODUCTION

It used to be assumed that deformation mesostructures recognized in the Paleozoic pre-Variscan rocks of the Pyrenees mainly derive from Variscan deformation. Recently, the presence of pre-Variscan folds, Mid Ordovician in age, has been documented in the southern slope of the Canigó massif (Casas, 2010). This work seeks to provide new insight into the structure of the Upper Ordovician rocks in the La Cerdanya area. Moreover, new data on the stratigraphy of the Upper Ordovician succession is also provided. Data were collected north of Bellver de Cerdanya, between the towns of Cortás, Eller, Ordén and Talltendre (Fig. 1) during detailed geological mapping (1/5.000) and structural analysis.

GEOLOGICAL SETTING

In this area, the upper part of the pre-Upper Ordovician rocks crops out extensively. It is a rather monotonous succession, composed of an unfossiliferous succession of rhythmic alternation of sandstones, siltstones and argillites. Layers vary in thickness from 1 mm to several cm and range in colour from grey to light green or light brown. Owing to its monotonous character, it is not easy to determine its lower limit and thickness, although a thickness of about 1500 m has been proposed for this upper part of the succession. This succession is classically known as Cambro-Ordovician, and corresponds to the "schistes de Jujols" established by Cavet (1957). Recent acritarch data (Casas and Palacios, pers. comm.) indicate that the uppermost part of this succession has a Late Cambrian (Furongian)-Early Ordovician (Tremadocian) age. The well dated Upper Ordovician succession (Cavet, 1957; Hartevelt, 1970) lies unconformably over the former unit (Santanach, 1972; García-Sansegundo et al., 2004; Casas and Fernández, 2007). The Upper Ordovician rocks constitute a fining upwards sequence similar to that described by Hartevelt (1970) in the Segre valley, in which this author defined five main siliciclastic stratigraphic formations. The Rabassa Conglomerate Formation, which constitutes the lowest part of the

succession, is made up of red-purple, largely unfossiliferous, conglomerates and microconglomerates that range in thickness from a few to 200 meters. Hartevelt (1970) attributed the Rabassa conglomerates to the Caradoc. The Rabassa conglomerates are overlain by the Cava Formation that varies in thickness from 100 to 800 m. Microconglomerates and feldspatic sandstones predominate in the lower part, followed upwards by shales, siltstones and fine grained sandstones, typically green or purple in colour. Brachiopods and bryozoans are locally abundant. Gil Peña et al. (2004) attributed a late Caradoc–early Asgill age to this formation, which is Mid Late Ordovician. The Estana Formation lies above the Cava Formation and consists of limestones and marly limestones up to 10 m in thickness. This formation constitutes a good stratigraphic key level, with abundant fossils, conodonts and brachiopods, yielding a mid Ashgill age (Gil Peña et al., 2004). The Ansovell Formation overlies the Estana limestone and is made up of dark shales and siltstones with minor interbedded quartzite layers in the uppermost part. The Bar Quartzite Formation, located at the top of the Upper Ordovician succession, consists of a 5 to 10 m thick quartzite layer that overlies the Ansovell Formation. An Ashgill age was proposed for the Ansovell and Bar formations by Hartevelt (1970) although Gil-Peña et al. (2004) suggest that the Ordovician-Silurian boundary can be located within the Bar quartzite.

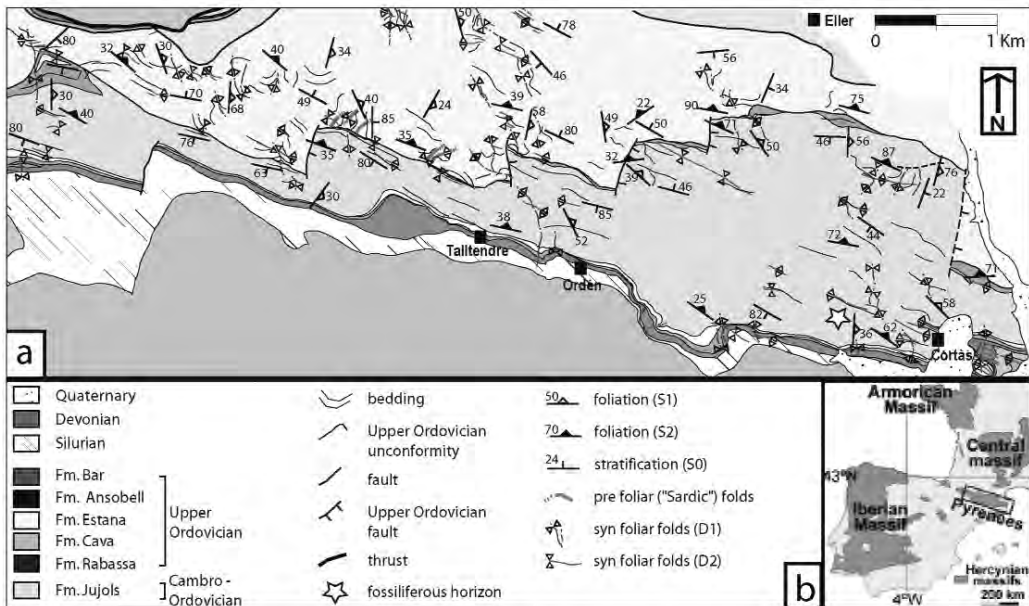


Figure 1. Geological map of the study area (a) with the location of the study area (b). The legend shows the main Variscan and pre-Variscan structures and the location of the fossiliferous horizon (W of Cortás) where the brachiopods were collected.

STRATIGRAPHY

In the stratigraphic section made near Ordén, the Upper Ordovician sequence exhibits a thickness of about 350 meters (Fig. 2). In this section the Cava Formation only presents three of the four members recognized by Hartevelt (1970): the basal member made of greenish to purple greywackes, sandstones, microcon-

glomerates, siltstones and slates, with rock fragments in its lower part ("a" member); the red and greenish silty slates ("b" member), and the siltstone ("c" member), which contains in its upper part some fossiliferous level with brachiopods, bryozoans, cystoids and rugose corals. It should be noted that, in this section, three brachiopod genera (*Porambonites* sp., *Eoanastrophia* sp., and *Dolerorthis* sp.) were collected, which have not yet been described in the Cava Formation (Fig. 3).

The new brachiopod fauna comes from the uppermost part of the "c" member above the "coquina" horizon described by Hartevelt (1970). However, the state of preservation of the fossils only allowed a generic assignment, which represents an intermediate fauna between the late Caradoc - early Ashgill brachiopods collected in the "coquina" horizon located in the upper part of this member (*Svobodaina havliceki*, *Rostricellula* sp., *Rafinesquina* sp.; Gil Peña et al., 2004) and the mid Ashgill brachiopods of the Estana Fm. (*Dolerorthis* sp., *Eoanastrophia pentamera*, *Iberomena sardoa*, *Leangella anaclyta*, *Longvillia mediterranea*, *Nicolella actoniae*, *Porambonites* (*Porambonites*) *magnus*, *Ptychopleurella villasi*; Gil Peña et al., 2004).

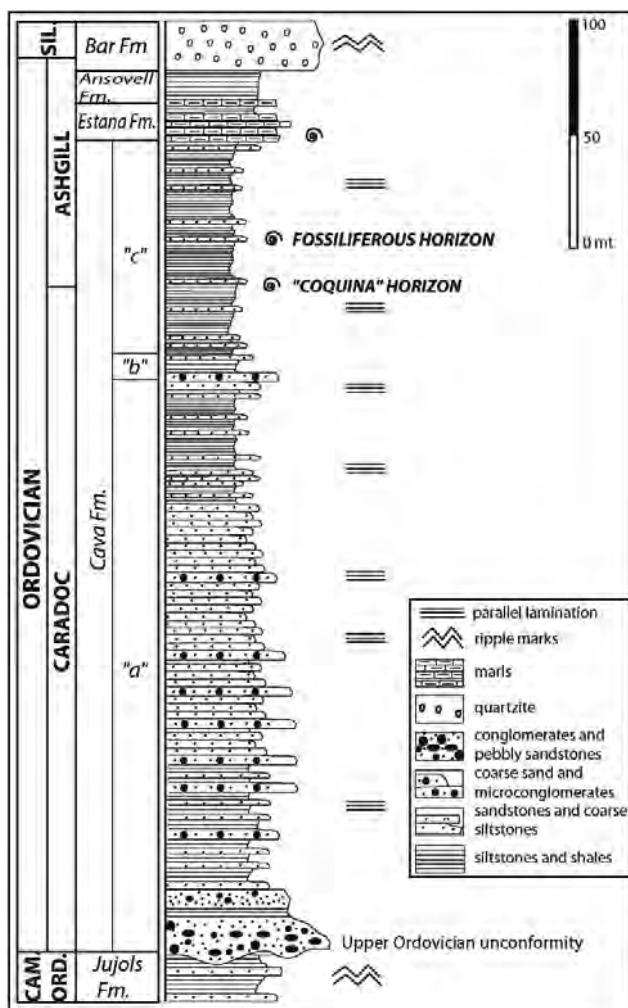


Figure 2. Stratigraphic section made near Ordén: the fossiliferous horizon with *Porambonites* sp., *Eoanastrophia* sp., and *Dolerorthis* sp. is located in the Cava Fm. between the "coquina" horizon of Hartevelt (1970) and the fossiliferous marls of the Estana Fm.

STRUCTURE

Different structures can be recognized in the study area: the Upper Ordovician unconformity, the Upper Ordovician normal faults, and three systems of folds: two of them of Variscan age and one of pre-Variscan age (Fig. 1a).

The Upper Ordovician unconformity, that separates the Upper Ordovician sediments from the underlying Cambro-Ordovician ones, can be identified from detailed mapping in several areas. The unconformity has a NW-SE trend and cuts the bedding of the pre-unconformity deposits at different angles ranging from a few to 90°.

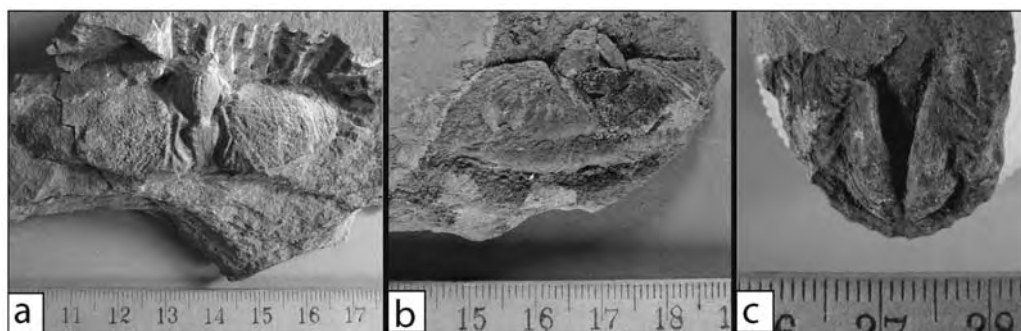


Figure 3. Brachiopods collected in the upper part of the “c” member of the Cava Fm.: a) *Porambonites* sp. (internal mould of ventral valve), b) *Dolerorthis* sp. (internal mould of ventral valve), c) *Eoanastrophia* sp. (internal mould of dorsal valve).

Several normal faults affect the Upper Ordovician succession, the Cambro-Ordovician sediments and the unconformity. The faults are steep and currently exhibit a broadly N-S to NNE-SSW cartographic trace. In most cases their hanging wall is the eastern block despite the presence of some antithetic faults. Displacement of some of these faults diminishes progressively upwards of the series and peters out in the upper part of the Upper Ordovician rocks, in the sediments of the Cava Formation, indicating that the faults became inactive during the Late Ordovician before deposition of the Ashgill metasediments (Fig. 1a).

Two systems of cleavage-related Variscan folds affect the pre- and the post-unconformity sediments, one with a N-S trend and the other with an E-W to NW-SE trend. Moreover, the pre-unconformity deposits are affected by another folding episode. This episode gave rise to metric to hectometric sized folds without foliation or metamorphism associated and oriented N-S to NE-SW. These folds were not recognized in the Upper Ordovician sediments and are sealed by the Upper Ordovician unconformity.

DISCUSSION

The brachiopods collected from the upper part of the Cava Formation represent an Ashgill fauna, intermediate between the late Caradoc–early Ashgill brachiopods of the “coquina” from the Cava Fm. and the mid Ashgill brachiopods of the Estana Fm. described by Hartevelt (1970) and revised by Gil Peña et al. (2004). This fauna marks a smooth transition between the *Svobodaina* fauna and the *Nicolella* one found in the Cava Fm. and in the Estana Fm. respectively, and represent a fauna similar to the one described in the Montagne Noire (France).

The similarities should be noted between the study area and the Iglesias and Sarrabus regions in the south of Sardinia, classic zones where an Upper Ordovician (“Sardic”) unconformity has been described. In Sardinia, a regional stratigraphic and angular unconformity separates the Cambro-Ordovician sequence from the underlying Upper Ordovician ones, as the stratigraphic gap marked by the “Sardic unconformity” is located between the Arenig?, dated by the youngest fossiliferous deposit under the unconformity (Pillola et al., 2007), and the late Caradoc, which is the age of the oldest fossiliferous horizon of the post-unconformity beds (Hammann, 1992; Leone et al., 2002). In Sardinia the pre-unconformity sequence is deformed by different E-W structures sealed by the unconformity and related to the “Sardic Phase” (Stille, 1939), in the form of metric to hectometric sized folds without cleavage, thrusts and thrust faults (Pasci et al., 2008).

The Upper Ordovician unconformity of the Pyrenees may be interpreted as equivalent to the “Sardic unconformity” and the pre-Variscan folds described in the Pyrenees may be equivalent to the “Sardic” folds of Middle Ordovician age. These folds were responsible for the deformation, uplift and erosion of the Cambro-Ordovician sediments and for the Upper Ordovician unconformity. Thus, the Pyrenees closely resemble the most external part of the Sardinian fragment of the Variscan orogen and exhibit marked differences with the rest of the Iberian Massif, where evidence of Ordovician deformation is limited.

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FINAL DESTINATION, FIRST DISCOVERED: THE TALE OF *OANDUPORELLA* HINTS, 1975

C.M.Ø. Rasmussen^{1,2}

¹ Center for Macroecology, Evolution and Climate, Natural History Museum of Denmark, University of Copenhagen,
Øster Voldgade 5–7, DK-1350 Copenhagen K, Denmark.

² Nordic Center for Earth Evolution (NordCEE), Natural History Museum of Denmark, University of Copenhagen,
Øster Voldgade 5–7, DK-1350 Copenhagen K, Denmark.
Christian@snm.ku.dk

Keywords: Upper Ordovician, brachiopod, *Oanduporella*, dispersal routes, palaeo-currents, evolutionary phylogeny.

INTRODUCTION

Oanduporella Hints, 1975 is a medium-sized, ventri-biconvex enteletoid brachiopod genus that originally was thought to have been a Late Ordovician Baltic endemic (Hints, 1975). However, since Havlíček and Branisa (1980) described the genus from the Lower Sandbian of the Bolivian Cordillera Oriental, several new occurrences have appeared in the literature. Today the genus is known to occur from the high southerly latitudes of Gondwana to the peri-Siberian Farewell Terrane in the northern hemisphere. Ironically, its youngest occurrences, are in fact those from Baltica (Hints, 1975; Paškevičius, 1994) and thus, the once assumed centre of origin, now appears to have been the final destination of the genus.

The Sandbian–early Katian interval was a crucial period in rhynchonelliformean brachiopod evolution, as it marks the pinnacle of the Great Ordovician Biodiversification Event. A culmination of a radiation event that started some 50 myr. earlier in the late Cambrian (Harper, 2006). By the Sandbian, the rhynchonelliformean brachiopods dominated most of the marine environment (Patzkowsky, 1995). Probably as a result of high plate tectonic activity, global sea level was possibly at its Phanerozoic peak with estimates of up to 400 metres above present day sea level (Hallam, 1992; Nielsen, 2004). In addition, the rapid drifting of continents and terranes further provided multiple mid-ocean refugia for benthos such as the brachiopods. A combination that together with the high global sea level, and the fact more shallow-water provinces entered warmer equatorial regions, led to a dramatic increase in brachiopod diversity. These Late Ordovician sea-level fluctuations were crucial for the distribution of substrate dependent taxa, like *Oanduporella*.

The current study surveys the global occurrences of this distinctive genus and analyzes morphological adaptations within the different species that are assigned to *Oanduporella*. These are discussed and placed within a phylogenetic evolution that, in turn, is used to elucidate how this enigmatic genus dispersed up through the Sandbian-lower Katian interval. This is achieved by applying previously reconstructed palaeo-oceanic current models to the most up-to-date palaeogeographic reconstructions and comparing this to the phylogenetic evolution of *Oanduporella*.

THE DISTRIBUTION OF *OANDUPORELLA*

At least three species are assigned to the genus. The type species, *O. reticulata* Hints, is from the Lower Katian (Oandu Regional Stage) Hirmuse Formation of Estonia. The type species is, in addition, known from coeval beds of Lithuania (Paškevičius 1994). Moreover the same species has been reported from older strata in the Midland Valley Terrane; Candela and Harper (2010) reported *O. cf. reticulata* from the Kirkcolm Formation of Wallace's Cast, southern Scotland, in beds assigned to the lower Katian (early Cheney, lower *clingani* Zone) and Candela (2003) reported *O. cf. reticulata* from Members II and III of the Bardahessiagh Formation in Pomeroy, Northern Ireland. These members are latest Sandbian–Early Katian in age (Candela, 2002). From Avalonia (eastern Ireland), *O. cf. reticulata* was reported from even older beds within the upper Sandbian (Soudleyan) by Harper et al. (1985) and Parkes (1994).

Accordingly, within the Late Sandbian–Early Katian interval, the type species seem to have occurred on both sides of the Iapetus Ocean (Kilbucho and Pomeroy being Laurentian inliers of the Midland Valley Terrane), as well as in mid-oceanic settings of the microcontinent Avalonia.

The oldest occurrences of the genus are known from Gondwana, peri-Siberia and possibly Laurentia. *O. alamensis* was described by Benedetto (1995), from the black shales of the upper Sandbian Las Plantas Formation of Northern Precordillera, Argentina. Rasmussen et al. (in press) described *O. kuskokwinensis* from allochthonous carbonate turbidites deposited in slope settings in the White Mountain area of west-central Alaska. Based on the association of conodonts and brachiopods the fauna was referred to the Early Sandbian (Rasmussen et al., 2011, see also Rigby et al. (1988) and Potter and Boucot (1992). This area is part of the Farewell Terrane, which is believed to have been positioned relatively close to Siberia in the Late Ordovician. Further, Potter and Boucot (1992) reported the genus from the Jones Ridge area, close to the Alaskan–Yukon border, in beds that are probably contemporaneous with the Farewell material. This is based on faunal affinities between the two regions, as well as faunas from the Eastern Klamath Terrane (Potter, 1990). The Jones Ridge material, however, has never been described, nor illustrated. Finally, Havlíček and Branisa (1980) reported *Oanduporella* sp. from the lower Sandbian Anzaldo Formation of the Bolivian Cordillera Oriental. This, the oldest occurrence of the genus, is relatively well stratigraphically constrained as it is found associated with the trilobite *Huemacaspis bistrami* that can be assigned to the Lower Sandbian (Suárez-Soruco, 1992; Waisfeld and Henry, 2003).

PHYLOGENETIC EVOLUTION – ARE THEY ALL REALLY *OANDUPORELLA*?

The question remains whether all of these widely distributed occurrences truly belongs to *Oanduporella sensu stricto*, i.e. does the genus have a mono- or polyphyletic origin? Therefore, I will briefly discuss some apparent evolutionary adaptations seen in the different species within the genus. It should be noted, however, that it is difficult to assess solely from available published images. Not only due to the different quality of the images, but also, as both the material from Avalonia and that from Wallace Cast, appear to be juvenile specimens.

It appears to be a general trend regarding the exterior of the *Oanduporella*-shells, that the very characteristic pitted microstructure changes from being both on the costae and lower order costellae, as well as in the interspaces between them in the Gondwanan and Avalonian material, to being positioned solely in the interspaces on the peri-Siberian and Baltic material of the genus (see Plate 1). However, Harper et al. (1985) only mentioned that the microstructure occurs in the interspaces, although some

microstructure is apparently developed both on the costae, as well as in the interspaces (figs. 28, 31 and 35 in Harper et al. (1985)).

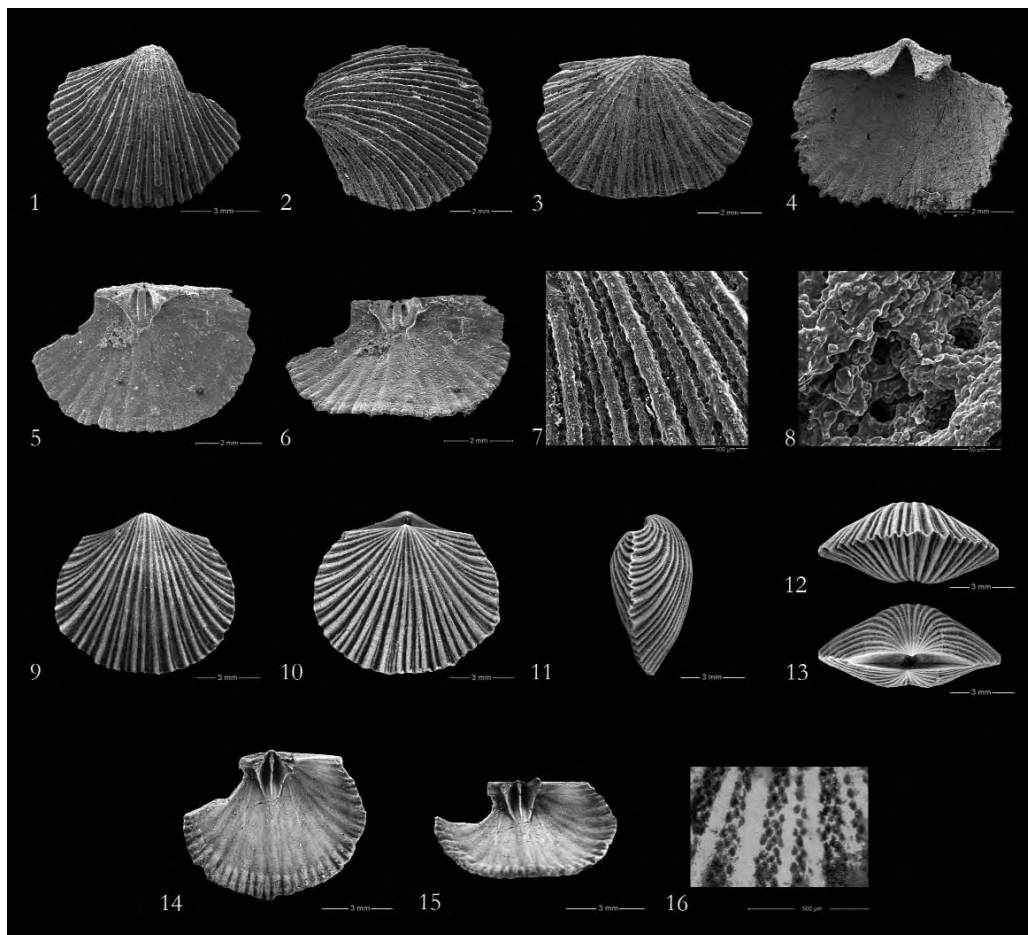


Plate 1. Different species of *Oanduporella* showing the intercostellate pitted microstructure and the differences in the dorsal cardinalia (compare with Figure 1). Figs 1–8: *O. kuskokwimensis* Rasmussen, Harper and Blodgett from the Lower Sandbian of west-central Alaska. 1, 2: Perpendicular, oblique views of ventral exterior. 3: Exterior view of dorsal valve (holotype). 4: Perpendicular view of ventral interior. 5, 6: Perpendicular and anterior views of dorsal interior (holotype). 7, 8: Close-ups on intercostellate pitted microstructure. Images are taken from Rasmussen et al. (in press) and the material is repositied at the Natural History Museum of Denmark. Figures 9–16: *O. reticulata* Hints (the type species), from the Lower Katian Hirmuse Formation. 9–13, exterior views of conjoined specimen (Br 4182): Ventral, dorsal, lateral, anterior and posterior views. 14, 15: Perpendicular and anterior views of dorsal interior (Br 4185). 16: Close-up on the intercostellate pitted microstructure (Br 4183). Images are taken from Hints (1975)/ www.geokogud.info with permission.

Interiorly, another interesting difference is the apparent shortening of the sub-parallel brachioophores and shaft-like cardinal process. Deduced from Havlíček and Branisa's Plate VI, fig. 19, it appears that in the Bolivian species the brachioophores project much farther anteriorly, than the cardinal process (see Fig. 1).

The same is seen in the species from the Northern Precordillera where the cardinal process is very short with brachioophores projecting farther anteriorly. In the Avalonian species the brachioophores and the

cardinal process are almost equally long, although much shorter than seen in the species from the Farewell Terrane and Baltica. In the Alaskan species (see Fig. 1 and Pl. 1, fig. 5, 6), the brachiophores and the cardinal process terminate at the same position anteriorly on the notothyrial platform. This is also seen in the Baltic type species (Fig. 1 and Pl. 1, fig. 14, 15). But, whereas the cardinal process is more rod-like and the brachiophores thicker and less elevated in the Alaskan species, the Baltic type species has a thin blade-like cardinal process with relatively thin, more elevated brachiophores. This difference may however be a taphonomic bias, as the Alaskan material is silicified.

Thus, again there is a difference between the specimens from the Northern Precordillera and Avalonia compared to those of peri-Siberia and Baltica (the material from the Midland Valley Terrane does not include any dorsal interiors; Y. Candela, pers. com. 2011). Figure 1 illustrates these transitions in the dorsal cardinalia of *Oanduporella* through the Sandbian–Lower Katian.

Although the species from Avalonia and in the Midland Valley Terrane differ from the type they are retained within *Oanduporella*. The same is the case with most of the other species (not considering the un-described one from the Jones Ridge), although *O. alamenensis* is more questionable. At least one of the illustrated dorsal interiors (Pl. 1, figs. 25, 26 in Benedetto, 1995) does not look like a draboviid cardinalia. Rather, it resembles a dalmanellid such as *Dalmanella* or *Paucicrura*. However, the other illustrations are probably a species of *Oanduporella*, although its ornamentation differentiates it from the other species of the genus. Most notably compared to the Baltic type species, as well as *O. kuskokwimensis*, both possess much thicker primary costae and in general the lower-order costellae are more widely spaced (and less numerous), compared to *O. alamenensis*. To me, the microstructure seen in *O. alamenensis* is more similar to that seen in the Avalonian material, whereas a true intercostellate microstructure is seen only in the Alaskan *O. kuskokwimensis* and in the Baltic occurrences of the type species (see Plate 1).

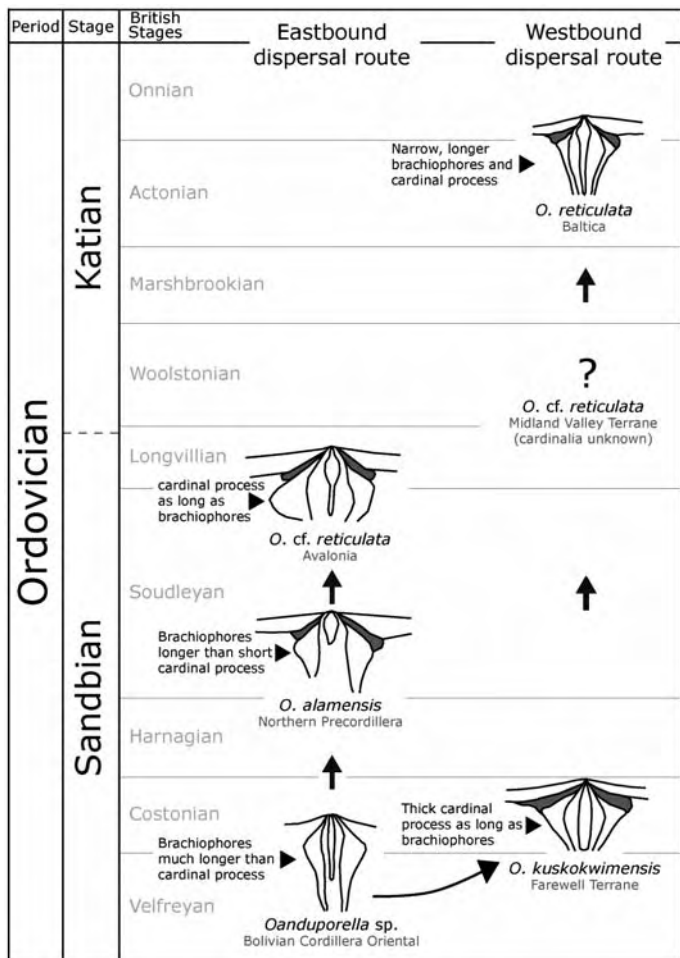


Figure 1. Morphological adaptations through time in *Oanduporella*. Arrows depict morphological adaptations through the interval, as well as the suggested dispersal routes discussed in the text.

Thus, it is argued here that the origin of *Oanduporella* is not polyphyletic. Instead, the above two mentioned morphological adaptations may have been achieved through two different dispersal routes without any mixing. Below, these phylogenetic implications have been applied to known palaeo-current models for the late Ordovician.

A POSSIBLE DISPERSAL PATTERN

As shown in Figures 1 and 2, the oldest known occurrences of *Oanduporella* are from the Bolivian Cordillera Oriental and the Farewell Terrane (and possibly the Jones Ridge area in cratonic Laurentia). Stratigraphically these earliest global occurrences are an interesting puzzle that only becomes more challenging when the above analyzed phylogenetic evolution is incorporated.

Following the various published palaeo-ocean current models for the Ordovician, i.e. Wilde (1991), Christiansen and Stouge (1999), Poussart et al. (1999), Herrmann et al. (2004), it seems likely that *Oanduporella* followed the cool-water south Panthalassic convergence gyre of Herrmann et al. (2004) towards the Northern Precordillera, where *O. alamensis* is found in upper Sandbian beds. From then on it likely followed the cold-water current along western Avalonia, where it also settled in the late Sandbian. It may have continued northwards towards the equator along the western margin of Baltica. However, this seems less likely as earlier occurrences should then have been described from the Baltic margin. In addition, as shown above, the Baltic type species is different from the Avalonian species.

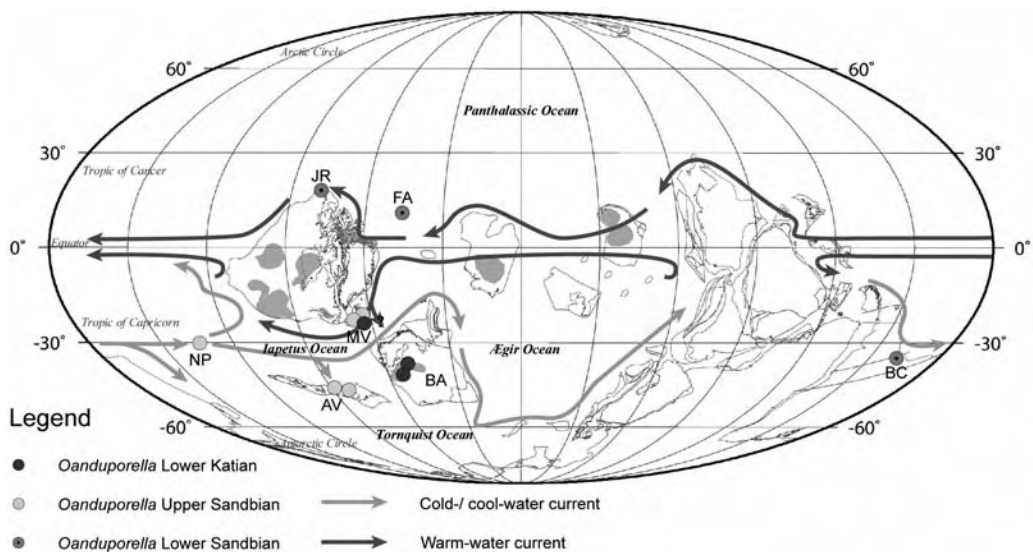


Figure 2. Mollweide palaeogeographic reconstruction for the Late Ordovician (Sandbian) showing the known occurrences of *Oanduporella*. The shown palaeocurrents are based on the literature (see text for references). Only those relevant for this study are shown. Derivations: AV – Avalonia, BA – Baltica, BC – Bolivian Cordillera Oriental, FA – Farewell Terrane, MV – Midland valley Terrane, NP – Northern Precordillera and JR – Jones Ridge, see legend for further explanation of symbols. Mollweide projection provided by Trond Torsvik, Norwegian Geological Survey.

Thus, a far longer path is required to explain how *Oanduporella* dispersed from its possible Bolivian centre of origin to reach its peri-Siberian and Laurentian occurrences. Either it dispersed along the Gondwanan margin towards the equator, or, it was carried from its Bolivian origin by the cool-water gyre, towards the equator and then dispersed along the equator towards Laurentia. If it followed the Gondwanan margin round the Tethys and Ægir oceans, it would have dispersed within its favorite substrate. However, large parts of these regions are only rudimentarily sampled, possibly explaining why the genus has not yet been described from these regions. The latter route, which would have taken it on a rather northerly path, indicates that *Oanduporella* should occur in deep water successions of Australia, South China and Siberia. South China is likely the best sampled of these regions, but was dominated by extensive carbonate deposition. Thus, arguably a substrate dependent genus like *Oanduporella* may be difficult to find there.

When the genus finally arrived on to a continental platform, it was probably because it entered the platform in a transgressive pulse succeeding a huge regression that had increased the siliciclastic content on the platform (Hints, 1998; Ainsaar and Meidla, 2001). This large regression may have aided migration of the genus from peri-Laurentia across the Iapetus Ocean and on to the Baltic platform.

CONCLUSIONS

Morphological adaptations in *Oanduporella* suggest that the genus expanded to its known extent via two main dispersal pulses; one eastbound towards Avalonia and one westbound towards peri-Siberia, Laurentia and Baltica. Thus, even though Avalonia and Baltica were relatively close to each other during the Sandbian–Lower Katian interval, this study suggest that faunal exchange may still have been limited, at least for taxa that were heavily substrate dependent.

Oanduporella was confined to relatively deep water environments, usually near the shelf-margin break or slope. The preferred biofacies of *Oanduporella* increased its chances for migration, especially as its main dispersal took place in the early Sandbian, during one of the peak eustatic transgressions in the Ordovician.

As speculated by Patzkowsky (1995) increased cosmopolitanism over time may simply be a result of high global sea level and preservation of more pandemic faunas onto the cratons. This could be the case with the initial migration and dispersal of *Oanduporella*. Thus, it is probably not a result of increased plate tectonics and the changing geography. Rather, it was favored by the very high global sea level, as well as the configuration of palaeo-currents. Ultimately, however, the dispersal of the genus seems to be controlled by its substrate dependency. These conclusions add faunal support to recent palaeo-ocean current modelling for the Late Ordovician and further gives a rare glimpse as to how deep-water sessile taxa dispersed during the Late Ordovician.

Acknowledgements

I would like to thank Dave Harper and Arne T. Nielsen, both the Natural History of Denmark, for enlightening discussions on this matter. Yves Candela, Edinburgh, is thanked for going through his old material on *Oanduporella* and Olle Hints, Tallinn, is thanked for allowing the use of pictures from the online Estonian database on geocollections (www.geokogud.info). Finally, I would like to acknowledge the Danish National Research Foundation for support to the Center for Macroecology, Evolution and Climate, as well as support to the Nordic Center for Earth Evolution (NordCEE).

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AN UNUSUAL MID-ORDOVICIAN ISLAND ENVIRONMENT ON THE WESTERN EDGE OF BALTICA: NEW PALAEOECOLOGICAL AND PALAEOBIOGEOGRAPHICAL DATA FROM HARDANGERVIDDA, SOUTHERN NORWAY

J.A. Rasmussen, A.T. Nielsen and D.A.T. Harper

Natural History Museum of Denmark (Geology), University of Copenhagen, DK-1350 Copenhagen K, Denmark. janr@snm.ku.dk,
arnet@snm.ku.dk, dharper@snm.ku.dk

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INTRODUCTION

The lithological composition of the Lower and Middle Ordovician metasediments of the Hardangervidda plateau, south-western Norway, is very different from the coeval Ordovician succession known elsewhere from Baltica, mainly because of an unusually high content of coarse-grained, siliciclastic material. The overall stratigraphy and lithology was described by Andresen (1978), who subdivided the Ordovician succession into five formations (Fig. 1).

Fossils are rare (Andresen, 1974) and predominantly deformed due to substantial tectonism during the Silurian Caledonian orogeny. Most fossils have been reported from the highly strained and folded, chlorite-rich limestones of the Bjørnaskalle Formation, although basal Ordovician graptolites of the *Rhabdinopora* group occur within black shales of the Bjørno Member (Størmer, 1940). The macrofossil content, principally trilobites and brachiopods, was described in detail by Bruton et al. (1985), while microfossils (conodonts) are here reported for the first time.

The unusual sedimentary succession has been related to the existence of a supposed adjacent land area during the Ordovician, which was named 'Telemark Land' (Skjeseth, 1952; Størmer, 1967). 'Telemark Land' was believed to cover an area somewhere between the southwestern part of the Oslo Region (Langesund, Kjekling) in the east to the west coast of southern Norway. An important aim of the present work is to test if this hypothesis is supported by palaeontological evidence.

GEOLOGICAL SETTING

The high plateau Hardangervidda in south-western Norway is the westernmost area with extensively-preserved Lower Palaeozoic platform sediments in Baltica. Relatively few studies have focussed on this tectonically disturbed succession (Brøgger, 1893; Reusch et al., 1902; Andresen, 1974, 1978; Bruton et al.,

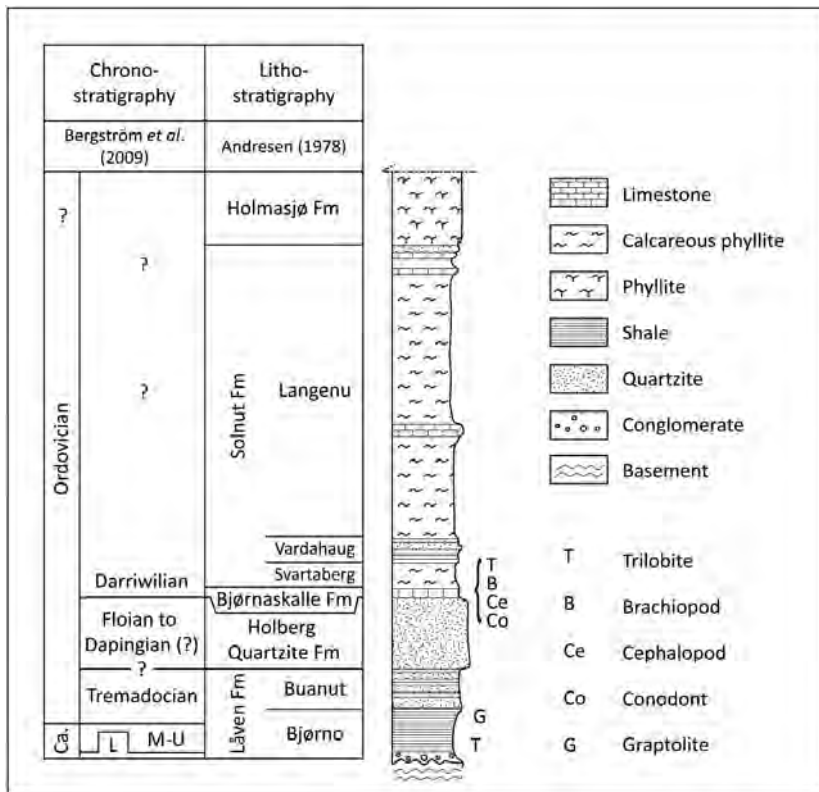


Figure 1. Stratigraphy and sedimentary succession of the Hardangervidda Group (modified from Andresen, 1978 and Bruton *et al.*, 1985). Fossil-bearing levels discussed in the text are indicated. The sedimentary log is not to scale.

1985). Despite the rather strong tectonic overprinting the local succession is considered to be part of the autochthon-parautochthon by previous authors. The Lower Palaeozoic rests unconformably on peneplained Precambrian basement and, locally, Proterozoic supracrustals.

A local lithostratigraphy was established by Andresen (1978). The Cambro-Ordovician (even Silurian?) strata, with an estimated thickness of some 400 m, were assigned to the Hardangervidda Group comprising the Låven, Holberg Quartzite, Bjørnaskalle, Solnut and Holmasjø formations (Fig. 1). The lowermost Låven Formation was subdivided into the Bjørno and Buanut members, of which the former corresponds to the Alum Shale Fm of the rest of Scandinavia (Nielsen and Schovsbo, 2006). Poorly preserved Middle Cambrian trilobites and early Tremadocian graptolites have been reported from the 10–60 m thick unit (Bruton *et al.*, 1985). A very thin Lower Cambrian sandstone unit (unnamed) is preserved locally at the base of the Låven Formation. The Buanut Member contains sandstone beds, intercalated with shale, and it may be characterised as a transitional unit continuous with the overlying Holberg Quartzite Formation, a 15–60 m thick quartz sandstone unit. In Baltoscandia only Hardangervidda is known to comprise Early Ordovician quartzites of such a considerable thickness. No fossils have been found in the latter unit so far. It is in turn unconformably overlain by the 4–8 m thick Bjørnaskalle Formation, which typically consists of strongly tectonized, impure limestone beds. Trilobites, brachiopods

and orthoconic cephalopods have been found in this unit, indicating an early Darriwilian age (Andresen, 1974; Bruton et al., 1985). We here report conodonts confirming this age. The Bjørnaskalle Formation is overlain by the c. 200 m thick Solnut Formation with a gradational boundary. It is subdivided into three members. From the base it consists of a c. 15 m thick dark-grey phyllite (Svartaberg Member), a c. 15 m thick grey quartz sandstone (Vardahaug Member), in turn overlain by the heterolithic Langenu Member, dominated by green phyllite with intercalations of limestone, thin sandstone beds and blackish shales. The youngest unit preserved below the thrust sheets is the 25–250 m thick Holmasjø Formation; its lower boundary is transitional but often tectonized. The unit is dominated by strongly deformed phyllites grading into quartz schist. Layers of sandstone and limestone, up to 2 m thick, also occur. No fossils have been reported from the unit so far. The Lower Palaeozoic strata are overthrust by cover rocks typically metamorphosed to greenschist facies (Bryhni and Sturt, 1985).

The quartzitic Holberg Formation thins out from 40–50 m in the southern part of the area to zero in the northern part (Andresen, 1978). The sandstone content in the underlying Buanut Member of the Låven Formation likewise diminishes in the same direction and the overlying Bjørnaskalla Formation grades into a green calcareous phyllite in a north-westwards direction. Hence it is obvious that the basin deepened towards the northwest and the hypothetical 'Telemark Land' must have been located somewhere southeast of Hardangervidda. It is possible that 'Telemark Land' reflects a late Tremadocian/early Floian updoming in response to Caledonian stress, maybe even a foreland bulge. The Solnut-Holmasjø formations in any case signal a change to foreland setting with much higher depositional rates than before. The same phenomenon is seen in the Oslo Region above the Svartodden Member of Huk Fm, which broadly speaking corresponds to the Bjørnaskalla Formation (compare Owen et al., 1990).

THE FOSSIL FAUNAS OF THE BJØRNASKALLE FORMATION

Fossils are generally sparse within the Bjørnaskalle Formation. The brachiopod fauna is poorly known and severely deformed (Bruton et al., 1985). It was collected 3.5 km SE of Haukelisetser and occurs in calcareous sandstones. It is moderately diverse containing at least three species of orthidine brachiopods including a coarse-ribbed form similar to either *Orthambonites* or *Paralenorthis*. Better preserved is a member of the *Antigonambonites plana* species group (Öpik, 1934) that suggests a late Dapingian-early Darriwilian age for this part of the formation; this species group is also described from the Huk Formation (Lysaker Member) of the Oslo Region (Öpik, 1939). Finally the fauna also contains a large species that Harper (in Bruton et al., 1985) tentatively assigned to the Alimbellidae, an aberrant group of orthidines.

One out of three processed samples from the Bjørnaskalle Formation contained conodonts. About 120 conodont elements were recovered from a relatively pure limestone sample collected 40 cm above the base of the Bjørnaskalle Formation near Locality B of Andresen (1974). The small conodont fauna is poorly preserved. It is characterized by *Drepanoistodus basiovalis*, *D. stougei*, *Baltoniodus medius*, *B. norrlandicus* (?), *Protopanderodus rectus*, *Lenodus* cf. *variabilis*, *Semiacontiodus davidi* and *Scalpellodus gracilis* (Fig. 2). Rare *Periodon macrodentatus*, *Dzikodus*? sp., *Microzarkodina* sp. (with one anterior denticle on the P-element) and *Paraistodus originalis* also occur. The species are common across Baltoscandia (Rasmussen 2001), and it correlates with the interval from the upper part of *Lenodus variabilis* Zone to the *Yangzeplacognathus crassus* Zone of the Baltic platform (e.g. Löfgren and Zhang, 2003; Stouge and Nielsen, 2003) suggesting an early Darriwilian age (latest Dw1 to early Dw2, most likely the latter), which narrows previous biostratigraphical assessments (Andresen, 1974; Bruton et al., 1985).

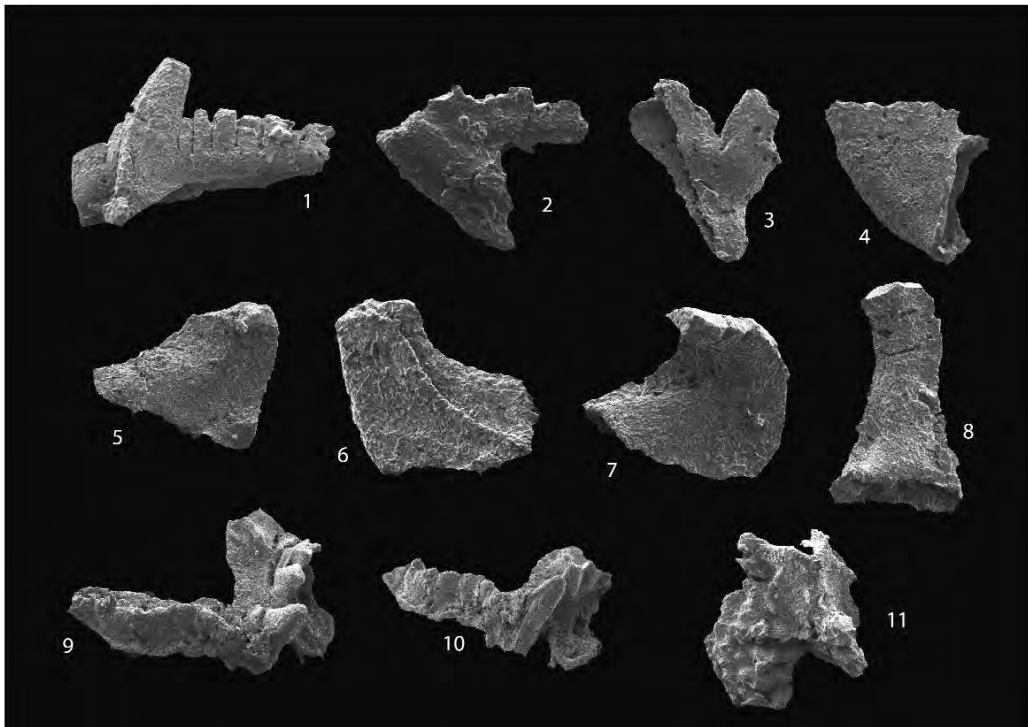


Figure 2. Conodont elements of the Bjørnaskalle Formation. Size of elements; figs. 1–8, 11, c. 0.2 mm; figs. 9–10, c. 0.4 mm. Repository: Natural History Museum, Oslo, Norway (PMO). 1–3, *Baltoniodus medius*; 1, Pa element (PMO 220.354/01); 2, Sb element (PMO 220.354/02); 3, M element (PMO 220.354/03). 4, 7–8, *Scalpellodus gracilis*; 4, short-based drepanodiform element (PMO 220.354/04); 7, short-based drepanodiform element, strongly deformed (PMO 220.354/05); 8, scandodiform element (PMO 220.354/06). 5–6, *Protopanderodus rectus*; 5, scandodiform element (PMO 220.354/07); 6, symmetrical acontiodiform element (PMO 220.354/08). 9–10, *Lenodus cf. variabilis*; 9, Pb element, upper view (PMO 220.354/09); 10, same as 9, oblique view. 11, *Dzikodus* sp. Pa element (PMO 220.354/10).

An orthoconic cephalopod belonging to the family Ormoceratidae was figured by Andresen (1974). The c. 6 cm long, broken specimen is rapidly expanding (12° according to Andresen) with slightly compressed siphuncle segments (SCR about 1.2 following the terminology of Frey 1995). The ratio between cameral length and diameter is moderate to judge from the illustrated specimen (c. 0.25). It probably belongs to the genus *Adamsoceras* Flower. The concavity of septae, the expansion rate of the shell and the SCR-ratio indicate further that it may be referred to *Adamsoceras holmi* (Troedsson), which has been described from the Darriwilian Hølen and Segerstad limestones of Öland, Sweden and equivalent units in Estonia (Troedsson 1926). However, more and better preserved specimens are needed to confirm this determination. Additional, poorly preserved orthoconic cephalopods, usually 30–40 cm long, have been observed within the Bjørnaskalle Formation. No siphuncle details were visible, but field observations showed that they are moderately expanding (6–7 degrees) with a short to moderate ratio between cameral length and diameter (c. 0.20). Orthoconic cephalopods are abundant within the Darriwilian Svartodden Member of the Huk Formation in the Oslo Region and the equivalent upper part of the Stein Formation in the Scandinavian Caledonides, which may give further evidence for correlation with these units.

PALAEOECOLOGY

The predominance of large orthidine brachiopods together with *Antigonambonites* suggests moderate water depths associated with predominantly hard substrates (Rasmussen et al., 2009).

Multivariate analyses of the conodont fauna from the Bjørnaskalle Formation and equivalent faunas from the Huk Formation in the Oslo Region (Slemmestad; Rasmussen, 1991) and the Stein Formation of the Scandinavian Caledonides (Stein, Andersön-B, Røste and Jørnolia; Rasmussen, 2001) were run to investigate the possible similarities between the six localities based on the occurrence of the ten selected generic units, the *Lenodus-Eoplacognathus* group, *Semiacontiodus*, *Baltoniodus*, *Microzarkodina*, *Scalpellodus*, *Drepanoistodus*, *Protopanderodus*, *Periodon*, *Costiconus* and *Nordiora*, using Correspondence Analysis (see Rasmussen and Stouge, 1995). The Correspondence Analysis Q-mode plot (Fig. 3) shows that the relative abundance of these genera in the Hardangervidda conodont fauna is most similar to the Slemmestad fauna. They are both situated in the left part of the diagram, while the outer platform localities from the Scandinavian Caledonides are located in the central and right parts of the diagram. This location indicates that the Hardangervidda fauna represents a relatively shallow or nearshore palaeoenvironment, probably slightly deeper than that of the contemporary Slemmestad locality farther to the east, but significantly more shallow or nearshore than the faunas described from the allochthonous Stein Formation situated in the Scandinavian Caledonides northeast of the study area. Accordingly, both conodonts and brachiopods indicate moderate water depth during deposition of the Bjørnaskalle Formation.

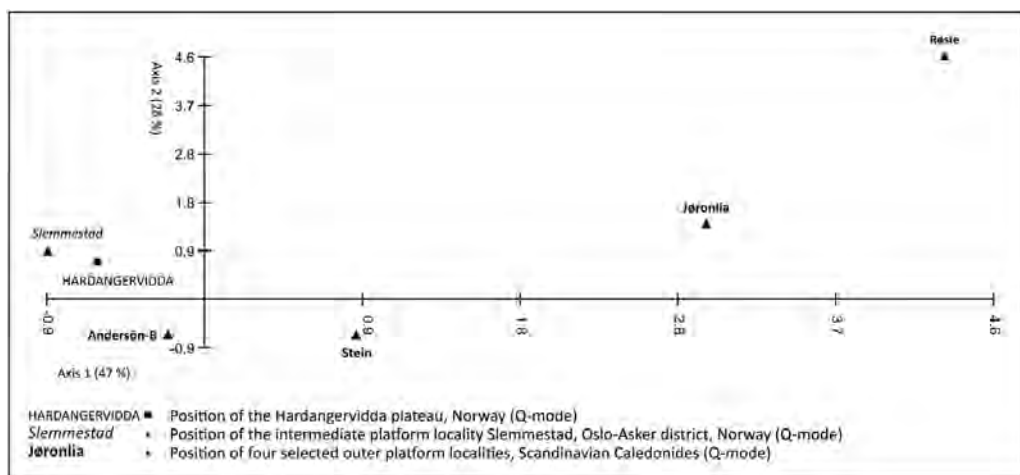


Figure 3. Correspondence Analysis Q-mode scatter-plot showing the relative position of the analysed localities. The analysis is based on the total number of specimens within the selected genera (*Lenodus-Eoplacognathus* group, *Semiacontiodus*, *Baltoniodus*, *Microzarkodina*, *Scalpellodus*, *Drepanoistodus*, *Protopanderodus*, *Periodon*, *Costiconus* and *Nordiora*) from Hardangervidda, Slemmestad, Andersön-B, Røste, Stein and Jørnolia.

Palaeobiogeography

With exception of the possible alimbellid, the brachiopod fauna of the Bjørnaskalle Formation is typical of the Baltic fauna (Rasmussen and Harper, 2008), however alimbellids and *Antigonambonites* are also known from localities belonging to the Celtic Province (Bruton and Harper, 1985), consistent with the

marginal setting of the Hardangervidda fauna. The biogeographic affinities of the brachiopod fauna are poorly constrained but the brachiopods have links with both the Baltic and Celtic provinces at this time.

The conodont fauna of the Bjørnaskalle Formation has a characteristic Baltic affinity. The relatively shallow or near-shore location of the analysed Hardangervidda fauna indicates that it was situated close to the shorelines of the emerged area 'Telemark Land'. The supposed proximal position on the northwestern side of 'Telemark Land' gave rise to deposition of sediments with an, for Baltica, abnormally high siliciclastic sandstone content (e.g. the Holberg Formation) in the southern and central parts of Hardangervidda. As pointed out above, the thickness of the Holberg Formation thins out towards the north and northwest, suggesting gradually deeper palaeoenvironments in this direction (Andresen, 1978; Bryhni and Sturt, 1985).

The Hardangervidda area was located between the intermediate platform areas of the Baltic craton and the lapetus island arcs and ocean islands palinspastically placed north of the Baltic craton in the Mid Ordovician (Fig. 4). The strong Baltic component in the conodont and brachiopod faunas favours that the sedimentary succession was transported only a small distance during the Caledonian orogeny. Severe tectonic deformation of the Ordovician sediments is in favour of a parautochthonous rather than an autochthonous position of the succession.

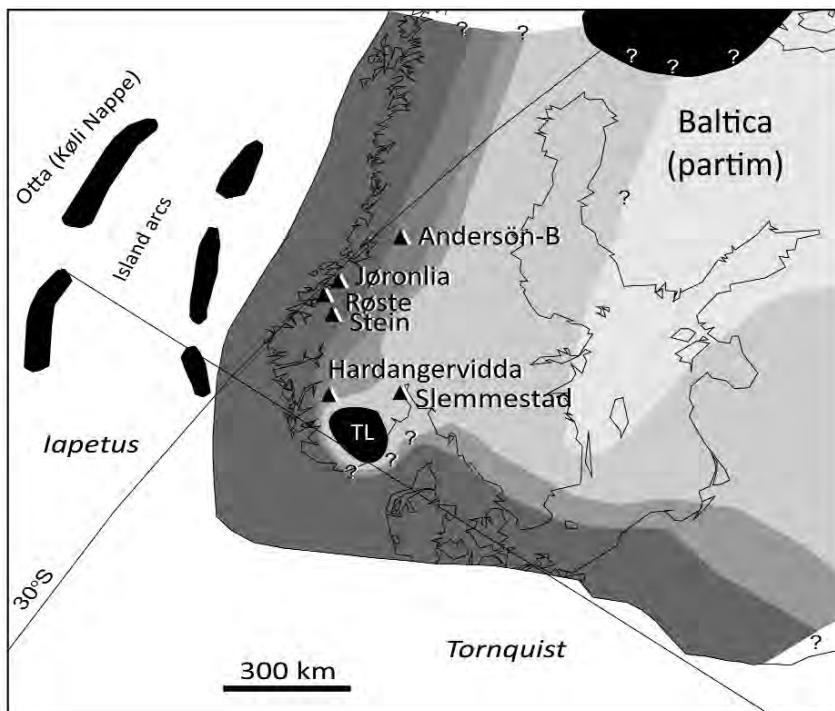


Figure 4. Darriwillian (c. 465 Ma) palaeogeographic position of Baltica with the palinspastic position of the analysed localities indicated (modified from Cocks and Torsvik 2005 and including information from Bruton and Harper 1988). Black areas indicate land areas or islands. TL='Telemark Land'. Conodont biofacies belts are marked with various grey colours, the most nearshore in light grey and the most distal in dark grey (based on Rasmussen and Stouge, 1995).

THERMAL MATURATION

The significant textural alteration of the conodont elements suggests the influence of regional metamorphism (Rejebian et al., 1987), which is in accordance with earlier interpretations (see e.g. Bryhni and Sturt, 1985) suggesting a Caledonian overthrust cover, metamorphosed in the greenschist facies. The conodont colour alteration index is CAI 5–6, which indicates a substantial post-sedimentary overburden of 9 km or more.

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BIOSTRATIGRAPHY OF THE MIDDLE ORDOVICIAN BRACHIOPODS FROM CENTRAL SPAIN

J. Reyes-Abril¹, J.C. Gutiérrez-Marco² and E. Villas³

¹ Escuela de Ingeniería Geológica, Grupo TERRA, Universidad de Los Andes, Núcleo Pedro Rincón Gutiérrez s/n, 5101 Mérida, Venezuela. jreyes@ula.ve

² Instituto de Geociencias (CSIC-UCM), Facultad CC. Geológicas, José Antonio Nováis 2, 28040 Madrid, Spain. jcgrapto@geo.ucm.es

³ Departamento de Ciencias de la Tierra, Facultad de Ciencias, Universidad de Zaragoza, Pedro Cerbuna s/n, 50009 Zaragoza, Spain. villas@unizar.es

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INTRODUCTION

A recent taxonomic study on the rhynchonelliformean brachiopods from the Darriwilian dark shales of the Central-Iberian Zone, Central Spain (Reyes-Abril, 2009; Reyes-Abril et al., 2010) has considerably increased the number of orthides and strophomenides known across the whole Iberian Peninsula. The studied brachiopods were collected in 58 localities of 6 provinces from the regions of Castilla-La Mancha, Andalucía and Extremadura. Middle Ordovician brachiopods previously known in the area were mainly derived from upper Darriwilian rocks, while most of the new studied brachiopods are from middle Darriwilian rocks. A total of 21 genera are now known from the whole Darriwilian strata of Iberia including the strophomenide genera *Aegiromena* and *Dactylogonia*, as well as 19 orthide genera, including *Almadenorthis*, *Gutiorthis*, *Orthambonites*, *Paralenorthis*, *Sivorthis*, *Apollonorthis*, *Atlantida*, *Brandysia*, *Eodalmannella*, *Howellites*, *Cacemia*, *Heterorthina*, *Tissintia*, *Mcewanella*, *Crozonorthis*, *Nocturnellia* and *Lipanorthis*, plus two new genera of the families Cremnarthridae and Harknessellidae.

Considering the regional biostratigraphical and biochronological context, in which the Ibero-Armorican brachiopod biozones can be easily correlated with biozones based on other fossil groups, we have followed the Mediterranean stratigraphic scale, originally proposed in Bohemia (Havlíček and Marek, 1973) and updated and completed in Iberia (Gutiérrez-Marco et al., 1995, 2002). Thus we are referring the stratigraphic range of the studied brachiopods to the Mediterranean regional stages Oretanian and Dobrotivian, approximately correlatable to the middle and upper stage slices (Dw2-3) of the global Darriwilian stage. Exception made with the uppermost Dobrotivian, which extends into the Upper Ordovician, correlating with the lowermost Sandbian (Gutiérrez-Marco et al., 2008; Bergström et al., 2009).

Brachiopods were first used in the late nineteenth century to subdivide the Ordovician succession in Iberia (Delgado, 1897). Since then, different brachiopod biozones have been proposed for the Middle Ordovician rocks of Spain (Born, 1918; García Alcalde and Arbizu, 1982; Villas, 1985; Young, 1985; Gutiérrez-Marco et al., 1984, 2002). Recent improvements in the detailed stratigraphical knowledge of

many successions in the Central Iberian Zone, and on the taxonomy of important brachiopod species, have allowed the redefinition and updating of previous brachiopod biozones, which are examined in the present note.

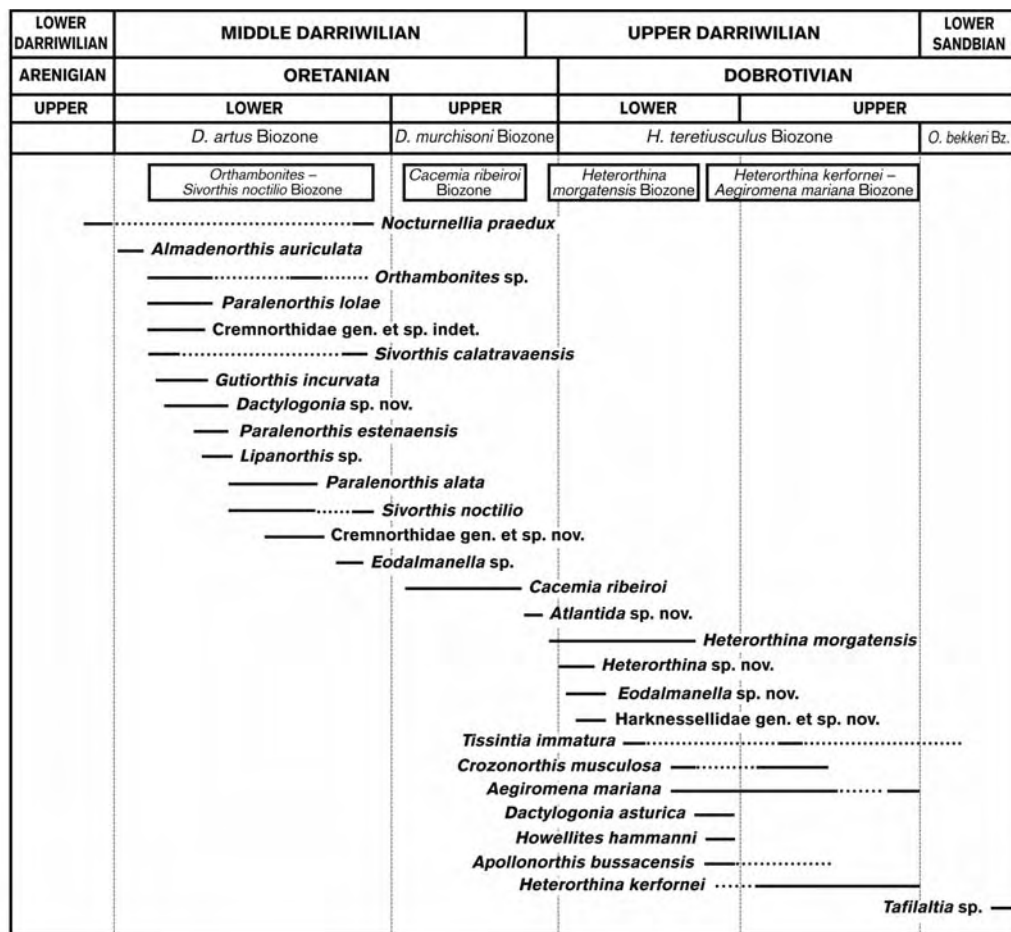


Figure 1. Stratigraphic correlation of the proposed Middle Ordovician brachiopod biozones, with the vertical range of known brachiopod species from Central Spain.

STUDIED BIOZONES

The stratigraphic range of the 27 brachiopod species identified in a wide area of the southern Central Iberian Zone (Reyes-Abril, 2009; Reyes-Abril et al., 2010) has allowed to review the biostratigraphical units defined upon brachiopods by previous authors. Our contribution recognizes, from base to top, the following four biozones: 1, *Orthambonites*–*Sivorthis noctilio* Partial Range Biozone (lower Oretanian; base of the Darriwilian 2); 2, *Cacemia ribeiroi* Taxon Range Biozone (upper Oretanian; upper Darriwilian 2); 3, *Heterorthina morgatensis* Taxon Range Biozone (uppermost Oretanian to lower Dobrotivian; lower Darriwilian 3), and 4, *Heterorthina kerfornei*–*Aegiromena mariana* Partial Range Biozone (lower to lower

upper Dobrotivian; uppermost Darriwilian 3). Although a pre-Oretanian species, *Nocturnellia praedux* (Havlíček in Arbin et al., 1978), occurs below the base of the lowest of the four biozones considered, its extreme rarity and its scattered occurrence in the lower Oretanian precludes its consideration as a biostratigraphic marker. The regional brachiopod biozonation can be applied to the Portuguese part of the Central Iberian Zone, but not to areas in northern Spain, where the lower and upper Dobrotivian assemblages include some taxa not recovered from Central Iberia. Some common brachiopods in the Middle Ordovician of the studied area, including the index-species of the proposed biozones, are illustrated in Plate 1.

Orthambonites–Sivorthis noctilio Biozone

This is a Partial Range Biozone, with the base defined by the lowest record of the genus *Orthambonites* on a wide area of the Central Iberian Zone, mostly coinciding in vertical range with the genus *Sivorthis*, which appears slightly higher in the succession. The top of the biozone is marked by the disappearance of *Sivorthis noctilio* (Pl. 1: G), the species most abundant, ubiquitous and easily recognizable in the whole Ibero-Armorican area. Among the associated species exclusive to the biozone are *Sivorthis calatravaensis*, *Paralenorthis estenaensis*, *Paralenorthis lolae*, *Gutiorthis incurvata*, a new genus and species of the family Cremnorthidae, as well as a new species of *Dactylogonia*. *Paralenorthis alata*, a species known from the British Arenigian (Bates, 1969), presents here its youngest occurrence. Punctual records are also known of *Nocturnellia praedux*, *Eodalmanella* sp. and *Lipanorthis* sp. The occurrence of *Almadenorthis auriculata*, restricted originally to its type locality, is now found below the base of the biozone, although it could be extended up through it in future studies. The upper boundary of the biozone coincides with a barren interzone for rhynchonelliformean brachiopods, which precedes the first record of the nominal form of the overlying biozone.

The *Orthambonites–S. noctilio* Biozone spreads through the southern part of the Central Iberian Zone (lower beds of the Navas de Estena Shales and the Navatrasierra Shales; lower and middle beds of the Río Shales), as well as through the middle of the Valongo and Moncorvo formations in Portugal (Sá, 2005). In the Armorican Massif (France) the biozone can also be recognized in the lower part of the Traveusot Formation in the synclines south of Rennes (Tromelin and Lebesconte, 1876; Pillet et al., 1990), where *S. noctilio* and the same new genus and species of the family Cremnorthidae are also represented.

The biozone proposed herein substitutes in range and meaning the old "*Monorthis*" *noctilio* or "*Orthis*" *noctilio* Biozone, considered by Gutiérrez-Marco et al. (1984, 2002) and San José et al. (1992), after the taxonomic revision of the nominal species and the redefinition of its base. This way, the first record of such regionally characteristic genera as *Orthambonites*, *Paralenorthis* or *Gutiorthis*, are included in the Biozone. Considering the brachiopods represented within the association, the biozone is also equivalent to the upper part of the "*Didymograptus* Shales" and to the "*Orthis noctilio* Shales" of Delgado (1908). The characteristic brachiopods of the "*Orthis calligramma* Zone" (Born, 1918) and the "*Hesperorthis* Biozone" (García Alcalde and Arbizu, 1982) occur also in the revised biozone. The occurrence of some index fossils of the biozone such as "*Orthis*" *noctilio* and "*Orthis*" *miniensis* in the Upper Ordovician of Sardinia (Meneghini, 1857; Vinassa de Regny, 1927; Leone, 1998) is unlikely and both were probably confused with younger forms. In this sense, and according to Havlíček et al. (1987), the Sardinian specimens identified as *O. noctilio*, *O. noctilio novata*, and even as *O. calligramma* by Vinassa de Regny (1927), really belong to *Nicolella actoniae*, a characteristic species for the late Katian (Upper Ordovician) of the Mediterranean Region.

From a chonostratigraphic point of view, the dating of the *Orthambonites*–*S. noctilio* Biozone as lower Oretanian, made by Gutiérrez-Marco et al. (1984, 2002), is corroborated with the record of graptolites from the *Didymograptus artus* Biozone throughout the whole range of the unit. The lower Oretanian is correlated to the first half of the Darriwilian 2 of the Global Scale.

Cacemia ribeiroi Biozone

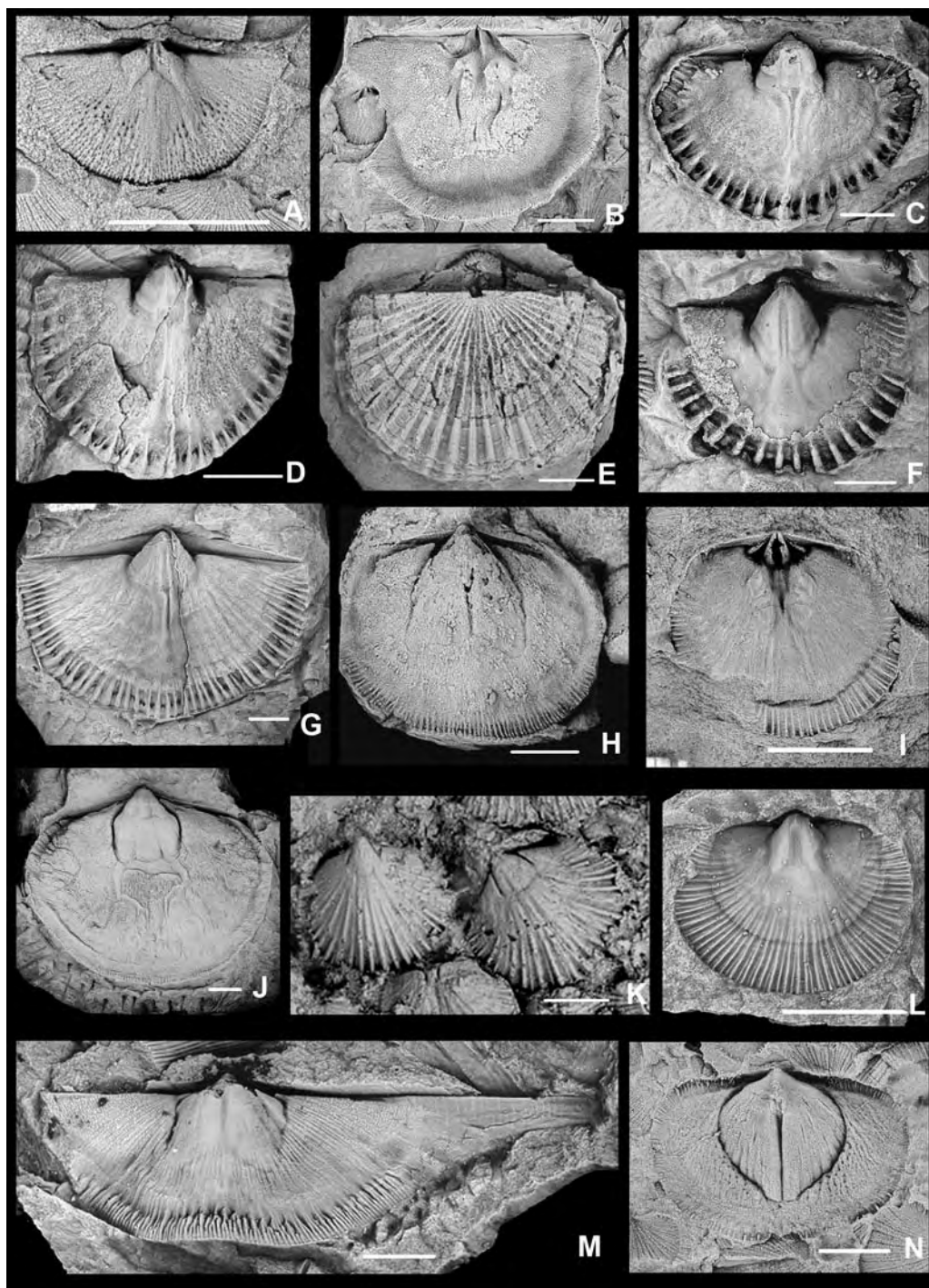
This biozone coincides with the vertical range of *Cacemia ribeiroi*, besides which, only a new species of *Heterorthina* occurs in its uppermost beds. The unit is widely spread through the whole Ibero-Armorican area, where the record of the graptolite *Didymograptus murchisoni*, which pre- and postdates this biozone, indicates an late Oretanian age (latest Darriwilian 2 substage).

Cacemia ribeiroi is a frequent and easily identifiable species, because its auriculated outline and relatively fine costellation (Pl. 1: M). Several authors have pointed out its stratigraphic value and used it to propose the "*Orthis Ribeiroi* Shales" in Bussaco (Delgado 1897, 1908), the "*Orthis ribeiroi* Zone" (Born, 1918) and the "*Cacemia* Biozone" (García-Alcalde and Arbizu, 1982) in Almadén, and the "*Cacemia ribeiroi* Biozone" in the southern Central Iberian Zone (Gutiérrez-Marco et al., 1994, 2002). However, mixed collections have been detected for the establishment of some of these units in the works of Delgado (1908) and Born (1918), with specimens coming from younger beds but also, occasionally, from lower strata, such as when *C. ribeiroi* was confused with other auriculated brachiopods from the lower Oretanian (*Sivorthis*, *Paralenorthis*). This is also the case of "*Orthis vespertilio*" Sow., a species repeatedly confused with *C. ribeiroi* in the Central Iberian Zone, following an erroneous identification and illustration by Verneuil and Barrande (1855) and Mallada (1875), in which part of the citations correspond with lower Oretanian localities.

The *C. ribeiroi* Biozone has been recognized in the Navas de Estena and Navatrasierra formations and in the upper part of the Río shales of the studied area. It is also represented in the Luarca Shales of the West Asturian-Leonese Zone (Gutiérrez-Marco et al., 1999). Outside Spain, the nominal species has been described from the upper part of the Brejo Fundeiro Formation (Cacemes Group) of Bussaco (Portugal), as well as in the lower part of the Postolonnec Formation of the Armorican Massif, France (Mélou, 1976). It has been also identified, with doubts, in the Algerian Sahara (Mélou et al., 1999).

Above the *C. ribeiroi* Biozone, an upper Oretanian graptolitic interval was recorded, yielding scarce trilobites and molluscs of broad vertical range. Nevertheless, in a Montes de Toledo section, these beds yielded a new species of the plectorthid *Atlantida*, a genus so far restricted to the upper Darriwilian of Morocco (Havlíček, 1971) and, in consequence, of little biostratigraphic value.

Plate 1. Some Middle Ordovician brachiopods from the Central Iberian Zone, including the index species of the proposed brachiopod biozones. Scale bars 5 mm, except where otherwise indicated. A, *Aegiromena mariana* Drot, 1969, ventral internal mould, MGM-6477-O, scale bar 2 mm, Calzada de Calatrava. B, *Dactylogonia asturica* (Villas, 1989), dorsal internal mould, MGM-6453-O, La Alameda. C, *Gutiorthis incurvata* Reyes-Abril and Villas (in Reyes-Abril et al., 2010), ventral internal mould, MGM-5968-O, Navas de Estena. D-E, *Orthambonites* sp., ventral internal mould, MGM-5999-O, Navas de Estena (D) and latex cast of dorsal exterior, MGM-6007-O, Ventas con Peña Aguilera (E). F, *Paralenorthis estenaensis* Reyes-Abril and Villas (in Reyes-Abril et al., 2010), ventral internal mould, MGM-6072-O, Navas de Estena. G, *Sivorthis noctilio* (Sharpe, 1849), ventral internal mould, MGM-6249-O, Ventas con Peña Aguilera. H, *Heterorthina morgatensis* Mélou, 1975, ventral internal mould, MGM-6792-O, Retuerta del Bullaque. I, L, *Heterorthina kerfornei* Mélou, 1975, dorsal internal mould, MGM-6758-O, Calzada de Calatrava (I) and ventral internal mould, MGM-6729-O, Calzada de Calatrava (L). J, *Apollonorthis bussacensis* (Sharpe in Ribeiro et al., 1853), ventral internal mould, MGM-6509-O, Calzada de Calatrava. K, *Nocturnellia praedux* (Havlíček in Arbin et al., 1978), internal moulds of two ventral valves, (left) MGM-6937-O, (right) MGM-6938-O, scale bar 2mm, Solana del Pino. M, *Cacemia ribeiroi* (Sharpe in Ribeiro et al., 1853), ventral internal mould, MGM-6335-O, Helechosa de Los Montes. N, *Crozonorthis musculosa* Mélou, 1976, ventral internal mould, MGM-6988-O, Calzada de Calatrava.



Heterorthina morgatensis Biozone

It is defined by the vertical range of its nominal species, and thus coincides with the *H. morgatensis* Biozone proposed by Villas (1985) in the Iberian Chains (NE Spain) and by Young (1985) in the Serra do Bussaco (Portugal). Besides *H. morgatensis*, also *Crozonorthis musculosa*, *Aegiromena mariana* and an undetermined dalmanellidine occur in the upper beds of the biozone. In its lower beds a new species of *Eodalmarella* has been recorded (Reyes-Abril, 2009).

The *H. morgatensis* Biozone is widely recognized throughout the Ibero-Armorican region, having been identified in the middle part of the Navas the Estena and Navatrasierra Shales, as well as in the El Caño Alternation of the Central Iberian Zone. Within the Navatrasierra Shales, common records of *H. morgatensis* lie below and above the laterally discontinuous Los Rasos Sandstones Member, being sometimes recorded in coquinoid beds within these sandy tempestites, or in its Eastern Sierra Morena equivalent El Caño Alternation (Mélou, 1975). In the Portuguese extension of the Central Iberian Zone, the *H. morgatensis* Biozone can be identified in Penha Garcia and in the Serra do Bussaco, at the top of Brejo Fundeiro Formation and in the base of the Fonte da Horta Formation (Henry et al., 1976; Young, 1985). It has also been recognized at the Iberian Chains (Villas, 1985) and the Cantabrian Zone of the Iberian Massif (Gutiérrez-Marco et al., 1996, 1999; Gutiérrez-Marco and Bernárdez, 2003). In the Armorican Massif, *H. morgatensis* is typically recorded in the Postolonnec Formation, within the shales that overlie the Kerarvail Sandstones, a local equivalent to the Los Rasos and Monte da Sombadeira formations (Mélou, 1975; Henry et al., 1976), as well as in the sandstone beds at the base of the Mont de Besneville Formation, in Normandy. The record of the species by Mélou (1975) in another Armorican locality, Andouillé-La Touche, at the top of the Andouillé Formation, is very unlikely. Those beds correspond to the trilobite biozone of *Placoparia borni* (Subzone of *Marrolithus bureaui*), which has been dated with graptolites and chitinozoans as lowermost Sandbian (Upper Ordovician). The form identified there by Mélou (1975) may coincide with another species from La Touche, determined as *Heterorthina* sp. by Young (1985: Pl. 21, figs. 8-15), but that could belong to a different genera like those occurring on a similar stratigraphic position at the top of the Postolonnec Formation (Botquelen and Mélou, 2007).

Chronostratigraphically, the base of the *H. morgatensis* Biozone can be referred to the Upper Oretanian based on the occurrence of graptolites from the *Didymograptus murchisoni* Biozone at the Cantabrian Zone and in a few Central Iberian localities (Gutiérrez-Marco et al., 1994, 1999, 2002). Nevertheless, most of its development corresponds to the lower Dobrotivian, and the biozone can be correlated, in a broad sense, with the lower half of the Darriwilian 3 substage.

Heterorthina kerfornei–*Aegiromena mariana* Biozone

This Partial Range Biozone is characterized by the almost total concurrent range of *Heterorthina kerfornei* and *Aegiromena mariana*. The base of the biozone coincides with the acme of *Aegiromena mariana*, which frequently crowds many bedding planes. In Central Iberia, the lower part of the biozone can be correlated with the *Morgatia hupei* trilobite Subzone, where the brachiopod assemblage reaches its highest diversity. Besides the two nominal species, *Howellites hammani* and Harknesellidae gen. et sp. nov. occur frequently, in addition to the local acme of *Crozonorthis musculosa* and the earliest record of *Apollonorthis bussacensis*. In higher beds of the unit, yet coinciding with the *Placoparia* (*Coplacoparia*) *borni* trilobite biozone, the brachiopod association is dominated by *H. kerfornei* and *A. mariana*, with the appreciable decrease of *C. musculosa* (restricted to a few horizons), and the sporadic record of

Dactylogonia asturica or *Apollonorthis bussacensis*. The disappearance of *Heterorthina kerfornei*, slightly higher than that of *A. mariana*, marks the biozone top, which coincides with the lower boundary of the *Lagenochitina ponceti* chitinozoan biozone, but not with the disappearance of the trilobite *Neseuretus tristani*.

The re-definition of the *H. kerfornei*–*A. mariana* Biozone (name adapted from Gutiérrez-Marco et al., 2002) is intended to solve the identification problems of other Dobrotivian brachiopod biozones overlying the *H. morgatensis* Biozone. Different solutions have been given by other authors to the difficulty of evaluating the biostratigraphic meaning of local occurrences of very conspicuous species such as *C. musculosa* or *A. bussacensis*. This is especially complicated in the absence of an effective biostratigraphic control provided by graptolites, chitinozoans or trilobites. The *H. kerfornei*–*A. mariana* Biozone, such as it is considered in this study, is a stratigraphical equivalent of the *A. mariana* and *E. musculosa* biozones of García-Alcalde and Arbizu (1982), and also equates the combination of the *A. mariana*–*E. musculosa* Concurrent Range Biozone and the *H. kerfornei* Taxon Range Biozone defined by Villas (1985). It is also coincident with the *H. kerfornei* Partial Range Biozone plus the ranges of those brachiopods occurring in the *Placoparia* (*Coplacoparia*) *borni* Taxon Range Biozone, as considered by Young (1985). The unit also includes the *H. kerfornei* Biozone of Gutiérrez-Marco et al. (1984, 1995) and San José et al. (1992).

In the study area, the *H. kerfornei*–*A. mariana* Biozone is represented in the upper part of the Navas de Estena Shales and in the Navatrasierra Shales; it also typically occurs in the Guindo Shales and in the lower part of the Botella Quartzite of the southernmost part of the Spanish Central Iberian Zone. In the Portuguese part, the biozone extends through most of the Fonte da Horta Formation, the Cabril Formation and probably, also, through part of the Carregueira Formation at Bussaco, Penha Garcia and Dornes (Henry et al., 1976; Young, 1985). The base of the unit could be represented in the upper third of the Valongo Formation of the Valongo-Arouca region (Couto et al., 1997), as well as in the Moncorvo Formation of Trás-os Montes (Sá, 2005), although detailed taxonomic studies are needed to verify it. Different occurrences of the biozone at the Iberian Peninsula are also known from the Iberian Chains (Villas, 1985) and the Cantabrian Zone (Truyols et al., 1996; Gutiérrez-Marco et al., 1999).

Outside the Iberian Peninsula, the nominal brachiopods are recorded in different localities from the Andouillé, Postolonnec and Travesout formations of the Armorican Massif (Mélou, 1973, 1975), with the exception of *A. mariana* in the Postolonnec Formation. Occasionally, the biozone can be characterized in North Africa, according to the study by Mélou et al. (1999) who record the co-occurrence of the two index species in Algeria, in beds yielding also *A. bussacensis* and *Tenuiseptorthis niliensis*.

Depending on the region considered within the studied area, the upper part of the *H. kerfornei*–*A. mariana* Biozone can be incomplete, since the pelitic beds characterizing it change laterally in the upper Dobrotivian to alternations of sandstones and shales, culminating in massive quartzites (Retuerta Sandstones, La Cierva or Botella Quartzites). In these coarse-grained units, corresponding to shallow and turbulent environments, the records of the nominal species are interrupted. The only brachiopods occurring there are forms restricted to sandy facies, particularly *Tissintia immatura*, a species geographically widespread that also occurs in older units (Los Rasos Sandstones). The reference by Mélou (1975) to the occurrence of *H. morgatensis* in the Botella Quartzite at Sierra Morena, could actually correspond also to *Tissintia*. *T. immatura* is also known from the Dobrotivian sandstones of the Sierra de San Pedro (Elice Formation). However, *Tissintia* is not the only brachiopod occurring in the sandy facies from the Iberian Dobrotivian, since *Tafilaltia* has also been cited from the top of the Retuerta Sandstones at the Toledo Mountains (Montero, 1989) and with certainty it is yielded by correlative sandstones at the Hesperian Chains of the Iberian Cordillera. This is why San José et al. (1992) and Gutiérrez-Marco et al. (2002) came

to propose the erection of a *Tafilaltia* Sub-biozone within the uppermost Dobrotivian sandstones.

The base of the *H. kerfornei*–*A. mariana* biozone is placed within the Gondwanan range of the *Hustedograptus teretiusculus* graptolite biozone, and its top lies within the *Oepikograptus bekkeri* Biozone, paralleled by the *Nemagraptus gracilis* Biochronozone. Therefore, its total range extends from the high-lower Dobrotivian to the terminal-upper Dobrotivian, correlatable in global terms to the upper half of the Darriwilian 3 and the basal beds of the Sandbian 1, bridging the boundary between the Middle and the Upper Ordovician series. However, most of the Central Iberian fossiliferous localities included in this biozone are from the uppermost Darriwilian, and have not been detected yet within the sandy and quartzitic facies that locally characterize the lowermost Sandbian.

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STRATIGRAPHY AND STRUCTURE OF THE UPPERMOST PART OF THE LUARCA FORMATION IN ALTO BIERZO, LEÓN (ORDOVICIAN, NW SPAIN)

M.A. Rodríguez Sastre¹ and L. González Menéndez²

¹ Quiñones 5, 24240 León, Spain. marsast@gmail.com

² Instituto Geológico y Minero de España, Parque Científico de León, Avda. Real, 24002 León, Spain.
l.gonzalez@igme.es

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INTRODUCTION

Recently excavated rock cuttings during quarrying activities in the (Middle Ordovician) rocks of the Alto Bierzo area of NW Spain allow a detailed study on the petrography, stratigraphy and tectonic history of the upper part of the Luarca Formation (Middle Ordovician) and its transitional beds to the Agüeira Formation (Upper Ordovician). The focus of this research is to characterise the presence of slate beds of commercial quality in a detailed 147 m long section mainly located in the Luarca Fm.

The quality of slate deposits is controlled by basin stratigraphy and tectonic features, and the study of both parameters can be used for designing mining activities around the Jola quarry in the Bildeo Mountain, 4.5 km west of Páramo del Sil (N42°44'0.5", W7°00'4.5": Fig. 1).

GEOLOGICAL SETTING

The Alto Bierzo area is located in the León province within the Navia Alto Sil domain of the West Asturian-Leonese Zone (WALZ) of the Iberian Massif (Pulgar et al., 1982; Pérez-Estaún et al., 1982a, 1982b; Navarro, 1982; Matas and Fernández, 1982; Suárez et al., 1992; Vera, 2004) (Fig. 1). The study area lies in the northern flank of the Vega de Espinareda Synform, where the Ordovician outcrops mainly belong to the Luarca Formation, which affected by at least three phases of Variscan deformation, resulting in widespread folds, cleavage, and localized faults, shear zones, kink bands and cataclases.

The thickness of the Luarca Fm in the Vega de Espinareda Synform is estimated to range between 500 and 1,200 m. The massive slates are black and contain sulphides, that are interpreted to indicate euxinic depositional environments. Interbedded volcanic rocks have been recently recognized in the surrounding area of Lago-Fontarón and in the Alto Sil domain (Villa et al., 2004).

From a stratigraphic point of view, the Luarca Formation depicts a gradational basal contact with the middle Cambrian to Lower Ordovician Los Cabos Group, which mainly consists of sandstones, quartzites,

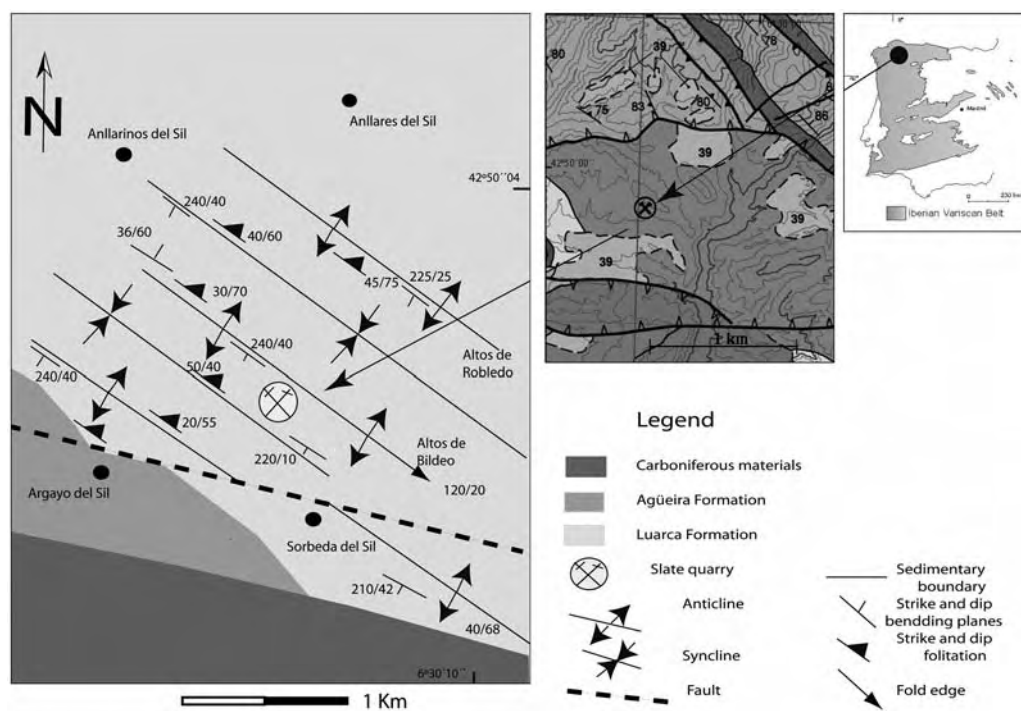


Figure 1. Geological maps of the study area (left, modified from Rodríguez Sastre and Gutiérrez Claverol, 2009; center, after Alonso et al., in press).

siltstones and slates. The Luarca Formation is conformably overlain by the Agüeira Formation (Upper Ordovician) with an apparent gradual and continuous transition between the two formations (Pérez Estaún and Marcos, 1981), perhaps involving a possible regional gap in sedimentation (Gutiérrez-Marco et al., 1999).

The classic description of the Luarca Fm was made in its type section near Luarca (Asturias) (Barrois, 1882). According to Marcos (1973) and other authors (in Vera, 2004), the total thickness is in excess of 500 m, and can be subdivided into three sequences: (1) the lower one, consists of 260 m of black pyritic shales with several interbedded ironstones and quartzites; (2) the intermediate sequence consists of the 50 to 80 m thick Sabugo Quartzite; and an upper sequence (3), represented by a thick massive unit of black shales and slates (250 m thick).

The most recent interpretation of the setting of the Ordovician sequence in the NW Iberian Massif indicates terrigenous marine shelf facies (shales, siltstones and sandstones) containing low diversity benthic assemblages of trilobites, ostracods, brachiopods, molluscs and echinoderms. This has been interpreted as cold-water faunas, indicative of high palaeolatitudes (Gutiérrez-Marco et al., 2002). In the studied area, Gutiérrez Marco et al. (1999) recorded the trilobites *Colpocoryphe* cf. *grandis* and *Neseretus?* sp. in a slate quarry near Anllares del Sil (Fig. 1), placed near the top of the Luarca Fm.

Depositional environment for the Luarca Fm in the Alto Sil domain was a relatively deep open shelf settled on a passive margin, with shallower environments towards the east, in the Cantabrian Zone (Vera, 2004).

Tectonic features of the southern branch of the WALZ (in Vera, 2004) consist of asymmetrical and angular hinge folds in fold trains developed during the first Variscan deformation phase (D1). The regional bedding attitude dips to the southwest as well as the axial planes of the identified folds. The folds are related to the first Variscan tectonic phase, with an well-developed axial plane parallel cleavage (S1). Subsequently to the D1 deformation, thrusts were developed (D2) accompanied by ductile deformation imparting a tectonic foliation (S2) and local cataclasites and associated D2 minor folds with subhorizontal limbs and curved hinges.

Finally, D1 folds were refolded by open, rounded folds with vertical axial planes developed during the third Variscan tectonic phase (D3). Incipient crenulations subvertical crenulations cleavage was developed (S3), Local kink-bands and faults are related to late orogenic phases.

STRATIGRAPHY

The stratigraphic sequence of the upper part of the Luarca Formation crops out in the upper 80 m of the trench section made in the Bildeo Mountain. Field observations were completed with data coming from five nearby prospecting boreholes. Several slate beds, with slightly different lithologies were recorded in a NE-SW section and are shown in Figure 2. In the outcrops, the original sedimentary features of the slates are difficult to recognise due to their metamorphic grade and the intense fracturation that affected the formation. The stratigraphic characterisation of commercial quality slate beds is very important for the planning of future quarrying activities.

In the studied area, the stratigraphic section (Fig. 2) roughly consists of a coarsening upwards (negative) sequence that, in descending order (from top to bottom), can be summarized as:

- 10 m: of sandstone interbedded with minor slates that might be related with the upper part of the transitional beds to the Agüeira Formation.
- 8 m of thin laminated siltstones and fine sandstones (lower part of the aforementioned transition) bearing pyritized ellipsoidal nodules 2-7 cm in diameter, flattened as a consequence of strong compaction and regional metamorphism. Some of these nodules are fossiliferous and have provided some remains of indeterminate trilobites.
- 20 m of silty slates showing smooth graphitized bright surfaces, which are oriented obliquely to the main D1 foliation planes.
- 0.5 to 5.25 m of silicified slates with segregated quartz, forming a reaction zone.
- 20-25 m of fine grained slate with some remains (<2 cm in size) of marine fossils such as crinoids and molluscs.
- 19 m of silty slate with black graphitic smooth surfaces.
- 3.6 m of slate with quartz cement from slate segregation, defining a reaction zone.
- > 1 m of fine grained slate, only partly observed in the core logging.

METAMORPHISM AND GEOCHEMISTRY

Regional metamorphism is characteristically of low-grade regional type (green schist facies), and the main paragenesis found is quartz+chlorite+sericite+muscovite. This assemblage, together with the absence of biotite, is indicative of metamorphism temperatures reaching between 200-430°C with burial

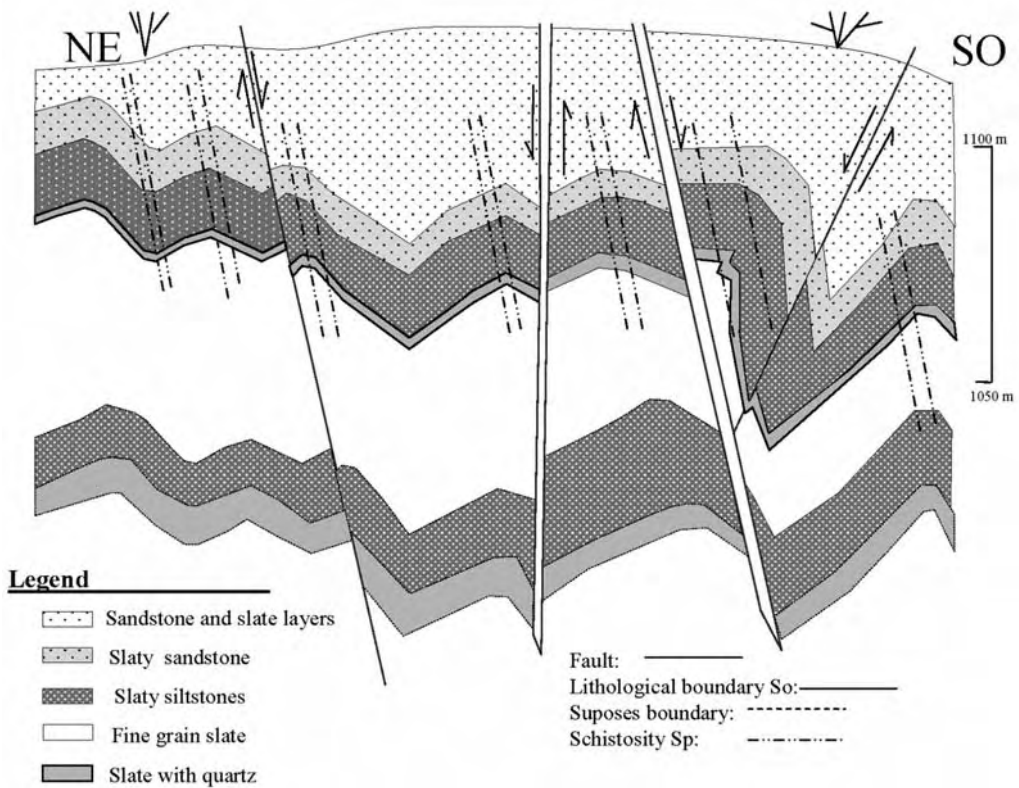


Figure 2. Geological cross-section around the Jola quarry showing the main lithostratigraphic and structural features of the area (modified from Rodríguez Sastre, 2008). The two upper sandy units represent the transitional beds from the Luarca Formation (Middle Ordovician, below) to the base of the Agüeira Fm (Upper Ordovician, not shown in the section).

depths of up to 10 km and pressures about 2-3 kb (Winkler, 1970). Chlorite is the index-mineral and its formation temperature was calculated by us using general geothermometer equations from microprobe analyses and structural formulae on the basis of 28 oxygen samples in four slate samples from NW Spain (Gómez Fernández and Bauluz Lázaro, 2009). The calculated temperature enters in the ranges between $384 \pm 32^\circ\text{C}$ (Cathelineau, 1988), $313 \pm 21^\circ\text{C}$ (Cathelineau and Nieva, 1985), $357 \pm 20^\circ\text{C}$ (Kranidiotis and Maclean, 1987) and $317 \pm 41^\circ\text{C}$ (Xie et al., 1997).

The influence of strain, combined with low to moderate temperature results in a mineral orientation perpendicular to the maximum pressure imparting cleavage to these rocks. Micrographs from the three thin sections of rocks from the studied section (Figure 3) show: A, section parallel to the foliation plane; and B and C sections perpendicular to the foliation plane. Chlorite and sericite-muscovite minerals grow in the cleavage surfaces at right angles to the direction of maximum shortening, while pre-synectonic minerals develop pressure or recrystallization (e.g. quartz) parallel to cleavage (Blyth and De Freitas, 2003). These examples correspond to fine grained slates of lepidoblastic texture. The foliation is a slaty cleavage when observed under the microscope (Fig. 3).

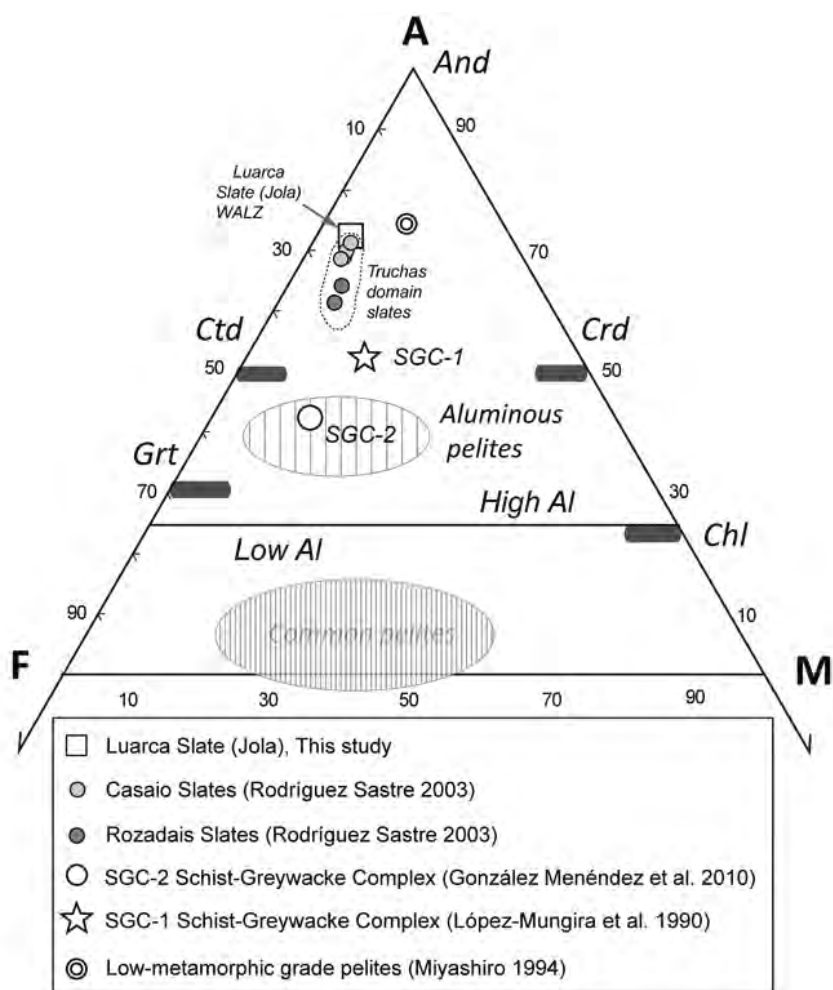


Figure 3. Micrographs of some thin sections on fine-grained slate from the upper sequence (PF1) of the Luearca Formation: A, C, PPL (x 2.5); B, XPL (x 2.5). AFM diagram plot of the study sample (Luearca Slate, Jola quarry). Samples from other sites of the Truchas domain (Casaio and Rozadais formations) and from the Schist-greywacke Complex (South Central-Iberian Zone) are shown for comparison. Mineral and pelite compositional fields taken from Pattison (2001) and Miyashiro (1994).

A= Al_2O_3 ; F= Fe_2O_3 ; M= MgO .

The typical chemical composition of the Luarca Formation in the Alto Bierzo area consists of 55.61% SiO₂, 23.52% Al₂O₃, 7.49% Fe₂O₃ and 1.63% MgO. This composition was measured from a sample of fine-grained slate. The analyses were undertaken on bulk-rock samples by the technical services of the Oviedo University. A projection on the AFM diagram of a representative sample of the fine-grained slate facies is also shown in Figure 3 and is compared with aluminous and common pelite fields (Pattison, 2001). The Luarca sample from the Jola locality can be classified as a high to very high-Al₂O₃ pelite-derived slates. Slates from other southern zones of the Variscan massif, such as the Truchas domain's: Casaio and Rozadais formations, are similar in composition to the studied sample, although their Al₂O₃ content is lower and its Mg/(Mg+Fe) ratio is slightly higher. The comparison with other slates and shales such as those derived from the Cambrian Schist-Greywacke Complex (Central-Iberian Zone) shows differences in the Al₂O₃ content (higher in the Luarca formation samples) and also in the MgO/(MgO+Fe₂O₃) ratio (lower in the Luarca Fm). The higher Al₂O₃ and lower Na₂O contents of the Luarca slates at the study zone (Jola quarry) increases the chemical index of alteration ($CIA = (100 \times Al_2O_3) / (Al_2O_3 + CaO + Na_2O + K_2O)$) to values of 81, significantly higher than other slates and shales (Truchas domain and the published samples from the Schist-Greywacke Complex).

TECTONICS

In this small area of the Alto Bierzo, a metric scale antiformal structure has been identified. It consists of asymmetrical and angular folds belonging to a larger antiform fold system. The axial plane dips to the south-west, and these folds are related to the first Variscan tectonic phase (D1), with a well-developed axial plane cleavage (S1=Sp). Sp shows a strike of 30° and dips 70° towards NE; while the sedimentary bedding (So) has a strike and dip of 240/40° towards SW. The main hinge orientation 120/20° plunging to the southeast. The main structures described are folds, faults and veins.

These features show different deformation styles, brittle and ductile, in accordance with the different rheological behaviour of the rocks in the studied sequence. In the upper sequence, brittle conditions were dominant, whereas down the sequence in the finer grained slate layers, more ductile conditions were evident. Some ductile deformation areas are spatially related to brittle cataclasites, and are related to the ductile deformation during the D2 phase. These structures appear to be related with reverse faults. Later structures include kink-bands developed during the latest phases of the Variscan orogeny deformation.

Characteristic features of the area are the following:

Folds

The main tectonic structures are meso-scale (average wavelength between 10 and 20 m) fold trains related to D1 during the Variscan orogeny, which are better observed in the sandy layers. The fold geometries are symmetric in section and their axial trace strike and dip is 120/20° towards NE. The axis plunge to the east, and the attitude of the fold is upright (S1 dip direction and dip is 30°/75°). These measurements were made in the northern limb of the Bildeo Mountain antiform (Fig. 1). The interlimb angle is 140°, which is classified as gentle. Towards the south, geometry variations were observed in the folds, which can be regarded as close- fold type. The southern limb of this S1 antiform shows similar orientation (dip direction and dip 34°/68°).

From the geological cross section (Fig. 2) it becomes apparent that folds are slightly disharmonic with adjacent beds having different wavelengths. Towards the north, an “M” fold group is recorded, indicating that it is the anticline core and to the south a “Z” shaped fold group occur. During the latest phases of deformation, kink-bands and crenulations were generated with direction and dip inclination of $358^{\circ}/70^{\circ}$.

Faults

Several groups of faults were recorded as normal and reverse faults. Normal faults form step faults with little throw in the northern area of the geological cross section related to D1, whereas in the southern area minor reverse faults not represented in Fig. 2 are dominant, depicting small slips and showing cataclases. They are related to D2. Main strike of fault planes is 100° and the dip of the fault is 80° N in the normal faults, and 80° S in the reverse faults, although one fault related with “Z” folds has a dip of 55° N. Gouge material was found in some faults with width varying from several centimetres to up 70 cm.

Quartz veins

Three main types of extensional quartz veins have been recognized during field observations. They are related with the mobilization of quartz during D1 folding of the slate sequences; with depth, they tend to disappear due to the fine grain lithology:

- Millimeter-wide veins filled with quartz are perpendicular to the axis of main D1 folds and present high continuity. They are recorded as dip joints and may be related with the direction of the maximum stress during folding in the Variscan Orogeny.
- Vertical quartz veins show a pattern oriented as strike joints. Widths vary between 10 and 30 cm with high continuity and filled with quartz and sulphide minerals. This type of vein is concentrated in the hinge area of the fold where extension is greatest during folding ($30/90^{\circ}$). Small displacements have been observed when they cross-cut the type 1 veins.
- Veins of quartz and other minerals in widths up to 10 cm with “Z” shaped occur in the main syncline area of the central part of the rock cutting in the NE-SW section.

CONCLUSIONS

Although there is an excellent knowledge of the WALZ geology due to the number of studies undertaken during the eighties and nineties of the last century, this is the first time that the main stratigraphic sequence in the Bildeo Mountain and its distinct fossil content, has been described. More specifically, it is the first time that the uppermost beds of the Luarca Formation at the Alto Bierzo area are described in detail, which should be useful to develop the mining activities that are planned in this area.

The stratigraphy shows differences in slate lithology between layers in this sequence, and the structure consists of a fold system affected by faults. In the Bildeo Mountain area, a reverse-graded slate sequence from the Middle Ordovician was folded in an antiform during Variscan times. The main metamorphic mineral index observed is chlorite, formed at a temperature of $317 \pm 41^{\circ}\text{C}$ following Xie et al. (1997). The metamorphic conditions range from $T \cong 200\text{--}430^{\circ}\text{C}$ and $P \cong 2\text{--}3$ Kbar. The geochemistry of the studied slates shows distinctive elements when compared with slates from other geological zones, such as the Truchas domain.

The folds correspond to minor features in the northeastern limb of the Vega de Espinareda Synform, attributed to the first Variscan deformation phase. Several tectonic features, like minor folds, faults and quartz veins, mainly related to the first and second phases of deformation, were recorded during fieldwork at the antiform. Its vein infilling corresponds to segregated quartz from the fine-grained slate beds. Normal and reverse faults were described. The reverse ones are filled with cataclasites and are related to D2. The geological cross-section shows the anticline and the movements of the associated faults in the Bildeo Mountain. The improved knowledge of the Ordovician stratigraphy and the tectonic history will assist the planning of new mining activities and profits could increase when mining operations focus on the most productive layers within the slate units of the Lueza formation.

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ORDOVICIAN VS. "CAMBRIAN" ICHNOFOSSILS IN THE ARMORICAN QUARTZITE OF CENTRAL PORTUGAL

A.A. Sá^{1,4}, J.C. Gutiérrez-Marco², J.M. Piçarra^{3,4}, D.C. García-Bellido², N. Vaz^{1,4} and G.F. Aceñolaza⁵

¹ Departamento de Geologia, Universidade de Trás-os-Montes e Alto Douro, Apartado 1013, 5001-801 Vila Real, Portugal.
asa@utad.pt, nunovaz@utad.pt.

² Instituto de Geociencias (CSIC-UCM), Facultad de Ciencias Geológicas, José Antonio Novais 2, 28040 Madrid, Spain.
jcgrapto@geo.ucm.es, diego.gbc@geo.ucm.es.

³ Laboratório Nacional de Energia e Geologia (LNEG), Ap. 104, 7801-902 Beja, Portugal. jose.picarra@lneg.pt.

⁴ Centro de Geociências da Universidade de Coimbra, FCTUC, Largo Marquês de Pombal, 3000-272 Coimbra, Portugal.

⁵ Instituto Superior de Correlación Geológica (UNT-CONICET), Miguel Lillo 205, 4000 Tucumán, Argentina.
acecha@webmail.unt.edu.ar

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INTRODUCTION

The Armorican Quartzite is one of the most characteristic units of the Paleozoic of SW Europe, being represented in the Lower Ordovician succession of Brittany and Normandy (western France), and also over most of the Hesperian and Iberian massifs of the Iberian Peninsula (in the clarified sense of San José, 2006), with the exception of the Ossa-Morena and South-Portuguese zones (Gutiérrez-Marco et al., 2002; Vera, 2004; Ribeiro, 2006). In Portugal and from north to south, the Armorican Quartzite facies is equivalent to the Marão Formation of Trás-os-Montes (Sá et al., 2005), the Santa Justa Formation of the Tabagón-Valongo-Tamames domain (Romano and Diggens, 1974), the Armorican Quartzite Formation of the Buçaco and Amêndoa-Mação areas (Young, 1988; Romão, 2000a) and the Serra do Brejo Formation in the Dornes area (Cooper, 1980). In spite of the generalized absence of biostratigraphical ties for correlation other than ichnofossils and a few chitinozoan or graptolite data, the latter generally coming from the overlying shales, the Armorican Quartzite in Portugal have been considered as involving a diachronism in sedimentation from Arenig to Llandeilo, becoming younger from west to east (Ribeiro, 1974) according to regional data from the Valongo to Trás-os-Montes areas. These data have been compiled in some syntheses (Hammann et al., 1982; Romano, 1982; Oliveira et al., 1992). However, the single paleontological argument in support of such diachronism, a Llandeilian trilobite found in the middle part of the Marão Formation at Moncorvo (Teixeira and Rebelo, 1976) was later reviewed by Gutiérrez-Marco et al. (1995), and Sá et al. (2003, 2009), who demonstrated that the supposed trilobite was in reality the trace fossil *Rusophycus carleyi* (James), also recorded in other Gondwanan areas within the Arenigian succession (Seilacher, 1970; Gibb et al., 2010). No other authors were able to demonstrate the claimed diachronism in the sedimentation of the Armorican Quartzite, whose deposit took place entirely in the *Eremochitina brevis* chitinozoan biozone (Paris, 1981, 1990; Paris et al., 1982, 2007), regarded as

“early-mid Arenigian” or as late Floian according to the global scale (Paris et al., 2007; Videt et al., 2010).

Romão et al. (2010) recently questioned the current age of the Armorican Quartzite in the southern Central Iberian Zone, and supported a local late Cambrian age for this formation in the Amêndoa-Carvoeiro synform based in a couple of ichnological data, a single U-Pb dating, and some highly speculative tectonostratigraphic inferences which in our opinion are far from being demonstrated. Also with reference to this area, Romão et al. (2010) envisaged the Armorican Quartzite as a highly diachronic late Cambrian to Early Ordovician unit for the Iberian Peninsula. This statement is refuted here with the presentation of new ichnologic evidence that supports the previous Early Ordovician dating of the Armorican Quartzite in the Amêndoa-Carvoeiro synform.

ICHNOFOSSIL DATA

Romão et al. (2010) mentioned the occurrence of *Cruziana* cf. *ománica* (sic) and *Cruziana? barbatarugosa* (sic) in the base of the Armorican Quartzite of the Amêndoa-Carvoeiro synform, located in the southern Central Iberian Zone (Fig. 1). With reference to the data of Seilacher (2007), they believe that these ichnospecies will support a late Cambrian depositional age for the Armorican Quartzite in the studied area. As the unit is dated as Arenigian in other parts of central and northern Iberia, Romão et al. (2010) concluded that the Armorican Quartzite represents a diachronous facies “between Upper Cambrian and Arenig from SW to NE across the Iberian Terranes (...) consistent with a foreland to the NE, in the basement of the Cantabrian Zone”.

According with the *Cruziana* stratigraphy for Paleozoic sandstones presented by Seilacher (1970, 1992, 1994, 2007), *Cruziana omanica* Seilacher 1970 is a late Cambrian form characterized by endopodal scratches reflecting a trifold leg with a stronger claw in the middle. *Cruziana barbata* Seilacher, 1970 and *C. rugosa* d’Orbigny, 1839 are two different ichnospecies, being the first exclusive of middle Cambrian beds (Seilacher, 1970, 2007) and the second ranging from Early to Upper Ordovician strata (Seilacher, 2007; Egenhoff et al., 2007). So far, no transitional specimens between *C. barbata* (middle Cambrian) and *C. rugosa* (Ordovician) have been described in the ichnological literature.

The ichnotaxa identified by Romão et al. (2010), although presented in open nomenclature, are clearly insufficient to confirm a particular age of the Portuguese occurrence. The recorded samples may correspond respectively to a *Cruziana* isp. (poorly preserved?) that may resemble *C. omanica*, and a “transitional” *Cruziana?* isp. (perhaps a single lobe of the trace?) between *C. barbata* and *C. rugosa*. The first identification, even if cautionary, is very strange, because the peculiar tricuspidate claw formula of *C. omanica* serves to recognize it even with a few isolated scratches. On the other hand, the single ichnotaxobase shared in common by *C. barbata* and *C. rugosa* is the existence of transverse markings to the bilobed trace, formed by the front legs of the tracemaker when digging in a procline position (Seilacher, 1970, 2007), which may be confused in eroded specimens.

In order to obtain an independent confirmation of the ichnological age given by Romão et al. (2010), our fieldwork in the area lead to the discovery of a very prolific ichnofossil site located in the Armorican Quartzite of the northern flank of the Amêndoa-Carvoeiro synform (Fig. 1). The fossiliferous outcrop is situated in the northern side of route EN 224 southeast of Amêndoa (GPS WGS84 coordinates N 39° 39' 15.52", W 8° 03' 48.78"), and was mapped as belonging to the upper part of the Armorican Quartzite by Romão (2000a, 2000b, 2006). In the laterally continuous quartzite beds, dipping south, we have collected abundant ichnofossils arranged in parallel to the bedding plane (ichnogenera *Cruziana*, *Monomorphichnus*,

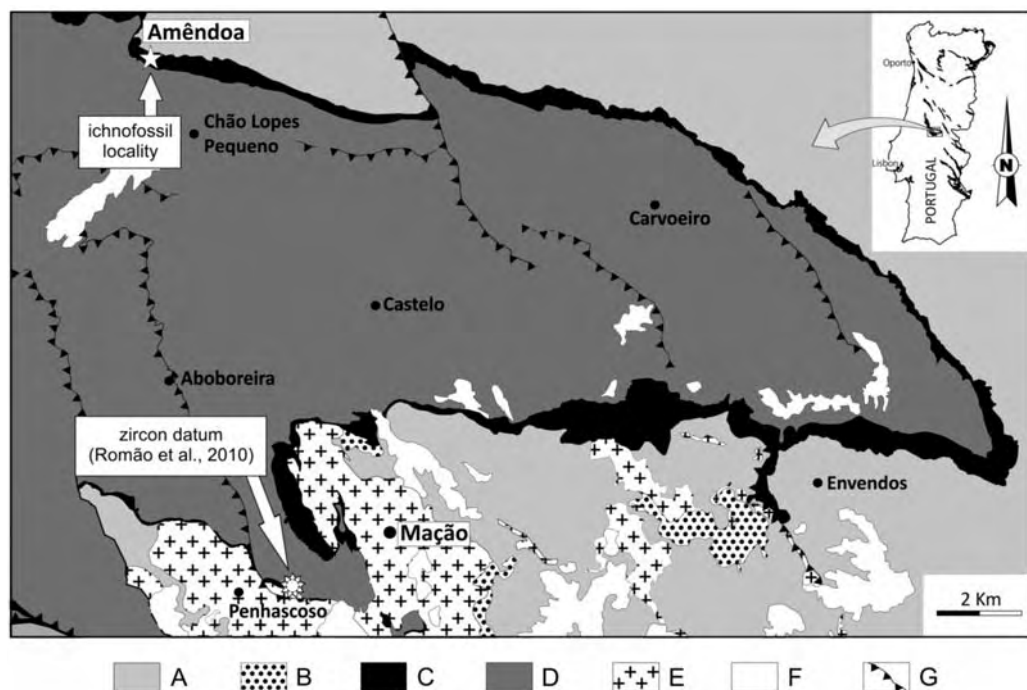


Figure 1. Geological sketch map of the Amêndoa-Carvoeiro syncline (adapted from Romão, 2000b) showing the position of the studied ichnofossil locality bearing *Cruziana imbricata* Seilacher (upper left) and the zircon sample on the Mação-Penhascoso laccolith yielding an Ordovician age (lower left). Inset map shows the location of the area in central Portugal. A, pre-Ordovician basement (Beiras Group); B, Vale do Grou Group (Tremadocian); C, Armorican Quartzite (Floian); D, Other Paleozoic (Darrivillan to Lochkovian) sediments; E, Paleozoic granites (Mação-Penhascoso laccolith); F, Post-Paleozoic cover; G, Traces of the main Variscan thrusts.

Rusophycus, *Arthropycus*, *Crossopodia*, *Protovirgularia*, *Palaeophycus*, *Planolites*, *Phycodes* and *Teichichnus*) as well as others oriented at high to perpendicular angles to it (*Lockeia*, *Lingulichnites*, *Arenicolites*, *Monocraterion*, *Skolithos*, *Daedalus*). A large part of this assemblage is represented in the Armorican Quartzite of central and northern Portugal (Cooper, 1980; Romano, 1991; Neto de Carvalho et al., 2003; Sá, 2005; Sá et al., 2003, 2006, 2007; Neto de Carvalho, 2006), which has been known since Delgado (1886, 1887, 1904).

Among the frequent traces of the *Cruziana rugosa* group, we have collected a well-preserved specimen of *Cruziana imbricata* Seilacher, 1970, a typical Arenigian form previously recorded in Portugal from the Armorican quartzite of the southeastern end of the Vilha Velha do Ródão syncline (locality Serra de São Miguel in Delgado, 1886, pl. 34, fig. 1-3: specimens of "*Rysophycus* cfr. *Rouaulti*, Lebesc." partly redrawn by Seilacher, 1970, fig. 7-14). Our material (Fig. 2) is a well preserved arched "bathtub" variant lacking the characteristic prominent endopodal scratches typical of the ichnogenus, which in this ichnospecies are replaced by scale-like "segments" shingling towards the front end, the anteriormost having a lobate aspect. These front-leg markings are difficult to interpret in terms of the digging action by the tracemaker. Seilacher (2007, p. 194) cannot refer them to flaplike appendages resembling the abdominal legs of chelicerates, because their shingling is not orientated towards the narrower rear part, as usually occurs in

these arthropods. However, on the right lobe of our non eroded specimen, three of the combed dig-marks show indications of up to six very faint rounded scratches—only noticeable with very low angle light—, which suggest a producer more related with the *Cruziana* tracemakers. In the same sense, Neto de Carvalho (2006, p. 257) cited a possible gradational specimen between *C. rugosa* and *C. imbricata* coming from the Armorican Quartzite at Penha Garcia (Penha Garcia-Monfragüe syncline), but apparently confused the imbricate seleniform dig-marks of *C. imbricata* with the true scratches.

Romano (1991) situated the Central Iberian occurrences of *Cruziana? imbricata* (sic) in the upper part of the Armorican Quartzite both in Portugal and Spain. Sá et al. (2006) added the occurrence of *C. cf. imbricata* in the Santa Justa Formation at Arouca. The single record of *C.? cf. imbricata* reported from Salamanca by Pickerill et al. (1984, fig. 2d) probably doesn't belong to the ichnospecies but resembles an ill-defined specimen of *Monomorphichnus* or *Rusophycus*. Outside the Central Iberian Zone, *C. imbricata* has also been reported from the Armorican Quartzite of the Iberian Cordillera (Kolb and Wolf, 1979).

According to Seilacher (1990, 1992, 1994, 2007), *Cruziana imbricata* is a typical Lower Ordovician ichnospecies on Ibero-Armorica and north Africa, and their record in the Amêndoa-Carvoeiro synform also matches with the ichnological data presented by Cooper (1980) from the Serra do Brejo Formation, a lateral equivalent of the Armorican Quartzite in the close paleogeographic vicinity of the Dornes area. Both occurrences contradict the existence of a late Cambrian *Cruziana* assemblage in the studied area, as stated by Romão et al. (2010), and agrees with the Arenigian age of the Armorican Quartzite as is being currently considered in SW Europe based on a number of geological and paleontological evidences.

THE ARMORICAN QUARTZITE AND THE ORDOVICIAN MAGMATISM

The available biostratigraphic information has demonstrated that the deposition of the Armorican Quartzite took place entirely within the Early Ordovician (essentially during the Floian) over a large area of NW Europe. In the northern Central Iberian Zone, the unit immediately postdates an important Cambrian–Early Ordovician magmatic activity (“Ollo de Sapo” belt and related rocks: Díez-Montes et al. 2010; Navidad and Castiñeiras, 2011 and references therein). This makes the Ordovician magmatism older than in other Iberian places outside the depositional area of the Armorican Quartzite facies (Casas et al., 2011).

Modern geochronometric dating of the upper part of the Armorican Quartzite in northern Iberia revealed the presence of age clusters in detrital zircons ranging between 550–800 and 2500–2800 Ma (Fernández-Suárez et al., 2002), as well as a 477.47 ± 0.93 Ma age from magmatic zircons in a single K-bentonite bed (Gutiérrez-Alonso et al., 2007), the latter establishing an absolute minimum age for the rifting that led to the opening of the Rheic Ocean in this peri-Gondwanan section. An age of 470.1–474.6 Ma was recently obtained for magmatic zircons in the volcanoclastic Ordovician unit underlying the Armorican Quartzite in northern Portugal (Gomes et al., 2009), roughly coincident with some imprecise U-Pb zircon and Rb/Sr whole rock dating from similar units in the western Armorican Massif (Bonjour et al., 1988; Bonjour and Odin, 1989).

In the southwestern border of the Central Iberian Zone, Romão et al. (2010) report the existence of the Mação-Penhascoso microgranite, which intrudes in the lower portion of the Armorican Quartzite of the southern part of the Amêndoa-Carvoeiro synform (Fig. 1). It is a pre-orogenic tabular body (a laccolith), more than 80 m thick (far from the kilometeric thickness illustrated by Romão et al., 2010, fig. 1B), that also shows intrusive contacts with the pre-Ordovician basement (Beira Group) and with the basal units of

the Armorican Quartzite (Vale do Grou Group), including the two "Sardic" unconformities that separate them. The laccolith grades into a northern rim of subvolcanic rhyolite textures with eruptive breccias, and it was affected, together with the sedimentary Cambrian-Ordovician host rocks, by three phases of Variscan deformation, being the D1 and D2 accompanied by cleavage and the D3 leading to the formation of the Amêndoa-Carvoeiro synform.

According to Romão et al. (2010), the pre-tectonic morphology and contact relationships of the Mação-Penhascoso laccolith contradicts its previous early-Variscan age assignment, dated in 402 ± 15 Ma by the Rb/Sr method after six whole-rock samples (Abranches and Canilho, 1982). The study of an additional sample from a fresh microgranite collected by Romão et al. (2010) below the Armorican Quartzite east of Penhascoso (Fig. 1), provided zircon grains for geochronological studies. Among them, the U-Pb (ID-TIMS) analysis of an individual prismatic zircon resulted in an Ordovician age of "ca 483 Ma" (sic) for this sample, which is considered by these authors as coincident with the age of the laccolith intrusion.

On the basis of the single evidence provided by this zircon datum (the uncertainty range and the number of analyzed zircons were never specified), Romão et al. (2010) hypothesized a sequence of facts leading to the present cartographic expression of the Mação-Penhascoso laccolith. These authors argue that the laccolith is posterior to the sedimentation, the slight "Sardic" tectonism and also the partial erosion that affected the "Upper Cambrian" Vale do Grou Group. And that it intrudes as a "relatively shallow" magmatic body the quartzite and conglomerate beds of the Armorican Quartzite before this unit ends its sedimentation in a coastal environment. As the intrusion of the laccolith was supposedly coeval to the sandy deposition leading to the Armorican Quartzite, which already needed to be of "some thickness and be compacted", the age of this unit should be estimated as "prior to ca 483 Ma (base of Tremadoc and Upper Cambrian)".

Following the sequential model of Romão et al. (2010), the Armorican Quartzite was "bent by the intrusion" of the mushroom-shaped granitic laccolith before completing its sedimentation. This fact supposedly generates a rising area above the intrusion, composed of poorly consolidated strata, which

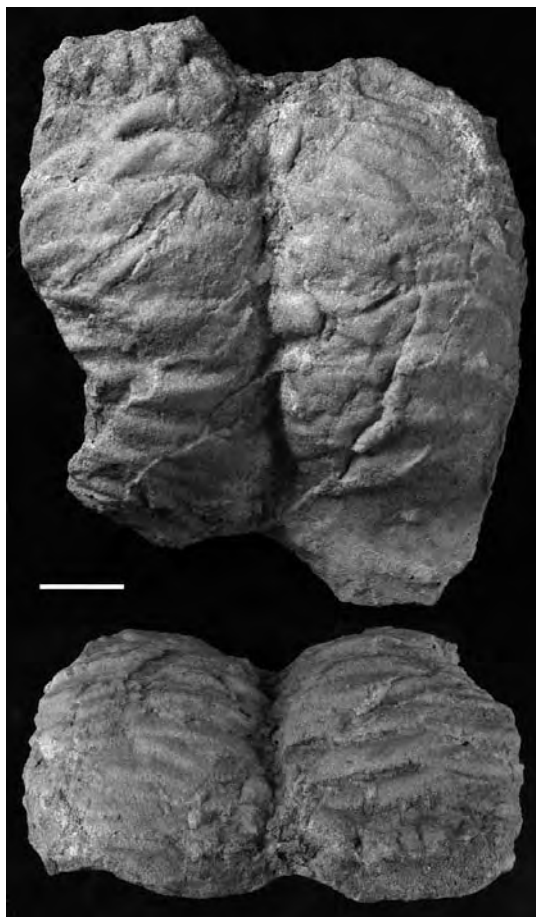


Figure 2. *Cruziana imbricata* Seilacher, 1970. A well preserved specimen from the upper part of the Armorican Quartzite at Amêndoa, in lower (above) and oblique-anterior views. Scale bar=1 cm.

were “rapidly eroded by marine erosion” or “may eventually emerge”. This complex process may explain, in their opinion, the variations in thickness of the Armorican Quartzite (from 0-15 m above the laccolith to about 40-50 m around it), and furthermore, that the microgranite shows in some places an erosive contact with the Middle Ordovician shales of the Brejo Fundeiro Formation, in absence of any thermometamorphic effect in the latter unit.

The highly speculative hypothesis of Romão et al. (2010) for the Mação-Penhascoso granite implies the assumption, not only of a surprising dating for the Armorican Quartzite in a really complicated intrusive setting, but also an age-readjustment for the remaining units and the angular unconformities involved in the area. Thus, the improperly-considered “major Sardic s.l.” phase on the Central Iberian Zone is dated by Romão et al. (2010) as “intra-Cambrian (base of the Upper Cambrian to Middle Cambrian?)”, when the same Toledanian Unconformity in Central Iberia is simply described as pre-Ordovician (Gutiérrez-Marco et al., 2002). However, the true Sardic Unconformity in Sardinia and the Pyrenees is formed by the angular contact between middle Berounian rocks over a Cambrian to middle Arenigian basement (Leone et al., 2002, and references therein). A second consideration about the diachronism of the Armorican Quartzite from SW to NE across Iberia, is that Romão et al. (2010) situated its equivalent in the Ossa-Morena Zone between middle Cambrian sediments and below the Barrancos Formation, but this particular facies has never been described from this zone. Third, a shallow emplacement of the laccolith during the sedimentation of the Armorican Quartzite in a marine environment necessarily would involve a phreatomagmatic volcanism, similar to the one observed in the Armorican Quartzite of the Cantabrian Zone, because the homogeneous and well calibrated sands that typify the Armorican facies could have remained porose and uncemented during millions of years. Fourth, the pair formed by the Lower Ordovician Armorican Quartzite (dated by *Cruziana imbricata* and the remaining traces) plus the Middle Ordovician fossiliferous shales are the same in the Amêndoa-Carvoeiro synform as in other Central Iberian Portuguese areas located further north (Buçaco, Valongo, Moncorvo). The interpretation of a local uplift followed by the partial erosion of this unit and the laccolith, if occurred before the deposition of the Brejo Fundeiro Formation, should involve an hiatus only explained by the authors at a local scale, i.e. between an Armorican Quartzite (of upper Cambrian-basal Tremadocian age for these authors) and the lower Oretanian shales. In the northern part of the Amêndoa-Carvoeiro synform, for instance, the sequence from the Armorican Quartzite to the Brejo Fundeiro Formation seems to be continuous, without any evidence of such long hiatus in sedimentation as if the lower unit were really older than in other Portuguese sections. In fifth and last place, the existence of an erosive contact between the base of the Brejo Fundeiro Formation and the Ordovician granites, as well as an eroded top for the Armorican Quartzite, needs to be adequately demonstrated through detailed sedimentological studies, lacking at present. In our opinion the tectonic nature of these contacts cannot be disregarded.

FINAL REMARKS

The discovery of *Cruziana imbricata* in the Armorican Quartzite of the northern part of the Amêndoa-Carvoeiro synform supports an Early Ordovician age for the unit, instead of the late Cambrian age suggested by Romão et al. (2010) through some misidentified trace fossils and from an absolute Tremadocian age derived from a tabular microgranite intruding the basal part of the quartzite. Both data have introduced a highly speculative model about the supposedly notorious diachronism of the Armorican Quartzite at an Iberian scale, independently regarded as of uniform age, indicated by the widespread Mid

Ordovician shales that overlie the quartzites. The uplifting of the area during the sedimentation of the Armorican Quartzite and the erosive processes that affected this unit and to the granitic laccolith before the Middle Ordovician are very unlikely and need a detailed demonstration.

The geochronometric Ordovician age of "ca 483 Ma" (sic) derived from a single zircon sample on the granite laccolith may relate its intrusion with the generalized Early Ordovician magmatism recorded in the northern Central Iberian Zone prior or partly simultaneous with the deposition of the Armorican Quartzite in some areas. Interestingly, the laccolith intrusion followed the tectonism that affected the Early Ordovician successions below the Armorican Quartzite in many places of the southern Central Iberian Zone, and whose volcanic influence is substantiated in Spain as well as Portugal: Vale de Bojas and Eucísia formations in Trás-os-Montes (Sá et al., 2005), "Montalto unit" of Valongo anticline (considered as "Proterozoic/Cambrian?" by Couto, 1993) or Serra Gorda Formation of Penha García Syncline (Sequeira, 1993). In any case, the present data are inconclusive until more analyses of magmatic zircons and the isotopic signatures of the Mação-Penhascoso microgranite body are completed. For the moment, the more conservative working hypothesis seems to relate this magmatism and their subvolcanic textures with the same volcanic arc affinity of the Early Ordovician Ollo de Sapo magmatic rocks in the northern part of the Central Iberian Zone.

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ORDOVICIAN GEOSITES AS THE BASIS OF THE CREATION OF THE EUROPEAN AND GLOBAL AROUCA GEOPARK (PORTUGAL)

A.A. Sá¹, D. Rocha² and A. Paz²

¹ Departamento de Geologia, Universidade de Trás-os-Montes e Alto Douro, Ap. 1013, 5001-801 Vila Real, Portugal; and Centro de Geociências da Universidade de Coimbra, Largo Marquês de Pombal, 3000-272 Coimbra, Portugal. asa@utad.pt

² AGA – Associação Geoparque Arouca, Rua Alfredo Vaz Pinto, 4540-118 Arouca, Portugal.
drocha@geoparquearouca.com. apaz@geoparquearouca.com

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INTRODUCTION

The scientific work carried out since 2004 in the “Valério’s quarry”, located in the Arouca municipality (NW Portugal, Fig. 1), has highlighted the importance of the Ordovician geological heritage found in this paleontological site and its environs, especially the Middle Ordovician “giant trilobites” (Sá and Gutiérrez-Marco, 2006; Sá et al., 2007, 2008; Gutiérrez-Marco et al., 2009). The construction of the Geological Interpretation Centre of Canelas (GICC), supported by funds from the Ardósias Valério & Figueiredo, Ltd. Co. and by the European LEADER+ programme, was the first step in conserving this heritage and bringing it to a wider audience. As well as the ‘giant trilobites’ Arouca has a rich and diverse geological heritage including the nodular granite of Castanheira (Variscan granite, locally known as “rocks that give birth”) and the Mizarela waterfall, which were recognized nationally as being of great importance for geo-conservation. It was felt to be important to create something which built on and sustained the initial public interest generated in the Middle Ordovician trilobites and added value to the other geo-heritage assets of Arouca. It was this which led towards the creation of a European Geopark, an initiative which was taken forward with considerable local political and public support. The aim was



Figure 1. Location map of the Arouca Geopark (AG).

to highlight the Europe-wide geological importance of Arouca and use it as the basis for the creation and delivery of a local sustainable development strategy, backed-up by a strong management structure and working with others to generate external resources supported by European funding programmes. These objectives were achieved in April 2009, with the confirmation of the Arouca Geopark as a member of both the European Geoparks Network and the Global Geoparks Network, under the auspices of UNESCO.

ORDOVICIAN GEOSITES OF THE AROUCA GEOPARK

In order to generate a more complete knowledge of the geological heritage of the municipality of Arouca, intensive field work was carried out. This work created an inventory, characterisation and evaluation of the geosites of the entire territory (Rocha, 2008; Sá et al., 2008). At the end of this task, 41 geosites had been catalogued including nine geosites associated with Ordovician rocks. In the outcrops of the Santa Justa Formation (Floian, Lower Ordovician) seven geosites were identified:

- i) the "Paiva library" characterised by the upright strata of the "Amorican Quartzite". Local people refer to these formations as 'the books on the library shelves';
- ii) the "Gralheira d'Água" quartzite ridge, in the Santa Justa Formation outcrop in the neighborhood of the "Valério quarry". This is one of the iconic sights of the territory, where it is possible to gain an insight into the entire area and its geology and geomorphology;
- iii) the ichnofossils of the Paiva river valley, located and studied in five different geosites of the territory (Vila Cova, Vilarinho, Cabanas Longas, Meitriz and Mourinha). These outcrops of the Santa Justa Formation quartzites have brought to light exceptionally well-preserved trace fossils of the *Cruziana* and *Skolithos* ichnofacies, assessed as having international significance, under the evaluation method proposed by Brilha (2005).

The "Valério's quarry," and the paleontological collection obtained from there and housed in the Geological Interpretation Centre of Canelas, constitutes another geosite, the interest of which is centred around the slates of the Valongo Formation (Darrivillan, Middle Ordovician) and their trilobite specimens. The location has yielded several thousands of trilobite fossils of 21 different species, including specimens of some asaphids and dikelokephalinids which are the world's biggest trilobite specimens (Sá et al., 2005; Sá and Gutiérrez-Marco, 2006; Gutiérrez-Marco et al., 2009) and have been assessed as having international significance (Rocha, 2008; Rocha et al., 2008). This collection is also composed of fossils of bivalves, rostroconchs, gastropods, cephalopods, brachiopods, crinoids, diploporite cystoids, hyolitids, conularids, ostracods, graptolites and ichnofossils (Sá and Gutiérrez-Marco, 2008; Sá et al., 2008a, 2008b).

The Upper Ordovician is represented in this geosites list by the glaciomarine diamictites of the Sobrido Formation, with its characteristic dropstones, that constitute the evidence in the territory of the late Ordovician glaciation. This geosite was assessed as having national importance (Rocha, 2008; Rocha et al., 2008).

ORDOVICIAN TRILOBITES AND SUSTAINABLE DEVELOPMENT

The establishment of the European Geoparks as a new heritage brand in Europe in 2000 created a new paradigm of conservation for our common geological heritage. Today, geoparks are those special places across the Earth that not only preserve our geoheritage but also use it for the sustainable development of local communities (Mc Keever, 2010). In this sense, in a European and Global Geopark,

alongside the protection and promotion of the geological heritage, one of the main pillars is the idea of all-embracing participation of local communities in the development of the territory. The ultimate goal is that local communities are able to develop, sustain and live in their territory with a sense of ownership for it and that it is developed in ways which assert its regional identity (Eckhardt, 2010). In this sense, the Ordovician geological heritage of the Arouca Geopark, particularly that which is linked to giant trilobites, assumes a role as an engine of the local territorial development strategy. As an example, in 2006 coinciding with the inauguration of the GICC, local people began to sell cinnamon butter cookies in the shape of trilobites (Fig. 2A). These were locally called “Trilobite de Canela” (Canela = cinnamon in Portuguese) making a local pun with “Trilobites de Canelas” (Canelas is the name of the small village where the quarry is located). Later, during the Pre-Conference field-trip of the 4th International Trilobite Conference in June 2008, the Mayor of Arouca, together with the international scientists present, inaugurated a “Monument to the Trilobites”, sited on a roundabout on the access road to Canelas (Fig. 2B). Today this monument is emblematic of the territory encouraged the people of Arouca Geopark to assume a sense of ownership towards the giant trilobites and thereby believe in their importance and the need for their protection. On the same day an exhibition in the Arouca Geopark headquarters with paintings, drawings and sculptures from Radko Šarič (Czech Republic) and Carlos Dias (Portugal), was inaugurated; this has been visited by more than 3.000 people (Fig. 2C). During the school years 2008 and 2009, as part of the celebrations of the International Year of Planet Earth, the teachers of all the primary schools and kindergartens of the Arouca Geopark worked with their pupils on the theme of “Geological heritage of Arouca Geopark”. The result was an exhibition with works focusing mainly on the Middle Ordovician fossils (Fig. 2D) and a carnival parade, notable in which were many trilobite masks (Fig. 2E).

In another initiative, from December 20th, 2008 to March 20th, 2009, different exhibitions comprising sculptures, drawings and aquarelles by Carlos Dias, an artist permanently inspired by and in love with trilobites, were created in twelve restaurants across the area (Fig. 2F). This brief series of exhibitions was enjoyed by 52,440 people (Rocha et al., 2010), and was widely reported on in the national and regional newspapers and by national television.

The image of the trilobite has been used as a signifier for, and visual identity of, the territory at tourism fairs, like BTL (Lisbon) or FITUR (Madrid), (Fig. 2G). These images and the products behind them have proved attractive to tour operators. Since 2009 the Arouca Geopark Association (AGA), the management structure of the Geopark, has also offered a variety of educational programmes, in which 13,102 students and 1,204 teachers have participated to date. Many of these school and college visits were to look at and study the fossils of the Ordovician and to follow the “Route of the Paleozoic,” a walking trail that allows visitors to observe several aspects of stratigraphy and paleontology of rocks of the lower Paleozoic. To date the GICC has received more than 45,000 visitors, confirming the value and importance that our Ordovician geological heritage has for the sustainable development of the Arouca Geopark.

CONCLUSIONS

In successfully managing a European and Global geopark, and in delivering a development strategy that actively includes the local communities, the Arouca Geopark Association has become a good example of this new form of sustainable development based on geological heritage of international significance, actively supported by UNESCO. Through highlighting the importance of the paleontological heritage found in “Valerio’s quarry” and making a case for its conservation, it was possible to improve the awareness of



Figure 2. Sense of ownership about the Ordovician "giant trilobites" in the Arouca European and Global Geopark. A, "Trilobites de canela" = cinnamon butter cookies in the shape of trilobites; B, "Monument to the trilobites", placed on the main roundabout in north of Arouca city; C, Trilobites iron sculptures by Carlos Dias; D, A trilobite viewed by a kindergarten class; E, Trilobites in the schools' carnival parade; F, Trilobite lamps in the restaurant of the Arouca Camping Park; G, Arouca Geopark mascot at the Lisbon Tourism Fair (BTL).

local and national politicians and generate their support for the creation of this Geopark. So, after three years of intense work, with tangible outcomes resulting from many initiatives, the nickname given to the Arouca Geopark by Fortey (2008) – the “Geopark of the giant trilobites” – is seen to convey the essence of the place. Today the municipality of Arouca is nationally known and referred to as “the land of trilobites”, which has grown in importance as a positive signifier for the area and which even overrides well-known elements of Arouca’s rich cultural and architectural heritage. Finally, we believe that the fact that many of the educational activities developed in the territory lead to an increase in knowledge about the rocks and fossils from the Ordovician, in the near future the Arouca Geopark will be recognized internationally as an “Ordovician Territory”, a place where the rich geological heritage of the Ordovician is studied, conserved, showcased and celebrated.

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GRAPTOLOID EVOLUTIONARY RATES: SHARP CONTRAST BETWEEN ORDOVICIAN AND SILURIAN

P.M. Sadler¹ and R.A. Cooper²

¹ Department of Earth Sciences, University of California, Riverside, California 92521. peter.sadler@ucr.edu

² GNS Science, PO Box 30368 Lower Hutt, N.Z. r.cooper@gns.cri.nz

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Recent research (summarised in Calner, 2008; Munnecke et al., 2010) has overturned the traditional view of the Silurian as an interval of warm equitable climate like the Ordovician. Instead, the Silurian is emerging as a period with a highly volatile ocean-atmosphere system (Munnecke et al., 2010). Graptolites provide a rare opportunity to study the changing diversity pattern and evolutionary dynamics of an entire class-level clade through the Ordovician and Silurian in detail. As a planktonic group with relatively short-ranging species of wide geographic distribution, they have been extensively used for biocorrelation and zonation, and the stratigraphic distribution of species is well recorded in the literature. The graptolite clade spans the evolutionary explosion in global marine biodiversity that took place in the Ordovician (GOBE) and one of the five great mass extinction events, at the end of the Ordovician (Sepkoski, 1985; Webby et al., 2004). Here we examine in high-resolution the diversity and species turnover rate time series for the entire graptoloid clade and their relation to these two major faunal revolutions. In particular we contrast the graptoloid evolutionary rate signatures of the Ordovician and Silurian and their possible drivers.

METHOD

A global composite sequence has been built from 2094 species in 518 stratigraphic sections worldwide by the constrained optimisation (CONOP) procedure. The method and database are those used to build the Ordovician and Silurian time scale (Sadler et al., 2009; Gradstein et al., in prep.), which is used herein, and are fully described by Sadler and Cooper (in press). We exclude from the rate estimates 253 species which have a duration of zero in our composite. Mostly they are recorded at one level in one section, are extremely rare, and are therefore most likely to be severely under-sampled. Our data include none-the-less many short-ranging taxa, found in only a few stratigraphic sections, or confined to one region. For this reason, our mean species durations are considerably less than found by Cooper and Sadler (2010) who used a subset of the data with only widespread taxa, for which facies preference could be inferred, and which are likely to have relatively long stratigraphic ranges.

The resulting diversity and rates curves are of high resolution; originations and extinctions are recorded at 2112 levels through the graptolite clade (74.55 myr), averaging about 30 kyr between levels. The rates are effectively instantaneous rates, here smoothed with a moving 1 myr smoothing window. The troublesome biases arising from time-binning the data (Foote, 2000; Alroy, 2008) are therefore avoided. We follow Foote and Miller (2007) in scaling the turnover rates to lineage million years (Lmy) which takes account of both the number of lineages (diversity) and the amount of time they are at risk of extinction. The CONOP output enables us to calculate this precisely. Because CONOP finds and uses the highest and lowest occurrences of taxa worldwide the effects of local sampling incompleteness on global taxon stratigraphic ranges are reduced. Also, by using species rather than higher-order taxa, variance in taxonomic practice is minimised.

RESULTS

The main features of the species richness curve (Fig. 1A) have been discussed by Sadler and Cooper (in press). Briefly, the clade expanded rapidly in the Floian with the diversification of the Dichograptidae, suffered a marked depletion in the late Darriwilian, and a major extinction at the end of the Ordovician. The Hirnantian diversity minimum was severe but very short-lived, and a strong rebound in the Llandovery returned species richness to Ordovician levels but a series of sharp diversity depletions progressively reduced the overall richness of the clade until, by the middle Pridoli, it reached a barely sustainable level. Interestingly, the stratigraphic levels at which glacial events have been inferred (Díaz-Martínez and Grahn, 2007; Lehnert et al., 2010 and references cited therein) and indicated by 'G' in Fig. 1, are marked by depleted species richness.

Mean standing species richness in the Ordovician (53.1 ± 1.3 , standardised to per-unit time) was significantly higher than in the Silurian (29.55 ± 1.1 , including the early Devonian) yet the number of species per myr was lower in the Ordovician (28.83) than in the Silurian (32.39).

This is because Ordovician species lived longer than those of the Silurian. The median duration of Ordovician graptoloid species (1.27 myr) is significantly greater than that of Silurian species (0.69 myr; Mann Whitney-U $p = <0.00$). The higher extinction probability in Silurian graptoloids is reflected in the higher overall species turnover rate. The change in turnover rate pattern starts at about 448 Ma in the Late Katian, rather than at the period boundary (445.13 Ma; Fig. 1B). If we ignore the extreme values at top (Devonian) and base (Tremadoc) of the graptolite clade where diversity is very low, and take the break in pattern at 448 Ma as the boundary between an 'Ordovician' pattern and a 'Silurian' pattern, we can contrast the two 'periods'. For these rate measurements we have re-sampled the composite at 100 kyr intervals to derive a regular per unit time rate. Mean species turnover rate for the Silurian (1.66 species per Lmy) is significantly faster than in the Ordovician (0.86 species per Lmy; $p = <0.000$). Further, it fluctuates much more strongly ($\sigma = 0.72$ for the Silurian and 0.33 for the Ordovician; $p = <0.000$, $df = 1$). The species diversification rate shows a similar pattern to turnover (Fig. 1C). Per Lmy deviation in species richness in the Silurian ($\sigma = 0.45$) is more volatile than in the Ordovician ($\sigma = 0.23$, $p = 0.009$, $df = 1$).

In addition to the well known spike in species turnover in the late Katian and Hirnantian, driven by a mass extinction, turnover rates peak in the Rhuddanian (Rh 2-3), Aeronian (Ae 2), early Sheinwoodian (Sh 1, Ireviken extinction event), lower Homerian (Ho 1; *lundgreni* extinction event), and the early Pridolian where sustained rapid turnover accompanies the demise of the graptoloid clade. In contrast, species turnover rates throughout most of the Ordovician are more uniform and conform with a pattern expected in times of 'background extinction'.

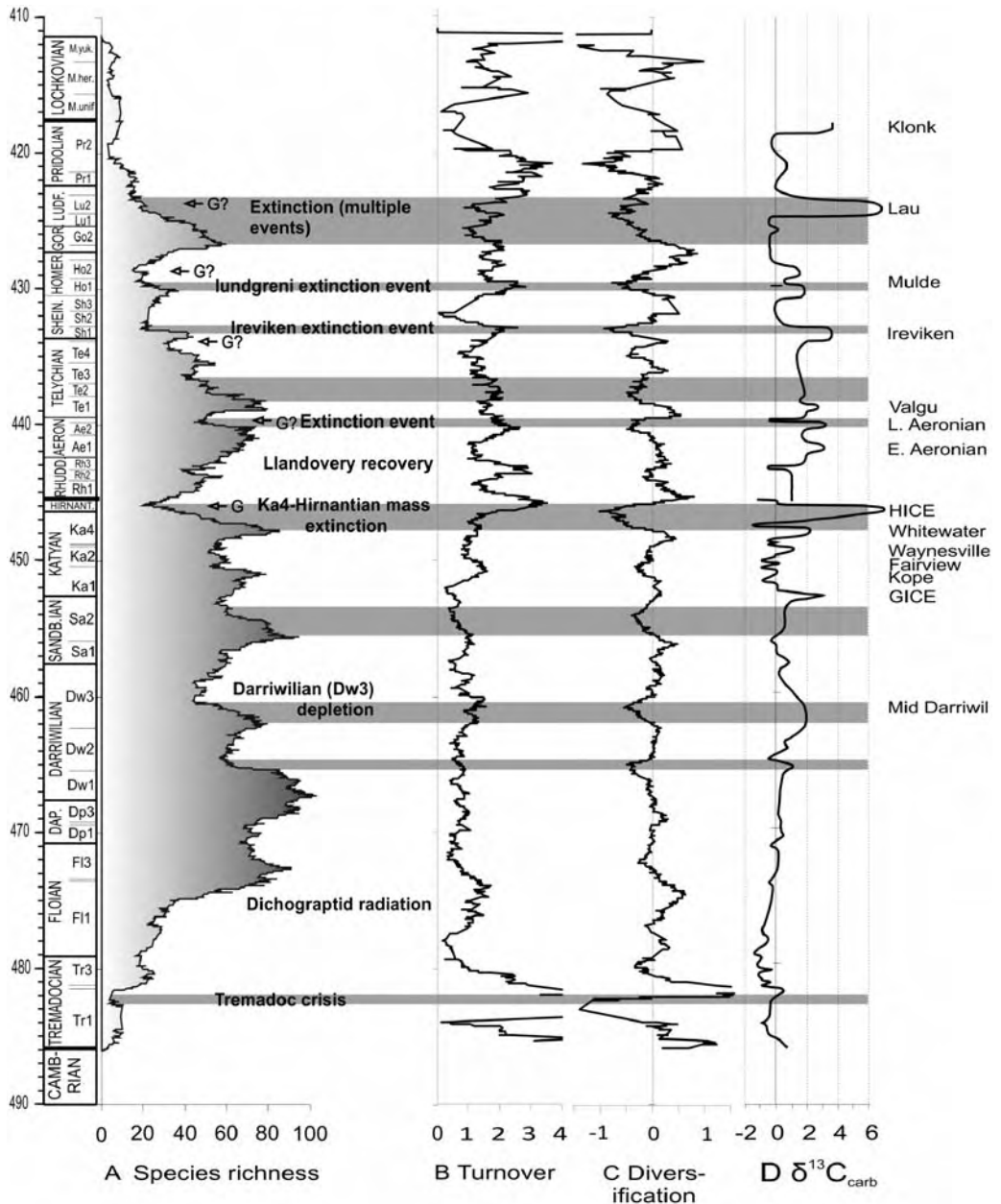


Figure 1. A, species richness of the graptoloid clade and main evolutionary events; B, species turnover (originations+extinctions per Lmy); C, species diversification (originations–extinctions per Lmy); D, $\delta^{13}\text{C}_{\text{carb}}$ curve and main isotopic events. [Carbon isotope curves and 'stage slices' are from Cramer et al. (2010) and Bergström et al. (2009)]. Intervals of sharp diversity loss are shaded. G, glacial event.

The late Katian-Hirnantian mass extinction greatly reduced graptoloid diversity and extinguished many major Ordovician groups, including the DDC (diplograptid-dicranograptid-climacograptid) fauna. There was thus an almost complete turnover across the transition to the Silurian. The question therefore arises, was the higher extinction probability of Silurian graptoloids the result of a new and different taxonomic composition? That is, is extinction probability an intrinsic property of the taxonomic groups themselves, rather than a result of an extrinsic influence such as environmental perturbations for example. To test for this possibility we use the normalograptids, the only group to be present in both the Ordovician and Silurian in substantial numbers. Ordovician normalograptids have a median duration of 1.20 myr, compared with 0.69 myr for the Silurian. Although the smaller sample size ($N = 76$ and 51 respectively) reduces statistical significance ($p = 0.20$) the normalograptids reflect closely the pattern for all taxa, and suggest that the higher extinction probability of Silurian species is unlikely to be an intrinsic property of the taxonomic groups present in the two time periods.

DISCUSSION AND CONCLUSIONS

The high-resolution turnover rate curve (Fig. 1) shows that evolutionary rates constantly changed with time, especially in the Silurian. Although they fluctuated, they are not pulsed (*sensu* Foote, 1994) and peaks do not obviously correlate with stage boundaries.

The generalised delta ^{13}C curves for the Silurian (Cramer et al., 2010) and Ordovician (Bergström et al., 2009) are shown in Figure 1D. The alignment of many of the intervals of sharp diversity loss through the Ordovician and, particularly, the Silurian (grey bands in Fig. 1) with positive excursions of the delta ^{13}C curve can be seen. It is noticeable that the much more strongly fluctuating delta ^{13}C curve in the Silurian is matched by similar volatility in the graptoloid species turnover rate and diversification rate curves (Fig. 1). The change in the carbon isotope pattern takes place in the early Katian raising the suggestion that the graptolite evolutionary rates may also change at this level. Further analysis of the individual extinction and origination rates and isotope data, currently being undertaken, should illuminate this question.

Our graptoloid evolutionary rate curves appear to provide a sensitive proxy for ocean climate and biogeochemistry, probably operating through the microphytoplankton, believed to be the main food resource of graptolites. Our analyses are consistent with a model of strong environmental influence on graptolite evolutionary rates and with a shift in oceanic chemistry, circulation and marine climate patterns, from relatively uniform or gradually changing through most of the Ordovician to fluctuating and more extreme in the latest Ordovician and Silurian. This interpretation agrees with models that predict a greenhouse Ordovician climate, gradually cooling through the Late Ordovician, to become an icehouse climate with strong latitudinal temperature gradients and a fluctuating, unstable ocean-atmosphere system through the late Katian and Hirnantian and, at least intermittently, through the Silurian (Saltzman and Young, 2005; Trotter et al., 2008; Munnecke et al., 2010; Lehnert et al., 2010; Vandenbroucke et al., 2010; Ainsaar et al., 2010)

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A BRIEF SUMMARY OF ORDOVICIAN CONODONT FAUNAS FROM THE IBERIAN PENINSULA

G.N. Sarmiento¹, J.C. Gutiérrez-Marco², R. Rodríguez-Cañero³, A. Martín Algarra³ and
P. Navas-Parejo³

¹ Departamento de Paleontología, Universidad Complutense de Madrid, José Antonio Novais 2, 28040 Madrid, Spain.
gsarmien@geo.ucm.es

² Instituto de Geociencias, Consejo Superior de Investigaciones Científicas-Universidad Complutense de Madrid,
José Antonio Novais 2, 28040 Madrid, Spain. jcgrapto@geo.ucm.es

³ Departamento de Estratigrafía y Paleontología, Facultad de Ciencias, Universidad de Granada, 18071 Granada, Spain.
charor@ugr.es, agustin@ugr.es, png@ugr.es

Keywords: Ordovician, conodonts, biostratigraphy, palaeobiogeography, reworked faunas, Spain, Portugal.

INTRODUCTION

Ordovician conodont studies in the Iberian Peninsula were initiated by Fuganti and Serpagli (1968), who recognized 21 morphospecies included in 15 morphogenera in the Upper Ordovician Urbana Limestone from a single locality in the Central Iberian Zone. Two years later Boersma (in Hartevelt, 1970) identified several morphotaxa in the Upper Ordovician Estana Formation of the Central Pyrenees. In the type section of the Upper Ordovician Cystoid Limestone of the Eastern Iberian Cordillera, Carls (1975) recognised 31 conodont morphotaxa. These pioneer findings were followed by the contributions of Kolb (1978), Hafenrichter (1979), Robert (1980), Robardet (1982) and Sanz (1988), who increased the number of taxa and localities with Katian conodonts, mostly attributed to the *Amorphognathus ordovicicus* Zone.

For twenty years, our knowledge on Ordovician conodonts came only from the single ubiquitous limestone unit that occurs in the upper part of many Iberian successions. Nonetheless, these are predominantly composed of terrigenous rocks (shales, siltstones and sandstones) which were deposited at high Gondwanan paleolatitudes near the South Pole (Gutiérrez-Marco et al., 2002, 2004). Then, some of these clastic deposits (siltstones, shales and storm-induced coquinoïd lenses, sometimes calcareous) were also sampled for conodonts: while siltstones and shales produced only fragmentary specimens, bioclastic beds in tempestites yielded usually fragmentary, but recognisable, elements.

EARLY TO EARLY-MID ORDOVICIAN CONODONTS

The oldest Ordovician conodont assemblage from the Iberian Peninsula was found by Sarmiento and Gutiérrez-Marco (1999) near Adamuz (Fig. 1, loc. 17) in the tectonically-complex boundary area between

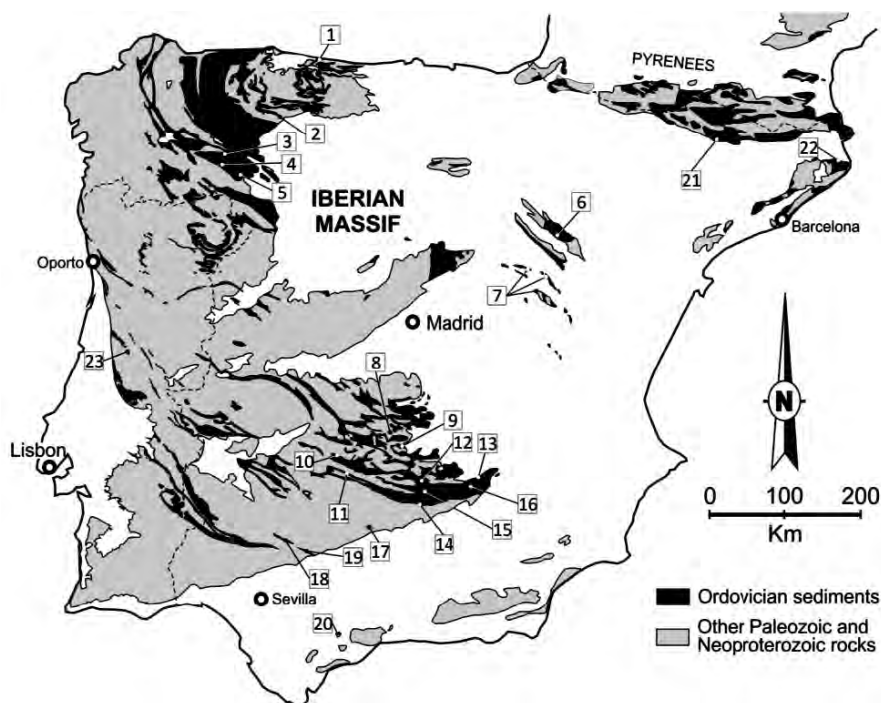


Figure 1. Map of the main Ordovician outcrops in the Iberian Massif, with locations of mentioned conodont localities in Spain and Portugal. 1, Sueve and Ordovician tunnel; 2, Portilla de Luna; 3, La Aquiana; 4, Casaio; 5, Rozadais-Truchas; 6, Fombuena; 7, Aragoncilo, Ojos Negros and Checa areas; 8, Porzuna; 9, Corral de Calatrava; 10, north of Chillón; 11, Alamillo; 12, Calzada de Calatrava; 13, Villamanrique-Terrinches; 14, El Centenillo; 15, Huertezuelas and Viso del Marqués; 16, Aldequemada; 17, Adamuz; 18, Cazalla de la Sierra; 19, Constantina; 20, Ardales; 21, Els Castells; 22, Les Gavarres; 23, Buçaco.

the Central Iberian and Ossa-Morena zones. Conodonts were obtained from a few decalcified (iron-rich) limestone cobbles embedded as olistholits in a Mississippian olisthostrome unit from the Guadalmellato domain, Los Pedroches Basin (Cózar et al., 2004; Armendáriz Dufur, 2009 and references therein). Besides some indeterminate brachiopod and trilobite fragments, the transported limestone contains a diverse and partially reworked conodont assemblage that includes a mixture of Lower and Middle Ordovician taxa. Sarmiento and Gutiérrez-Marco (1999) emphasized the importance of the reworked conodonts to infer the existence of unknown units with Early Ordovician limestones in an area exclusively characterized by siliciclastic deposits at that epoch. The existence of the so-called “phantom formations” (Branson and Mehl, 1940) can be assumed to have occurred in some original areas located westwards, that were tectonically juxtaposed to the southern border of the Central Iberian Zone during the Variscan collision. These would have been partially eroded and resedimented in a typical syn-orogenic foredeep basin during the Mississippian, as can be reconstructed by the presence of a continuous record of Early Ordovician to Devonian fossiliferous olistoliths.

Most of the recovered conodont elements from the older limestone cobbles are poorly preserved, preventing an accurate taxonomic identification. Preliminary studies reveal the presence of *Cordylodus* sp., *Paltodus* cf. *deltifer* (Lindström), *Paltodus* cf. *subaequalis* Pander, *Drepanodus* cf. *arcuatus* Pander, *Drepanodus* spp., *Scolopodus striatus* Pander, *Teridontus* spp., *Acanthodus*? sp., *Acodus* spp.,

Hammannodus sp., *Protopanderodus*? sp., *Gothodus* cf. *costulatus* (Lindström), *Baltoniodus* cf. *triangularis* (Lindström), and *Baltoniodus* sp., beside other coniform elements yet to be identified (Pl. 1, fig. 1-5). The Lower Tremadocian *Paltodus deltiifer* Zone can be inferred by the presence of *Paltodus* cf. *deltifer* (Lindström) and *Hammannodus* sp., but the record of *Cordylodus*? sp. with robust elements of *Acanthodus*? sp. and *Teridontus*? sp. does not exclude the existence of an even older reworked conodont fauna in the assemblage. We propose here a correlation of these “phantom limestone formations” with the Upper Cambrian to Lower Ordovician sequence of the southern Montagne Noire (France), which occasionally incorporates thin limestone beds formed in temperate waters of the Gondwanan margin (Álvaro et al., 2003). This may be confirmed by the Spanish record of *Hammannodus*, a genus so far only known from the Saint Chinian Formation of the southern Montagne Noire (Serpagli et al., 2007).

The younger elements recorded in the mixed Early Ordovician conodont sample from Adamuz may compose an assemblage of elements doubtfully related with *Baltoniodus triangularis* (Lindström), *Baltoniodus* sp., *Drepanodus arcuatus* Pander, *Scolopodus striatus* Pander and possible representatives of the genera *Drepanoistodus* and *Protopanderodus*, among other taxa. This association can be tentatively assigned to the *Baltoniodus triangularis* Zone, broadly representative of Lower Dapingian strata (lowermost Middle Ordovician), which must be simultaneous with the sedimentation of the local limestone also bearing the previously mentioned reworked conodonts.

The only Floian conodonts known from the Iberian Peninsula occur in a lingulid shell-bed located near the top of the Barrios Formation, a local equivalent of the Armorican Quartzite in the Cantabrian Zone of northern Spain (Fig. 1, loc. 1). The assemblage is currently being studied, but preliminary data indicate the presence of coniform elements of the genera *Drepanodus*, *Drepanoistodus* and *Protopanderodus* (Gutiérrez-Marco and Bernárdez, 2003).

MID TO EARLY-LATE ORDOVICIAN CONODONTS

Conodont assemblages representative of the late Darriwilian–early Sandbian interval are assigned to the Dobrotivian stage in Mediterranean regional chronostratigraphy (Gutiérrez-Marco et al., 2008; Bergström et al., 2009), and were reported by Sarmiento et al. (1995a). These occur in thin lenses of calcareous coquinas intercalated in sandy tempestites, or interbedded with quartzite and micaceous shale alternations. All of them belong to a thick siliciclastic group dominated by dark shales and sandy tempestites, broadly known in central Spain as the “Tristani Beds” (San José et al., 1992, with previous references). These have similar counterparts in the Iberian Cordillera and also in the French Armorican Massif, where related conodonts were described by Lindström et al. (1974: age reviewed by Lindström, 1976).

A stratigraphically older conodont assemblage (early Dobrotivian) was recognized in the Central Iberian Zone, as occurring near the top of the El Caño Formation (loc. CC-II), southwest of Calzada de Calatrava (Ciudad Real) (Fig. 1, no. 9). Some other stratigraphically younger conodonts (late Dobrotivian) come from the lower part of the Botella Quartzite in a nearby outcrop (loc. VM-X), and from the La Cierva Quartzite (loc. POR-V) near Porzuna (Ciudad Real) (Fig. 1, no. 8). Similar Dobrotivian conodonts from the Iberian Cordillera have been identified near Fombuena (Zaragoza) in a calcareous intercalation occurring in the middle part of the Sierra Member of the Castillejo Formation (FB-II: Fig. 1, loc. 6).

Most of the conodont taxa recorded from these localities were identified in open nomenclature due the scarcity of elements and their poor preservation. The presence of *Amorphognathus* aff. *inaequalis*

Rhodes, *Amorphognathus* sp., *Baltoniodus* aff. *variabilis* (Bergström), *Coelocerodontus* sp., *Complexodus* sp., *Drepanoistodus* sp., *Icriodella* cf. *praecox* Lindström, Racheboeuf and Henry, *Plectodina* cf. *flexa* Rhodes, among other indeterminate specimens (Pl. 1, figs. 6-16), suggests a biostratigraphical interval below or within the basal part of the *Amorphognathus tvaerensis* Zone. The existence of some reworked conodonts from older conodont biozones occurring in the assemblages from both the Central Iberian and the Eastern Iberian Cordillera localities cannot be ruled out. However, the exact dating of these sequences is firmly established by a combination of graptolite and chitinozoan data, as well by the local trilobite and brachiopod biozones (see Gutiérrez-Marco et al., 2002).

Bedding plane assemblages of lower Dobrotivian conodonts occur in shales from the Sueve Formation of the Cantabrian Zone (Fig. 1, loc. 1). A preliminary identification of these (Gutiérrez-Marco and Bernárdez, 2003) reveals several forms of *Drepanoistodus*, *Panderodus* and *Semiacontiodus*, that are being presently studied.

LATE ORDOVICIAN CONODONTS

The presence of mid Berounian (= late Sandbian to early Katian) conodonts in the Central Iberian Zone was first mentioned by Sarmiento (1993) as coming from an ironstone level at the base of the Cantera Shales near the El Centenillo, Jaén (Fig. 1, no. 14). Only *Amorphognathus* sp., *Panderodus* sp. and *Icriodella* sp. (Pl. 1, figs. 17-18) were identified from this level. This finding allows a chronostratigraphical correlation with equivalent levels of the Piedra del Tormo Member of the Fombuena Formation, Eastern Iberian Cordillera, where Kolb (1978) previously identified and illustrated one fragment of *Icriodella* sp. The mid Berounian age of these conodonts was provided by their association with index species of brachiopod and trilobite biozones.

Plate 1. Some Ordovician conodonts from Spain. 1-3, Early Ordovician reworked specimens from Adamuz; 3-4, Early Mid Ordovician autochthonous assemblage from Adamuz; 6-16, Dobrotivian (late Darriwilian-early Sandbian) beds from Central Iberian localities; 17-18, mid Katian specimens (?*Amorphognathus superbus* Zone) from the Central Iberian Zone; 19-34, mid-late Katian conodonts (*Amorphognathus ordovicianus* Zone) from Central Iberian localities; 35-39, Hirnantian assemblage (upper *A. ordovicianus* Zone) from the Malaguide Complex.— 1, *Teridontus*? sp. ADZ-OI-9875; 2, *Drepanodus*? sp. ADZ-OI-9359; 3, *Cordylodus*? sp. ADZ-OI-9859; 4-5, *Baltoniodus* cf. *triangularis* (Lindström) [4, Pb element ADZ-OI-9374; 5, Pb element ADZ-OI-9375]; 6, *Complexodus*? sp. Pb element POR-V-134A; 7, *Icriodella* aff. *praecox* Lindström, Racheboeuf and Henry. S element POR-V-0834; 8-9, *Amorphognathus* aff. *inaequalis* Rhodes [8, Pa element POR-V-091A; 9, M element POR-V-093A]; 10, *Plectodina* sp. Sa element, CC-II-164A; 11-13, *Plectodina* cf. *flexa* (Rhodes) [11, Pa fragmentary element POR-V-163A; 12, Sa element POR-V-162A; 13, Sc element POR-V-161A]; 14-15, *Baltoniodus* aff. *variabilis* (Bergström) [14, Sa element POR-V-103A; 15, M element POR-V-108A]; 16, *Baltoniodus* sp. Sa element POR-V-104A; 17, *Icriodella* cf. *superba* Rhodes, fragmentary Pa element LC-IV-109; 18, *Panderodus* sp. LC-IV-102; 19-20, *Hamarodus europaeus* (Serpagli) [19, Sc element HZ-IA/6-6567L; 20, M element HZ-IA/7-2735G]; 21, *Scabbardella altipes* (Henningsmoen), a element HZ-IA/6-2547F; 22, *Panderodus gracilis* (Branson and Mehl), ?graciliform element RN-X-6564L; 23-26 and 32, *Sagittodontina robusta* Knüpfers [23, Pb element CO-B/5-2073F; 24, Sb element HZ-IA/6-1708D; 25, indeterminate element HZ-IB/III-1709D; 26, Pb? element HZ-IA/6-1711D; 32, Pb element CT-III/7-1]; 27-30, *Amorphognathus ordovicianus* Branson and Mehl [27, M element HZ-IB/VIII-2743G; 28, Pb element (left) HZ-IA/6-885H; 29, Pb element (right) HZ-IA/6-951H; 30, Sb element HZ-IA/6-965H]; 31, *Eocarniodus gracilis* (Rhodes), CO-A/8-2293F; 33-34, *Istorinus erectus* (Knüpfers) [33, CT-I/1-36; 34, CS-VIII/A-6-7]; 35-39, *Walliserodus amplissimus* (Serpagli), elements in inner and outer lateral views [35, a element O3A-69-58; 36, b element O3A-69-60; 37, c element O3A-69-50; 38, d element O3A-69-40; 39, e element O3A-69-57]. Scale bars=100 µm (figs. 1-34) and 200 µm (figs. 35-39)



Amorphognathus superbus Zone

In the Ossa-Morena Zone, conodonts that are attributed with doubts to the *A. superbus* Zone were obtained from a redeposited horizon occurring in the basal part of the Pelmatozoan Limestone Formation from the northern flank of the Valle syncline, west of Cazalla de la Sierra, Seville (Robardet et al. 1998, Sarmiento et al., 2000c). The conodont assemblage includes few but well preserved elements of *Icriodella* sp., *Amorphognathus* aff. *complicatus* Rhodes, *Plectodina* sp. and *Aphelognathus*? sp.

In the Central Iberian Zone, Del Moral (2004, 2007) described and illustrated conodonts of the *Amorphognathus superbus* Zone coming from the upper beds of the Bancos Mixtos Formation in the sections of Corral de Calatrava, Huertezuelas and Viso del Marqués (Fig. 1, loc. 15). The conodont record includes *Amorphognathus superbus* (Rhodes), *Sagittodontina robusta* Knüpf, *S. cf. robusta*, *Dichodella*? sp., *Icriodella superba* Rhodes and *I. cf. superba*.

Amorphognathus ordovicicus Zone

Pioneering conodont studies developed in Spain before 1990 (see Introduction) were centered in a single Ordovician limestone deposit found throughout the Iberian Peninsula, and that was correctly attributed to the middle to upper Katian (Ka3-4) *Amorphognathus ordovicicus* Zone, and referred to the Kralodvorian stage of the Mediterranean regional scale.

The common assemblage from this Zone in Spain (Pl. 1, figs. 19-34) is represented by *Amorphognathus ordovicicus* Branson and Mehl, *Eocarniodus gracilis* (Rhodes), *Hamarodus europaeus* (Serpagli), *Istorinus erectus* Knüpf, *Panderodus gracilis* (Branson and Mehl), *Sagittodontina robusta* Knüpf, *Scabbardella altipes* (Henningsmoen), and several species identified in open nomenclature belonging to the genera *Drepanoistodus*, *Nordiodus*, *Panderodus*, *Protopanderodus*, *Pseudooneotodus* and *Walliserodus*?

In NW Spain, records of this conodont zone occur in an unnamed limestone near Portilla de Luna (Fig. 1, loc. 2) in the Cantabrian Zone (Del Moral, 2003; Del Moral et al., 2003), as well as in the La Aquiana and Casaio formations of the Ollo de Sapo domain of the northern Central Iberian Zone (Fig. 1, locs. 3-5), also from limestone pebbles redeposited in the Hirnantian shales of the Rozadais Formation (Sarmiento, 1993; Sarmiento et al., 1999).

In the Iberian Cordillera conodonts of the *A. ordovicicus* Zone were identified in the Cystoid Limestone (Fig. 1, loc. 6) and Ojos Negros formations (Fig. 1, loc. 7) by Sarmiento (2002) and Del Moral (2007), as well as from limestone pebbles and dropstones incorporated into the Hirnantian Orea Formation.

In the Central Iberian Zone, the Urbana Formation is a characteristic and widespread limestone unit, in spite of its lenoid character, the scarcity of its outcrops and its reduced and highly variable thickness. More than twenty sections belonging of twelve localities, mainly in the southern Central Iberian Zone (Fig. 1, locs. 9-16), were carefully sampled for conodonts since 1990, which constitutes a very important increase in our knowledge of Katian conodont faunas in the Iberian Peninsula (Sarmiento, 1990, 1993; Sarmiento et al., 2000b; Del Moral, 2002a, 2002b, 2007; Del Moral and Sarmiento, 2008) (Pl. 1, figs. 19-34). In the Portuguese part of the Central Iberian Zone, conodonts from the *A. ordovicicus* Zone have been also recorded in the Poiães Member of the Ferradosa Formation, Serra do Buçaco (Sarmiento et al., 2000a, 2001).

In the southern border of the Central Iberian Zone, a few conodont elements obtained from rare Katian pebbles in a Mississippian olisthostrome from the Adamuz area (Córdoba: Fig. 1, loc. 17) were assigned, with doubts, to the *A. ordovicicus* Zone.

In the Ossa Morena Zone, conodonts indicative of the *A. ordovicicus* Zone have been recorded in the Pelmatozoan Limestone Formation of the Valle and Cerrón del Hornillo synclines, northern Seville province (Fig. 1, loc. 18-19) by Sarmiento (1993) and Sarmiento et al. (2008).

In NE Spain, conodonts of the *A. ordovicicus* Zone have been identified in the Pyrenees (Fig. 1, loc. 21) and in the Catalanian Coastal Ranges (Fig. 1, loc. 22), occurring in the El Baell Formation of the Freser valley (Sanz-López and Sarmiento, 1995) and in the Madremanya limestones of Les Gavarres massif (Sarmiento et al., 1995b), respectively.

All the mentioned Late Ordovician conodont occurrences of the *A. ordovicicus* Zone, with the exception of that from the Pyrenees and others with very low number of specimens, have been ascribed to the Mediterranean Province of the North Atlantic Realm, by the presence of the typical genera *Sagittodontina* and *Istorinus* (Sweet and Bergström, 1984).

Recently, a peculiar conodont fauna dominated by simple cones of *Walliserodus amplissimus* (Serpagli) and *Scabbardella altipes* was found in the Malaguide Complex of the Betic Cordillera (Fig. 1. loc. 20; Pl. 1, figs. 35-39), and referred to the extension of the *A. ordovicicus* Zone into the Hirnantian stage. This conodont association significantly differs in composition from older assemblages (Ka2-3) of the same Zone occurring in the remaining areas of the Iberian Peninsula, but it is very similar to the assemblage recorded from the upper levels of the Uggwa limestones of the Carnic Alps, thus suggesting close palaeogeographical relationships between the Malaguide Complex and the Alps (Rodríguez-Cañero et al., 2010). This Betic fauna represents the youngest Ordovician conodont record in SW Europe and the first Ordovician conodonts found in the Western Mediterranean Alpine Orogen.

CONCLUSIONS

Ordovician conodonts from the Iberian Peninsula account for the following record: early Tremadocian *Paltodus deltifer* Zone, the early Dapingian *Baltoniodus triangularis* Zone, some Floian forms are under study, successive poorly known assemblages from the latest Darriwilian to early Sandbian, and late Sandbian to early Katian intervals, the Katian *Amorphognathus superbus* Zone and the mid-late Katian to Hirnantian *Amorphognathus ordovicicus* Zone. The very sporadic and incomplete record of conodont associations previous to those from the *A. ordovicicus* Zone is common in all areas placed at high latitudes near the south polar margin of the Gondwanan continent. Nevertheless, these occurrences can be used for regional correlations and can provide palaeogeographical inferences for this characteristic domain of the North Atlantic Conodont Realm. CAI data derived from Ordovician conodonts occurring in several places of the Iberian Massif were summarized by Sarmiento and García-López (1996) and Sarmiento et al. (1999).

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THE LATE ORDOVICIAN GLACIAL EVENT IN THE CARNIC ALPS (AUSTRIA)

H.P. Schönlaub¹, A. Ferretti², L. Gaggero³, E. Hammarlund⁴, D.A.T. Harper⁵, K. Histon², H. Priewalder⁶,
C. Spötl⁷ and P. Štorch⁸

¹ Austrian Academy of Science, Center for Geosciences, Dr. Ignaz Seipel-Platz 2, 1010 Vienna, Austria. hp.schoenlaub@aon.at

² Dipartimento di Scienze della Terra, Università di Modena e Reggio Emilia, Largo S. Eufemia 19, 41121 Modena, Italy.
ferretti@unimore.it; hiscat@interfree.it

³ Dipartimento di Studio del Territorio e delle sue Risorse, Università degli Studi di Genova, Corso Europa 26, 16132 Genova, Italy.
gaggero@dipteris.unige.it

⁴ Nordic Center for Earth Evolution (NordCEE), Institute of Biology, University of Southern Denmark Campusvej 55,
5230 Odense M, Denmark. emma@biology.sdu.dk

⁵ Natural History Museum of Denmark (Geological Museum), University of Copenhagen,
Øster Voldgade 5-7, DK-1350 Copenhagen K, Denmark. Dharper@snm.ku.dk

⁶ Geological Survey of Austria, Neulinggasse 19, 1030 Vienna, Austria. Helga.Priewalder@geologie.ac.at

⁷ Department of Geology and Palaeontology, University of Innsbruck, Innrain 52, 6020 Innsbruck, Austria.
Christoph.Spoetl@uibk.ac.at

⁸ Institute of Geology, Academy of Sciences of the Czech Republic, Rozvojova 269, 165 02 Prague 6, Czech Republic.
storch@gli.cas.cz

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INTRODUCTION

The Carnic Alps of Southern Austria and Northern Italy (Fig. 1) represent one of the very few places in the world where an almost continuous biostratigraphically well-constrained succession of Lower Paleozoic rocks is preserved and is as such a key locality along the Northern Gondwana Margin regarding Lower Paleozoic correlation. For example, the world-famous Cellon Section has been utilized as a geographic reference district (RD) for both Silurian conodont correlation studies (Kleffner, 1995) and for the evaluation of Silurian global eustatic changes (Brett et al., 2009; Johnson, 2010) for the North Gondwana region and across the peri-Gondwana Terranes.

However, studies on the Ordovician succession in the Carnic Alps date from the early 1960s to the 1980s and need revision in light of recent research trends (e.g. Bergström et al., 2009; Cramer et al., 2010; Finnegan et al., 2011) to define relationships between the new global series and stages on a regional basis for improving global correlation. The data from current research projects on the Late Ordovician - early Silurian interval of this middle latitude temperate sector are summarised here with regard to identification of global signals in the Carnic Alps.



Figure 1. Main regions of anchizone to lower greenschist metamorphosed fossiliferous Paleozoic strata in the Eastern Alps. Note the Periadriatic Line separating the Carnic Alps and the Karavanke Mountains (Southern Alps) from other Alpine Paleozoic remnants belonging to the Eastern Alps. Enlarged map shows the localities referred to in the text.

LITHOSTRATIGRAPHY

Due to tectonic deformation disparate facies are juxtaposed in close proximity with a series of distinctive paleogeographic/paleoenvironmental settings being represented in each of the different nappes or thrust sheets of the Carnic Alps.

The Middle to Upper Ordovician series of the Central and Western Carnic Alps are divided into a tripartite sequence of rocks with various clastics to volcanoclastics at the base, overlain by a limestone dominated succession with a few meters of sandstones at the top (Fig. 2). Fossils occur in all three levels although the equivalents of the Darriwilian and Sandbian stages have not yet been recognized. In particular, this applies to the fossiliferous Uggwa Shale in the Central and Eastern Carnic Alps from which

a rich brachiopod fauna has been collected in recent years. Mapping in the Western Carnic Alps clearly indicates that the lateral equivalents of this formation are the Himmelberg Sandstone, Fleons Graywacke and Comelico Porphyry. In the Central Carnic Alps, apparently coeval sections at the base of Mount Cellon, Nölblinggraben, Rauchkofel South, Rauchkofel Boden, Oberbuchach, Hoher Trieb and Valbertad are well known; however, most have only been studied to a minor degree. Complementary sections occur in the Uggwa Valley of the Eastern Carnic Alps and at Feistritzgraben in the Western Carnic Alps.

Two major facies associations are displayed in the Late Ordovician of the Central Carnic Alps: massive cystoid-rich limestones (Wolayer Limestone Formation), quartz arenites and graywackes representing the shallow-water environments and shales and bedded wackestones representing more basinal settings (Uggwa Limestone Formation). In deeper water settings the Hirnantian Plöcken Formation, belonging to the *Normalograptus persculptus* graptolite Zone, succeeds the latter. Periglacial deposits which clearly reflect the diamictite nature of part of the Plöcken Fm. thus provide unequivocal evidence of the Hirnantian glaciation in this region. The Bischofalm Quartzite succession represents the lateral equivalent of the above formations. The data presented here focuses on the former three of the sections being studied and a brief overview of the lithological successions for each area is given below.

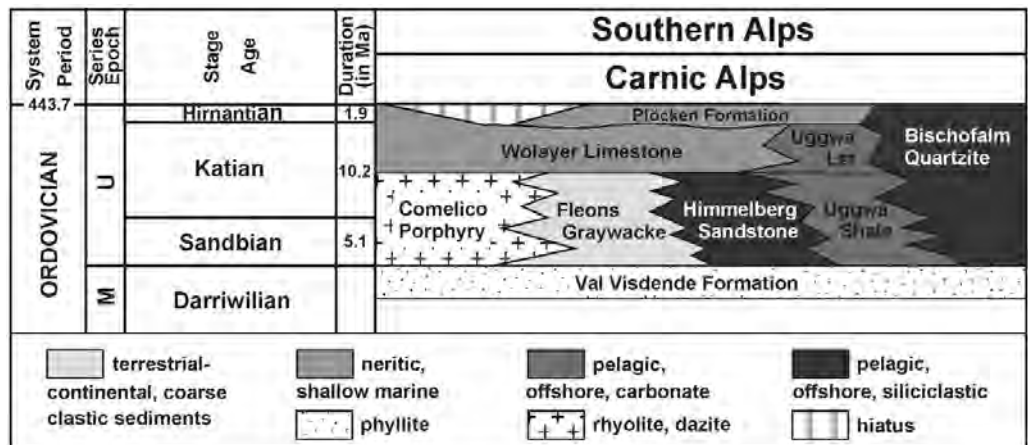


Figure 2. Middle to Upper Ordovician stratigraphy of the Carnic Alps.

Cellon Section

The section is exposed in the Cellon avalanche gully near Plöcken Pass at an altitude of 1500 m, approximately 1 km from the Austrian–Italian border. The succession forms part of the Cellon Nappe and is dominated by shales, siltstones and bedded wackestones representing a more basinal setting (Fig. 3). Sampling (see below) for brachiopods, graptolites, conodonts as well as chemostratigraphy and chronostratigraphy is in progress.

Uggwa Shale. At the base of the Cellon Section the Uggwa Shale attains a thickness of at least 100 m. The greenish to grayish shales mainly comprise claystones to siltstones which grade into the overlying marlstones and argillaceous limestones attributed to the Uggwa Limestone.

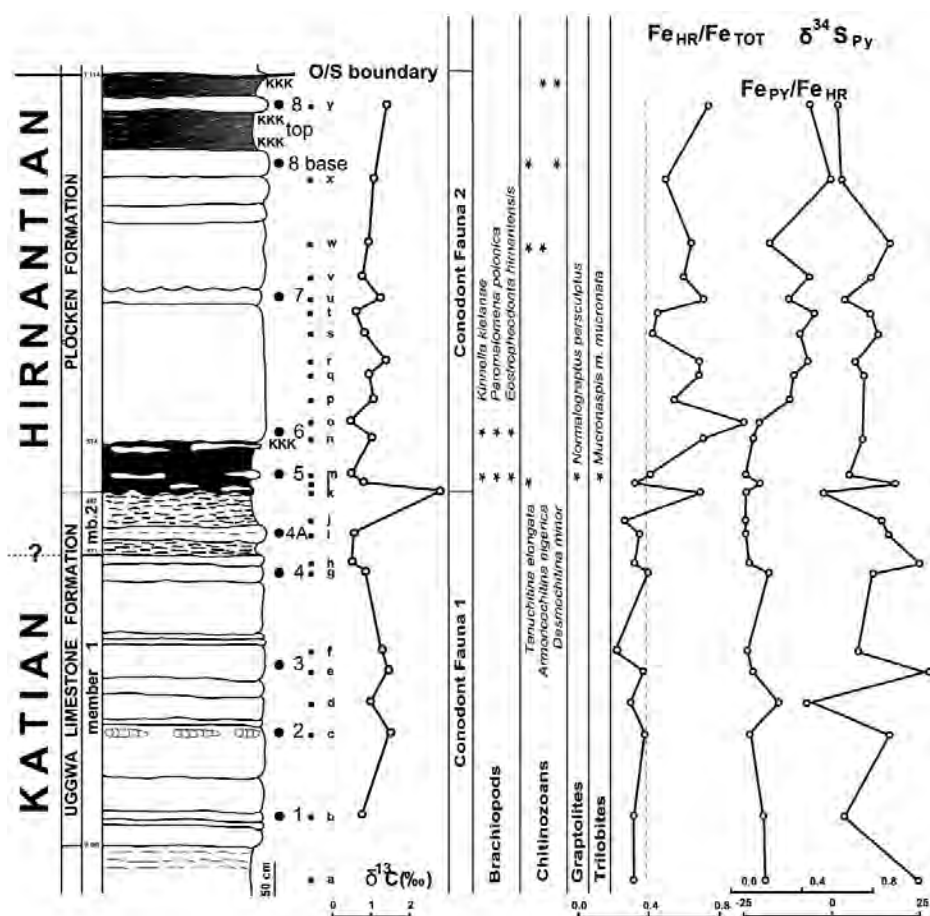


Figure 3. Late Ordovician interval of the Cella Section. Lithostratigraphic column based on new field measurements by HPS and AF, bed numbers mainly after Walliser (1964). Vertical trends in key geochemical parameters (Iron, sulphur and carbon isotopes) across the Hirnantian glacial event are illustrated; letters a-y indicate sampling points. Letters KKK indicate position of K-bentonite levels sampled for radiometric dating. New and revised biostratigraphical data indicate the standard *Normalograptus persculptus* Graptolite Zone, the *Amorphognathus ordovicicus* Conodont Zone, the *Tanuchitina elongata* Chitinozoan Zone and the diagnostic *Hirnantia* brachiopod fauna. Trilobite faunas are also indicative of this interval.

Uggwa Limestone. The bedded continuous wackestone known as the Uggwa Limestone. has a thickness of 4.11 m (Bed nos. 1 – 4: after Walliser, 1964). It is overlain by 20 cm of greenish siltstones and 25 cm of argillaceous lime- to marlstones (Bed 4A). This more compact bed is succeeded by 40 cm of unfossiliferous greenish siltstones. In our view the whole package represents the Uggwa Limestone Formation which can be subdivided into two members: Member 1 represents the compact limestone unit (Bed nos.1 - 4), Member 2 the overlying greenish siltstones including Bed 4A.

Plöcken Formation. With a distinct change in colour, this sequence of strata is followed by grayish siltstones with intercalations of impure bioclastic limestone lenses containing fossil remains of the *Hirnantia* brachiopod fauna and of poorly preserved and rare graptolites. In total this shaly horizon

comprises 0.77 m (Bed no. 5). It is overlain by 5.40 m of more massive impure pyritiferous limestones and sandstones (Bed nos. 6 - 8). This 6.17 m-thick rock sequence is attributed to the Plöcken Formation.

Nöblinggraben Section

This section is located c. 10.6 km to the northeast of the Cellon locality at an altitude of 1110 m and belongs not only to another tectonic unit but also represents another facial development. The Late Ordovician to Silurian succession is dominated by black graptolitic shales and cherts with sporadic limestone intercalations. It thus represents a deep-water setting. Sampling (see below) for graptolites and chitinozoans as well as chronostratigraphy is in progress. Dropstones have been identified in the Plöcken Fm. and are biostratigraphically constrained by graptolites within the Hirnantian.

Uggwa Limestone. Overlying unfossiliferous shales of the Late Ordovician are the equivalents of the Uggwa Limestone and these closely resemble the corresponding calcareous Member 1 at the Cellon Section although this interval is here only 1.30 m thick. Member 2 is represented by greenish siltstones with intercalations of argillaceous limestone lenses showing a thickness of 5.20 m (Jaeger and Schönlaub, 1977).

Plöcken Formation. Member 2 of the Uggwa Limestone is unconformably overlain by a 0.15 m-thick pyritic, pebble-bearing sandstone bed which grades into 1.60 m-thick grayish siltstones with intercalations of impure bioclastic layers. Of particular interest are clasts of exotic crystalline rocks including plagioclases, silicites and granites showing dimensions as large as 5 cm in diameter (Schönlaub and Daurer, 1977) that are being reanalyzed for provenance. Sedimentary debris comprises clay- and siltstones, micaceous sandstones and quartzites. A probable K-bentonite level overlying the clast layer was sampled for radiometric dating and a poorly preserved chitinozoan fauna from the bioclastic intervals is being studied.

The Plöcken Fm. is succeeded by a 1.80 m-thick package of laminated quartzites with interbedded black schists. Age-diagnostic graptolites have yet not been found in this horizon but occur in the overlying black schists (Jaeger and Schönlaub, 1977) indicating a *C. vesiculosus* Zone age within the middle Rhuddanian at its base. Consequently, the quartzites below may represent the *P. acuminatus* or even the *A. ascensus* Zone of the base of the Silurian.

Rauchkofel South Section

This section is exposed on the southern flank of Mount Rauchkofel at an altitude of 2000 m (Schönlaub, 1971, 1985, 1988). The slightly overturned succession starts with the Uggwa Shale and is followed by the Uggwa Limestone, the Plöcken Fm. and various Silurian (Brett et al., 2009) to Lower Devonian limestones.

Plöcken Formation. Of particular interest is the sharp boundary between the Uggwa Limestone and the overlying 9 m-thick, pebble-bearing blackish sandstone of the Plöcken Fm. It is composed of medium-grained sandstones with rounded and angular, irregularly distributed clasts of limestones, quartzites and quartz. These are interpreted as dropstones and reinforce the evidence found at other sections for the diamictite nature of part of the Plöcken Fm. and the waning effects of the Hirnantian glaciation in the region. Analyses of the clasts are in progress and graptolites biostratigraphically constrain the interval.

BIOSTRATIGRAPHY

Graptolites

Uppermost Ordovician graptolites were first reported in the Carnic Alps by Jaeger et al. (1975) from the Feistritzgraben Gorge (see also Schönlaub, 1988), from the lowermost Plöcken Fm. just above the Ugghwa Limestone. Abundant graptolite rhabdosomes, confined to black slate rich in globular pyrite and affected by tectonic strain, have been tentatively assigned to *Normalograptus persculptus* (Elles and Wood). Poorly preserved specimens of the upper Hirnantian zonal index graptolite *Normalograptus persculptus*, however, can be easily misidentified with its likely ancestor *Normalograptus ojsuensis* (Koren and Mikhaylova) which is common in the lower Hirnantian *Normalograptus extraordinarius* Zone. Hence, the Late Hirnantian age of this graptolite occurrence must be considered with some reservation.

Uncommon, but better preserved specimens, assignable to *Normalograptus persculptus*, co-occur with the *Hirnantia* fauna in grayish siltstones of the lowermost Plöcken Fm. at the Cellon Section (Fig. 3). This association is assignable to the upper Hirnantian *Normalograptus persculptus* Zone.

In the Nölblinggraben (or Bischofalm) Section, silty-shaly intercalations high in the Bischofalm Quartzite yielded several poorly to moderately well preserved rhabdosomes of *Normalograptus* ex gr. *normalis*. The biostratigraphic significance of this monospecific *Normalograptus normalis* s.l. assemblage is limited, although closely similar occurrences are known from shaly interbeds within the post-glacial, late Hirnantian through to Llandovery Los-Puertos Quartzite (Gutiérrez-Marco et al., 1998) and Criadero Quartzite (Štorch et al., 1998) in Spain. At Nölblinggraben, the highest quartzite bed in the sequence is overlain by a silty black shale in which H. Jaeger found the mid-Rhuddanian index graptolite *Cystograptus vesiculosus* and abundant graptolites of the lowermost Aeronian *Demirastrites triangulatus* Zone (Jaeger and Schönlaub, 1977; Schönlaub, 1985).

At the Waterfall Section near Zollnersee Hütte a definite graptolite assemblage of early Rhuddanian (earliest Silurian) age is identified from an apparently overturned succession of black lydites and siliceous shales, c. 1 m below the massive Bischofalm Quartzite. The assemblage includes *Parakidograptus acuminatus* (Nicholson), *Normalograptus normalis* (Lapworth), *Normalograptus mirnyensis* (Obut and Sobolevskaya), *Glyptograptus* aff. *tamariscus* (Nicholson), *Neodiplograptus bifurcus* (Ye) and *Neodiplograptus lautus* Štorch and Feist and indicates the lower part of the *Par. acuminatus* Zone. The assemblage, however, is rather peculiar since some taxa typical of coeval faunas on other Peri-Gondwanan sections are missing [*Neodiplograptus lanceolatus* Štorch and Serpagli and *Normalograptus trifilis* (Manck)] whereas *Nd. bifurcus* of Chinese provenance is new to Europe.

At the Rauchkofel South Section, barely identifiable monograptid rhabdosomes were found in heavily cleaved black slates just above the diamictite succession. Either a tectonic contact, prominent stratigraphic unconformity, or both separate the two units. Therefore, the Rhuddanian and at least a substantial part of the Aeronian are likely omitted in the succession.

Conodonts

In the Late Ordovician of the Carnic Alps the conodont biostratigraphy is based on the pioneering study of Walliser (1964) at the Cellon Section who documented elements from the "Bereich I" (bed nos. 1 to 8) and on that of Serpagli (1967) on the "Tonflaserkalk" at the Rifugio Nordio and Monte Zermula Sections in the Italian Carnic Alps. Both works give a complete overview of the Late Ordovician conodont fauna

from the area. Subsequent papers dealing with the Upper Ordovician successions of the Carnic Alps did not include age-diagnostic conodonts. Ferretti and Schönlaub (2001) documented the *Amorphognathus ordovicicus* conodont Zone both in the Uggwa Limestone and Wolayer Limestone Formations with the finding of the “holodontiform element”, critical for species differentiation within the genus. Together with *Amorphognathus ordovicicus* Branson and Mehl, elements of *Amorphognathus lindstroemi* (Serpagli) were also documented within the Uggwa Limestone Fm. Faunas from both formations yielded abundant representatives of *Hamarodus europaeus* (Serpagli) and *Scabbardella altipes* (Henningsmoen), as well as of *Walliserodus amplissimus* (Serpagli) in some levels, and belong to the HDS (*Hamarodus europaeus*, *Dapsilodus mutatus*, *Scabbardella altipes*) biofacies of Sweet and Bergström (1984). The latter has been documented to date along the peri-Gondwana sector only in Sardinia (Ferretti and Serpagli, 1999) and in lower latitude areas of Avalonia and Baltica.

A slightly younger fauna occurs in the overlying Plöcken Fm. at the Cellon Section (Fig. 3 – Conodont Fauna 2), representing the only Hirnantian conodont fauna described to date along the northern Gondwana area. The fauna has a moderate diversity being composed of some twenty species. The association consists of small and fragmentary elements, documenting the first appearance of *Sagittodontina* Knüpfer and *Istorinus* Knüpfer, taxa common in older horizons of colder regions in the Mediterranean Province (such as Thuringia, Spain, NW France and Libya). Elements of “*Dichodella-Birkfeldia*”, which possibly correspond to the Gamachian genus *Gamichignathus* McCracken, Nowlan and Barnes, are abundant. *Amorphognathus* cf. *Amorphognathus ordovicicus* Branson and Mehl and *Amorphognathus lindstroemi* (Serpagli) were documented by Ferretti and Schönlaub (2001).

High resolution sampling from the Cellon Section, initially limited to bed 4 of the Uggwa Limestone Fm. and to beds 7 and 8 of the Plöcken Fm., will facilitate a more precise conodont based biostratigraphic control of the Late Ordovician interval.

Palynomorphs

At the Cellon Section 16 samples ranging through the Late Ordovician interval from the base of the Uggwa Limestone Formation to the top of the Plöcken Formation were prepared palynologically and examined for palynomorphs and the results are outlined briefly here. Analyses of samples from the Nölblinggraben Section are in progress.

Acritarchs

Ten of the analysed samples yielded only poorly preserved acritarchs that could not be determined (Priewalder, 1987).

Chitinozoans

The chitinozoans, in most cases similarly badly damaged, are present in only four samples from the Plöcken Formation: the first association occurs in siltstones just below bed 5, the second in bed 7, the third in bed 8 and the fourth community is derived from sandy shales above bed 8, i.e., just below the Ordovician/Silurian boundary (Priewalder, 1997).

Besides representatives of the genera *Calpichitina* Wilson and Hedlund, *Conochitina* Eisenack, *Rhabdochitina* Eisenack and *Spinachitina* Schallreuter, and a few Ancyrochitininae, three stratigraphically

important taxa could be identified on the basis of several unequivocal specimens: *Armoricochitina nigerica* (Bouché) (late Katian - Hirnantian), *Tanuchitina elongata* (Bouché) (terminal Katian - Hirnantian) and *Desmochitina minor* Eisenack (long-ranging, but not crossing the Ordovician/Silurian boundary).

The chitinozoans therefore indicate a Hirnantian age for the Plöcken Fm. (Fig. 3). The poor state of preservation of the chitinozoans (as well as the acritarchs), however, indicate a high energy sedimentary environment which probably led to selective preservation of the identified chitinozoan taxa. Hence, the *Tanuchitina elongata* chitinozoan biozone (base - late Hirnantian) is stated with some reservation. Finally, *Armoricochitina nigerica* (Bouché) and *Tanuchitina elongata* (Bouché), two typical North Gondwanan taxa and for the first time described from Niger, suggest a close relationship between the two depositional areas.

In the Nölblinggraben Section, a few badly preserved representatives of the Conochitinidae are present in the impure bioclastic intervals of the Plocken Fm., which are quite similar to those documented from the same interval at the Cellon Section.

Brachiopods

Faunas occur at three key horizons within the Upper Ordovician succession. An abundant and diverse brachiopod fauna has been described from the Himmelberg Sandstone and Uggwa Shale (Havlíček et al., 1987). The fauna is unusual, being characterized by a number of typical Gondwanan taxa, with links to Bohemia and Morocco, but with immigrants from Avalonia and possibly elsewhere. Higher in the succession, green siltstones within the upper member of the Uggwa Shale contain elements of the widespread deep-water *Foliomena* fauna (Harper et al., 2009; Rong et al., 1999). The terminal Ordovician *Hirnantia* fauna (Fig. 3) has been recorded from the Plöcken Formation (Jaeger et al., 1975). The fauna is typical of the Kosov brachiopod province (Rong and Harper, 1988) and there is clearly a depth gradient across the region from shallower-water facies at Hoher Trieb to deep water at the Cellon Section.

CHEMOSTRATIGRAPHY

Iron and Sulfur

Geochemical signals reveal a dynamic ocean chemistry during the Hirnantian in the Cellon Section (Fig. 3). By using the ratio of highly reactive iron over total iron contents in the sediment we get an estimation of the reducing conditions in the water column (Raiswell and Canfield, 1998; Poulton and Canfield, 2004). The late Katian and earliest Hirnantian has unequivocal values, just below the conventional threshold for anoxic values at 0.38. Unless turbidities affected the clastic input and diluted an iron signal of anoxic conditions, the Uggwa Limestone appears to have been deposited within an oxic water column. However, moving into the Hirnantian and the Plöcken Fm., there is a clear enrichment of reactive iron. The pyrite content, in the reactive iron, is at first, present but modest and increases towards the end-Hirnantian. It seems that the Plöcken Fm. and *Normalograptus persculpus* interval of the Hirnantian had a reducing water column. The conditions were at first ferruginous and later on richer in sulphide, however, not euxinic. We also note a heavy composition of sedimentary pyrite sulphur in the late Katian and presumably early Hirnantian. This could be an indication of low sulphate concentrations (Habicht et al., 2004) not only at the Cellon Section, but globally. A limited and depletable sulphate pool in the global ocean might give us

an indication that euxinia has increased in deeper parts of the ocean, burying excessive carbon and pyrite. This would contradict that deep ocean ventilation increased in the early Hirnantian. It would also demand a process that can mute the effect from cooling in terms of the sea water hosting more dissolved oxygen, as increased euxinia would mean less oxygen at least in some parts of the ocean.

Carbon Isotope Chemistry

The stable isotopic values of carbon at the Cella Section straddle around +1‰ throughout the Uggwa Limestone (increased from a value of -1.1 for carbonate in the underlying Uggwa Shale) and show a prominent excursion of +2.8‰ precisely at the unconformity with the overlying Plöcken Fm. (Fig. 3). If confirmed by high-resolution sampling this excursion coincides with the prominent peak in carbonate- $\delta^{13}\text{C}$ at the Katian-Hirnantian boundary (HICE - Bergström et al., 2009). The remainder of the Plöcken Fm. shows again consistently low values with a slight trend toward increasing values upsection.

CHRONOSTRATIGRAPHY

K-bentonites

The K-bentonite levels found in the Upper Ordovician of the Carnic Alps are quite rare and have relatively few equivalents elsewhere in Europe with the exception of beds reported from the British Isles, Baltoscandia, Poland and Lithuania (Histon et al., 2007). One of the four horizons (base Bed no. 6) noted in the Cella Section occurs within the *Hirnantia* fauna interval (Fig. 3) and this level is also found at the Hoher Trieb Section. Three levels (Bed 8) occur higher in the *Normalograptus persculptus* graptolite Zone at the Cella Section, one of which may be correlated with the single horizon noted at the Oberbuchach Section within this interval. Two lower levels at the Oberbuchach Section may be correlated with that found in the *Amorphognathus ordovicicus* conodont Zone at the Valbertad Section.

These data reinforce the notion that explosive volcanism associated with the amalgamation of pre-Alpine segments was not simply collisional in nature but represented a variety of source materials and tectonic settings. The K-bentonites belong to a tectonically active terrane dominated by calc-alkaline mafic lavas and pyroclastics in the Late Ordovician, Silurian and Early Devonian which was either situated north or south of the Carnic Alps but separated from the latter by an oceanic realm or at least an open sea of unknown width. However, the K-bentonite horizons in the Carnic Alps range from a few millimeters to 2-3 centimeters in maximum thickness indicating that the volcanic source area must have been quite distant. Histon et al. (2007) concluded that the majority of the K-bentonites found in the Carnic Alps were derived from neighbouring peri-Gondwanan terranes rather than from far distant sources at the eastern margin of the closing Iapetus Ocean.

Radiometric dating

Initial sampling of the K-bentonite levels identified from the Upper Ordovician successions of the Cella Section and Nöblinggraben Section for further analyses and radiometric dating was carried out in September 2010. The levels consist of yellow to dark brown clays, in general with a putty-like texture; the mineral composition is dominated by authigenic clay minerals and goethite, together with quartz, albite, ilmenite, magnetite, Ca-F apatite, F-apatite, indicating that a pristine igneous component is preserved.

According to the preliminary data in Histon et al. (2007), the andesite bulk composition of the K-bentonite levels indicates that U-Pb radiometric dating by SHRIMP on zircons is feasible. The phase separation is currently in progress.

On the whole, the radiometric dating will constrain the volcanic processes at the Ordovician – Silurian boundary, and may allow geotectonic inferences at a regional scale to be drawn.

DISCUSSION

The new chronostratigraphic classification of the Ordovician System presented by Bergström et al. (2009) with biostratigraphical standard zonations has made it essential to identify the $\delta^{13}\text{C}$ excursion (HICE) with precision in the Upper Ordovician interval of the Carnic Alps; this will permit recognition and subdivision of the Hirnantian Stage. Data integrated from multidisciplinary studies by our international team focussing on different aspects of lithostratigraphy, biostratigraphy, chemostratigraphy and chronostratigraphy as outlined briefly above have highlighted further evidence for the Hirnantian Stage based on the identification of the $\delta^{13}\text{C}$ Excursion (HICE) in the Cellon Section, although additional high resolution sampling is required to fully confirm this during the next field season. Evidence for paleoenvironmental and climatic/oceanic signals from a variety of isotope analyses has improved our knowledge of small scale perturbations within the marine succession which will allow high resolution correlation with other sectors. Sedimentological evidence recording the cold water influx of the Hirnantian glaciation event in the form of diamictites within the Upper Ordovician successions at the Rauchkofel South and at Nöblinggraben sections is now precisely constrained biostratigraphically thus adding further data for the timing of this event along the North Gondwana Margin.

New collections of graptolites, conodonts and chitinozoans have identified the index fossils for the global standard biostratigraphic zonations from a variety of sections and correlation of brachiopod faunas has documented distinct facies related assemblages recognized globally. These new results are complimentary to the faunal record documented previously and add a further recalibration of the latter biostratigraphic data. To date, the index graptolite for the lower Hirnantian, *Normalograptus extraordinarius* has not been found in the Carnic Alps. We conclude, however, that the siltstones of Member 2 of the Uggwa Limestone Fm. at the Cellon Section may correspond to this level (Fig. 3). The other possibility is that the unconformity separating Member 2 and the Hirnantian Plöcken Fm. encompasses the index graptolite zone for the basal Hirnantian. Finally, radiometric dating of interbedded volcanic layers will add precise time lines within which to collate the overall data set emerging for the Late Ordovician interval in the Carnic Alps. Thus, correlation of this pivotal sector as a regional reference for the North Gondwana area is now more feasible within a global context.

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INTENSE VOLCANISM AND ORDOVICIAN ICEHOUSE CLIMATE

B.K. Sell

Section of Earth and Environmental Sciences, University of Geneva, Rue des Maraîchers 13, 1205 Geneva, Switzerland.
Bryan.Sell@unige.ch

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INTRODUCTION

Volcanic episodes represented by Large Igneous Provinces are often evoked as possible causes of global climate change and major biological disturbances (Wignall, 2001; Cather et al., 2009). These provinces represent globally significant volcanism, however, they do not represent the only direct evidence of large-scale volcanic episodes in the past (Kennet and Thunell, 1975; Sigurdsson, 2000). Evidence of explosive volcanism is preserved as tephra in sedimentary rocks of all kinds and represents an equally important record of volcanism. Although volcanic event beds have low preservation potential in sedimentary rocks (Wheatcroft, 1990), previous global tephra compilations (Kennet and Thunell, 1975; Sigurdsson, 2000; Cambray and Cadet, 1996; Straub and Schmincke, 1998) show temporal clusters that indicate widespread explosive volcanism. To date, the tephra record has not been compiled beyond 100 million years ago, thus it is not known whether such large-scale volcanic events are a persistent feature of the Earth.

It is generally understood that widespread explosive volcanic episodes occurred during the Ordovician-Silurian (e.g. Kolata et al., 1996) and the Carboniferous-Permian (Bohor and Triplehorn, 1993), however, this tephra (K-bentonite, tonstein, tuff, etc.) record has never been quantified. Four episodes of explosive volcanism (Kennet and Thunell, 1975; Sigurdsson, 2000; Straub and Schmincke, 1998) are identified on the basis of tephra in marine drill cores and are interpreted to have occurred during the Pliocene (0–2 Ma), Miocene (14–16 Ma), Eocene (37–42 Ma), and the Late Cretaceous (65–99 Ma). These anomalously tephra-rich intervals in the sedimentary record correlate with known glacial episodes and have been implicated as being causally related, among other mechanisms, to climate cooling (e.g. Prueher and Rea, 2001). It does not seem likely that a single Cenozoic eruption could cause long-term climate effects, but, multiple eruptions over short period of time could be causally related to climate cooling (e.g. Jicha et al., 2009; Pollack et al., 1976).

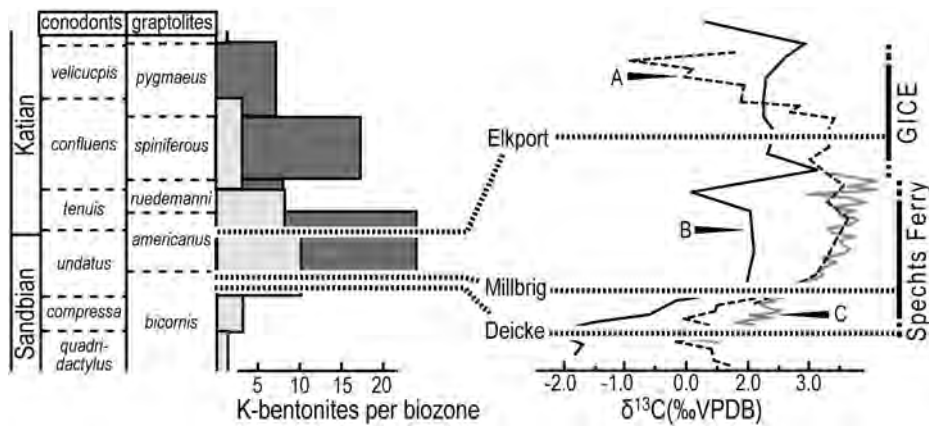


Figure 1. The Guttenberg carbon isotope excursion (GICE) and Spechts Ferry carbon isotope excursion. The Spechts Ferry excursion (Ludvigson et al., 2004) is usually considered part of the GICE (e.g. Bergström et al., 2010). Different chemostratigraphic and volcanogenic apatite trace element studies from A, West Virginia (Young et al., 2005; Sell and Samson, 2011), B, Iowa (Ludvigson et al., 2004; Emerson et al., 2004), and C, Pennsylvania (Patzkowsky et al., 1997; Carey et al., 2009; Sell and Samson, 2011) show that the approximate beginning of the GICE is likely coincident with the Deicke K-bentonite. The Deicke K-bentonite could not be located at Dolly Ridge and its stratigraphic position is inferred here.

It has been argued that a single eruption could have initiated Late Ordovician climate cooling (Buggisch et al., 2010). Also, a previously known carbon isotope excursion (GICE) was argued to not begin with the cooling of the Late Ordovician climate because it clearly post-dates the Millbrig and Deicke K-bentonites and $\delta^{18}\text{O}$ excursion. This Late Ordovician climate cooling argument places more emphasis on SO_2 from the Deicke eruption rather than cooling resulting from CO_2 drawdown that may be evidenced by the GICE. Contrary to these arguments, previously established tephra (K-bentonite) correlations and consideration of the Phanerozoic tephrostratigraphic record suggests that a volcanic episode composed of many large eruptions is coincident with long-term trends in both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ chemostratigraphy (Fig. 1).

TEPHRA CORRELATIONS AND CHEMOSTRATIGRAPHY

The ability to correlate highly altered tephra is invaluable for investigating proposed correlations (e.g. Delano et al., 1994; Haynes et al., 1996; Samson et al., 1995; Emerson et al., 2004; Mitchell et al., 2004; Carey et al., 2009; Sell and Samson, 2011). These single-crystal geochemical correlations have unparalleled stratigraphic resolution that are critical for interpreting relationships among the notoriously complex rocks of the Late Ordovician, which are likely plagued with poorly-known stratigraphic gaps and time-condensed stratigraphic intervals, both of which could lead to misinterpretations. Such problems are apparent with the GICE, which is widely used for stratigraphic correlation on a global scale (Bergström et al., 2010). For example, the widely cited Millbrig and Deicke K-bentonites occur within the GICE in Pennsylvania (Patzkowsky et al., 1997) and below the GICE in some other sections around the eastern U.S. (Young et al., 2005). Part of the reason for this discrepancy may be due to miscorrelated or unidentified K-bentonites (Carey et al., 2009). A miscorrelation of K-bentonites is possible due to the numerous K-bentonites that span the Sandbian-Katian stage boundary (e.g. Kolata et al., 1996).

The published $\delta^{13}\text{C}$ carbon isotope excursion and apatite trace element data for the Sandbian-Katian boundary interval shows that the GICE and the positive oxygen isotope excursion associated with the Deicke K-bentonite are likely correlative (Figure 1), which is contrary to the findings of Buggisch et al. (2010). Also, it was suggested that the Millbrig and Kinnekulle K-bentonites would not have a similar climate cooling effect because the beds are the result of multiple smaller eruptions, thus explaining the absence of a positive $\delta^{18}\text{O}$ excursion. Whether these beds represent multiple eruptions would not be of consequence because the volcanic process that produced the Millbrig K-bentonite had the same aerial extent (Mitchell et al., 2004), i.e. a similar magnitude, and similar SO_2 content (Kolata et al., 1996) as the eruption that produced the Deicke K-bentonite. Also, the scale of a single eruption does not match the scale of the expected effect from SO_2 flux to the stratosphere (Pollack et al., 1976). More stratigraphic sections need examined for chemostratigraphy with an emphasis on the appropriate sampling interval and stratigraphic gaps within the context of all correlated K-bentonites. Also, it may be worth considering that most volcanic ash that is associated with the GICE is dispersed within the same sedimentary rock such that the clay mineralogy and elemental composition of the host rock would be useful for determining the timing of potential volcanic influence (e.g. Do Campo et al., 2010; Straub and Schmicke, 1998).

VOLCANISM AND CLIMATE COOLING

I propose an alternative explanation for the relationship between Late Ordovician K-bentonites and climate cooling that appears to be consistent with the other Phanerozoic cool modes (Frakes et al., 1992) and the long-term oxygen isotope record (Veizer et al., 1999). The explanation is based upon the abundance of tephra found in the geologic record. I use the term tephra to refer to any volcanically derived layer of ash-fall, bentonite, tuff, tonstein, and some volcanoclastic or pyroclastic materials enclosed within sedimentary rocks. Tephra reported in geologic publications were compiled into a list that spans the past 542 million years. Tephra reported in separate locations and not proven to be correlative were counted as a unique tephra occurrence. Duplicate entries were avoided by excluding multiple reports from the same outcrop. One entry was made into the compilation where tephra details were ambiguous with respect age and quantity. Other tephra compilations derived from marine drill cores are included in this analysis (Kennet and Thunell, 1975; Sigurdsson, 2000). All of the tephra are binned in time by epoch following Gradstein et al. (2005). Clusters are tentatively defined for the Phanerozoic as maxima represented in the lower portion of Figure 2 and are regarded as evidence of explosive volcanic pulses (i.e. flare-ups or episodes). It is not possible to quantify the uncertainties of what defines a cluster at this time, however it appears likely that the Phanerozoic tephra show a non-random distribution with respect to time.

It is difficult to assess a monographic bias within the tephra compilation because testing for such effects is typically applied to diversity within a given group of organisms. However, a first-order examination of the compilation shows that the data are biased with respect to a few research articles (e.g. Kolat et al., 1996; Bohor and Triplehorn, 1993; Kennett and Thunell, 1975). Even so, the discussion here is mostly binary – tephra are either present or not. A potential monographic effect that likely influences the interpretations here is that of tephra thickness. If tephra thickness is to be analogous to biotic diversity, then it seems likely that only the thickest tephra dominate the compilation, which is similar to how only the largest fossils will sometime receive most of the attention in a given time and place. An examination of the available thickness data shows that average thickness of reported tephra for any given period is not less than two centimetres. This monographic effect appears to be acceptable for this discussion because the focus is on large-scale and

widespread eruptions, which would be expected to produce these thicker beds in distal deposits. If the tephra record is to be further quantified or be used to measure relative intensities of different volcanic episodes, then any monographic biases with respect to thickness will need to be assessed.

Tephra preserved in pre-Cenozoic sedimentary rocks are typically a result of relatively large eruptions because the beds are relatively thick and distal from their eruptive volcanic center. The average thickness of tephra in this study is 18 centimetres. Event beds less than approximately 10 centimetres thick that were deposited between the Late Triassic to the present have low preservation potential (Brandt, 1986) because of bioturbation. This mixing zone decreases rapidly to two centimeters for event beds deposited prior to the Late Triassic and for the remaining portion of the Phanerozoic. The average reported thickness and variation in bed thickness for tephra broadly reflect this trend, however thickness is often not reported in the literature. This broad and incomplete tephra thickness record might suggest that the majority of smaller eruptions are mixed into their host rocks and that macroscopic beds represent exceptionally large

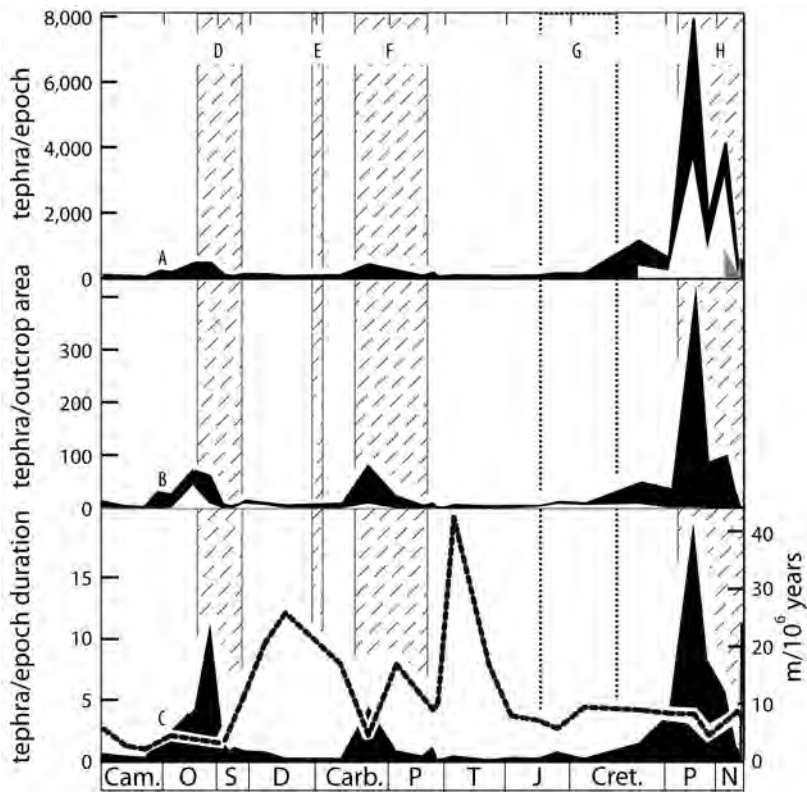


Figure 2. The distribution of tephra throughout the Phanerozoic. A, the black plot shows the uncorrected number of tephra per epoch for the Phanerozoic. The small white plot shows the tephra compilation from Sigurdsson (2000), and the small grey plot shows the tephra compilation of Kennett and Thunell (1975). B, the black plot shows the total number of tephra per epoch divided by global outcrop area from Wilkinson et al. (2009) with the fraction of tephra from North America in the white plot. C, the black plot is the result of correcting the tephra in B for the duration of each epoch. Superimposed on C, as a dashed line, is the Phanerozoic continental volcanic intensity curve of Ronov (1976), which is in meters per million years. D, F, and H are the major glacial episodes of the Phanerozoic. E, represents a short period with evidence for minor glaciation and G represents a cool period without direct evidence for glaciation.

eruptions, which is consistent with observed tephra preservation in modern ocean sediments (Kennett, 1981). Also, the preserved Palaeozoic tephra record shows that the large eruptions tend to cluster in time (e.g. Millbrig, Deicke, Elkport, and Kinnekulle K-bentonites), which is also expected when considering recent studies of large eruptions in the Cenozoic (Gusev, 2008; Mason et al., 2004).

Preservation should not be affected greatly by foreland basin development as this tectonic activity appears to be necessarily related in to increased levels of volcanism, subduction, and other orogenic processes. That fewer tephra appear to be preserved on passive margins is to be expected and may be consistent with the interpretations presented here. Alternatively, it may be that foreland basin development may decrease preservation potential of macroscopic ash-fall beds because of increased levels of sediment slumping on foundering platforms. In either case, more consideration is needed toward testing for this potential bias. Likewise, subduction of oceanic crust does not appear to bias the tephrostratigraphic record (Kennet and Thunell, 1977), although this is un-testable with rocks in the Paleozoic because all oceanic crust is virtually absent.

Epoch-scale preservation biases can be examined via global sedimentary map areas (Wilkinson et al., 2009), sedimentary rock formation names (Peters, 2006), and sea-level changes, e.g. the epeiric sea effect (Peters, 2007). Because the tephra compilation presented here has to be binned by geologic epochs of various durations, the bin size must be taken into consideration (Fig. 2). A sampling area bias is apparent when the Paleozoic and Mesozoic tephra record is combined with the tephra records derived from modern ocean sediment cores. Correcting for outcrop area and time smooths the general trend, amplifies older tephra clusters from the Mesozoic and Paleozoic, and depresses the apparent Cenozoic tephra clusters. Three other minor tephra clusters become apparent in the late Silurian to Early Devonian, Late Permian,

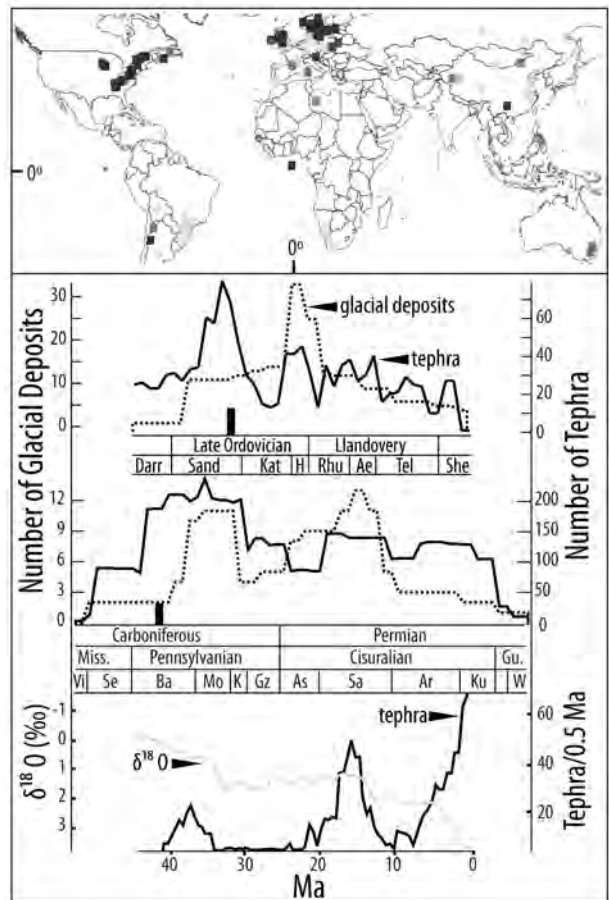


Figure 3. Cool modes (Frakes et al., 1992) and tephra abundance. Tephra (solid line) and glacial deposit (dashed line) distributions for the Carboniferous-Permian and Ordovician-Silurian glaciations. The world map shows locations of tephra and other explosive volcanics; squares for Ordovician-Silurian and circles for Carboniferous-Permian periods. The vertical black bars show the approximate timing of the thickest and most widespread tephra. The Cenozoic tephra abundance and $\delta^{18}\text{O}$ curve is modified from Sigurdsson (2000). All are plotted at approximately the same time scale for comparison.

and Middle to Late Jurassic. Sea-level and the diversity of sedimentary rock packages are not mutually exclusive phenomena. Correcting tephra abundance for the influence of sea-level and the number of sedimentary rock formation names (Peters, 2006) has little effect on removing apparent clusters of tephra.

The distribution of tephra shown here is expected on the basis of the Phanerozoic record of continental volcanism (Ronov, 1976) and orogenic activity (Hain and Seslavinskii, 1991). Although the interpretations of Ronov (1976) and Hain and Seslavinskii (1991) may be dated, the map units with their age assignments are broadly correct. Some continental flood basalts are related to the opening of oceans and other tectonic processes not related to large-scale explosive volcanism, thus the ratio of basaltic to sub-aerial volcanism would be expected to be higher (Fig. 2). Admittedly, this is an overly simplistic interpretation, but these broad relationships suggest that there may be alternating periods where one type of volcanic process is dominant. Also, if this temporal clustering of relatively large eruptions is a real phenomenon, then it is not readily explained by plate tectonic theory (Gusev, 2008; Mason et al., 2004).

Paleozoic tephra clusters appear to correlate with climate cool modes (Figs. 2 and 3) and this evidence of volcanism appears to be globally distributed (Fig. 3). The preserved tephra record, after accounting for some biases, appears to faithfully record the presence of volcanic episodes at multiple temporal and spatial scales. This interpretation of the record is consistent with the Cenozoic record and suggests that large-scale episodic explosive volcanism is a persistent feature of the Earth associated with cool climate modes. While the temporal relationship between volcanism and climate appears to be clear, the implicit causal relationship is likely complex and difficult to discern. For example, in order for volcanism to have such a long-term effect on albedo requires relatively constant level of large magnitude explosive volcanism. However, Mason et al., (2004) cautions against estimating total explosive volcanic output on the basis of a record that only records the largest eruptions. Also, it seems unlikely that decreased albedo via volcanic aerosols alone would have a long-term climate effect such that other processes affected by increased levels of sub-aerial volcanism need explored.

The data presented here appear to be consistent with weathering induced glaciation hypothesis (Kump et al., 1999). An increase in the level of volcanism would mean that more volcanic centers are available or exposed for silicate weathering. The CO₂ drawdown resulting from weathering coupled with increased levels of sub-aerial volcanism would have an even greater effect on climate cooling. At present, it does not seem likely that increased abundances of ash-fall beds are simply coincident with weathering of volcanic rocks.

A CONSTANT RAIN OF ASH?

Proposed mechanisms for global climate cooling from sustained explosive volcanism (Jicha et al., 2009) suggest that very large eruptions during the height of a volcanic episode could occur on a decade to century scale, thus maintaining a flux of SO₂ and Fe to the stratosphere and ocean surface waters, respectively. Sulfur species, mainly SO₂, ejected into the stratosphere combines with OH and H₂O over a period of weeks to create H₂SO₄, which has a strong effect on planetary albedo (Robock, 2000; Zeilinski, 2000). In addition to the albedo change, a sustained flux of Fe to the ocean via volcanic glass would increase CO₂ drawdown. The effect of volcanic ash (Fe-fertilization) on ocean primary productivity could be more significant than previously thought (Duggen et al., 2010; Bains et al., 2000). Direct empirical evidence comes from pronounced decreases in atmospheric CO₂ after two of the largest eruptions between 1958 and 1997 (Cather et al., 2009).

On the basis of SO_2 flux to the stratosphere, global average temperature can be expected to decrease by 1°K for 1–2 years after a large eruption, which could be sustained on decade scales for large eruptions that are closely spaced in time (Pollack et al., 1976). Radioisotopic ages from four volcanic arcs in the northern Pacific Ocean suggest a minimum of one relatively large eruption every 13,000 years between 28 and 35 Ma during the Eocene volcanic pulse (Jicha et al., 2009). Tephra clusters from other arcs (Kennett and Thunell, 1975; Sigurdsson, 2000) suggest that the Eocene-Oligocene flare-up was longer. Adding four more hemispheres of a similar amount of volcanic arcs suggest that at least one very large explosive eruption happened every 3,300 years. This frequency likely underestimates total explosive volcanic output because it appears that more than 75% of volcanic ash near some volcanic arcs is dispersed while only voluminous eruptions that yield high sedimentation rates are recorded as discrete tephra layers (Kennett, 1981). The ability to gauge the extent of explosive volcanism is dramatically reduced and suggests that far more eruptions may have happened during flare-ups. Recent field-work in the Ordovician, Silurian, and Jurassic appears to support this view (unpublished data) in that many unreported ash-fall beds have been discovered by the author and colleagues that are macroscopic (>1 centimetre) and smaller (< 1 centimetre), which indicates that there are many more tephra to be found.

CONCLUSIONS

This is the first attempt at quantifying the Paleozoic tephra record, which is critical for evaluating volcanic-climate hypotheses. Preliminary observations suggest that each volcanic long-term pulse during the Phanerozoic appears to correlate with long-term $\delta^{18}\text{O}$ seawater trends (Veizer, 1999) that suggest a long-term relationship that may extend beyond the Phanerozoic. The tephrostratigraphic record, however incomplete, should be considered in concert with other cooling mechanism that include enhanced volcanic rock weathering and changing orbital parameters as well as a broad array of tectonic and biotic processes. If the record means nothing more than increased volcanic activity, then the evidence presented here is a reminder that the Earth's climate system is complex and likely involves many yet to be discovered processes. The incomplete rock record likely contains new clues about links between seemingly disparate processes. The tephra record of the Phanerozoic, perhaps, may inform us about the “volcanic winter to snowball Earth hypothesis” (Stern et al., 2008). The intensity of both the climate cooling and volcanism may be related, which suggests there may be larger flare-ups yet to be discovered in pre-Phanerozoic sedimentary rocks. Also, the information compiled here may be useful for evaluating models that invoke glaciation as a cause of explosive volcanism (e.g., Huybers and Langmuir, 2009). The long-term tephra record presented here appears to suggest that volcanic episodes predate glacial events with volcanism maintained for the duration of the climate cooling.

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NEW U-Pb ZIRCON DATA FOR THE GSSP FOR THE BASE OF THE KATIAN IN ATOKA, OKLAHOMA, USA AND THE DARRIWILIAN IN NEWFOUNDLAND, CANADA

B.K. Sell¹, S.A. Leslie² and J. Maletz³

¹ Section of Earth and Environmental Sciences, University of Geneva, Rue des Maraîchers 13, CH-1205 Geneva, Switzerland.
Bryan.Sell@unige.ch

² Geology and Environmental Science, James Madison University, 395 South High Street,
MSC 6903, Harrisonburg, Virginia, USA 22807. lesliesa@jmu.edu

³ Department of Geosciences, Colorado State University, 322 Natural Resources Building, Fort Collins,
Colorado, 80523-1482, U.S.A. yorge.maletz@colostate.edu

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INTRODUCTION

Numerous Ordovician ash-fall beds, often called K-bentonites because of post-deposition alteration (e.g. Kolata et al., 1996), present a relatively unique situation in terms of radioisotopic dating potential compared to other periods in the Palaeozoic. Even so, there are relatively few radioisotopic dates, such that a small number of dated ash-fall beds can cause relatively large shifts in the early Palaeozoic numerical timescale (e.g. Bowring and Erwin, 1998). In our ongoing efforts to enumerate the Ordovician timescale we selected several beds from the late Sandbian at the Katian Global Stratotype Section and Point (GSSP) in Oklahoma, USA and the Darriwilian, in Newfoundland, Canada. These ash-fall beds are of immense stratigraphic importance because each location is associated with three different biostratigraphic schemes, chitinozoans, conodonts, and graptolites. Having such biostratigraphic data is important because other important locations are often restricted to one fauna group because of lithofacies. For example, it is difficult to make precise biostratigraphic correlations from graptolite-dominated black shale to conodont-dominated limestone. The purpose here is to 1) briefly discuss the interpretation of radioisotopic age determinations of K-bentonites in these intervals and 2) show some new preliminary U-Pb zircon data for these Middle to Late Ordovician beds that are critically important for constraining the timescale.

PREVIOUS AGE DETERMINATIONS

The various age determinations of late Sandbian K-bentonites are understandably misleading (Bergström et al., 2004; Huff, 2008) because of the differences in isotopic systems and methods, and their respective results. No one method or isotopic system is inherently better at determining the age of a K-

bentonite. However, details of the method, precision required by the geologic questions, and behaviour of the isotopic system in a given crystal with respect to its post-depositional environment should dictate which method is best suited for age determination (see review by Bowring and Schmitz, 2003). In the case of highly altered tephra from the Palaeozoic, U-Pb zircon via isotope dilution thermal ionization mass spectrometry (ID-TIMS) is the preferred method because of the achievable precision (Parrish and Noble, 2003). Over the past few decades, radioisotopic age determinations have generally become more precise, which has been partially driven by geologic questions that demand greater precision and accuracy. It was also partially driven by the realization that many igneous rocks contain multiple populations of crystals with subtle age differences – this situation requires the analysis of single crystals. It is possible that multiple crystal populations within Sandbian K-bentonites have affected the different age determinations, which have been historically variable. In the case of Sandbian K-bentonites, overlapping ages could be an indication of inherited older crystals as well as a lack of desired precision. However, it has not been possible until recent years to analyze single crystals with the desired precision, such as U-Pb in single zircon crystals, because of methodological limitations. Other single-crystal methods have been applied to the Sandbian K-bentonites, yet the application of these excellent methods, such as Ar-Ar, may not be appropriate to determine age differences between closely spaced beds, such as the Deicke, Millbrig, and Kinnekulle K-bentonites (Schoene et al., 2006). For example, while the Ar-Ar biotite data of Min et al. (2001) appear to demonstrate that the Millbrig, Deicke, and Kinnekulle K-bentonites are not the same age, the exact duration between the three beds is not obvious because of the less than desirable analytical precision. Other potassium rich crystals, such as sanidine, can be analyzed (e.g. Chetel et al., 2004), however it is uncertain how these types of crystals have been affected by post-depositional conditions and the relatively large external errors due to inter-laboratory bias and decay constant uncertainty would greatly expand the total age error. The Sandbian K-bentonites are highly altered tephra and have likely been exposed to post-depositional conditions that can easily affect the potassium-rich minerals. All of this would lead one to conclude that recent radioisotopic dating attempts have not been sufficiently precise in showing differences between Sandbian K-bentonites (Bergström et al., 2004; Huff, 2008), which is a valid first-order conclusion. However, the various radioisotopic dating techniques applied to the Sandbian K-bentonites have shown an increase in precision with each successive age interpretation. Likewise, overlapping errors of age determinations, such as that for the Deicke, Millbrig, and Kinnekulle K-bentonites, are not useless. Various statistical tests, i.e. the students t-test, applied to both the U-Pb and Ar-Ar data from these three beds shows that age equivalency can be rejected (p -value = 0.00).

U-Pb ZIRCON

Zircon crystals are robust because they resistant to complete alteration in the various post-depositional environments that host the Sandbian K-bentonites. At least part of the crystal remains a closed system with respect to radioisotopes. Alteration of parts the original zircon often results in lead-loss, but air (Krogh, 1982) or chemical abrasion techniques (Mattinson, 2005) can be applied such that the unaltered portion of the zircon can be analyzed. Testing the whether a zircon crystal has been compromised is possible by the ability to simultaneously measure two decay systems, $^{238}\text{U} \rightarrow ^{206}\text{Pb}$ and $^{235}\text{U} \rightarrow ^{207}\text{Pb}$, which is the greatest advantage for using zircon as a chronometer. Previous U-Pb zircon age determinations of the Sandbian K-bentonites are useful despite larger uncertainties (Tucker and McKerrow, 1995). In these previous age determinations the crystals were carefully selected so as to avoid obviously older crystals and

inclusions that can affect the age determination. Typically, slender and clear zircon crystals yielded the youngest ages. These age determinations made at a time of significant methodological advances in decreasing common lead (Pb^{204}) contamination (Davis et al., 2003), but prior to recent advances in removing damaged areas of the zircon crystals that have lost Pb. These problems combined with inherited zircons that are slightly older may obscure the approximate eruption or minimum age of some Sandbian K-bentonites. New analyses with the CA-TIMS method (Mattinson, 2005), which is an adaptation of the aforementioned ID-TIMS method, may help reveal more complexity in the zircon populations of these beds and decrease age uncertainty. As this method continues to evolve, much more will be learned about timescales with increasingly greater precision.

Previously, comparing U-Pb zircon age determinations among different labs that use the same method has been difficult. The U-Pb zircon community has recently made attempts at inter-calibration between laboratories with good success. At the center of this effort are new isotopic tracers used for isotope dilution and standard U-Pb solutions (see www.earth-time.org). These efforts have removed a significant amount of uncertainty in the method. This does not mean that previous age determinations are rendered useless or invalid, rather that these data will need to be reinterpreted with respect to new findings.

K-BENTONITE AND ZIRCON SAMPLES

Ash-fall bed samples for U-Pb zircon analyses have been collected from several Katian-Sandbian and Darriwilian locations in Oklahoma and western Newfoundland, respectively. Two beds were sampled at 4.3 and 5.0 meters below the top of the Womble Shale, which is just below the GSSP for base of the Katian in Atoka, Oklahoma. These two K-bentonites are within the *Climacograptus bicornis* graptolite and the *Amorphognathus tvaerensis* conodont Zones (Goldman et al., 2007). Other clay bed samples were collected from the Atoka section as well as outcrops near Fittstown, Oklahoma and a roadcut on U.S. Interstate Highway 35. These other samples did not yield zircon crystals, although several beds were unsampled. Two Darriwilian samples yielded zircon from suspected K-bentonites at 2.5 and 41 meters above the base of the Mainland Section in the Cape Cormorant Formation on the on Port au Port Peninsula, western Newfoundland Canada (Albani et al., 2001). Four suspected K-bentonites yielded zircon from the Table Head Group in the West Bay Centre quarry section also on the Port au Port Peninsula (Albani et al., 2001 and references therein). Although all samples yielded zircon, one bed from each section was selected for preliminary analyses (Mainland 41m and West Bay Centre Quarry D). The Mainland section sample is within the *Pterograptus elegans* graptolite Zone and the West Bay Centre Quarry sample is within the *Holmograptus spinosus* graptolite Zone.

The bentonite samples were disaggregated with an electric laboratory mixer, washed to remove the clay fraction, sieved using 250-micron mesh, and placed in heavy liquids to separate the dense zircon fraction. All zircon crystals were selected under a reflected light microscope. The zircon samples were annealed and then chemically abraded in hydrofluoric acid following a modified method from Mattinson (2005). The samples were cleaned and then dissolved in Teflon® microcapsules with EarthTime 2535 U-Pb tracer solution and analyzed using a thermal ionization mass spectrometer. More details of the exact analytical methods used can be found in Schaltegger et al. (2008).

NEW AGE DATA

We analyzed single zircon crystals from one bed of the Mainland Section at 41 meters (Albani et al., 2001), and one bed from the West Bay Centre quarry. These new data (Fig. 1) are preliminary and should be interpreted with caution although the results are in good agreement with previous age estimates by others (Tucker and McKerrow, 1995). The Womble #2 and Womble #1 K-bentonite samples yielded weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 452.76 ± 0.19 Ma (MSWD = 1.08) and 453.53 ± 0.28 Ma (MSWD = 1.2), respectively. The samples from the Mainland Section at 41 meters and West Bay Centre Quarry D yielded weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 464.5 ± 0.4 (MSWD = 1.5) and 464.57 ± 0.95 (MSWD = 1.09), respectively.

INTERPRETATIONS

The new U-Pb data together with the biostratigraphic data, whole rock chemical data (Leslie et al., 2008), and chemostratigraphy (Goldman et al., 2007) suggest that either of the two K-bentonites in the Womble Shale could be identified as the Millbrig K-bentonite. However, there are other K-bentonites in this interval elsewhere, such as the Elkport K-bentonite, that may have a more similar age. It seems likely that either of the Womble K-bentonites are not correlative with the Deicke K-bentonite. The preliminary U-Pb zircon age of the Womble K-bentonite #2 is 452.76 ± 0.19 Ma and overlaps in error with that determined for the Millbrig K-bentonite at 453.1 ± 1.3 Ma (Tucker and McKerrow, 1995), however the Millbrig age also overlaps in error with Womble #1. The mean $^{206}\text{Pb}/^{238}\text{U}$ ages for the Womble #2 (452.76 ± 0.19) and Womble #1 (453.53 ± 0.28) do not overlap with that of Deicke K-bentonite at 454.5 ± 0.5 Ma or the potentially synchronous Kinnekulle K-bentonite in Scandinavia at 456.9 ± 1.8 (Tucker and McKerrow, 1995). More precise U-Pb analyses of a known Millbrig K-bentonite will be required for comparison. These new preliminary ages are more precise and indicate that K-bentonites in this interval can be distinguished with an adequate number of U-Pb zircon analyses that yield non-overlapping ages (Fig. 2). Also, the data here appears to support the suggestion (Schoene et al., 2006) that the previously determined Ar-Ar biotite ages (Min et al., 2001) for the Millbrig, Deicke, and Kinnekulle K-bentonites are likely too young (Fig. 2).

The Sandbian-Katian stage boundary age as suggested by Sadler et al. (2009) may be too old by at least four million years (Fig. 2). This is a significant timescale adjustment, however it may not drastically affect any stratigraphic interpretations. The age interpretations suggest a shorter duration for the Katian stage and this may have implications for the timing of causal explanations regarding the Late Ordovician mass extinction and ice age(s). The younger age appears to be more consistent with Bergström et al. (2008), who correlate the base of the Katian with the later portion of the Caradoc. Sadler et al. (2009) correlate the base of the Katian stage with the earlier portion of the Caradoc. Age data from other K-bentonites, such as the Dickeyville and Elkport, which are both stratigraphically higher, may help constrain the age of the Sandbian-Katian stage boundary and serve as a test of the interpretations presented here. Initial data from the Elkport K-bentonite sampled from the Curdsville Member of the Lexington Limestone in Kentucky indicates an age of 452.40 ± 0.86 on the basis of three single-crystal U-Pb zircon analyses. This data together with the age data from the Womble #2 K-bentonite suggests that the age of the base of the Katian is approximately no older than 452 Ma.

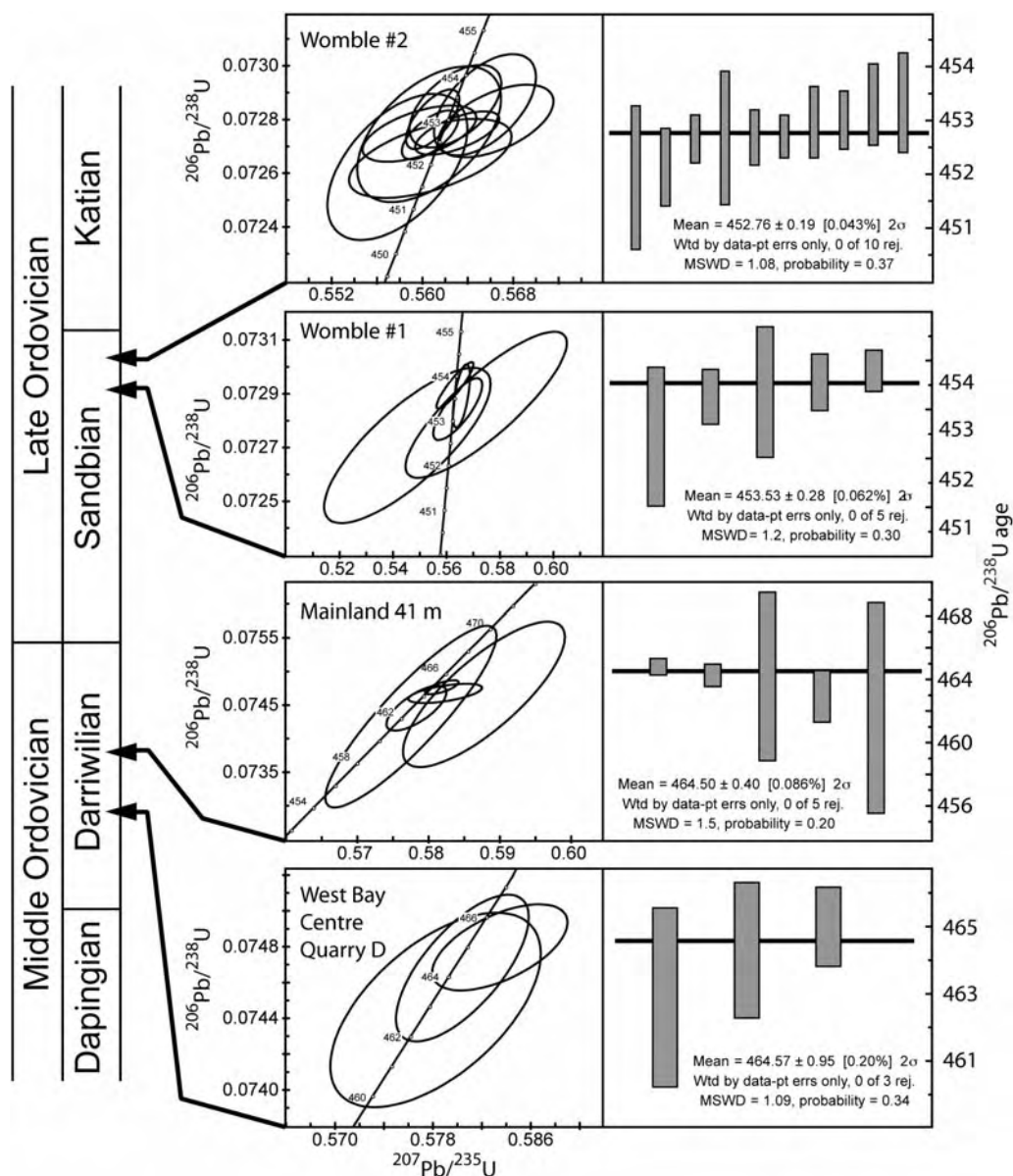


Figure 1. U-Pb zircon Concordia and weighted mean of $^{206}\text{Pb}/^{238}\text{U}$ ages. Data plotted with ISOPLOT (Ludwig, 1991). The mean square of the weighted deviates (MSWD) is a probability statistic gives an indication of whether the data points are equivalent (see Ludwig, 1998 for explanation and references). Error ellipses and error bars include internal errors only and are presented at 2σ . Sample locations and ash-fall beds are illustrated in Goldman et al. (2007) and Albani et al. (2001) except for the West Bay Centre Quarry sample.

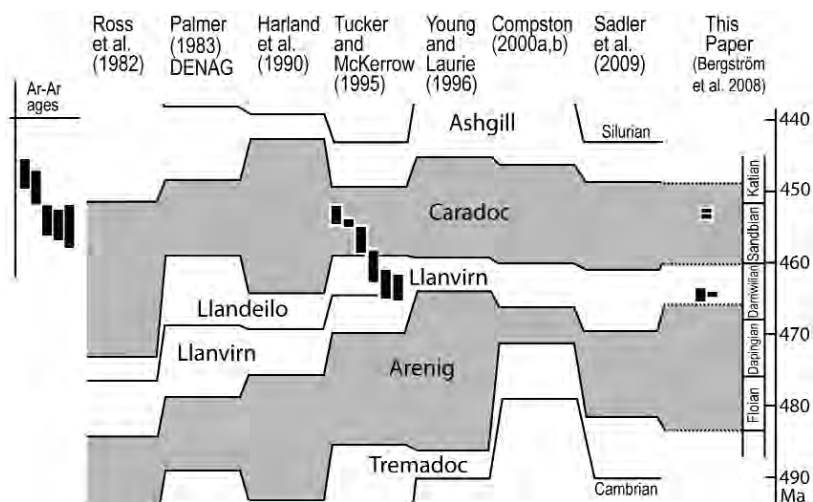


Figure 2. New U/Pb zircon ages suggest a shift in stage boundaries for the Middle to Late Ordovician timescale (modified after Sadler et al., 2009 and Bergström et al., 2008). Prior to this study there were scant few U-Pb ages for the Ordovician, which all showed overlapping internal errors. Note that our new measurements show greatly improved internal precision compared to previous ages in the same interval. The black vertical bars represent internal age uncertainties and do not include tracer calibration and decay constant errors. The black bars under the Tucker and McKerrow (1995) heading represent U-Pb ages and includes the age from Huff et al. (1997). The black vertical bars under the Ar-Ar heading include ages from Min et al. (2001) and Tucker and McKerrow (1995). The references for Ross et al. (1982), Palmer (1983), Harland et al. (1990), Young and Laurie (1996), and Compston (2000a,b), while not discussed in the text, are included for comparison.

Although the Elkport K-bentonite has not been directly associated with an index fossil, the bed likely occurs near the boundary between the *Phragmodus undatus* and *Plectodina tenuis* Conodont Zones, which is just above the Sandbian-Katian boundary and within the approximate lower portion of the Guttenberg Carbon Isotope Excursion (GICE). The Elkport K-bentonite is known to occur in the Guttenberg Member of the Decorah Formation, which is within the lower portion of the GICE in Iowa (Ludvigson et al., 2004) and has been identified on the basis of apatite trace element composition (Emerson et al., 2004). The Elkport K-bentonite has been correlated from the Guttenberg Member to the Curdsville Member of the Lexington Limestone in Kentucky on the basis of apatite trace elements (Sell, 2010), which also places the bed in the lower portion of the GICE. The GICE largely occurs within the *P. tenuis* Zone with the lowest portion of the excursion extending into the *P. undatus* Zone (Young et al., 2005). The Sandbian-Katian stage boundary is just below the base of the *P. tenuis* Zone. If the Elkport K-bentonite is within upper portion of the *P. undatus* Zone and is the same age as the Womble #2 K-bentonite, then the base of the Sandbian-Katian stage boundary can be traced across much of eastern North America using apatite trace element chemistry of both the Elkport and Millbrig K-bentonites (e.g. Carey et al., 2009; Emerson et al., 2004; Mitchell et al., 2004). This could potentially cause some conflicts with respect to the M4-M5 sequence stratigraphic boundary because the Elkport K-bentonite is interpreted to be within the M5 sequence whereas the Womble #2 K-bentonite is within the younger M4 sequence. More biostratigraphic data, sequence stratigraphic analysis, and radioisotopic data will be needed in locations where the Elkport K-bentonite can be identified in order to test the age relationships suggested here.

Demonstrating whether K-bentonite beds in the Womble Shale will directly correlate with other K-bentonites outside of Oklahoma, i.e. determining a chemical fingerprint would greatly increase the utility

of the Katian GSSP. Apatite phenocrysts would be extremely useful for tephrostratigraphic correlation as it has been for many other beds in this interval (Sell and Samson, 2011; Carey et al., 2009; Emerson et al., 2004; Mitchell et al., 2004). However, K-bentonite beds from Oklahoma have not yielded apatite that would permit tephrostratigraphic correlation (Leslie et al., 2010). K-bentonite correlation may still be possible by analyzing glass inclusions in beta quartz (e.g. Delano et al., 1994). Also, trace elements (Schoene et al., 2010) and other isotopes in single zircon crystals, such as oxygen, could serve as a useful geochemical correlation tools. In the absence of such geochemical constraints, the most precise U-Pb zircon ages possible will be required.

The new ages for the Table Head Group K-bentonites from Newfoundland suggests that the boundaries of the Darriwilian stage (Fig. 2) may need to be adjusted by three to four million years compared to the most recently constructed timescale of Sadler et al. (2009). Only a few concordant analyses could be made at this time for the West Bay Centre D K-bentonite, however the preliminary ages of both Darriwilian K-bentonites are self-consistent. The Table Head Group K-bentonites suggests an age of approximately 464 Ma for the boundary between the Darriwilian 2 and 3 Australasian Zones, whereas Sadler et al. (2009) suggest an age of at 269.97 Ma. How the Darriwilian correlates with the British Series and international stages poses a conflict with different timescales. We illustrate the correlations of Bergström et al. (2008) that place the Darriwilian 2 and 3 within the Llanvirn British Series. Sadler et al. (2009) correlate the Darriwilian 2–3 boundary within the Arenig British Series.

Previous radioisotopic ages for the Middle Ordovician are few and relatively imprecise, such that the ages presented here represent an improvement in the precision of the numerical timescale. However, these new ages are very similar to previous U-Pb zircon ages from this time interval. Tucker and McKerrow (1995) determined a U-Pb zircon age from a bed in the *Didymograptus muchisoni* Zone at 464.6 ± 1.8 , which correlates with the upper portion of the *Pterograptus elegans* Zone and compares well with our 464.5 ± 0.4 Ma age. A similar a U-Pb zircon age reported by Huff et al. (1997) from early Darriwilian in Argentina is 464 ± 2 , which may be too young when compared to our data and that of Tucker and McKerrow (1995).

CONCLUSIONS

The new U-Pb zircon analyses of K-bentonites suggest that there is great potential for the Ordovician to possess the most highly resolved numerical timescale in the Palaeozoic. Combining the resolving power of U-Pb zircon from the numerous K-bentonites with that of biostratigraphy could significantly advance our understanding of causal mechanisms that lead to the Great Ordovician Biodiversification Event, end-Ordovician mass extinction, and end-Ordovician cool mode. Our understanding of the timing of possibly related chemostratigraphic phenomena such the middle Darriwilian $\delta^{13}\text{C}$ excursion (MDICE) and the late Sandbian to early Katian Guttenberg $\delta^{13}\text{C}$ Excursion (GICE) will certainly benefit from the age constraints provided here. More specifically, it now seems possible that we can distinguish the ages of the Deicke, Millbrig, and Kinnekulle K-bentonites with non-overlapping numerical ages. These K-bentonites represent some the largest eruptions in the Phanerozoic such that precise ages would greatly aid in understanding their impact on the global environment. Although the results presented here are preliminary, it is encouraging that these new U-Pb data are more precise and in agreement with previously determined U-Pb zircon ages. In the future, new age determinations and other isotope and trace element analyses from the same single-crystals, as well as other different crystals, should be combined to test proposed correlations between beds. These new approaches can advance our understanding of the similarities and differences of these enormous eruptive units while increasing their stratigraphic utility.

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ORDOVICIAN REGIONAL CHRONOSTRATIGRAPHIC SCHEME OF THE GORNY ALTAI

N.V. Sennikov, O.T. Obut and E.V. Bukolova

Trofimuk Institute of Petroleum Geology and Geophysics SB RAS, Acad. Koptug av. 3, 630090 Novosibirsk, Russia.
sennikovnv@ipgg.nsc.ru, obutot@ipgg.nsc.ru, bukolovaev@ipgg.nsc.ru

Keywords: Ordovician, regional chrono- and lithostratigraphical units, Gorny Altai.

INTRODUCTION

Ordovician strata are widely distributed on the territory of Gorny Altai. Moreover, the Altai Ordovician basin (Fig. 1) is rather well investigated by paleontological and biostratigraphical studies and has some of the best Siberian sections with a potential to check the definition and correlation of the new global stages of the International Chronostratigraphic Chart (ICC), as well as to trace global sedimentary and biotic events in particular sections.

Correlation of most sections is based mainly on trilobites and brachiopods that were sampled from lithologically diverse sedimentary facies, both carbonatic and siliciclastic. Orthostratigraphic pelagic graptolites, conodonts and chitinozoans are usually very rare (Obut and Sennikov, 1986; Sennikov, 1996; Sennikov and Obut, 2003; Paris et al., 2004; Izokh et al., 2006; Sennikov et al., 2008). The geological practices for regional correlation in Russia are based on a special set of chronostratigraphic units, defined as regional stages, also called "horizons" (Stratigraphic Code of Russia, 2006). Regional stages (horizons) coincide with historical-geological stages of development of each particular region, related with advances in the study of various marine groups.

Regional stages (horizons) embrace same-age formations (or parts of them) distributed within the region. Regional stages (horizons) are established and collectively approved at the special and periodical all-Russian stratigraphic meetings, usually held every 10-15 years, or occasionally 5-7 years apart. In the regional stratigraphic charts the sequence of the regional stages (horizons), along with regional zonation worked out on pelagic orthostratigraphic groups, serves as a base for regional correlation of local stratigraphic units (formations and groups) as also for interregional correlation with regard to other regional stages (horizons) from adjacent regions. In the Ordovician, some of the regional graptolite and conodont zones allow precise correlation of regional stages with the global series and stages of the ICC and the International Geologic Time Scale (IGTS).

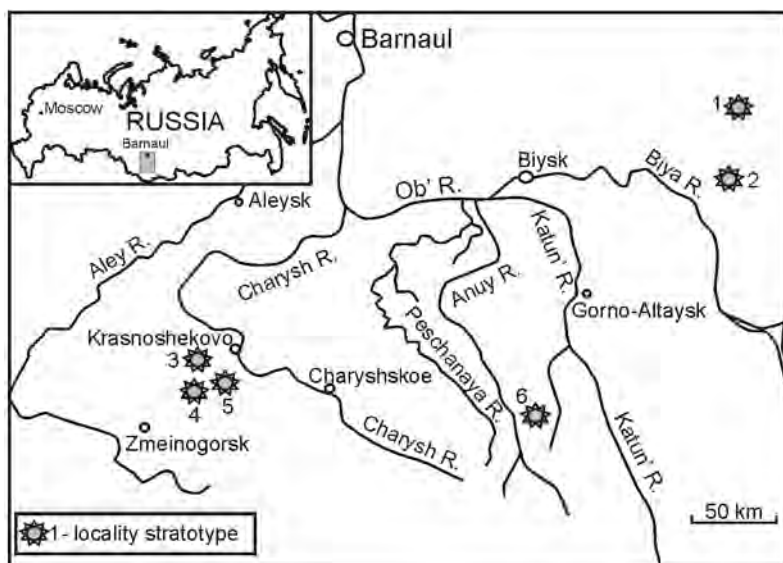


Figure 1. Location of the stratotype sections for the Ordovician Regional Stages of the Gorny Altai.

ORDOVICIAN REGIONAL CHRONO- AND LITHOSTRATIGRAPHIC UNITS OF THE GORNY ALTAI

The officially adopted Ordovician stratigraphic charts that are used today in Russian geological map legends were worked out more than 30 years ago (Decisions..., 1983; Sennikov et al., 1988). Recently an attempt to improve the old charts and produce new ones was made (Sennikov et al., 2008). Figure 2 introduces a simplified model of the Ordovician stratigraphic chart proposed for the Gorny Altai. The base for the Ordovician regional stages (horizons) is the successive change in trilobite and brachiopod assemblages. Thus these two faunal groups are not only a tool for intraregional correlation of sections, but rather a "basement" for the basis for recognition of regional stages (horizons). These regional stages based on trilobites and brachiopods, and also allow, in most cases, an inter-regional correlation with adjacent regions, for instance Gorny Altai with Salair, Kuznetsky Alatau, Siberian Platform, Taimyr, Kazakhstan and the Urals. Also widely distributed tabulate and rugose corals, ostracods, ichthyo-fauna, bryozoans, crinoids, nautiloids, gastropods and radiolarians are applied for the characterization and correlation of local sections, and for paleoecologic and paleogeographic reconstructions. These faunas are rarely used for intra-regional and inter-regional correlation. Real chronostratigraphic position of the regional stage boundaries relative to the respective global stage boundaries of the ICC could be evaluated and precisely validated only after the analysis of the graptolite and conodont zonal succession proposed for the Gorny Altai (Sennikov, 1996; Izokh et al., 2006; Sennikov et al., 2008).

Ordovician regional stages (horizons) for the Gorny Altai, Salair, Siberian Platform, Urals, Taimyr and Kolyma basins were established on the basis of most widely distributed fauna (trilobites and brachiopods), taking into account data from other benthic fauna (tabulate and rugose corals, ostracods, etc.). Pelagic associations (graptolites, conodonts, chitinozoans and radiolarians) are scarcely distributed in the above-mentioned basins, and were not used for definition of regional stages, but were used as additional elements inter-regional correlations. Before recent times it was methodologically reasonable that the series

International stratigraphic scale				Regional chronostratigraphic units				Local lithostratigraphic units						
SYSTEM	SERIES	STAGE	"SUB-STAGE"	REGIOSTAGE (Horizon)	Graptolite zones	Conodont zones	Chitinozoan zones	Western and north-western parts	Western, north-western and central parts	Northern part	North-eastern part			
O R D O V I C I A N	UPPER	HIRNANTIAN	Hi ₁	TEKHTEN'	persculptus		ordovicianus	Conochitina micracantha	SILICEOUS-TERRIGENOUS	TEKHTEN' FM.	CHEBOR FM.			
			Hi ₂		supernus	ornatus		Lagenochitina dalbyensis, Desmochitina lecaniella				?		
			Ka ₁		supernus	BULUKHTA FM.								
			Ka ₂		linearis									
		KATIAN	Ka ₃	KHANKHARA	quadrim-cronatus	KHANKHARA FM.		GUR'YANOVKA FM.						
			Ka ₄		caudatus									
			Ka ₅		bicornis									
			Sa ₁		multidens									
		SADBIAN	Sa ₂	BUGRYSHIKHA	petifer / antiquus lineatus	BUGRYSHIKHA FM.								
			Sa ₃		gracilis / serratus									
			Sa ₄		teretiusculus									
			Sa ₅		jakovlevi / coelatus									
	MIDDLE	DARRIWILIAN	Da ₁	KOSTINSKY	balhaschensis / kirgisicus	"suevicus-variabilis"	Cyathochitina calix			KARASU FM.				
			Da ₂		dentatus									
			Da ₃		austrodentatus									
			Da ₄		sinodontatus / Cardiodiaptus									
		DAPINGIAN	Dp ₁	LEBED'	caduceus imitatus		VOSKRESENKA FM.		TULOY FM.					
			Dp ₂		maximo-divergens									
			Dp ₃		gibberulus									
			Dp ₄		deflexus									
		FLOIAN	Fl ₁	TAYANZA	angustifolius elongatus / broggeri	evae		Conochitina raymondi			MARCHETA FM.			
			Fl ₂		densus									
			Fl ₃		balticus									
			Fl ₄		approximatus									
	LOWER	TREMADOCIAN	Tr ₁	TAYANZA	osloensis / hyperboreus	proteus			TALITSA FM.		ISHPA FM.			
			Tr ₂		tenellus / kieri									
			Tr ₃		?									
			Tr ₄		?									

Figure 2. Ordovician regional chronostratigraphical scheme of the Gorny Altai.

and stages of the redefined British scale (Tremadoc, Arenig, Llanvirn, Caradoc, Ashgill) were established on the basis of trilobites and brachiopods. Recently this paradigm changed and the earlier used "Stratotype of the stratum" principle was replaced by the new "Global Stratotype Section and Point" principle by the International Commission on Stratigraphy, with the need for allocation of reference sections of global stage boundaries with boundaries of regional zones. The Ordovician System experienced one more innovation: new stages were adopted and ratified, Tremadocian, Floian, Dapingian, Darriwilian, Sadbian, Katian and Hirnantian (Ogg et al., 2008). None of them were specifically defined by standard biostratigraphic units or single paleontological characteristics, so that most lack any reference to characteristic trilobites and brachiopods. The "GSSP principle" precisely defined and officially ratified chronostratigraphic position of the boundaries for new Ordovician stage, marked by the base (FAD) of particular graptolite or conodont zones. Thus, dissonance occurred –horizons (regional stages) in Early Paleozoic regional charts are defined by benthic associations, while ICC stages position their lower boundaries based on pelagic groups. So, two scales –global (ICC) and regional– are made using different parameters. It is evident that evolutionary rate of pelagic fauna is higher than that of benthic fauna and the chance that their boundaries will coincide at evolutionary stages of pelagic and benthic fauna is extremely low. So, a need of independent existence and specific mission of both charts (global ICC and regional) seems to be justified.

Boundaries of regional Ordovician stages in the Gorny Altai area could be aligned by means of graptolite and conodont zones with global stage and substage boundaries of the ICC (Bergström et al., 2009) as follows: 1) the lower boundary of the Tayanza Regional Stage is aligned with the lower Tremadocian boundary; 2) the lower boundary of the Lebed Regional Stage with the lower Floian

boundary; 3) the lower boundary of the Kostinsky Regional Stage with the lower boundary of the middle Darriwilian Substage; 4) the lower boundary of the Bugryshikha Regional Stage with the lower boundary of the upper Darriwilian Substage; 5) the lower boundary of the Khankhara Regional Stage with middle of upper part of the lower Sandbian Substage; 6) the lower boundary of Tekhten Regional Stage with the lower boundary of the second Katian Substage.

CONCLUSION

Integration of the regional stages (horizons), plus zonation on orthostratigraphic fauna (graptolites, conodonts, chitinozoans), allow the solving of specific problems. Among them: 1) stratification of local sections; 2) correlation of sections (intra-regional and partly inter-regional) of generalized sections from adjacent regions; 3) age dating (inter-regional correlation of sections of distant regions and global correlation including that with IGTS units); 4) definition of age of sedimentary and biotic events, such as black-shales, transgressions and regressions, reefal sediments, extinction and abrupt change of taxonomic composition; and 5) alignment of regional sedimentary and biotic events with levels of global events.

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TRACES OF THE GLOBAL AND REGIONAL SEDIMENTARY EVENTS IN EARLY ORDOVICIAN SECTIONS OF THE GORNY ALTAI (SIBERIA)

N.V. Sennikov¹, O.T. Obut¹, E.V. Bukolova¹ and T.Yu. Tolmacheva²

¹ Trofimuk Institute of Petroleum Geology and Geophysics SB RAS, Acad. Koptug av. 3, 630090 Novosibirsk, Russia.

SennikovNV@ipgg.nsc.ru, ObutOT@ipgg.nsc.ru, BukolovaEV@ipgg.nsc.ru

² VSEGEI, Sankt-Petersburg, Russia. Tatiana_Tolmacheva@vsegei.ru

Keywords: Early Ordovician, global and regional events, Gorny Altai.

INTRODUCTION

Successive short-term abiotic global-scale events are defined in different regions of the world in the Late Cambrian–Early Ordovician. They are detected in the lithological record in relatively short stratigraphic interval, usually combined with biotic events marked by structural and composition change in paleobiota (Walliser, 1986). The identification of events in the studied region that could be regarded as global, can only be done through the application of zonal stratigraphy of the local sequences.

GLOBAL EVENTS

Six global events of an eustatic nature have been established worldwide for late Cambrian–Early Ordovician epochs.

- (1) Terminal Late Cambrian Lange Ranch regressive event (LRRE): was designated on the basis of material from North America and China (Miller, 1984) and coincides with base of *Cordylodus proavus* conodont Zone.
- (2) First Early Ordovician Acerocare regressive event (ARE) (Erdtmann and Miller, 1981; Erdtmann, 1986): was defined in carbonate shelf sequences of North American, Siberian and Chinese platforms and aligned with the base of the *Iapetognathus fluctivagus* conodont Zone and the base of the *Rhabdinopora flabelliformis parabola* graptolite Zone. Previously the ARE event was believed to correspond to LRRE eustatic event (Erdtmann and Miller, 1981).
- (3) Black Mountain Transgressive event (BME): linked to the base of the *Cordylodus angulatus* conodont Zone and defined in Australia (Miller, 1984).
- (4) Peltocare Regressive event (PRE) (Erdtmann, 1986): was established in the Baltic paleobasin and correlated with the lower boundary of the *Adelograptus tenellus* graptolite Zone.
- (5) Kelly Creek Regressive event (KCE) (Nicoll et al., 1992): was designated in Australia below the base of the *Paroistodus proteus* conodont Zone.

(6) Ceratopyge Regressive event (CRE) (Erdtmann, 1986): was established in North America and positioned in the upper Tremadocian, middle part of the *Araneograptus murrayi* and *Paroistodus proteus* graptolite and conodont zones respectively.

At the Tremadocian/Floian (Arenigian) boundary a global biotic Basal Arenig Bio-Event (BAgB) event (Walliser, 1986) is defined. It is aligned with the base of the *Tetragraptus approximatus* graptolite Zone.

Graptolite and conodont data from the Cambrian/Ordovician boundary of Altai (Fig. 1) allow identify and on the basis of zones and subzones precisely calibrate such events, as well as align them with known global events (Sennikov, 1994; Iwata et al., 1997; Sennikov et al., 2003, 2004, 2008) (Fig. 2).

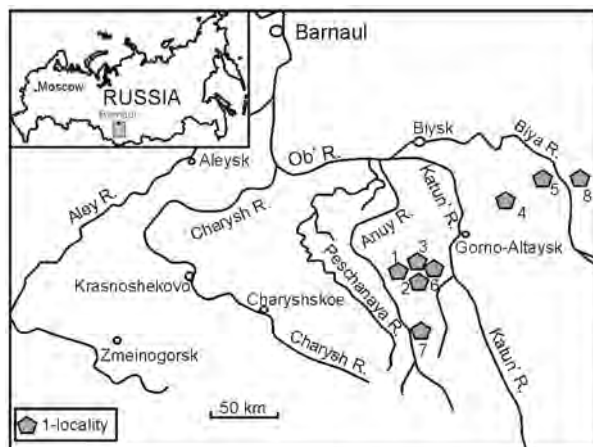


Figure 1. Localities of the Early Ordovician events in the Gorny Altai (for explanation, see text).

System	Series	Stage	Graptolite zones	Conodont zones
Ordovician	Lower	Floian	<i>P. angustifolius elongatus</i> / <i>E. broggeri</i>	<i>O. evae</i>
			<i>P. densus</i> / <i>A. ballicus</i>	<i>P. elegans</i>
		Tremadocian	<i>T. approximatus</i>	<i>P. proteus</i>
			<i>T. osloensis</i> / <i>A. hyperboreus</i>	No data
Cambrian	Upper	Batyrbaiian (Stage 10)	<i>C. tenellus</i> / <i>K. klaeri</i>	No data
			No data	<i>lapetognathus</i>
			No data	<i>C. proavus</i>
				No data
				<i>C. lindstromi</i>
				No data
				No data
				<i>C. minutus</i>
				<i>E. notchpeakensis</i>

Figure 2. Early Ordovician graptolite and conodont zones for the Gorny Altai.

In the Gorny Altai, the LRRE (end of Batyrbaian) is provisionally located at the base of lower Kamlak Subformation, where conodonts of *C. proavus* Zone were recovered (Fig. 1, loc. 1; Fig. 3). However, the lower boundary of this unit is tectonic and a lithological marker for the LRRE has not been found yet.

The second regressive event, ARE, is designated within the Kamlak Formation, in the northern part of Altai shelf basin, where it coincides with the lower/middle subformation boundary (Fig. 1, loc. 2; Fig. 3). At the base of the middle subformation there are three consecutive 30 m-thick conglomerate beds. These conglomerates contain middle-sized, poorly sorted, well-rounded pebbles, which compose 50% of the rock. Limestones from the middle part of middle Kamlak Subformation in the "Kamlak" section yielded conodonts of the *lapetognathus fluctivagus* Zone.

Presence of the next transgressive event, BME, in the Gorny Altai was not proved by paleontological data, since conodonts of the *C. angulatus* Zone are absent. But, within the coarse-grained grey-colored beds of the shallow-water upper Kamlak Subformation, as well as in the Choya Formation, deep-water black shales are observed, that could be regarded as BME traces (Fig. 1, loc. 3, 4; Fig. 3).

The regressive event PRE is defined in the Ishpa Formation (Fig. 1, loc. 5; Fig. 3). It could be recognized by the appearance of coarse-grained sandstones and mudstones in the upper part of formation overlying mainly organogenic limestones. Graptolites of the *Ad. tenellus* Zone were identified at the "Perevalnyi" section, in the upper part of Ishpa Formation.

The fifth regressive event, KCE, was defined in the upper Kamlak Subformation based on the presence of basal conglomerates up to 170 m thick (Sennikov et al., 2008) (Fig. 1, loc. 6; Fig. 3). Conglomerates

have large- and middle-size, poorly sorted, well-rounded pebbles, which constitute 60-70% of the rock. Graptolites of the *T. osloensis*/*Al. hyperboreus* Zone, aligned with the *Kiaerograptus* Zone, were recovered from the sandstones that overlie the conglomerates. The Baltoscandian standard *Kiaerograptus* Zone is below the base of the *Par. proteus* conodont Zone, which will make it possible to identify the KCE when graptolites of this zone are found.

CRE, the last regressive event in the Gorny Altai, is identified at the base of the Marcheta Formation by the appearance of red-color terrigenous rocks and red cherts (Fig. 1, loc. 7; Fig. 3). Terrigenous rocks from the underlying Talitsa Formation are mainly grey-colored, rarely lilac, and cherts are mainly violet and red-brown. In the "Marcheta-2" section, conodonts of the middle part of *Par. proteus* Zone were recorded.

In the Altai shelf basin, the biotic BAgB event and, thus, the global transgressive event, is identified in the lower part of the Tuloi Formation (Fig. 1, loc. 8; Fig. 3). This formation overlies, with dip and azimuthal unconformability and thick (up to 130 m) basal conglomerates, different Cambrian horizons: the Lower Cambrian in the "Lebed" section, the Middle Cambrian in the "Tagaza" section, and the Upper Cambrian in the "Tandoshka" section. It also overlies the Lower Ordovician (Tremadocian in the "Ishpa" and "Tuloi" sections). Conglomerates have large-, occasionally middle-size, poorly sorted, well-rounded pebbles, which occupy 80-89% of the rock. Graptolites of the *T. approximatus* Zone were recovered from overlying terrigenous strata at the lower part of the formation in the "Tuloi", "Lebed" and "Tagaza" sections.

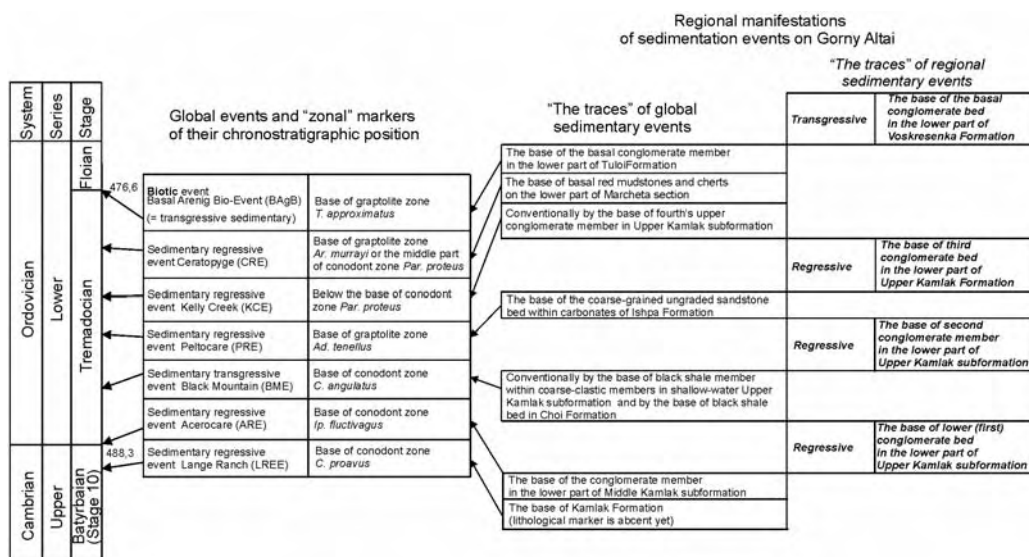


Figure 3. Early Ordovician global and regional events in the Altai paleobasin.

REGIONAL EVENTS

During the Ordovician, the Altai paleobasin developed at an active continental margin (Sennikov, 2003; Sennikov et al., 2008). This resulted in large-scale deepening and rising of the paleobasin. The 6 global events designated in the Gorny Altai sections and discussed above were subjected to regional movements (Fig. 3). Three additional Tremadocian-Floian regional regressions and one transgression, defined in the

Gorny Altai sections, between the global sedimentary events, reflect frequency of such regional dislocations.

As all known Early Ordovician global events (together with their biostratigraphic position and lithological manifestation) were recognized in the Gorny Altai, it is possible to suggest that rates of regional deepening (rising) of the Altai paleobasin were considerably less than those of global eustatics and could not grade its consequences.

CONCLUSIONS

Conodont and graptolite zonation established on the basis of faunal associations recovered in the Gorny Altai is composed of zones whose lower boundaries correspond to chronostratigraphic position of global sedimentary and biotic events.

The number of events, some of regional scale, recognized in the Altai Late Cambrian–Early Ordovician basin are larger than the globally known sedimentary events. This allows us to suggest that the Altai paleobasin at this time was subjected to much more large-scale episodic risings and deepenings than had been believed before. The reasons for these movements are still to be investigated.

Acknowledgements

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CONODONT BIODIVERSITY DYNAMICS FROM THE ORDOVICIAN OF BALTOSCANDIA

H.D. Sheets¹, D. Goldman², S.M. Bergström³ and C. Pantle²

¹ Department of Physics, Canisius College, Buffalo, New York, 14208. sheets@canisius.edu

² Department of Geology, University of Dayton, Ohio, 45469. dan.goldman@notes.udayton.edu

³ School of Earth Sciences, The Ohio State University, Columbus, Ohio, 43210. stig@geology.ohio-state.edu

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INTRODUCTION

Goldman et al. (this volume) used a quantitative stratigraphic correlation and seriation method, Constrained Optimization (CONOP9, Sadler et al., 2003), to construct a high resolution correlation model and composite range chart from the stratigraphic range data of 159 conodont species in 24 boreholes and outcrops around Baltoscandia (Fig. 1). We converted the CONOP9 generated composite section into a timescale by assigning the absolute ages of conodont biozone bases (from Webby et al., 2004) to the first appearance datums of the key conodont index taxa in the composite, and then scaling it appropriately (see Goldman et al., this volume, figure 2). We then tabulated presence/absence data for each species at every collection horizon in all 24 sections, and projected each of those collection horizons back into the composite timescale. Finally, we subdivided the timescale into 60 685 ky intervals (a temporal resolution approximately one fourth that of the median conodont zone duration) spanning the *Paltodus deltifer* through *Amorphognathus ordovicianus* conodont zones. Within this binned timescale we calculated conodont biodiversity, origination rates, and extinction rates from the middle Tremadocian to the Hirnantian. We refer the reader to Goldman et al. (this volume) for a brief history of conodont research in Baltoscandia, a short discussion of the Ordovician stratigraphic setting, and a description of CONOP9 methodology.

BIODIVERSITY CALCULATIONS

A range of different measurements drawn from the presence/absence data for every conodont species in each binned time interval within the composite time scale are useful in understanding the evolutionary history of conodonts within Baltoscandia. The simplest approach is simply to plot the number of observed taxa per bin, a simple biodiversity plot. In addition to the observed taxa within each interval, it is also possible to plot all extant taxa within each bin, as a number of species will range-thru an interval without being observed. The

difference between these two plots is informative as to the relative completeness of the sampling, or the fossil recovery rate. The more exhaustive and complete the sampling, the more similar these two plots will be, whereas gaps between the two indicate a failure in the process of preservation and recovery of multiple taxa. This failure may be due to a wide range of factors, including variation in collection effort, lack of appropriate depositional environments at some times, variations in water depth at one or more localities or shifts in ocean circulation patterns that might make individual localities inhospitable for certain taxa. It is simplest to refer to all of these factors as a single variable, fossil recovery rate. It is

quite clear that if the fossil recovery rate is low, or is highly variable over time, then estimates of extinction and origination that do not take recovery rate into account will confound failure to recover specimens with extinction or origination.

Rates of extinction and origination were thus calculated using both simple approaches which do not incorporate estimates of fossil recovery rate, using metrics discussed by Foote (2000), and also more complex approaches based on Capture-Mark-Recapture (CMR) models (Connolly and Miller, 2001; Chen et al. 1995; Liow and Nichols, 2010) as developed for use in wildlife monitoring (Pollock et al., 1990; Lebreton et al., 1992; Pradel, 1996) which produce simultaneous, independent estimates of extinction, origination and fossil recovery rate. CMR methods are widely used in modern ecological monitoring and adapt well to paleontological data, effectively separating fossil recovery rates from biodiversity estimates.

CMR models use maximum likelihood methods and require a series of steps to arrive at estimates of fossil recovery rate, extinction and origination. In our application, we consider a series of different CMR models, in which rates of recovery, extinction and origination can either be constant over all bins, or change from bin to bin. The first step in using such models is to determine if the most complex model (in which all parameters vary in each bin, a fully time dependent model) has enough descriptive power to describe the observed biodiversity pattern, an assumption that might not be true if there were two subgroups of taxa with very different biodiversity patterns, or if risk of extinction depended strongly on taxon duration. Goodness of Fit testing is done using a Monte Carlo simulation (Cooch and White, 2001) to determine if

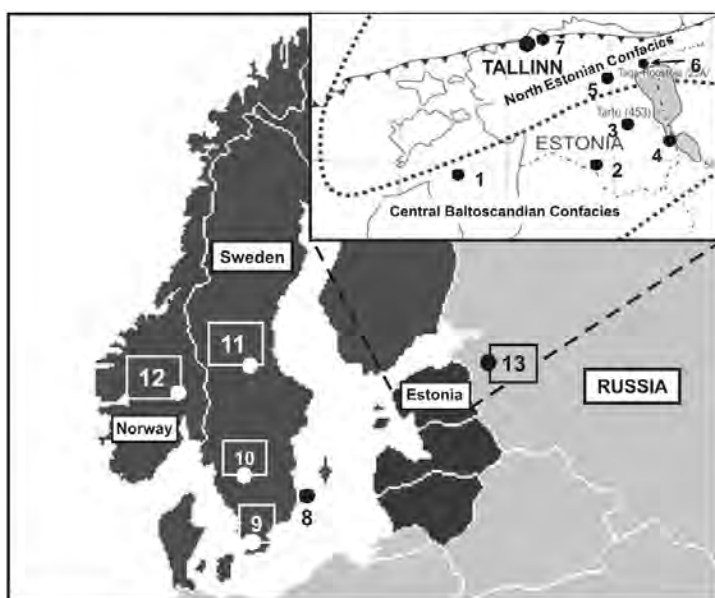


Figure 1. Locality map for outcrops and boreholes in Baltoscandia. Closed circles on the Baltic States inset map are borehole locations; saw-toothed line delineates the present extent of Ordovician carbonates; and the dotted lines represent boundaries of the confacies belts. Numbered boreholes are: 1) Ruhnu; 2) Valga; 3) Tartu; 4) Mehikoorma; 5) Kerguta; 6) Taga-Roostoja; 7) Maekelda. Numbered outcrops are: 8) Öland; 9) Scania; 10); Västergötland; 11) Siljan Region; 12) SC Norway; 13) Putilivo Quarry and Lava River, Russia. Estonia map modified from Modirirski et al. (2002).

the model fits the actual data as well as it fits simulated data generated using the model in a Monte Carlo process. If the model is a substantially worse fit to the real data than to simulated data, it is evidence of model failure.

If the fully time dependent model is judged to fit the data, it is then necessary to consider other possible models that describe the dynamics of biodiversity change. Recovery, extinction and origination might each vary with time, changing in each bin, or they might be constant. Statistical approaches to model choice are popular adjuncts to CMR methods, and although the mathematics can rapidly become complex, the basic ideas are straightforward (Chamberlin, 1890; Aikake, 1973; Burnham and Anderson, 1998). When we consider models of the physical world, there is generally a trade-off between descriptive power, meaning that the model closely fits a given data set, and generality, meaning that the model could be expected to fit well to newly added data, or a newly collected set of data. In a model ranking procedure, one generates a set of candidate models, in this case all combinations of constant and varying recovery, extinction and origination rates, and then ranks the models based on a measure of how effectively the model matches the data, balancing descriptive power and generality. Statistical model choice allows the simultaneous comparison and evaluation of a wide range of models, and can reject simpler models in favor of more complex models, as well as indicate when two or more models should be considered viable, outcomes which are not possible with simpler and more familiar statistical hypothesis testing approaches (Burnham and Anderson, 1998). The Aikake Information Criteria (AIC, Aikake, 1973) has been used in a number of paleontological studies (Wagner et al., 2006; Zambito et al., 2008; Handley et al., 2009), and the AIC_c measure, which provides a quantitative statistic for choosing among competing models and is effective at small sample sizes, was used here. Other approaches to model choice include cross validation and Bayesian Information methods (Ivany et al., 2009). The model choice procedure is useful in that it produces the best available estimates of rates (given the available models and data), and is informative as to whether rates are constant or variable.

Error bars for the rates of fossil recovery, extinction and origination are then estimated by bootstrapping (re-sampling with replacement) the presence/absence data of the individual taxa in the binned composite (Chen et al., 1995). The error bars thus reflects the variation due to collection density, but not the uncertainty in the composite time scale.

STATISTICAL RESULTS

An examination of the plot of observed taxa and standing biodiversity (Fig. 2A) indicates that the sampling in this data set was very complete in most intervals after the Tremadocian, as the curves for extant taxa and observed taxa are almost identical. Slight dips in sampling intensity are visible in the Floian, end-Darriwilian and Sandbian.

The most complex, fully time-dependent CMR model did satisfy the goodness of fit (GOF) test, with a *p* value of 0.90 (excluding the first 15 bins, which have extremely low sample sizes per bin) based on 100 Monte Carlo simulations. The AIC model choice procedure indicated that a model with a constant extinction rate, but time-varying origination and fossil recovery rates was the best description of the data (Table 1). The limited sampling in the first 15 bins indicates the CMR model should be viewed cautiously in this interval, as there was not enough data to determine if the model described the data well.

Extinction	Origination	Recovery	AIC	Deviation in AIC	AIC Wt.
Constant	Constant	Constant	3032.68	331.154	0
Constant	Varying	Constant	2887.04	185.509	0
Varying	Constant	Constant	2902.84	201.304	0
Varying	Varying	Constant	2879.59	178.063	0
Constant	Constant	Varying	2728.84	27.310	0
Constant	Varying	Varying	2701.53	0	1
Varying	Constant	Varying	2725.32	23.791	0
Varying	Varying	Varying	2762.23	60.694	0

Table 1. Results of AIC-based model choice. The model with the lowest AIC score is the closest to the observed data in terms of information content. The AIC weight (AIC Wt.) for each model is the relative probability that the model is true compared to the model with the lowest AIC value. In this case, a single model is overwhelmingly favored, as all other models have weights well below 1×10^{-4} .

The simple metric and the CMR model both indicated very similar patterns of origination rate (Fig. 2B). There is a strong apparent peak at the Tremadocian/Floian boundary, but this is probably due to the increased number of collections at this point, rather than a biological effect. Examination of the estimated extinction rate (Fig. 2C) shows the expected constant rate estimated via the CMR method and a slightly varying rate from the simple metric, which appears to oscillate slightly about the CMR estimated mean rate.

The recovery rate estimated from the simple metric (observed taxa/extant taxa) is strikingly similar to the CMR estimate of recovery rate, and indicates the high degree of uniformity in sampling, particularly after the end of the Tremadocian. There are some substantial gaps in sampling however, indicating the importance of using CMR methods to understand the interaction of recovery rates with the other biodiversity estimates.

DISCUSSION

Sweet (1988) produced a global Ordovician conodont biodiversity curve that serves as an interesting point of comparison for our Baltoscandian curve (Fig. 2D). In our analyses Baltoscandian conodont faunas exhibit low diversity in the Tremadocian with a rapid rise near the base of the Floian. As noted above, this dramatic increase may be due in part to large increase in the number of sections and collections that were available for inclusion in our analysis. Biodiversity continues to rise, although more slowly, from the *Prioniodus elegans* Zone to the lower Darriwilian. Generally, the curve exhibits a broad plateau from the middle Floian to the early Sandbian with a central peak in the uppermost *Baltoniodus norrlandicus* Zone (a maximum of 39 species). Conodont diversity then gradually declines from the lower Darriwilian throughout the rest of the Ordovician Period. During this long decline a nadir (5 species) is reached in the upper part of the *Amorphognathus tvaerensis* Zone (*Baltoniodus alobatus* Subzone) before rebounding slightly in the *Amorphognathus superbus* Zone (14 species). Interestingly, this upper *Amorphognathus tvaerensis* Zone diversity low is also approximately coincident with the Estonian Oandu Stage, an interval that also exhibits low diversity in organic walled microfossils (Kaljo et al., 1995). The decline in faunal diversity then continues through the *Amorphognathus ordovicicus* Zone.

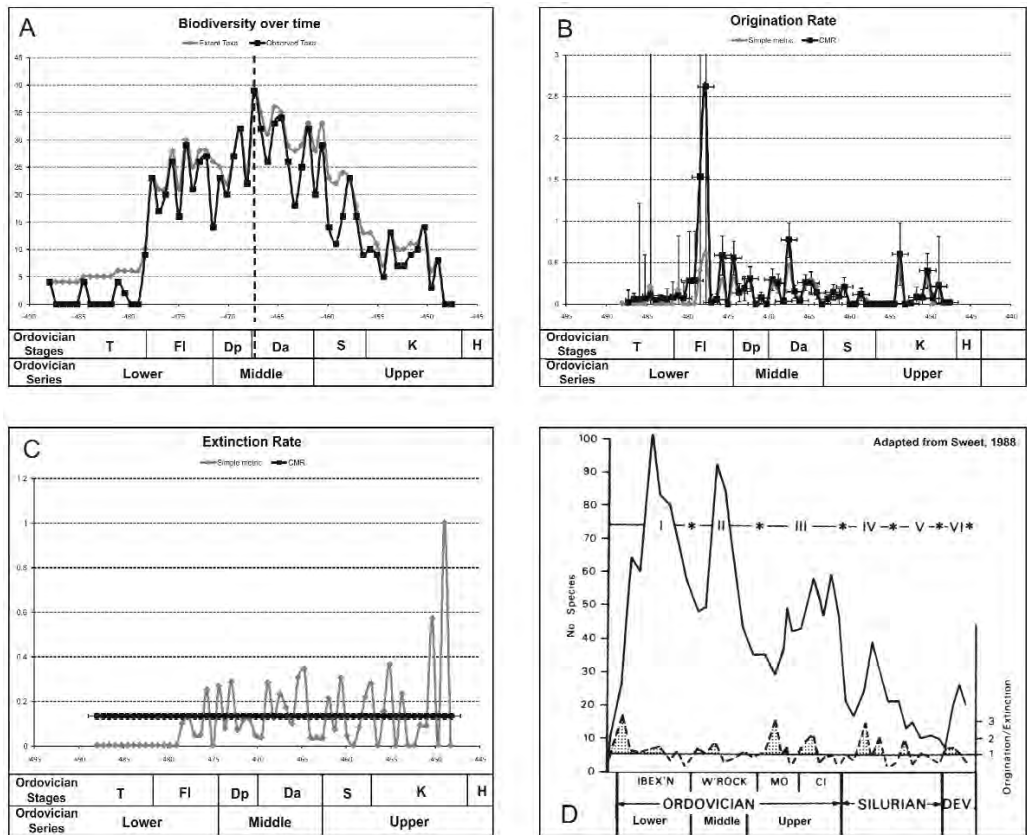


Figure 2. Biodiversity dynamics of Ordovician conodonts from Baltoscandia. Biodiversity, origination rates and extinction rates are calculated in 649 Ky bins. A) Biodiversity curve showing actual counts (diamonds) and range-through taxa (squares). The close correspondence of the two curves indicates relatively complete sampling in most intervals. B) Origination rates. Simple Foote (2000) metrics are represented by diamonds and CMR estimates by squares. The basal Floian peak is partially (but not completely) an artifact of an increase in collections at that point. Note that the rates fluctuate substantially over time and that the two curves show a close correspondence. The CMR model indicates that origination is time dependent, and declines through much of the late Middle and Late Ordovician. C) Extinction rates. Simple Foote (2000) metrics are represented by diamonds and CMR estimates by squares. The best CMR model indicates that extinction rates are generally constant across most of the Ordovician. D) Global Ordovician conodont diversity. Adapted from Sweet (1988). Note that the greatest global diversity occurs in the middle Ibeixian (middle Tremadocian) a time of low diversity in Baltoscandia.

Sweet's (1988) global biodiversity curve differs from ours in some significant ways (Fig. 2D). Baltoscandian biodiversity lacks the large Early Ordovician (mid Tremadocian) peak that is exhibited in Sweet's (1988) global curve, a reflection of the abundant and species-rich faunas that are found in Laurentia at this time (e.g., Repetski, 1982; Sweet and Tolbert, 1997; Ethington and Clark, 1964). Baltoscandian peak diversity is mirrored by the second major peak in Sweet's curve, although it is difficult to precisely assess the exact correspondence of the global and Baltoscandian peaks. Finally, Sweet's (1988) curve exhibits a substantial and relatively long lasting diversity rebound from the middle Mohawkian through the Cincinnati North American stages (late *Amorphognathus tvaerensis* Zone through the *A. ordovicicus* zones) a feature that is far more subdued and short-lived in our curve.

Hammer (2003) published an Ordovician Baltic biodiversity curve for conodonts that looked broadly similar to our curve but differed in some small but interesting ways. Hammer's (2003, text-fig. 3D) curve also had a single large diversity peak in the Darriwilian, but at a slightly younger time than ours (approximately 2 million years younger). Hammer's curve also shows a much steeper post-acme decline than does our curve, with diversity plummeting in the late Darriwilian. Hammer's (2003) curve also exhibits a rather large Katian diversity rebound, a feature that he notes, however, may be an artifact of sampling. We attribute the variation in our diversity patterns to the different ways we counted taxa and correlated the individual conodont successions.

Perhaps the most interesting outcome of our analyses was the fact that conodont diversity changes across the Middle and Late Ordovician appear to be driven by depressed origination rates. The estimated extinction rate derived from CMR methodology shows a constant value for most of the Ordovician (Fig. 2C), only rising substantially in the Hirnantian - a rise due in part to edge-effects, which are an artifact of reaching the upper margin of our data set. Both the simple metric calculations and the CMR model indicate, however, that origination rates are time-dependent, with high values in the Lower and lower Middle Ordovician that decline after the lower Darriwilian (Fig. 2B). Small increases occur in the early and late Katian, which account for the minor diversity rebound at this time. It is also interesting to note that this depressed origination begins just about the same time as a prominent mid-Darriwilian positive carbon isotope excursion (MDICE, Kaljo et al., 2007), although the relationship between conodont evolution and the isotopic excursion (or any possible environmental changes that it represents) remains unclear.

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THE DISTRIBUTION OF GONDWANA-DERIVED TERRANES IN THE EARLY PALEOZOIC

G.M. Stampfli¹, J. von Raumer² and C. Wilhem¹

¹ Institut de Géologie et Paléontologie, UNIL, CH-1015 Lausanne, Switzerland. gerard.stampfli@unil.ch

² Dept. of Geosciences, University of Fribourg, Ch. du Musée 6, CH-1700 Fribourg, Switzerland. juergen.vonraumer@unifr.ch

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INTRODUCTION

The present day Variscan basement areas of Europe have been recognized as generally derived from Gondwana, based on fauna, facies and detrital zircons distributions. These areas represent only a portion of Europe (Iberia, France, Central Europe), but it is obvious that similar Variscan basements are present in the whole Alpine and Mediterranean areas too, up to the Caucasus. In tracing these terranes back to their possible position around Gondwana, it became obvious that they could not all be positioned north of Africa, as generally shown on reconstructions. We developed the concept of a ribbon like Galatian superterrane that comprised most of these "European" Variscan elements. In Ordovician times, this superterrane extended from the north of South America to South China (located in continuity to Africa). Along such a length, the geodynamic evolution was not the same, but presents strong similarities. Geodynamic scenarios for the whole Paleozoic have been developed for the different segments, thus allowing us to re-distribute the subterranees in a coherent way. The diachronous openings of the Rheic *s.l.* ocean, then of the Paleotethys, represent the main reconstruction guidelines, together with major magmatic activity distribution in space and time.

LATE CADOMIAN CYCLE AND GEODYNAMIC CONSTRAINTS FOR THE NORTH CHINA DOMAIN

Before the Ordovician most of the Variscan areas had been affected by the Cadomian event, viewed as the accretion along the Gondwana margin of amalgamated arcs, some derived from Gondwana, some from the North China domain. This is supported by the numerous types of late Proterozoic magmatic rocks found in the Cadomian terrane assembly. This accretion to Gondwana was followed by a new pulse of magmatism all along the Gondwana margin in Late Cambrian Ordovician times, the margin evolving as a cordillera.

It is well accepted today that South China was a part of Gondwana since the Late Proterozoic (Wilhem, 2010). The overall geometry of Gondwana consisted of a right angle between the African segment and the

South Chinese segment (Fig. 1). Starting from the easternmost part of this cordillera, Cambrian opening of back-arc basin (Fig. 1A) resulted in the detachment from the South Chinese segment of the Qilian Terrane (e.g. thermal history, paleontologic and stratigraphic affinities between South China and Qilian) (Tung et al., 2007). In Late Cambrian (Figs. 1B and C), the arc migrated alone toward Baltica leaving the Qilian microcontinent behind surrounded by passive margins (e.g. platform deposits and North China type fauna) (Xu et al., 2006). During the Late Cambrian and Ordovician time, the migrating arc was colliding both with the Gondwana and North China margins, turning the latter margin from passive to active after subduction reversal (Sinian-Cambrian passive margin and beginning of arc magmatism in Late Cambrian) (e.g. Yang et al., 2001; Ratschbacher et al., 2003). A back-arc basin was then developed within the southern margin of North China (Figs. 1C and D) and formed the Qilian-Erlangping Terrane (Xia et al., 2003), which successively collided with the Qilian and future Hunian terranes (Figs. 1D and E) (formation of the Ordovician North Qaidam and North Qilian sutures) (e.g. Yang et al., 2006). This new-amalgamated terrane was accreted in turn to North China (Fig. 1F) in the Silurian (i.e. flysch, molasse, intrusions, deformation) and a new active margin took place under North China (i.e. Upper Silurian arc magmatism) (e.g. Ratschbacher et al., 2003).

FROM ACTIVE TO PASSIVE MARGIN, OPENING OF THE RHEIC

After accretion of the arc to North Gondwana, cordillera collapse took place north of Africa, whereas back-arc opening north of South-America had already triggered the detachment of Avalonia and the opening of the Rheic ocean *s.str.*. The margin setting along Gondwana (von Raumer et al., 2002) is characterised by an active margin since the early Ordovician, and the subsequent opening of the Rheic ocean, after a period of subsidence and rifting behind Avalonia-Hunia (e.g. von Raumer and Stampfli, 2008). Consequently, the basement areas of these regions show a strong activity of crustal extension and rifting during the early Ordovician, accompanied by the intrusion of granitoid rock series at different crustal levels (e.g. Ollo de Sapo granitoids; Montero et al., 2007; Bea et al., 2010). The detachment of Hunia from Gondwana gave birth to the eastern branch of the Rheic ocean, slightly younger (c. 460 Ma) (i.e. constraints coming from the geodynamic scenario for the North China Domain) than the western branch (c. 480 Ma). In the eastern part of the Gondwana margin, comprising among others the Alpine domain, the period of Ordovician active margin setting started later than in the west. The stepwise magmatic evolution in the Austroalpine basement with granitoids (orthogneisses) and mafic rock suites (metagabbros, metabasites, eclogites) (Schulz et al., 2008) began with a Cadomian (550-530 Ma) volcanic arc basalt mafic suite with Th/Yb typical of subduction-related magmatism, and subsequent 470-450 Ma old I- and S-type granitoid intrusions, followed by alkaline within-plate basalt to MORB-type mafic suites around 430 Ma (eastern Rheic spreading). In the external domain of the Alps, the magmatic evolution of the Ordovician active margin is equally documented by the intrusion of granitoids and mafic rock suites between 470-450 Ma (e.g. Bussy et al., this volume).

Thus, the overall geodynamic scenario can be followed through the cessation of magmatic activity north of Gondwana and the diachronous onset of passive margin settings during the Ordovician. Geometries and velocities of tectonic plates at that time are also strongly constraining the origin of Avalonia and Hunia. Avalonia had to be accreted to Baltica-Laurentia and Hunia to North China during the Silurian (Fig. 1).

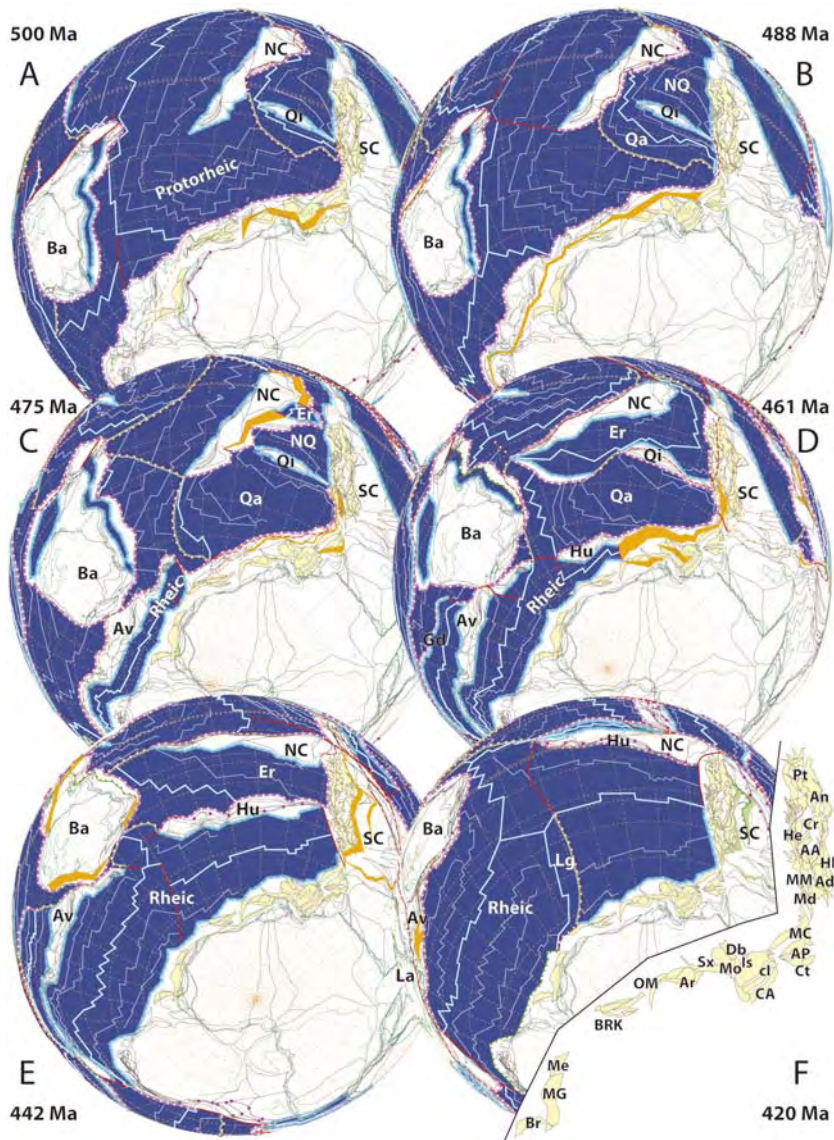


Figure 1. Global reconstructions for the Late Cambrian to the Silurian. Av, Avalonia; Ba, Baltica; Er, Erlangping; Gd, Gander arc; Hu, Hunian; Lg, Ligerian arc; NC, North China; NQ, North Qilian ocean; Qa, Qaidam ocean; Qi, Qilian; SC, South China. The Galatian superterrane (F inset) is made of 4 sub terranes, from bottom to top. The Meguma terrane: Br Brunswick; MG Meguma; Me, Moroccan Meseta. The Armorica terrane: BRK, Betics-Rif-Kabbilies; OM, Ossa Morena; Ar, Armorica s.str; Sx, Saxothuringia; Mo, Moesia; Db, Dobrogea; Is, Istanbul. The Ibero-Ligerian terrane: cl, central Iberia; CA, Cantabria; Ct, Cataluña; AP, Aquitaine Pyrenees and Corsica; MC, Massif Central; Md, Moldanubian. The intra-Alpine terrane: MM, Montagne Noire-Maures; Ad, Adria and Sardinia; AA, Austroalpine; He, Helvetic; Cr, Carpathian; HI, Hellenidic; An, Anatolic; Pt, Pontides. Along the Eurasian margin, the opening of the Rhenohercynian ocean has detached the Hanseatic terrane from the mainland: eM, eastern Meseta; Po, south Portuguese; Ch, Channel; MR, mid-German rise; CC, Caucasus. Major rifts are shown in darker grey.

OPENING OF PALEOTETHYS

We are departing here from our previous model (Stampfli et al., 2002) where Hunia was considered as the main ribbon like microcontinent leaving Gondwana during the opening of Paleotethys in the Silurian. The Silurian accretion of Hunia to North China implies that this accretion took place when the Paleotethys was not yet opened. Thus, Hunia represents a first train of terranes leaving Gondwana more or less at the same time than Avalonia (during the Ordovician). The second train of terranes leaving Gondwana in the Devonian has been called the Galatian superterrane (von Raumer and Stampfli, 2008).

In the late Ordovician, both western and eastern segments of the Rheic made a single oceanic domain. North of Africa, the passive margin of Gondwana became again an active margin during the Devonian. This followed the collision of the margin with an intra-oceanic arc (Ligerian) and the obduction of part of the back-arc oceanic crust, followed by subduction reversal (Fig. 2). This is well recorded by HP metamorphism corresponding to the eo-Variscan tectonic event (from c. 400 Ma to c. 370 Ma), and the emplacement of Devonian ophiolites along the Gondwana margin in Spain, France and Central Europe. In this suture are also found remnants of older oceans, either the Ordovician Rheic ocean (c. 460 Ma and younger) or older fragments (c. 500 Ma, e.g. Arenas et al., 2007, 2009) related to the Qilian arc and brought to the surface during the rifting and detachment of the Hun terranes, thus forming the toe of the Gondwanan Ordovician passive margin.

From the upper Ordovician to the Silurian, crustal extension is observed along the Gondwana margin through the sedimentary record (e.g. Schönlaub, 1997), the subsidence patterns, the interruption of sedimentation and the intrusion of basic volcanics at different places (von Raumer et al., 2008). New monazite age-data (Schulz and von Raumer, in press) confirm an early Silurian thermal event for the Aiguilles Rouges area. Located along the S-Chinese (Gondwana) margin this area is the witness of the transform type Rheic margin. The emplacement of 450 Ma gabbros at different places and the many early Silurian acidic volcanics of the Noric Terrane, again, are the signature of an extending crust in the Alpine domain; the older ones (450-420 Ma) are related to the eastern Rheic opening, the youngest (410-380 Ma) to the opening Palaeotethys (von Raumer et al., 2011).

THE GALATIAN TERRANE ACCRETION TO EURASIA

The Galatian superterrane was detached from Gondwana in segments, starting from the west, north of N-Africa with the detachment of the Armorica *s.l.* segment around 400 Ma, then the Ibero-Ligerian fragment after the eo-Variscan collisional event (c. 390 Ma) and the Intra-alpine segment just after (c. 380 Ma). A triple junction was established around the Arabian promontory, corresponding to the three branches of Paleotethys. The Iranian seaway separated the Iranian-Afghan domain from South China, the Sulu-Dabie seaway separated South China from the intra-Alpine terrane, and the N-African seaway separated Gondwana from Armorica-Iberia.

These oceanic branches were back-arc basins that merged to give the Devonian Paleotethys. During their drifting the Iberian-Intra-Alpine segment passed behind the Armorican one. This imbrication was even exaggerated when Armorica collided with the Hanseatic arc detached from Eurasia in the late Devonian. This arc extended from New-Foundland up to the Caspian area, the back-arc basin is represented by the Rhenohercynian oceanic domain in the west and the Paphlagonian pelagic domain in the east (e.g. Stampfli and Kozur, 2006). The Hanseatic arc is represented by terranes such as the S-Portuguese, Channel,

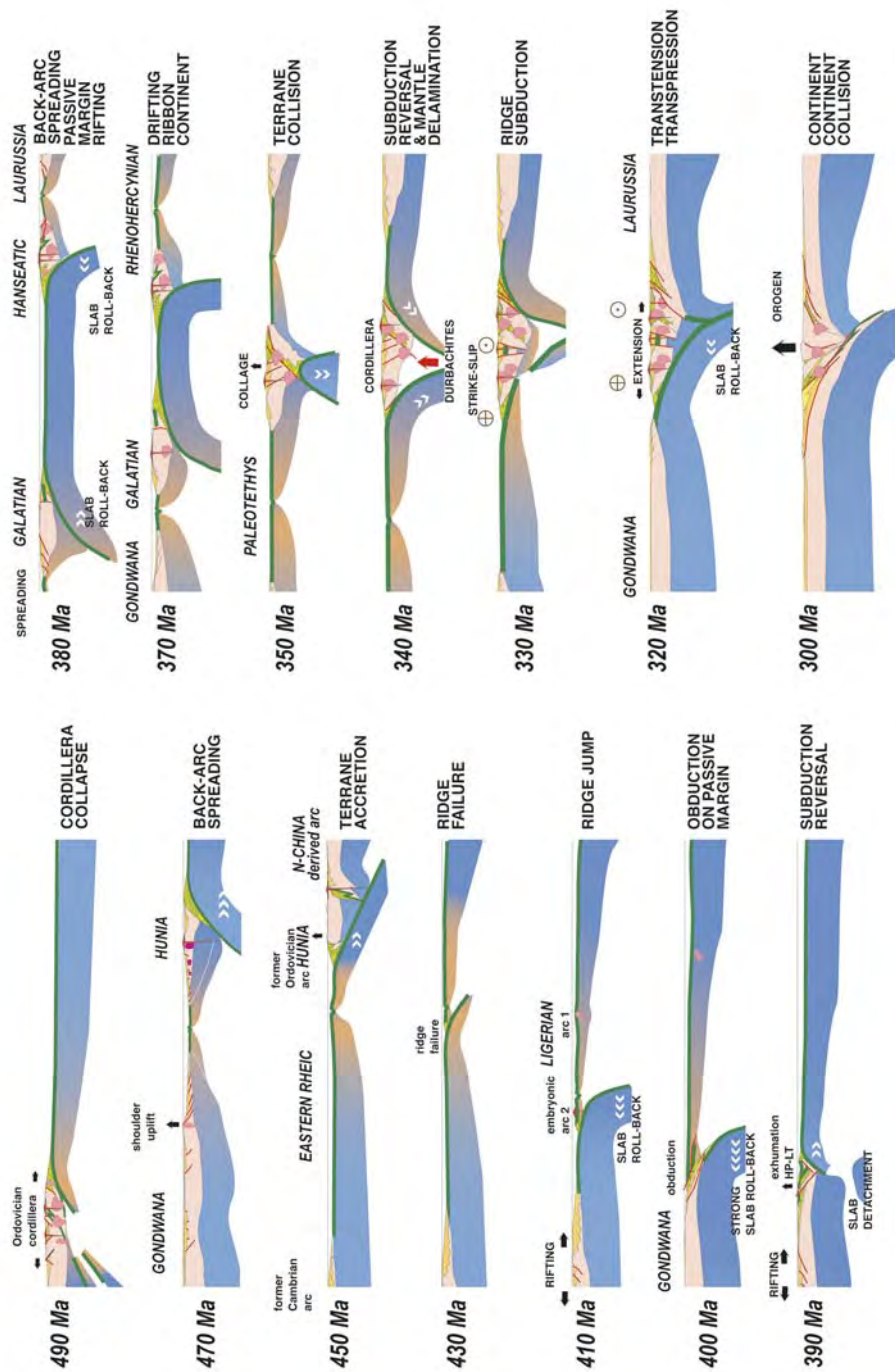


Figure 2. Cross section models of the evolution of the Gondwana margin from 490 to 300 Ma. The sections are tied to Gondwana, so the continent to the right is changing through times. For the left part of the figure North China elements are facing Gondwana, whereas for the right side of the figure it is Laurussia. The horizontal scale is not respected.

East-Meseta and Mid-German Rise, and part of the Caucasus and Black-sea in the east. The Hanseatic terranes were imbricated with fragments of the Armorican or Meguma terranes around the Iberian landmass. We follow here the imbrication model of Martínez Catalán et al. (2007), where the amalgamated Armorican and Rheohercynian terranes were indented by the Iberian promontory around 360-350 Ma. Terrane duplication took place further east too, due to the counter-clockwise rotation of Gondwana/Paleotethys with regards to Europe. Finally in the Late Carboniferous, Gondwana collided with the terranes accreted around Laurasia, given birth to the final Variscan tectonic event.

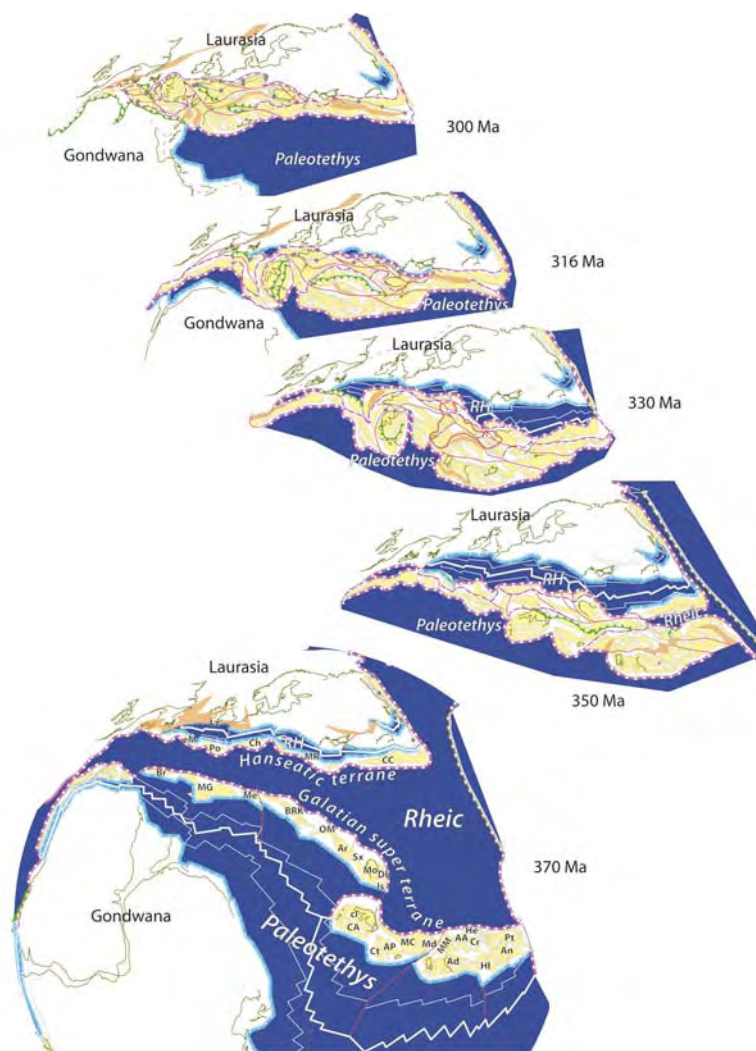


Figure 3. Reconstructions of the Variscan domain from 370 to 300 Ma, this model shows how the GDUs (light grey) implied in the Variscan collision can be restored into a single ribbon like superterrane before 370 Ma (see Fig.1). Main rift zones are shown in darker grey.

CONCLUSIONS

The duplication, rotation and oroclinal bending of the amalgamated Variscan terrane had to be deciphered before to be able to restore the Galatian superterrane geometry, then the Ordovician margin of Gondwana. Markers such as the Cadomian suture zones, the eo-Variscan obduction events, or the Paleotethys passive margin sequences were used to establish the former sub-linear geometry of the superterrane. Other aspects such as the distribution of Cambrian carbonates, the Hirnantian glacial deposits or the detrital zircon distribution were used to better constrain the position of the superterrane geodynamic elements (GDU) around Gondwana. The global reconstruction model and database elaborated at the Lausanne university (Hochard, 2009) using plate tectonic and synthetic isochrone principles (Stampfli and Borel, 2002) was of a great help in constraining geometries and plate velocities. These factors are fundamental when considering the wandering of a large plate such as Gondwana. But this would be useless without the repeated efforts of several generations of field geologists that gathered key information from the Variscan basement areas, their work is strongly acknowledged here.

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MIDDLE ORDOVICIAN BIVALVES FROM BOHEMIA, SPAIN AND FRANCE

M. Steinová

Czech Geological Survey, Klárov 3, Praha 1, 11821, Czech Republic.
marika.steinova@geology.cz

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INTRODUCTION

Middle Ordovician bivalves from Bohemia have been an overlooked group for a long time. In France Babin (1966, 1977, 1981), Bradshaw (1970), Babin and Robardet (1972), Babin and Mélou (1972) and Babin and Beaulieu (2003), and in Spain Babin and Gutiérrez-Marco (1985, 1991), and Gutiérrez-Marco et al. (1999) systematically revised the bivalves from the Ordovician. A modern revision of Middle Ordovician bivalves was also realized by Soot-Ryen and Soot-Ryen (1960) in Norway, Pojeta (1971) in North America, Babin and Destombes (1990) in Morocco, Sánchez (1990, 2003) in Argentina, Cope (1999) in Wales, Fang and Cope (2004), and Fang (2006) in China, and Sá (2008) in Portugal. In Bohemia they were described by Barrande (1881) and revised by Pfab (1934). Middle Ordovician bivalves from the Šárka Formation of Bohemia (lower and middle Darriwilian), are being revised by the author.

GEOLOGICAL SETTING

The Šárka Formation was deposited during the early and mid Darriwilian. The largest part of the Šárka Formation is developed in shale facies. Within the shales occur horizons with siliceous nodules containing a well-preserved and diverse fauna. Black shales prevail in the central parts of the Prague Basin, while sedimentary iron ores are present in onshore settings (Havlíček, 1998). Accumulations of volcanic and volcanoclastic rocks are also very common in the Šárka Formation.

MIDDLE ORDOVICIAN BIVALVES OF BOHEMIA

Barrande (1881) described nine species from the Šárka Formation (early to mid Darriwilian) of Bohemia. He allocated them to the genera *Nucula* and *Leda*, and defined the new genera *Babinka*,

Redonia and Synek. Pfab (1934) revised them and reassigned some of Barrande's species to other genera: *Praeleda* Pfab, 1934, *Praenucula* Pfab, 1934, *Pseudocyrtodonta* Pfab, 1934 and mainly to *Ctenodonta* Salter, 1852, and some of them were left in open nomenclature. Kříž (1995) described a representative of the genus *Coxiconchia* Babin, 1966 from the Šárka Formation.

Modern revision is in progress now and shows that in the Šárka Formation of Bohemia occur the following bivalve taxa (Plate 1): *Pseudocyrtodonta ala*, *Pseudocyrtodonta incola*, *Praenucula dispar*, *Praenucula bohémica*, *Praenucula applanans*, *Praeleda pulchra*?, *Babinka prima*, *Redonia deshayesi*, *Coxiconchia britannica holubi*. Occurrence of *Praeleda pulchra* in the Šárka Formation is uncertain, because this species is very similar to *Praenucula bohémica*, and *Praeleda* and *Praenucula* are not well defined yet. I consider that *Ctenodonta* does not occur in the Middle Ordovician of Bohemia. *Ps. ala*, *Ps. incola* and *Redonia deshayesi* belong to the Actinodontida; *Praenucula dispar*, *Pr. bohémica*, *Pr. applanans* and *Praeleda pulchra* belong to the Palaeotaxodonta; *Babinka prima* belongs to the Lucinida and *Coxiconchia britannica holubi* is presently classified as Lucinida (Bieler et al., 2010).

RELATIONSHIPS OF THE MIDDLE ORDOVICIAN BIVALVES OF BOHEMIA, SPAIN AND FRANCE

Babin and Gutiérrez-Marco (1991) described from the Middle Ordovician bivalves of Spain the species ?*Ctenodonta escosurae*, *Praenucula costae*, *Praenucula sharpei*, *Cardiolaria beirensis*, *Ekaterodonta hesperica*, *Myoplusia bilunata perdentata*, *Cadomia britannica*, *Goniophora (Cosmogoniophora) sp.*, *Modiolopsis elegantulus*, *Cyrtodontula sp.*, *Glyptarca? lusitanica* [now *Hemiprionodonta lusitanica*: see Cope (1996)], *Ananterodonta oretanica*, *Babinka prima*, *Coxiconchia britannica*, *Redonia deshayesi*, *Dulcinea manchegana*.

Hemiprionodonta lusitanica has its older record in the Lower Ordovician of Bolivia (Sánchez and Babin, 2005), and *B. prima* in the Lower Ordovician of France (Babin, 1977, 1982). *Praenucula costae* from Spain and *Praenucula bohémica* from Bohemia seem to be conspecific: they have similar hinges with concavodont teeth in posterior part of the shell, convexodont teeth in anterior part of the shell, rounded posterior adductor muscle scar and elongated anterior adductor muscle scar. *Praenucula costae* has more anteriorly elongated shell than *Praenucula bohémica*. *Praenucula costae* is also abundant in the Middle Ordovician of France. *Praenucula sharpei* (Spain) is probably a younger synonym of *Praenucula applanans* (Bohemia), which has a characteristic shape and hinge (concavodont teeth in the posterior part of the shell, convexodont teeth in the anterior part of the shell). *Coxiconchia britannica britannica* has larger shells, and deeper, relatively larger anterior adductor muscle scars than *Coxiconchia britannica holubi*. *Coxiconchia britannica holubi* differs from *Coxiconchia britannica guiraudi* in having smaller shell, reduced cardinal teeth, and smaller accessory muscle scars in a more dorsal position (Kříž, 1995). I agree with Babin and Gutiérrez-Marco (1991) that *Redonia bohémica* from Bohemia is conspecific with *Redonia deshayesi* from

Plate 1. Bivalves from the Šárka Formation, Bohemia. A, K, *Praenucula bohémica* (Barrande, 1881), articulated specimen, MBHR 14448; A, right lateral view (x 7.1); K, dorsal view (x 10.5). B, C, *Praenucula dispar* (Barrande, 1881), articulated specimen, MBHR 7982; B, dorsal view (x 7.6); C, left lateral view (x 4.6). D, F, *Pseudocyrtodonta incola* (Barrande, 1881); D, left valve, MBHR 12569, left lateral view (x 8.3); F, right valve, MBHR 12701, dorso-lateral view (x 11). E, *Babinka prima* Barrande, 1881, right valve, L 27086, right lateral view (x 2.3). F, *Pseudocyrtodonta incola* (Barrande, 1881), right valve, MBHR 13415, right lateral view (x 11). G, *Pseudocyrtodonta ala* (Barrande, 1881), right valve, MBHR 12701, dorso-lateral view (x 6.7). H, L, M, *Praenucula applanans* (Barrande, 1881); H, M, articulated specimen, MBHR 14619 in dorsal view (H, x 9.1) and right lateral view (M, x 6.9); L, two valves, MBHR 2395, lateral view (x 6.1). J, I, *Redonia deshayesi* Rouault, 1851, articulated specimen, L22656; J, dorsal view (x 4); I, right lateral view (x 4.5). N, *Coxiconchia britannica holubi* Kříž, 1995, right valve, MBHR 5498, lateral view (x 1.9).



Spain and France. I would like to mention here that also *Redonia anglica* (Salter in Murchison, 1859) from Wales, is most probably conspecific with *Redonia deshayesi*.

PALAEOECOLOGY

Middle Ordovician bivalves from Bohemia are relatively small (with *C. britannica holubi* as the only exception), and are regarded as infaunal deposit feeders or shallow infaunal filter feeders (Table 1). No isofilibranchs and pteriomorphs are known from Bohemia, similarly to the Ibero-Armorican area, where the isofilibranchs and pteriomorphs are scarce (Babin and Gutiérrez-Marco, 1991), being these forms frequent in Baltica (Soot-Ryen and Soot-Ryen, 1960).

Genera	Mode of life
<i>Praenucula</i>	infaunal deposit feeder
<i>Concavodonta</i>	infaunal deposit feeder
<i>Praeleda</i>	infaunal deposit feeder
<i>Pseudocyrtodonta</i>	shallow infaunal filter feeder
<i>Redonia</i>	shallow infaunal filter feeder
<i>Babinka</i>	shallow infaunal filter feeder
<i>Coxiconchia</i>	shallow infaunal filter feeder

Table 1. Mode of life.

Stratigraphical occurrence of the same species in Bohemia, Spain and France

Myoplusia bilunata perdentata is known from the Middle Ordovician (Darriwilian) of Spain, from the Upper Ordovician (Sandbian) of Bohemia, and from the Upper Ordovician (Katian) of France. *Babinka prima* is known from the Middle Ordovician (Darriwilian) of Spain, from the Middle Ordovician (Dapingian and Darriwilian) of Bohemia and from the Lower Ordovician (Floian) and Middle Ordovician (Dapingian) of France. *Praenucula applanans* is known from the Middle Ordovician (Darriwilian) of Bohemia and Spain. *Redonia deshayesi* is known from the Middle Ordovician (Darriwilian) of Bohemia, Spain and France. *Redonia anglica* occurs in the Lower Ordovician (Floian) and in the Middle Ordovician (Darriwilian) of Wales.

CONCLUSIONS

The study of the Bohemian Darriwilian bivalves from the Šárka Formation contributes to our knowledge of the Middle Ordovician fauna of the world. The Middle Ordovician Bohemian bivalves are closely related to those of Spain and France.

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MIDDLE ORDOVICIAN (DARRIWILIAN) GLOBAL CONODONT ZONATION BASED ON THE DAWANGOU AND SAERGAN FORMATIONS OF THE WESTERN TARIM REGION, XINJIANG PROVINCE, CHINA

S. Stouge¹, P. Du² and Z. Zhao²

¹ Geological Museum, University of Copenhagen, Øster Voldgade 5–7, DK-1350 Copenhagen K, Denmark. svends@snm.ku.dk

² Institute of Exploration and Development, Tarim Oilfield Company of PetroChina, P.O. Box 123 Korla, 841000, Xinjiang, P.R. China. dupinde@126.com, xjzxx@126.com

Keywords: Conodont zonation, Middle Ordovician, Dawangou Fm, Tarim Basin, China.

INTRODUCTION

The long ranging and pandemic conodont genus *Periodon* Hadding appeared first in the Floian global Stage. It flourished in the Dapingian, Darriwilian and Sandbian stages and disappeared in the latest Ordovician. Recent systematic work has recognized several species all of which have great potential for global or international correlation. In the Darriwilian global Stage the characteristic *Histiodella* lineage also becomes significant (Stouge, 1984, 2004; Chen et al., 2006; Stouge and Zhao, 2006) and combined, these two taxa provide the base for an important zonal system than can be used for global correlation specifically in the lower and middle Darriwillian (= stage slices Dw1 and Dw2 of Bergström et al., 2009).

During work on the Middle Ordovician succession, the shallow to deeper water limestone in the Tarim Basin has been found to contain a rather rich conodont association in which *Periodon* and *Histiodella* are characteristic (e.g. Zhang and Gao, 1991; Zhao et al., 2000; Wang et al., 2007; Stouge and Zhixin, 2006). A biostratigraphy, based exclusively on these taxa and their chronostratigraphical position in the Dawangou and the lower part of the Saergan formations, western Tarim, Xinjiang Province, China (Fig. 1), is here proposed for the middle Darriwilian (Middle Ordovician).

GEOLOGICAL SETTING

The Tarim cratonic basin is situated in northwestern China and covers an area of 560 000 km² (Fig. 1). It is bordered by the Tian Shan Mountains to the north, the Kunlun Mountains to the southwest and the Aejin Mountains in the southeast. The Tarim Basin was situated adjacent to East Gondwana and occupied the low palaeolatitudes of the southern hemisphere by the early Palaeozoic.

The Tarim plate was a stable craton during the Cambrian to Permian. The lower Cambrian to upper Ordovician succession experienced downwarp probably caused by thermal subsidence, and a vast thickness

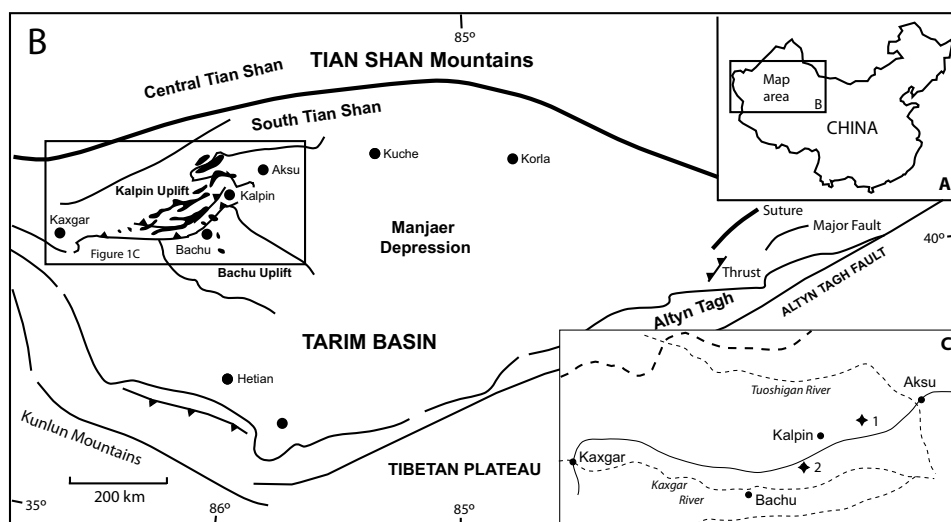


Figure 1. A, The Tarim Basin and the significant structural elements of the Kalpin and Bachu uplifts. B, the distribution of lower Palaeozoic strata and the location of the Yangjikan section (locality 1) from inner platform and more distally located Gowandou section (locality 2).

of sediments accumulated in the basin. The Middle Ordovician mainly carbonate deposits from the Kalpin area in the Tarim region of western Xinjiang, China (Fig. 1) accumulated in a predominantly shallow to deeper, subtidal environment that was interrupted by the short-lived pelagic episode at the mid to late Ordovician transition (Chen et al., 2006; Zhou et al., 1992; Hennissen et al., 2010). In the Late Ordovician, the mid Caledonian movements (Mid to Late Ordovician) became intense and caused tectonic deformation with NW–SE compression due to the subduction of the Tianshan Ocean under the Tarim Block (Windley et al., 1990; Carrol et al., 2001; He et al., 2009). The mid Caledonian movements activated the NE-extending Kalpin fault and the Kalpin uplift became separated from the Bachu uplift area (Fig. 1) and developed a major hiatus between the Late Ordovician and Early Silurian.

LOWER TO UPPER ORDOVICIAN STRATIGRAPHY

The Lower to lower Middle Ordovician Qiulitan Group, ca. 900 m thick, is composed of grey to light-grey, thick-bedded and massive limestone, dolomitic limestone and dolomite. The strata accumulated on a platform under shallow water conditions. Owing to the mid Darriwilian transgressive event the base of the overlying Dawangou Formation is composed first of grey, thick- to medium-bedded wackestone followed by deeper water mainly nodular-bedded wackestone and parted lime mudstone at the top. In addition, field investigation of the Darriwilian strata revealed that both lateral and vertical facies and thickness variability prevailed in the study area (e.g. Zhou et al., 1992; Zhang et al., 2003). Thus marginal to distal deposits of the Dawangou Formation are overlain by black shale of the Saergan Formation representing the maximum deep water, oceanic environment. The Kanling Formation (Upper Ordovician) conformably overlies respectively, the Dawangou and Saergan formations and is a nodular bedded, grey, purplish and red wackestone unit.

STUDIED SECTIONS

Yangjikan section (Fig. 1, location 1). This section in total extends from Lower Ordovician carbonates to Silurian argillaceous sediments. The Dawangou Formation, composed mainly by carbonates is ca. 42 m thick. It conformably overlies carbonates of the Upper Quilitag Formation (Lower to Middle Ordovician) and is overlain by the grey, yellow to red brown, nodular limestone of the Kanling Formation (Upper Ordovician).

Dawangou section (Fig. 1, location 2). The investigated part of the section exposes the Upper Qiulitag Formation and the Dawangou Formation, 22 m thick, which is conformably overlain by black, graptolitic shale (*Pterograptus elegans* to *Nemagraptus gracilis* biozones) and minor limestone beds of the Saergan Formation (ca. 13 m thick), followed by the Kanling Formation. The section serves as an auxiliary Global Stratotype Section and Point (GSSP) for the base of the Upper Ordovician Series (Bergström et al., 2000).

CONODONT BIOSTRATIGRAPHIC DESCRIPTION AND INTERPRETATION

The Darriwilian (Middle Ordovician) strata contain diverse assemblages of conodonts referable to the 'North Atlantic Province'. Through the succession species of *Periodon* and *Histiodella* constitute the main basis for biostratigraphic zonation, supported to lesser degree by *Dzikodus*, *Eoplacognathus*, *Paraistodus* and *Yangtzeplacognathus*. The scheme for the Darriwilian Stage slices Dw1 and Dw2 (Middle Ordovician, see Bergström et al., 2009) conodont zones is presented in Figure 2.

The following biostratigraphic zones/subzones in order of oldest to youngest are proposed and their correlation is discussed:

1. *Periodon macrodentatus* Zone, erected by Stouge (in press), comprises strata that contain a distinctive assemblage of the following taxa: *Periodon macrodentatus*, *Histiodella holodentata*, *Yangtzeplacognathus crassus* and *Histiodella* sp. A. It is divisible into lower *Histiodella holodentata* Subzone, a middle *Yangtzeplacognathus crassus* Subzone and upper *Histiodella* sp. A Subzone (Darriwilian, stage slices Dw1–2).
2. *Periodon zgierensis* Zone (new), divisible into a lower *Histiodella kristinae* Subzone and an upper *H. bellburnensis* Subzone (new) (mid Darriwilian, stage slice Dw2). The base of the zone is marked by the first appearance of *Periodon zgierensis* in the Yangjikan section and the top is at the first appearance of *Periodon aculeatus* and/or *Pygodus serra* in the Gowangou section. The base of the *H. bellburnensis* Subzone is marked by the first appearance of the nominal species in the Gowangou Formation at Yangjikan section.

The uppermost Middle Ordovician (upper Darriwilian, stage slice Dw3) strata are dominated by *Periodon aculeatus* and characterized by *Pygodus serra* and *P. anserinus* (Fig. 2). No new zones/subzones are proposed here for the Dw3 stage slice.

SIGNIFICANCE

The same faunal succession has been recorded from South China (Zhang, 1998) and the biozones established here can easily be recognized in South China. Likewise the same succession is found at the

System	Stage	Stage slice	Condont zones and subzones		Graptolite zones
ORDOVICIAN	Darriwilian	Dw3	(Pygodus serra)		Pterograptus elegans
			Periodon zgierensis	Histiodella bellburnensis	
				Histiodella kristinae	
		Dw2	Periodon macrodentatus	Histiodella sp. A	Nicholsongraptus fasciculatus
		Yangtzeplacognathus crassus		Acrograptus ellasae	
		Dw1		Histiodella holodentata	Undulograptus austrodentatus
				not (yet) distinguished	

Figure 2. The chronostratigraphical position of the *Periodon* biozones and *Histiodella* biosubzones.

GSSP section at Huangnitang, Changshang, Zhejiang, China (Chen et al., 2006) and the *Periodon macrodentatus* Zone and *Y. crassus* and *Histiodella kristinae* subzones are recognized.

The equivalent time interval to the Dawangou Formation of the Kalpin area in eastern North America is well known. *Periodon* and/or *Histiodella* faunas are common in the Appalachian region, in the Quebec Appalachians (Uyeno and Barnes, 1970; Poplawski and Barnes, 1973; Landing, 1986). Conodonts of late Mid Ordovician age were described from the Cow Head and Table Head groups of western Newfoundland reaching from the *Histiodella holodentata* to the *Histiodella bellburnensis* subzones (Stouge, 1984, 2001, in press; Stouge and Zhao, 2006).

Periodon-Histiodella assemblages are known from the succession in Utah-Nevada-California of USA (Harris et al., 1979). In Baltoscandia (Rasmussen, 2001; Stouge and Nielsen, 2003; Mehlgrén and Eriksson, 2010) the zones are represented and Rasmussen and Stouge (1995) demonstrated the correlative potential of these faunas across the Iapetus Ocean. The same succession is also well-known in the successions of the Pre-Cordilleran Argentina (see Albanesi and Ortega 2003, for a summary) and the *Periodon* zones and *Histiodella* subzones promoted here can be applied.

CONCLUSIONS

The conodont genera *Periodon* and *Histiodella* are represented by several species all with a characteristic range in the Darriwilian Stage. Biogeographically, the Darriwilian *Periodon-Histiodella* faunas

are pandemic in low to mid high latitudinal regions and typical of deep shelf, basin margin, and basin slope settings in western China and elsewhere. The association is uniformly distributed and continuous through the western Tarim succession, in South China, Pre-Cordilleran Argentina and Laurentian margin deposits. In northern Europe the presence of the taxa is more constrained to transgressive intervals and their occurrence may be incomplete in these regions.

The *Periodon* based biozones promoted here are less detailed than local biozonations usually recognized in a single richly fossiliferous section. This indicates that biostratigraphically important and global species may have greater stratigraphic ranges than are commonly appreciated.

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THE BASE OF THE ORDOVICIAN SYSTEM – A HORIZON IN LIMBO

F. Terfelt¹, G. Bagnoli² and S. Stouge³

¹ Lund University, Department of Earth and Ecosystem Sciences, Division of Geology, Sölvegatan 12, 223 62 Lund, Sweden.
fredrik.terfelt@geol.lu.se

² Dipartimento di Scienze della Terra, Università di Pisa, via Santa Maria 53, 56126 Pisa, Italy. bagnoli@dst.unipi.it

³ Natural History Museum of Denmark, Geological Museum, University of Copenhagen, Øster Voldgade 5-7,
DK-1350 Copenhagen K, Denmark. svends@snm.ku.dk

The base of the Ordovician System as now defined coincides with the first appearance datum of the conodont *lapetognathus fluctivagus* in the stratotype at Green Point, Newfoundland, Canada. A re-investigation of the conodont succession from the Green Point section revealed, however, that the primary correlation marker for the base of the Ordovician System, *I. fluctivagus*, is not present in the boundary interval of the stratotype section. In consequence, the section does not fulfil the biostratigraphic requirements of a GSSP. The horizon, as now defined, is in a position partway through the range of *lapetognathus preaengensis* in bed 23 of the Green Point section. The first occurrence datum of *lapetognathus fluctivagus* is recorded in bed 26, thus above the first appearance datum (FAD) of planktic graptolites, above the first occurrence of *Cordylodus lindstromi*, and well above the FAD of *I. preaengensis*. As a consequence of these problems, a restudy of the GSSP section and the other sections in the Cow Head Group is needed and a revised biostratigraphic definition of the boundary is necessary. The most favourable situation would be to retain the same point in time; however, this requires that a different primary marker taxon has to be selected from the same point. Unfortunately, there are no candidates fulfilling the biostratigraphic requirements for a primary marker among conodonts and other fossil groups in the current GSSP level. Based on our present knowledge of the fossil fauna, the only option is to move the boundary level to another horizon. The FADs of the following taxa have excellent potential as primary correlation tools for such a horizon: the conodonts *Cordylodus intermedius*, *Cordylodus andresi*, *Eoconodontus notchpeakensis*, and the agnostoid *Lotagnostus americanus*.

THE LOWER TO MIDDLE ORDOVICIAN CONODONT BIOSTRATIGRAPHY OF NORTHERN TIAN SHAN (WESTERN PART OF THE KIRGYZ RANGE), KYRGYZSTAN

T.Yu. Tolmacheva¹, K.E. Degtyarev², L.E. Popov³, A.V. Ryazantsev², A.B. Kotov⁴ and P.A. Aleksandrov⁵

¹ Russian Geological Institute, St. Petersburg, Russia. Tatiana_Tolmacheva@vsegei.ru

² Geological Institute RAN, Moscow, Russia. degtkir@ginras.ru, avryazan51@mail.ru

³ Department of Geology, National Museum of Wales, Cardiff, United Kingdom. Leonid.Popov@museumwales.ac.uk.

⁴ Institute of Precambrian Geology and Geochronology, Russian Academy of Science, 199034 St Petersburg, Russia.

⁵ Geological Faculty, Moscow State University, Moscow, Russia.

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INTRODUCTION

In Northern Tian Shan the Ordovician conodonts are widely applied for age determination in mapping and other geological activities for decades, but their published record remains very poor (e.g. Apayarov et al., 2008). This paper is a part of the larger study that documents structural position and characters of the Lower Paleozoic sedimentary and volcanogenic units exposed in the north-eastern side of the Makbal antiform in the western Kyrgyz Range (Taldybulak and Kentash river basins). The Ordovician sequence in the region comprise predominantly siliciclastic sediments including olistostrome horizons, subordinate volcanic and volcanoclastic rocks; the latter hosting the gold-molybdenum ore-deposit (Degtyarev et al., 2010). Limestone lenses and beds suitable for microfossil extraction by acid dissolution are extremely rare. Trilobites and graptolites, which have been used sporadically for biostratigraphical purposes (Lisogor, 1977), remain very poorly known and therefore not always reliable. The conodont biostratigraphy together with zircon U-Pb dating allow to establish for the first time a composite chronostratigraphical framework for previously insufficiently dated Ordovician lithostratigraphical units developed in the region. In addition, newly discovered conodont assemblages show distinct biostratigraphical signatures, suggesting palaeogeographical position of the Northern Tian Shan block in relative distance from the tropical Australasian sector of Gondwana.

REGIONAL GEOLOGICAL SETTING, STRATIGRAPHICAL FRAMEWORK

The studied Lower Paleozoic volcano-sedimentary complexes are considered to be deposited in island arc and backarc settings (Fig. 1). The Ordovician sequence in the area rests unconformably on the Makbal metamorphic complex and the middle to upper Cambrian siliciclastic and carbonate rocks of the Kenkol Series that

represent a source of olistoliths within the Ordovician olistostrome horizons. The Ordovician sedimentary succession is tectonically capped by the middle Darriwilian granitic intrusive rocks. The Ordovician succession in the backarc basin is subdivided into three lithostratigraphic units, including unnamed olistostrome unit, Taldybulak and Kyzylkainar formations. The substantially volcanic Kentash Formation was presumably formed in the island arc setting.

The Ordovician *olistostrome unit* consists of conglomerates, gritstones and breccias with shale and limestone clasts. Some horizons include large blocks and slices of limestone, chert and shale. In the Chungur Pass the unit comprises sandstone matrix with blocks of limestone. There are a few old reports on the occurrence of the early Tremadocian trilobites in limestone clasts (Lisogor, 1977), however it has not found confirmation

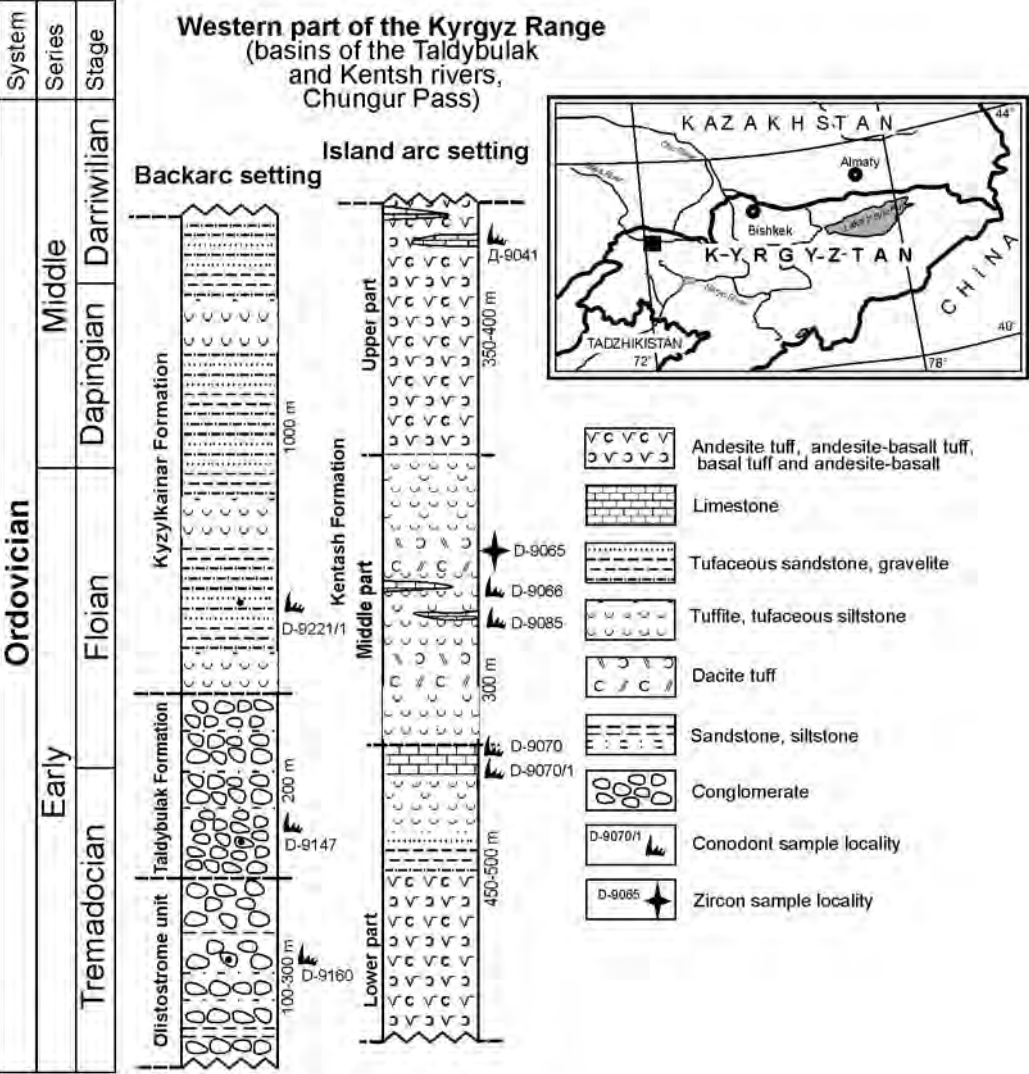


Figure 1. Scheme showing the area studied and stratigraphic columns with sampled levels.

in present study. Clasts of grey and yellow cherts in the area south-east of the Taldybulak ore deposit (Locality D-9160) yield poorly preserved Furongian conodonts, including *Cambroistodus* cf. *cambricus* (Miller), *Phakelodus tenuis* (Müller), *Furnishina* cf. *curvata* Müller and Hinz, *Furnishina furnishi* Müller, and *Prooneotodus rotundatus* (Druce and Jones). The age of the unit is presumably Lower Ordovician mainly basing on occurrence of Furongian and Tremadocian fossils in olistoliths. The real thickness of the olistostrome unit is unidentified because it is strongly dislocated and lack fossils in the matrix.

The olistostrome unit is conformably overlain by the *Taldybulak Formation* of coarse-grained conglomerate with clasts comprised chert, jasper, effusive and intrusive rocks, total thickness above 200 m. In the area south of Taldybulak River chert olistoliths in the lower part of the formation (localities D-9147, D-9148) contain Furongian (upper Cambrian) conodonts, including *Ph. tenuis*, *F. cf. curvata*, *P. rotundatus*, *F. furnishi* and *Furnishina* sp. (Pl. 1, fig. 1, 2).

The succeeding *Kyzylkainar Formation* comprises rhythmic intercalation of tuffaceous and siliciclastic rocks grading from siltstone to sandstone, conglomerate and breccia. Tuffs were sampled for U-Pb zircon dating. The obtained SIMS (SHRIMP-II) results disperse over the 470 - 510 Ma interval. The youngest data are clustered at c. 470 Ma indicating a probable eruption time.

In the Chungur Pass section an olistolith of grey limestone in tuffites yields a single element of *Cordylodus* sp. (Pl. 1, fig. 3). A few records on trilobites and graptolites were reported earlier (Lisogor, 1977; Stepanenko, 1959). The Taldybulak formation has a tectonic contact with granites of the Taldybulak and Almalý massifs.

The *Kentash Formation* was presumably formed within volcanic arc setting. It has faulted contacts with surrounding sedimentary units. There are conflicting reports on a possible age of the Kentash Formation which was assigned to the Lower, Middle and Upper Ordovician, or to the Lower Devonian (Apyarov et al., 2008). The Kentash Formation is up to 1500 m as thick and subdivided into three informal subunits.

The lower c.450-500 m thick part of the formation comprises coarse lithoclastic andesite and andesite-basalt tuffs with several horizons of tuffites and tuffaceous siltstones and rare carbonate lenses. The succession is topped by the 30 m thick bed of grey mainly algal limestone. The sample of bioclastic limestone taken from the middle part of the carbonate unit (Locality D-9070-1; N 42°30' 16", E 72° 54' 57") yielded numerous conodonts, including *Drepanodus arcuatus* Pander, *Paraistodus proteus* (Lindström), *Tropodus australis* (Serpagli), *Drepanoistodus* sp., *Scolopodus* cf. *houlianzhaiensis* An and Xu, *Drepanoistodus latus* Pyle and Barnes, *Acodus* sp. indet., and *Protoprioniodus* sp. The background assemblage is dominated by conodont elements assignable to *Acodus* sp. (Pl. 1, fig. 16), while the occurrence of biostratigraphically informative taxa, e.g. *P. proteus*, *T. australis* and *S. cf. houlianzhaiensis* indicates latest Tremadocian - earliest Floian age (*Paraistodus proteus* Biozone) of the fauna.

The middle part of the Kentash Formation, up to 300 m as thick, comprises tuffaceous sandstones, siltstones, gritstones and fine andesite-basalt tuffs with occasional clasts and lenses of limestone. A limestone sample (Locality D-9085; N 42° 30' 46", E 72° 55' 37") contains low diversity conodont assemblage, including *S. cf. houlianzhaiensis*, *Tropodus? sweeti* (Serpagli), *D. arcuatus*, *Bergstroemognathus extensus* (Graves and Ellison) and *T. australis*.

About thirty meters up the section, a thin lens (3-5 cm) of green-grey limestones in calcareous siltstones with numerous trilobites (Locality D-9066; N 42° 30' 46" E 72° 55' 31") contains abundant and diverse conodont fauna. Up to 23 conodont taxa have been recovered, including *B. extensus*, *Oelandodus elongatus* van Wamel, *D. arcuatus*, *P. proteus*, *Cornuodus longibasis* (Lindström), *Tropodus comptus*, *T. australis*, *Paracordylodus gracilis* Lindström, *Juanognathus* sp. *Stolodus* cf. *stola* (Lindström), *Drepanoistodus* sp., *S. cf. houlianzhaiensis*, *Prioniodus* sp., *Coelocerodontus* sp., *Polonodus? corbatoi* (Serpagli), *D. latus*, *T.? sweeti*, *Kalli-*

dontus cf. *serratus* Pyle and Barnes, *Protopanderodus* cf. *gradatus* Serpagli and *Protoprioniodus* sp. At least four species including two representatives of *Acodus* are new.

The assemblage is strongly dominated by two new *Acodus* species, *D. arcuatus* and *T. australis*, which make up to 80% of 1500 counted specimens. The upper ranges of most of the listed taxa do not exceed *Prioniodus elegans* Biozone, suggesting the early Floian age of the fauna. Only *P.?* *corbatoi* was originally reported from the *Oepikodus evae* Biozone of San Juan Formation of Argentinean Precordillera (Serpagli, 1974), but it is also present in the upper part of the *Prioniodus elegans* Biozone, in particular in the Cow Head Group of Western Newfoundland, as it was shown by Stouge and Bagnoli (1988).

Poorly preserved lingulate brachiopods were found in both conodont samples from the middle part of the Kentash Formation. The list includes *Ombergia* cf. *mirabilis* Holmer, Popov and Bassett, *Ottenbiella* sp. and *Scaphelasma* sp. *O. mirabilis* is widely spread in Baltoscandia in *Paroistodus proteus* and *Prioniodus elegans* biozones and also reported from the Lower Ordovician olistoliths in the Silurian Pul'gon Formation of the Alai Range in Kyrgyzstan (Holmer et al. 2000; Tolmacheva et al., 2001). The earliest occurrences of *Scaphelasma* in Baltoscandia and elsewhere are not older than Floian (*Prioniodus elegans* – *Oepikodus evae* biozones). Thus data from stratigraphical distribution of brachiopods are in a good agreement with conodont data and also suggesting the *Prioniodus elegans* Biozone.

The upper c.350–400 m of the Kentash Formation comprises mainly volcanic and volcano-sedimentary rocks of intermediate, intermediate-basic and basic composition. It is the best exposed on the right side of the Kentash River. Here it consists of dacite and rhyodacite tuffs interbedded with andesite tuffs and andesite lava flows; other lithologies include relatively thin (less than 10 m) layers and lenses of calcareous tuffites, calcareous sandstones, siltstones and grey bioclastic limestones with numerous brachiopods, gastropod, bryozoan and trilobite fragments. A single sample taken from bioclastic limestone (Locality D-9041: N 42° 31' 31"; E 72° 51' 56") yielded conodonts *Panderodus?* *nogamii* (Lee), *Ansella* cf. *robusta* (Ethington and Clark), *Juanognathus variabilis* Serpagli, *D. arcuatus*, *Histiodela holodentata* Ethington and Clark, *Prioniodus?* sp., *Drepanodus* sp. and new conodont genus et sp. indet. (total 35 elements). The age of the assemblage is assessed as the early Darriwilian, because of occurrence of *H. holodentata* and *A. cf. robusta*. Both species appears in North America not earlier than the lower Darriwilian (Ethington and Clark, 1981). The genus et sp. indet. 2 (Pl. 1, fig. 29) it is also biostratigraphically informative as it has been also found in the lower Darriwilian of Kazakhstan (Naiman Formation) and Gornyi Altai (Voskresensk Formation). The Kentash Formation is capped by grano-syenite massif with U-Pb ID-TIMS zircon age of 468 ± 4 Ma.

Plate 1. Selected conodonts from the sections studied. Scale bar for fig. 33, 35–38 = 200 µm; for fig. 34 = 20 µm. Conodonts in fig. 1 and 2 photographed in transmitted light. 1, *Furnishina* cf. *curvata* Müller and Hinz, 1991, D-9147, x110. 2, *Furnishina furnishi* Müller, 1959, D-9147, x80. 3, *Cordylodus* sp., D-9221/1, x42. 4, 5, *Bergstroemognathus extensus* (Graves et Ellison, 1941), D-9066; 4, Sa element, posterior view, x40; 5, Pb element, x55. 6, 7, 8, *Tropodus?* *sweeti* (Serpagli, 1974), D-9066; 6, M element, x50; 7, P element, x50; 8, Sd element, x60. 9, *Paroistodus proteus* (Lindström, 1971), D-9066: M element, x55. 10, *Paracordylodus gracilis* Lindström, 1955, D-9066: S element, x48. 11, *Oelandodus elongates* van Wamel, 1974, D-9066, S element, x65. 12, *Scolopodus* cf. *houlianzhaiensis* An and Xu, 1983, D-9066, x80. 13, 14, 18, 26, *Tropodus australis* (Serpagli, 1974), D-9066; 13 – P element, x45; 14, Sa element, x55; 18, Sd element, x60; 26, Sc element, x50. 15, *Drepanoistodus latus* Pyle and Barnes, 2003, D-9066, M element, x76. 17, *Kallidontus serratus* Pyle and Barnes, 2002, D-9066, Pa element, x46. 16, 19, 27, *Acodus* sp. indet., D-9066; 16, P element, x56; 19, Sc element, x45; 27, M element, x55. 20, *Polonodus?* *corbatoi* (Serpagli, 1974), D-9066, Sc element, x60. 21, 22, 23, Gen. et sp. indet. 1, D-9066; 21, tetracostate element, x85; 22, 23, tricostate element, x60; detail of the cusp, upper oblique view, x120. 24, *Protopanderodus* cf. *gradatus* Serpagli, 1974, D-9066, Sd element, x60. 25, *Tropodus comptus* (Branson and Mehl, 1933), D-9066, Sc element, x55. 28, *anderodus?* *nogamii* (Lee, 1975), D-9041, P element, x60. 29, Gen. et sp. indet. 2, S element, posterior view, D-9041, x84. 30, 31, *Ansella* cf. *robusta* (Ethington and Clark, 1981), D-9041; 30, Sc element, x50; 31, P element, x50. 32: *Histiodela holodentata* Ethington and Clark, 1981, P element, D-9041, x45. 33, 34, *Scaphelasma* sp., D9066; 33, incomplete dorsal valve, exterior; 34, dorsal umbo showing pitted microornament. 35, 36, *Ombergia* cf. *mirabilis* Holmer, Popov and Bassett, 2000, D9070; 35, incomplete ventral valve with broken pedicle tube, posterior view; 36, incomplete dorsal valve, interior. 37, 38, *Ottenbiella* sp., D9066; 37, incomplete dorsal valve interior; 38, incomplete ventral valve.

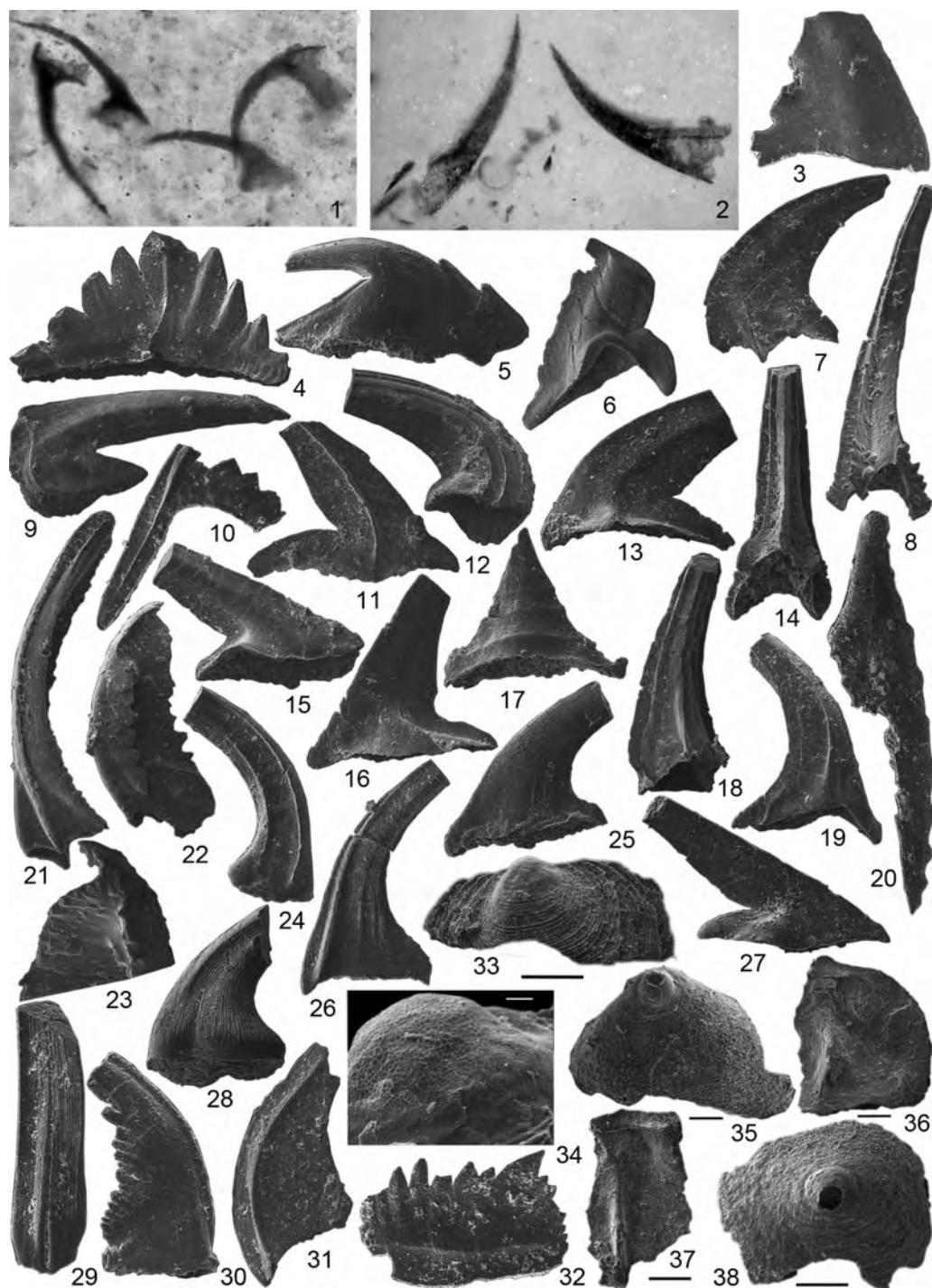


Plate 1

COMPARISON OF BIOSTRATIGRAPHICAL AND GEOCHRONOLOGICAL DATA

The sample of fine-grained dacite tuff (D-9065; N 42° 30' 45"; E 72° 55' 27") has been collected for U-Pb isotope dating in the middle part of the Kentash Formation just 20 m above the carbonate lens with numerous conodonts of the *Prioniodus elegans* Biozone (sample D-9066). It yields eu- and subhedral zircons 50-250 µm in size and 1.2-2.5 aspect ratio. Zircons are short-prismatic to prismatic, transparent and semi-transparent. ID-TIMS U-Pb dating has been carried out on the most transparent zircons of various appearance from size fractions <50, 85-100 and >100 µm. The analyzed zircon grains show the Concordia age of 474 ± 2 Ma that is considered as an age of the dacite tuff formation. It corresponds to the middle part of an estimated time span for the Floian Stage, which is according to Ogg et al. (2008) covers the time interval from 478.6 ± 1.6 to 471.8 ± 1.6 Ma. Thus the U-Pb date obtained for the middle part of the Kentash Formation is well supported by biostratigraphic dating.

AFFINITY OF CONODONT FAUNA AND ITS PALEOGEOGRAPHICAL SIGNIFICANCE

Since the studied Ordovician succession in North Tian Shan is dominated by siliciclastic and volcanoclastic rocks, it was impossible to establish continuous biostratigraphical sequence; however four successive conodont faunas spanning from the Cambrian (Furongian) to the Lower Darriwilian can be recognized.

The Cambrian (Furongian) conodont assemblage is represented by small and tiny paraconodont elements, which can be observed only on cleaved surfaces of cherts or in thin sections. Clusters of elements are common (Pl. 1, fig. 1) suggesting that there was no re-deposition. Identification of simple cone elements imbedded into the rock is problematic and it suffers also from a low number of recovered elements. The obtained conodont assemblage certainly includes species with wide geographic distribution that often occur in basinal setting. Similar faunas dominated by paraconodonts are commonly found both in radiolarian cherts and deep water carbonates in Kazakhstan (Dubinina, 2000).

A number of conodont elements recovered from the lower part of the Kentash Formation is too small for precise age determination. It is similar in taxonomic composition with conodont associations recovered from the middle part of the Kentash Formation, but dominant taxa are different.

Among 23 species recorded in the middle part of the Kentash Formation, 13 (including those that are provisionally identified) are cosmopolitan or have wide geographic distribution. Only *S. cf. houlianzhaiensis* is probably confined only to North China. Endemic species, including gen. et sp. indet. 1 and a few representatives of *Acodus* together constitute more than a half of elements total number in the assemblage. Abundance of *Acodus* species suggests that it was probably derived from a shallow water environment (Zhen et al., 2003). *Prioniodus* sp. and *P. gracilis* which are abundant in deep water sediments are represented just by few specimens.

The occurrence of *B. extensus*, *T. australis*, *S. cf. houlianzhaiensis* and *P. cf. gradatus* in the studied samples is indicative for a so-called Australasian Conodont Province (Webby et al., 2000), which is confined to the Australasian sector of Gondwana, Tarim, South and North China. Some cosmopolitan species recovered from the early Floian of the Kentash Formation are also common in contemporaneous Chinese and Australian conodont faunas. However, *P.?* *corbatoi*, *T.?* *sweeti* and *K. cf. serratus* characteristic of shallow water conodont assemblages of Argentinean Precordillera, Western Newfoundland and Northern Canada are absent in the 'East' Gondwanian faunas. Moreover, *Serratognathus* and specific

Rhipidognathus, which are typical endemic elements of the Australasian Province (Zhen et al., 2009), are absent in the studied assemblage. Relatively large number of endemic taxa in the assemblage from the middle part of the Kentash Formation alone and a presence of several species, which have not been reported from 'East' Gondwana, may suggest the position of the Northern Tian Shan terrain at a considerable distance from mainland Gondwana in the Early Floian.

The lower Darriwilian conodont assemblage from the upper part of the Kentash Formation is known only from a limited number of conodont elements representing eight different species; six of them are relatively common. Among them *P. nogami* is one of the most abundant. The geographical distribution of this species is confined to Gondwana and peri-Gondwanan terrains suggesting Australasian affinity of the assemblage. Relative proximity of Northern Tian Shan to Chingiz-Tarbagatai and Gornyi Altai terrains is indicated by occurrence in all these areas of yet undescribed conodont of new genus and species (Pl. 1, fig. 29).

Acknowledgements

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COMPARATIVE ANALYSIS OF THE EARLY ORDOVICIAN BALTOGRAPTID SPECIES OF NORTHWESTERN ARGENTINA, BALTOSCANDIA AND SOUTH CHINA

B. A. Toro¹, J. Maletz², Y.D. Zhang³ and J. Zhang³

¹ CONICET. Departamento de Paleontología, IANIGLA, CCT-CONICET Mendoza, C.C. 131, 5500 Mendoza, Argentina.
btorogr@mendoza-conicet.gob.ar

² Department of Geosciences, Colorado State University, 322 Natural Resources Building, Fort Collins, CO 80523-1482, U.S.A.
yorge.maletz@colostate.edu

³ LPS, Nanjing Institute of Geology and Palaeontology, CAS, Nanjing 210008, China. ydzhang@nigpas.ac.cn,
zhangju.0331@163.com

Keywords: Early Ordovician, *Baltograptus*, NW Argentina, Baltoscandia, South China.

INTRODUCTION

Deflexed two-stiped graptolite species from NW Argentina were considered as indicators of Arenigian age since the 1960' (Turner, 1960). However, only in recent years Toro and Brussa (2003) and Toro and Maletz (2007, 2008) reevaluated the records of a number of Argentinean deflexed baltograptids and pointed out the importance of these forms for the accuracy of the biostratigraphic scheme and the regional correlation previously proposed in the Central Andean Basin. Toro and Maletz (2007) analyzed a number of well-preserved specimens of the genus *Baltograptus* and related forms from several Floian sections in the Argentinean Eastern Cordillera. They highlighted the relevance of this group in the regional biostratigraphic framework as well as tackled remaining questions regarding the international distribution of *Baltograptus*. This paper deals with the analysis of additional graptolite material from the Lumara and Santa Victoria areas of Northwest Argentina and the equivalent graptolite faunas recently reviewed from SW China and Sweden. The main objectives of this research are the confirmation of previous regional results and the integration of the new information into the international biostratigraphic and taxonomic framework.

The analyzed material comes from the Early Ordovician successions of the Argentinean Eastern Cordillera (Acoite Formation), South China (Hungshihyen Formation) and a number of Swedish drill cores.

The Eastern Cordillera constitutes the southern portion of the Central Andean Basin (Fig. 1). According to recent studies it evolved as the forebulge depozone of the extended Ordovician foreland basin system in northwest Argentina (Bahlburg and Furlong, 1996; Astini, 2003). Early Ordovician platform sediments represented by black and grey siltstones and shales interbedded with fine to medium-grained sandstones toward the top of the sequence of the Santa Victoria Group (Santa Rosita and Acoite Formations) were deposited on a low gradient ramp, under the influence of a large scale prograding deltaic system in the east, and an active volcanic arc complex in the west.

A taxonomical revision was carried out on the material stored in the Nanjing Institute of Geology and Palaeontology and additional specimens were recently collected from a few thin layers of green-yellow shale in the Hungshihyen Formation of Ercun section, Kunming, and the Guihuaqing section, Luquan, Yunnan Province, SW China. These Chinese specimens were originally assigned to the *Corymbograptus deflexus* Biozone, which corresponds to the lower part of the *Didymograptus* (s.l.) *simulans* Biozone in Britain (Cooper et al., 1995) and Chewtonian of Australasia (VandenBerg and Cooper, 1992), and thus indicated a late Floian age. The age of the graptolite fauna from the Hungshihyen Formation, which is dominated by deflexed forms together with a few pendent forms (e.g. *Didymograptellus obesus*), was successively discussed by Zhang and Chen (2003) and Zhang et al. (2007a). Zhang et al. (2010) recently considered that it may not correspond to *B. deflexus* Biozone, but to the lower biozones *Acrograptus filiformis* or *Didymograptellus eobifidus*.

The discussion of these data is completed with the information based on slender species of *Baltograptus* preserved in relief in NW Argentina and Sweden. The Swedish material from drill cores in Scania (SW Sweden) includes numerous undistorted and well-preserved specimens of a number of biostratigraphically relevant species (see Maletz and Ahlberg, 2011).

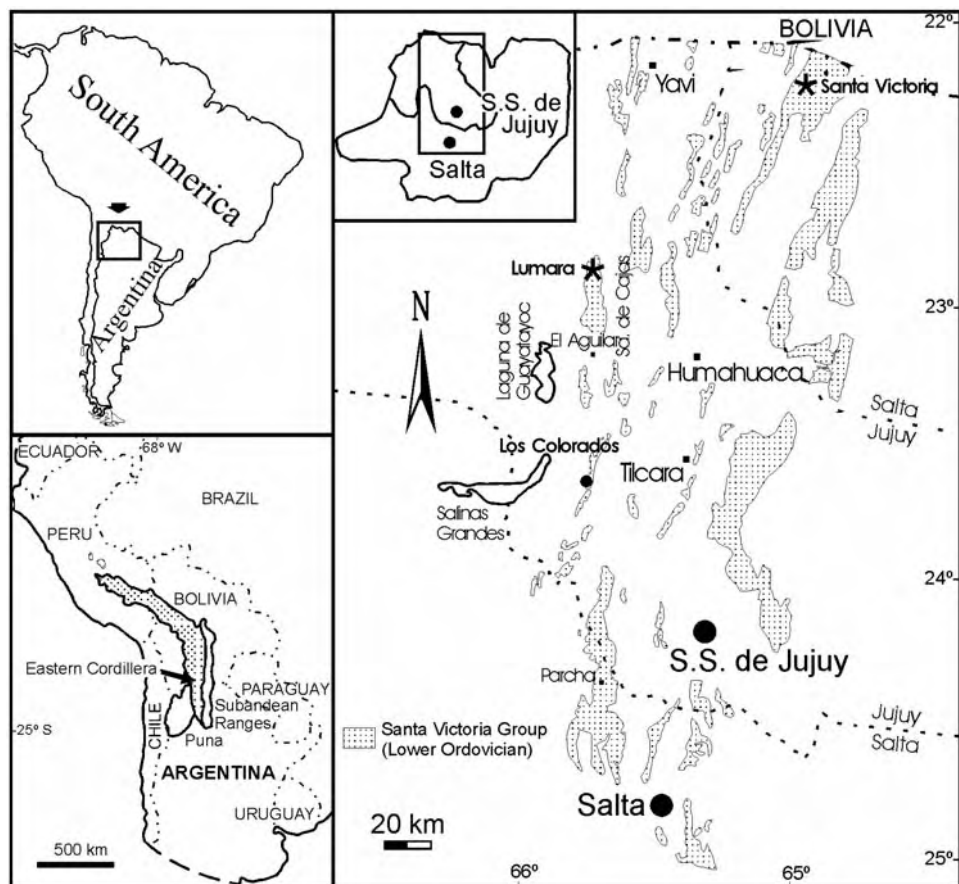


Figure 1. Location map of the Argentinean Eastern Cordillera. The asterisk corresponds to the site of the studied sections.

TAXONOMIC, BIOSTRATIGRAPHIC AND PALEOECOLOGICAL IMPLICATIONS

Toro and Maletz (2007) considered *B. geometricus* as the oldest species of the genus *Baltograptus*. It is widely distributed in the *Tetragraptus phyllograptoides* Zone of Sweden (Maletz et al., 1991) and was also identified in the NW of Argentina (Toro, 1997; Toro and Maletz, 2008) and in SW China (Pl. 1, fig. 11). This species exhibits an isograptid proximal development like *B. vacillans* (Pl. figs. 1, 5, 13), which is commonly recorded from the subsequent Early Floian levels in the three studied regions (Ortega and Rao, 1994; Zhang et al., 2007b; Maletz and Ahlberg, 2011). An important remaining aspect of this study is the taxonomical classification of the wide-stiped Chinese deflexed graptolites (Zhang et al., in prep.), previously assigned to a number of different species by Mu et al. (1979), and later termed the "*B. calidus*" and "*B. turgidus* group" by Maletz (1994) and Toro and Maletz (2007). Preliminary results of the comparative study allows to confirm the similarities of this material with the best preserved specimens from the early Floian strata of the Lumara and Río Mecoyita sections, in Northwestern Argentina. Toro and Maletz (2007) provided biostratigraphic columns of the Los Colorados and Santa Victoria areas. Although the biostratigraphic scheme for the Argentinean Eastern Cordillera is still under revision (Toro and Maletz, in prep.) preliminary graptolite distribution in the Lumara section is presented in Fig. 2. The Argentinean specimens are stored in the paleo-invertebrate repository of the IANIGLA, CCT-CONICET Mendoza (IANIGLA-PI) and Córdoba University (CEGH-UNC).

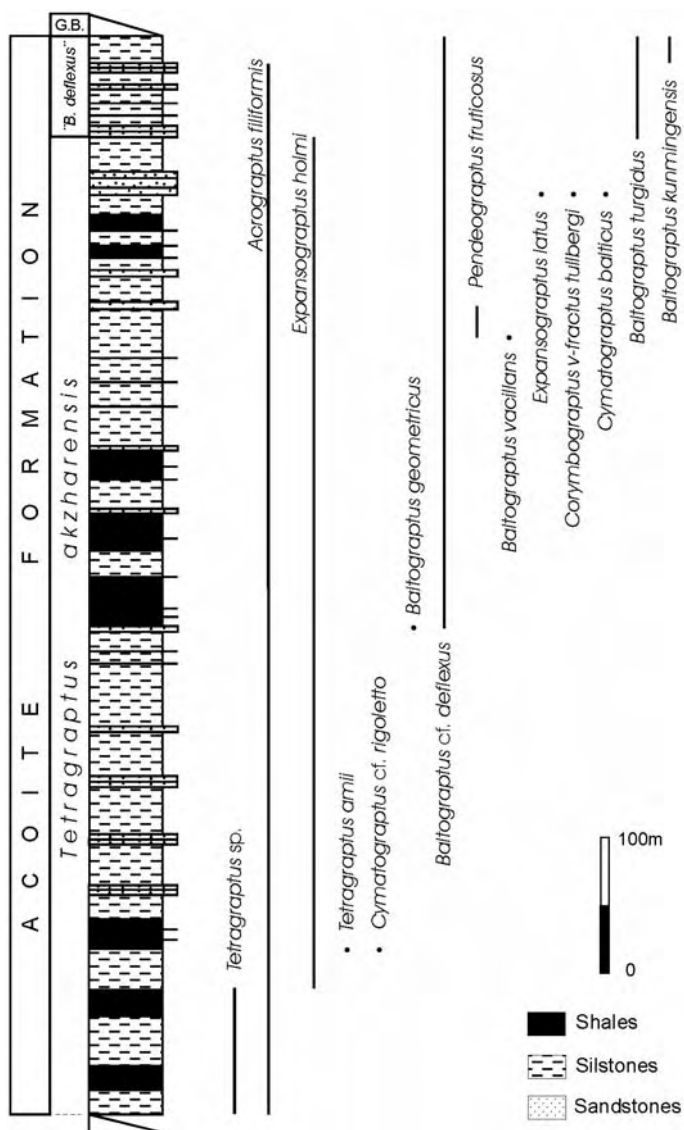


Figure 2. Stratigraphic range of select graptolites from Lumara area.

Toro (1996) described the wide-stiped rhabdosomes from the Lumara section, commonly filled with pyrite, for the first time in Argentina as *B. turgidus* and *B. kunmingensis* (Pl. 1, fig. 12). Better preserved material was subsequently obtained from different sections of Santa Victoria area, and was assigned to the *Baltograptus turgidus* group (Toro, 1998; Toro and Maletz, 2007: fig. 4 A-D, F-G). Additional specimens recently collected in the Río Mecoyita section, in Santa Victoria area were also studied. They are commonly well preserved in fine gray sandstones corresponding to the lower portion of the Acoite Formation. Some specimens are preserved in partial relief showing clearly an isograptid proximal development, with the first theca growing relatively low from the sicula (Pl. 1, figs. 4, 6). These characteristics perfectly agree with the original description of the genus *Baltograptus* (Maletz, 1994).

Similar characteristics are also observed in the Chinese specimens (Pl. 1, figs. 7-10) allowing their inclusion in the genus *Baltograptus*.

Specimens originally assigned to *B. triangulatus* can be included in the *B. turgidus* group. They reach a maximum width of less than 2 mm of and seem to be identical to the Argentinean material from Lumara and Santa Victoria sections (Pl. 1, figs. 9, 12).

Slender forms like *B. kunmingensis*, *B. wudingensis* and the *B. varicosus* group, approximately 1.5 mm of wide in average, can be related and they are very similar to the slender specimens recorded in Santa Victoria area, which show an isograptid type of proximal development (Pl. 1, figs. 1, 6, 7).

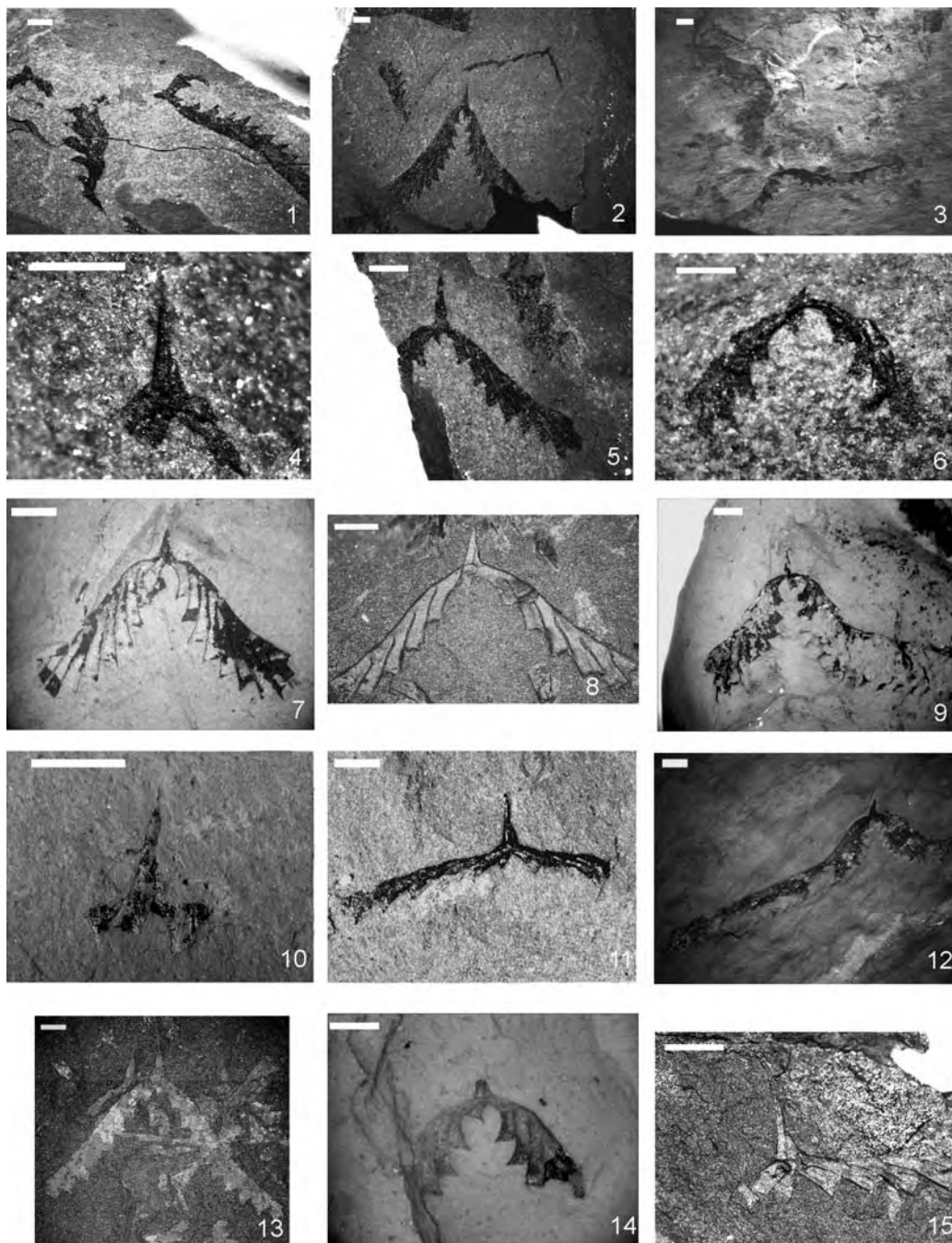
The widest Chinese forms are more than 2 mm wide (up to 2.5 mm maximum width) like *B. calidus* and *B. yunnanensis* may be synonyms. Given that the Argentinean material does not exceed 2 mm in width, the presence of these taxa has not yet been confirmed in NW Argentina.

Based on the analysis of robust specimens in the Lumara section, Toro (1996) proposed for the first time faunistic affinities with the robust deflexed species of South China. The best preserved material collected in different sections of Santa Victoria as well as the preliminary results of this comparative study confirm the faunal affinities between the graptolite faunas from both regions and contribute to the recent statistical analysis of the faunistic similarities recently presented for the early Floian graptolites (Vento et al., 2010; Vento et al., in prep.).

The studied graptolite faunas of NW Argentina, South China and Sweden were all recorded in clastic sedimentary successions, however some differences regarding the distribution of the various species of the genus *Baltograptus* can be recognized. The most robust forms, like *B. calidus* and *B. yunnanensis* groups have not been recorded from Sweden and are also apparently absent from NW Argentina. The association

Plate 1. Best preserved specimens of *Baltograptus* genus from Argentina, China and Baltoscandia. 1, 4, 6, 10, *Baltograptus kunmingensis*: 1, 4, 6 from the Acoite Fm, Mecoyita River (*T. akzharensis* Biozone), NW Argentina. 1, different growing stages specimens, IANIGLA-PI 2362; 4, early stage of development preserved in relief showing isograptid type and low origin of th¹, IANIGLA-PI 2364; 6, young specimen exhibiting the isograptid crossing canal in semi-relief, IANIGLA-PI 2366; 10, early flattened stage of development from the Hungshihyen Fm in the Ercun section, Kunming, Yunnan, SW China, NIGP 153679 (EC1).— 2, 9, 12, 14, *Baltograptus turgidus*: 2, mature specimen associated with *A. filiformis* in the Mecoyita River (*T. akzharensis* Biozone), IANIGLA-PI 2363; 9, mature specimen from the Hungshihyen Fm at Modaoqou, Kunming, Yunnan, NIGP 32079; 12, mature specimen filled with pyrite from the Acoite Fm, Lumara section, NW Argentina (*T. akzharensis* Biozone), CEGH-UNC 12208; 14, juvenile specimen from the Hungshihyen Fm in the Guihuaqing section, Luquan, Yunnan, NIGP 153680 (AGC-15).— 3, *Baltograptus* sp. from the Lumara section ("*B. deflexus*" Biozone), slender form associated to strong rhabdosomes of *B. turgidus* group, CEGH-UNC 12421.— 5, 13, *Baltograptus vacillans*: 5, from the Mecoyita River (*T. akzharensis* Biozone), IANIGLA-PI 2365; 13, from Kiviks Esperod, flattened specimen showing the general preservation of the material in Scania, LO 345t (syntype).— 7, *Baltograptus wudingensis* from the Hungshihyen Fm at Dalieshang, Kunming, Yunnan, complete specimen showing isograptid type of development, NIGP 32106.— 8, *Baltograptus* sp. 2 (*sensu* Maletz and Alberg, 2010) from Lerhamn drill core, Sweden, reverse view with artus type development and short sicula, LO 10582t.— 11, 15, *Baltograptus geometricus*: 11, complete specimen from the Dawan Fm at Shuangliuba, Shizhu, Chongqing, China, showing isograptid proximal development in relief, NIGP 32160; 15, latex cast of the specimen T 130 from the Toyen section, Oslo, Norway, showing the reverse side with an isograptid development and low origin of th¹.

Plate 1



of robust species, like the *B. turgidus* group, with more slender ones, like *B. kunmingensis* in South China or *B. cf. B. deflexus* and *Baltograptus* sp. (Pl. 1, fig. 3) in NW Argentina can be noted. These dissimilar associations may be explained by minor paleoenvironmental changes of oxygenation and energy in the sea floor of the studied regions.

Acknowledgements

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THE AGE OF THE *P. LINEARIS* GRAPTOLITE BIOZONE: A PROGRESS REPORT ON A POTENTIAL SOLUTION

T.R.A. Vandenbroucke¹, A.T. Nielsen² and J.K. Ingham³

¹ Université Lille 1, FRE 3298 du CNRS: Géosystèmes, bâtiment SN5, Avenue Paul Langevin, 59655 Villeneuve d'Ascq cedex, France. Thijs.Vandenbroucke@univ-lille1.fr

² The Geological Museum, University of Copenhagen, Øster Voldgade 5-7, DK-1350 Copenhagen, Denmark. Arnet@snm.ku.dk

³ The Hunterian Museum, Glasgow G12 8QQ, UK.

Rickards (2002) suggested that the *P. linearis* graptolite Biozone is of Rawtheyan age in the historical type area of the Ashgill Series (Cautley district, Cumbria, Northern England). However, this dating of the *P. linearis* graptolite Biozone is in contradiction with its age assignment in, amongst others, the classic Girvan area (South Scotland), where it straddles the Caradoc-Ashgill boundary, as indicated by shelly fauna data. In order to contribute to the solution of the existing correlation problem, we study graptolite-chitinozoan relations in three key sections through the discussed interval. The creation of a chitinozoan reference framework is useful as at least a part of the problem seems to revolve around graptolite biozonal definitions. The key sections obviously include the aforementioned Cautley and Girvan districts, as well as the Vasagård section on the Danish island of Bornholm. On Bornholm, the upper part of the Dicellograptus Shale represents the *D. clingani* and *P. linearis* zones, whereas the overlying Lindegård Mudstone contains graptolites indicative of the *D. complanatus* Zone.

Here we present the first detailed study of the rich and diverse chitinozoan fauna of the *D. clingani* and *P. linearis* zones from the Vasagård section. The chitinozoans are representative of the *Spinachitina cervicornis* to *Tanuchitina bergstroemi* biozones. A correlation with the chitinozoan biozonation in the Cautley district, that has a predominantly Baltoscandian signature (Vandenbroucke et al., 2005) and is well correlated with the graptolite and shelly fauna biozones described from the region, is entirely feasible.

We will also report on our first results from the Girvan district. The Whitehouse subgroup of the upper Ardmillan Group (Ingham, 2000), including the Penwhapple Formation with *P. linearis* biozone graptolites has now been carefully sampled for chitinozoans (80+ samples in total). Vandenbroucke et al. (2003) presented a reconnaissance study of the chitinozoans from the district; a full appraisal of this fauna represents work in progress. Comprehensive sampling through the Myoch, Mill and lower Shalloch formations should establish a tight link between chitinozoan faunas and the basal *D. complanatus* biozone faunas in the Lapworth member there and also with 'classical' *D. complanatus* faunas in the overlying strata, together with a good low Ashgill 'Pusgillian' shelly fauna from the Forge mudclast conglomerate Member of the Mill Formation. The preliminary data (including chitinozoans found in *C. pygmaeus* graptolite Biozone elsewhere on Laurentia) indicate a mix of Baltoscandian and Laurentian faunal elements, and suggest a good correlation potential with the Vasagård section and the Cautley district. The final goal of the project is to provide a detailed correlation of these sections, using graptolites, chitinozoans, and modern correlation techniques such as constrained optimization, in order to re-assess the age of the *P. linearis* graptolite biozone.

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POLAR FRONT SHIFT AND ATMOSPHERIC CO₂ DURING THE GLACIAL MAXIMUM OF THE EARLY PALEOZOIC ICEHOUSE

T.R.A. Vandenbroucke¹, H.A. Armstrong², M. Williams^{3,4}, F. Paris⁵, J.A. Zalasiewicz³, K. Sabbe⁶,
J. Nölvak⁷, T.J. Challands², J. Verniers⁸ and T. Servais¹

¹ Géosystèmes, FRE 3298 du CNRS, Université Lille 1, France. Thijs.Vandenbroucke@univ-lille1.fr

² PalaeoClimate Group, Department of Earth Sciences, Durham University, Durham, UK.

³ Department of Geology, University of Leicester, Leicester, UK.

⁴ British Geological Survey, Keyworth, UK.

⁵ Géosciences, UMR 6118 du CNRS, Université de Rennes I, Rennes, France.

⁶ Protistology and Aquatic Ecology, Department of Biology, Ghent University, Ghent, Belgium.

⁷ Institute of Geology, Tallinn University of Technology, Tallinn, Estonia.

⁸ Research Unit Palaeontology, Department of Geology, Ghent University, Ghent, Belgium.

Our data address the paradox of Late Ordovician glaciation under supposedly high $p\text{CO}_2$ (8 to 22x PAL: Pre-industrial Atmospheric Level) (Vandenbroucke et al., 2010). The paleobiogeographical distribution of chitinozoan ("mixed layer") marine zooplankton biotopes for the Hirnantian glacial maximum (440Ma) are reconstructed and compared to those from the Sandbian (460Ma): they demonstrate a steeper latitudinal temperature gradient, and an equator-wards shift of the Polar Front through time from 55-70°S to ~40°S. These changes are comparable to those during Pleistocene interglacial-glacial cycles. In comparison with the Pleistocene, we hypothesize a significant decline in mean global temperature from the Sandbian to Hirnantian, proportional with a fall in $p\text{CO}_2$ from a modeled Sandbian level of ~8x PAL to ~5x PAL during the Hirnantian. Our data suggest that a compression of mid-latitudinal biotopes and ecospace in response to the developing glaciation was a likely cause of the end-Ordovician mass extinction.

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CHITINOZOANS OF RIBEIRA DA LAJE FORMATION, AMÊNDUA-MAÇÃO SYNCLINE (UPPER ORDOVICIAN, PORTUGAL)

N. Vaz¹, F. Paris² and J.T. Oliveira³

¹ Trás-os-Montes e Alto Douro University, Ap. 1013, 5001-801 Vila Real, Portugal. nunovaz@utad.pt

² Géosciences, UMR 6118 du CNRS, Université de Rennes I, Rennes, France. florentin.paris@orange.fr

³ Laboratório Nacional de Energia e Geologia (LNEG), Estrada da Portela, Zambujal-Alfragide, Ap. 7586, 2720-866 Amadora, Portugal. tomas.oliveira@lneg.pt

Young (1985, 1988) established the lithostratigraphy of the Upper Ordovician of the Amêndua-Mação Syncline (central Portugal). In this region three formations were defined with, in ascending order, the Cabeço do Peão, the Ribeira da Laje and the Casal Carvalhal formations.

The type section for the Ribeira da Laje Formation was defined in the Ribeira da Laje valley, near Sanguinheira region, on the northern limb of the Amêndua/Mação Syncline. The lower part of this formation is composed of micaceous mudstones and bioturbated silty sandstones. Its upper part registers an increase, in frequency and thickness, of discrete sandstone beds, culminating with 9 m of quartzite beds, up to 1,5 m thick. These quartzites are overlain by 5 m of thinly bedded bioturbated sandstones (Young, 1985, 1988).

The Ribeira da Laje Formation contains very few macrofossils and Young (1985, 1988) proposed a middle Ashgill age based on correlation with the Buçaco Syncline. Ten samples were collected in the type section of the Ribeira da Laje Formation for chitinozoan investigation. In different samples, two key species were recovered, that allowed the determination of the diagnostic taxa, *Euconochitina tanvillensis* (Paris) (in Robardet et al., 1972) and *Acanthochitina barbata* Eisenack 1931 (Vaz, 2010). They document the *E. tanvillensis* Biozone (Paris, 1990, 1999) and *A. barbata* Biozone (Paris, 1999; Paris and Verniers, 2005) of early to mid Katian age, respectively.

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ORDOVICIAN COSMIC SPHERULES FROM THE CORDILLERA ORIENTAL OF NW ARGENTINA: PRELIMINARY SEM AND EDX INVESTIGATION

G.G. Voldman¹, G.L. Albanesi¹, C.R. Barnes², G. Ortega¹ and M.J. Genge³

¹ CONICET – Museo de Paleontología, Universidad Nacional de Córdoba, Casilla de Correo 1598, Córdoba X5000FCO, Argentina.
gvoldman@efn.uncor.edu, galbanesi@arnet.com.ar, gcortega@arnet.com.ar

² School of Earth and Ocean Sciences, University of Victoria, Victoria, BC, Canada V8W. crbanes@uvic.ca

³ Department of Earth Science and Engineering, Imperial College London, Exhibition Road, London SW7 2AZ, UK.
m.genge@imperial.ac.uk

Keywords: Microspherules, cosmic dust, Lower Ordovician, Cordillera Oriental, Argentina.

INTRODUCTION

Fluctuations in the influx of extraterrestrial materials to Earth play an important role in the weak equilibrium between the oceans, atmosphere, climate, and life (e.g., Álvarez et al., 1980). Extraterrestrial flux is assumed to have been more or less constant with a few peaks in the accretion rates, such as the K/T boundary. The discovery of numerous fossil meteorites in marine limestones from southern Sweden reflects an extraordinary increase in the flux of extraterrestrial matter to Earth during the Mid-Ordovician (Schmitz et al., 2001). A Mid-Ordovician increase in the meteorite flux is further supported by an iridium anomaly, osmium isotope data and by the distribution of sediment-dispersed extraterrestrial (ordinary chondritic) chromite grains from Sweden and central China (Cronholm and Schmitz, 2010). Accordingly, Dredge et al. (2010) determined a flux of micrometeorites one to two orders of magnitude greater than at present in Arenig limestone samples from the Durness Group, in Scotland.

Micrometeorites are extraterrestrial particles between 10 µm and 1 mm in size recovered from the Earth's surface (Rubin and Grossman, 2010). Melted micrometeorites formed as molten droplets during atmospheric entry are known as cosmic spherules (Genge et al., 2008). The discovery of magnetic spherules in acid-insoluble residues from conodont samples encouraged a systematic search for Ordovician micrometeorites from Northwestern Argentina. An important depocenter containing Ordovician fossiliferous rocks from the Central Andes of Argentina is presently analyzed (Fig. 1). In the Central Andean Basin, Ordovician strata are superbly exposed at Cordillera Oriental, a thick-skinned mostly east-vergent thrust system, limited to the west by the Puna plateau and to the East by the Sierras Subandinas (Ramos, 1999). The stratigraphy of the Cordillera Oriental reflects relatively shallow environments ranging from outer shelf to shoreface rarely dominated by tidal complexes, in contrast to the deep water setting of the Puna (Astini, 2003). In particular, at the Zenta Range the Lower Ordovician strata are over 3000 m in thickness (Santa Victoria Group) (Astini, 2008).

In this region, the Santa Rosita Formation is represented by a thick succession consisting of monotonous alternating series of shales and sandstones with subordinated calcareous concretions, coquinas and cal-

carenites. The productive samples Z4 and Z8 were obtained from two localities along the road from the Zenta path to the Santa Ana path, in both flanks of a local anticlinal structure (Fig. 1). Sample Z4 was taken from a mudstone level located in the eastern flank of the anticline at Santa Ana path, 15 m below sample Z5, a calcareous coquina that yielded conodonts of the *Acodus deltatus* – *Paroistodus proteus* Zone, and close to a shaly interval characterized by the graptolite *Araneograptus murrayi* (J. Hall). Sample Z8 was from a calcareous coquina located in the western flank of the anticline at ca. 4000 m altitude, which is intercalated with brown shales bearing graptolites of the *Hunnegraptus copiosus* Zone; i.e., late Tremadocian (Albanesi et al., in press).

METHODOLOGY

All of the rock samples were processed in a clean laboratory following the standard techniques employed to recover conodonts (Stone, 1987). The insoluble residue was then separated in a sieve size 200 (75 µm) and inspected for microfossils and spherules under the binocular microscope in the Laboratory of Micropaleontology at Universidad Nacional de Córdoba, Argentina. The surface texture and composition of the spherules was analyzed using a SEM Hitachi S-4800 coupled with a Bruker EDX detector at the Advanced Microscopy Facility of the University of Victoria, Canada.

RESULTS

A total of ~220 spherules, generally ranging ~75–250 µm in diameter, were recovered from the Tremadocian samples Z4 and Z8 of the Zenta Range. SEM imaging of the particles reveal hollow, massive, spherical or drop-shaped forms, with well developed dendritic and polygonal crystalline textures (Fig. 2). Similar surface textures have been described by numerous authors (e.g., Wang and Chatterton, 1993; Szöör et al., 2001; Stankowski et al., 2006; Korchagin et al., 2010; Dredge et al., 2010), which related them to

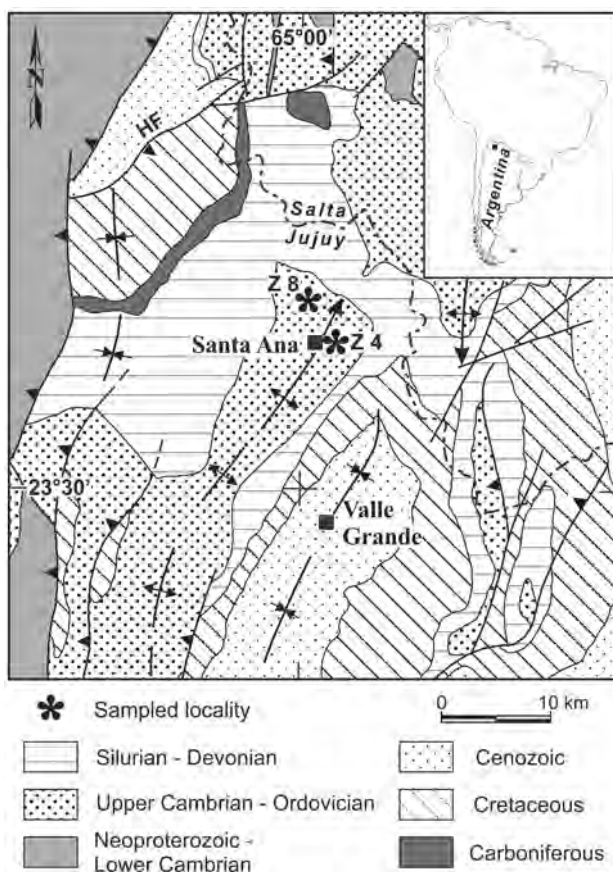


Figure 1. Geologic map from the southern sector of the Zenta Range including the sampled localities (adapted from Albanesi et al., in press).

rapid cooling of micrometeorites from high temperatures. EDX analysis of the particles suggests they are composed principally of iron oxides consistent with I-type spherules, containing mainly magnetite and/or wustite with rare Fe-Ni metal droplets (Genge et al., 2008). Surface EDX suggests up to 1 wt% Al, Si, Ca, Mg, Ni and Cr. Although Ni enrichment is usually considered marker for an extraterrestrial origin, only minor amounts of Ni were detected in the spherules. This is in accordance with the heating during the

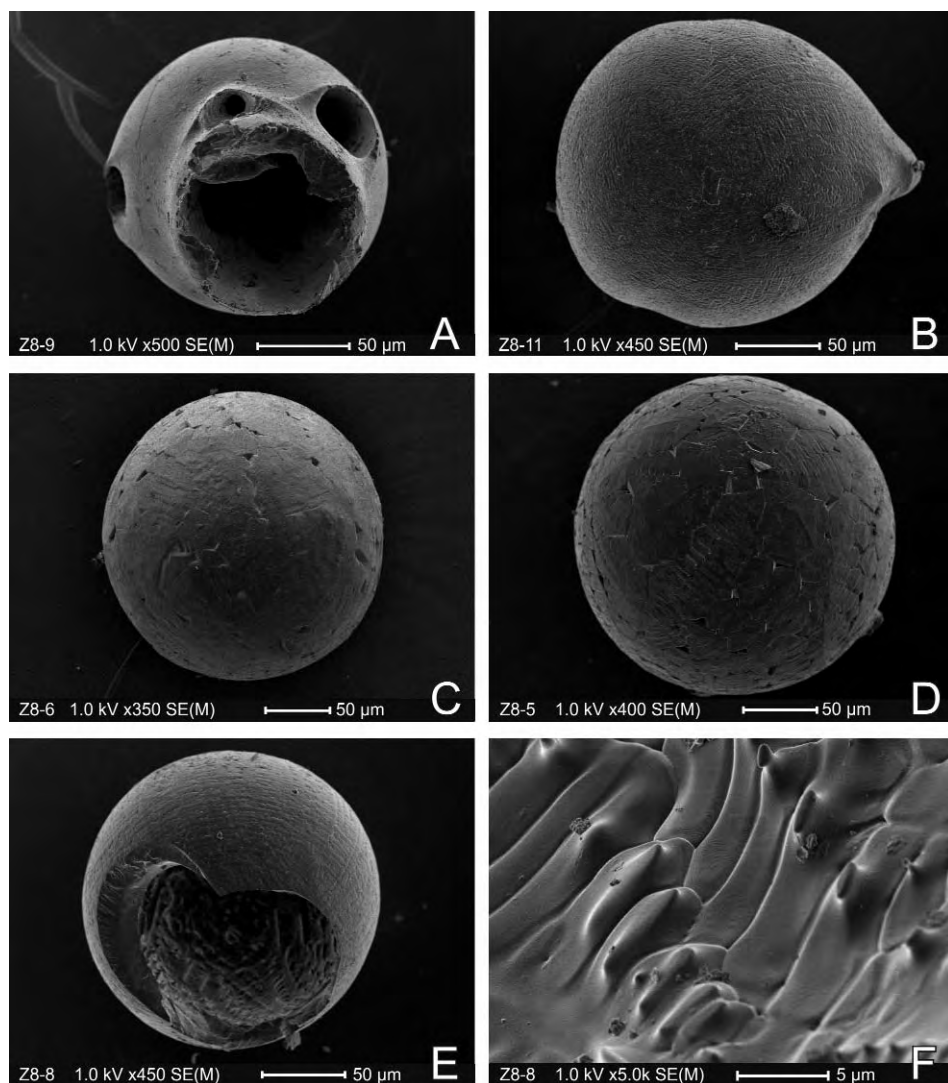


Figure 2. SEM secondary images of I-type spherules from the Santa Rosita Formation (late Tremadocian) recovered from the Zenta Range in Cordillera Oriental. A) Spherule with dendritic texture and escape structures. B) Drop-like spherule with coarse dendritic texture and protruding knobs. C) Spherule with coarse polygonal texture. D) Spherule showing brick-work texture superposed on a polygonal pattern. E) Hollow spherule displaying dendritic texture. F) Detail of the inner side of spherule E showing a fine dendritic pattern.

atmospheric entry: as the micrometeorites melt to form silicate and metallic melts, these are subsequently separated into individual spherules due to density difference (Brownlee et al., 1984). Eventually, the iron in the metallic spherules oxidizes, leaving a high Ni or Pt group metallic core. Depending on the deceleration experienced, metallic cores of the spherules migrate to the front of the particle and separate, leaving a crust that can constitute a Fe-oxide spherule free of Ni (Bi et al., 1993; Yada et al., 1996). Conversely, droplets produced during meteorite ablation form at lower altitudes and contain higher Ni concentrations, in response to the higher oxygen fugacity (Genge and Grady, 1999). Alternatively, volcanogenic magnetic spherules tend to be rich in Ti (El Goresy, 1968; Szöör et al., 2001), however Ti only constitutes a trace component in the Argentinian particles. Furthermore, thermodynamic considerations, the morphology and the absence of inclusions with magmaphile elements rule out a volcanogenic genesis for our spherules (del Monte et al., 1975; Iyer et al., 1999). An anthropogenic contamination can be discounted since these spherules were diagenetically linked to the host sediment.

DISCUSSION AND CONCLUSION

The micrometeorite bearing strata in Argentina span the late Tremadocian, which equates with part of a period of elevated flux of extraterrestrial material, as recorded several thousand kilometres apart from coeval horizons in Scotland, Sweden and central China. Parnell (2009) related the enhanced Middle Ordovician meteorite flux with global scale deposition of olistostromes by destabilisation of continental margins following meteorite impacts. This author proposed that up to 500 impactors of 100 m in diameter, including 250 impactors if only landward impacts are considered, fell within about 30 km of the 20,000 km long lapetus coastline. Alternatively, Meinhold et al. (2011) challenged the idea that mass wasting was mainly produced by meteorite impacts over a period of almost 10 Ma, and proposed an earthquake driven mechanism related to plate-tectonic processes, possibly magnified during a period of global sea-level lowstand.

The particles recovered in the current study are I-type spherules which represent only a small fraction ~1% of the current day micrometeorite flux (Genge et al., 2008) and are formed from extraterrestrial dust rather than large objects such as meteorites. The abundances observed suggest an elevated extraterrestrial dust flux in Northwestern Argentina but does not necessarily imply that large impact events occurred at that time since the continuous flux of dust falling onto the Earth unrelated to significant impact events (e.g., Brownlee, 1985; Taylor et al., 1998). Moreover, the occurrence of small meteorite fragments (~500 m) could record the infall of larger objects, big enough to produce abundant microparticles but too small to produce large craters (French and Koeberl, 2010).

Impact, extraterrestrial and volcanic spherules are increasingly used for interpreting geological correlation and palaeogeography. Future geochemical, petrographical and palaeontological studies would provide evidence of the true magnitude and geographical distribution of these cosmic events during the early Phanerozoic history of the Earth and their role in the explosion of biodiversity during the Ordovician Period (e.g., Schmitz et al., 2008).

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BIODIVERSITY PATTERNS AND THEIR IMPLICATIONS OF EARLY-MIDDLE ORDOVICIAN MARINE MICROPHYTOPLANKTON IN SOUTH CHINA

K. Yan¹, J. Li¹, and T. Servais²

¹ Nanjing Institute of Geology and Palaeontology, Chinese Academy of Sciences, State Key Laboratory of Palaeobiology and Stratigraphy, East Beijing Road, Nanjing 210008, China. yankuiboy@gmail.com, junli@nigpas.ac.cn

² FRE 3298 du CNRS, Géosystèmes, Université de Lille 1, SN5, USTL, F-59655 Villeneuve d'Ascq, France.
Thomas.Servais@univ-lille1.fr

Since the 1980s, about 30 sections from South China have been investigated for Early-Middle Ordovician acritarch studies (Yan et al., 2011). Based on these literatures, several papers discussed the biodiversity of the South Chinese phytoplankton in the Ordovician. In the last decade (e.g. Li et al., 2007).

In the present study, more than 100 samples are collected from six Early-Middle Ordovician sections located in different lithofacies of the South China as well as acritarch assemblages described from literatures of South China for diversity analysis. Several diversity curves are presented, including measurements of total and mean diversity, as well as origination and extinction rates.

The total acritarch diversities in South China increase until the *suecicus* graptolite biozones indicating a major acritarch radiation in the Ordovician. The Ordovician acritarch origination curves also infer that this acritarch radiation event happened with increasing disparities of acritarch forms in the lower Floian in South China. The variation of the Ordovician acritarch diversity changes from the six sections investigated here suggests that the acritarch diversity changes would be related to local environment and sea-level changes.

Ordovician acritarch diversity changes are also been studies in other palaeocontinents, such as Baltica (Hints et al., 2010), England (Molyneux, 2009), North Africa (Servais et al., 2004), northern Gondwana margin (Vecoli and Le Hérissé, 2004). The variation of acritarch diversity change patterns in different palaeocontinents implies the different ecological evolution patterns in the GOBE (The Great Ordovician Biodiversity Event).

We select acritarch assemblages from four sections to analyses the relationship between acritarch diversity changes and local sea level changes. The acritarch diversity changes from four sections can partly be compared to the local sea-level changes in South China but additional research is needed to understand the pattern, including the inshore - offshore trends. The diversity of acritarchs increased rapidly during the Early-Middle Ordovician, perhaps because of the spreading of continental masses and increasing habitat space with rising sea levels.

The phytoplankton curves are compared with the diversity changes of different invertebrate fossil groups, including chitinozoans (Paris et al., 2004), conodonts (Wang and Wu, 2007; Wu et al., 2010), graptolites (Zhang et al., 2007), trilobites (Zhou et al., 2007), and brachiopods (Zhan et al., 2005). It can be assumed that the Ordovician phytoplankton radiation paralleled a long-term increase in sea level with an accompanying expansion of flooded continental shelf areas. The availability of increased quantities of

phytoplankton in the Lower-Middle Ordovician of the Yangtze Platform allowed the radiation of zooplanktonic groups, and at the same time accelerated the rise of suspension feeders.

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EARLY-MIDDLE ORDOVICIAN ACRITARCH ASSEMBLAGE FROM CHENGKOU, CHONGQING CITY, SOUTH CHINA

K. Yan¹, J. Li¹ and T. Servais²

¹ Nanjing Institute of Geology and Palaeontology, Chinese Academy of Sciences, State Key Laboratory of Palaeobiology and Stratigraphy, East Beijing Road, Nanjing 210008, China.
yankuiboy@gmail.com, junli@nigpas.ac.cn

² FRE 3298 du CNRS, Géosystèmes, Université de Lille 1, SN5, USTL, F-59655 Villeneuve d'Ascq, France. Thomas.Servais@univ-lille1.fr

Keywords: Acritarchs, Early-Middle Ordovician, Chengkou, Dacao Formation, Yingpan Formation, China.

INTRODUCTION

Although Chinese Paleozoic acritarch research developed fairly late, more than 100 scientific articles dealing with this topic have been published. Half of them described Ordovician acritarch assemblages from South China (Li et al., 2002a). Most of these studies were focused on acritarch taxonomy and biostratigraphy, while some publications discussed palaeogeographical and palaeoenvironmental implications.

The Houping section is located near Chengkou County. The Rock of Ordovician System is complete in Chengkou area. The Dacao Formation consists of dark grey lenticular limestone, oolitic limestone, and bioclastic limestone, with a total thickness reaching 33 m. And the overlying Yingpan Formation consists of dark grey to black or grey-green shales, intercalated with bioclastic limestone lenses, reaching a total thickness of 37 m. Biostratigraphical investigations in the Chengkou area are mostly based on conodonts and graptolites. In this study, the most recent graptolite biozonation, established by colleagues from the Nanjing Institute of Geology and Palaeontology (Zhang pers. comm., 2004; Zhan et al., 2005), is used.

MATERIAL AND METHODS

Fourteen palynological samples have been collected for acritarch analyses in the upper Dacao and Yingpan formations from the Houping section in Chengkou, Chongqing city (Fig. 1). The samples AFI2025, AFI2030, and AFI2033 were collected from the *approximatus* graptolite Biozone in the upper part of the Dacao Formation, and the sample AFI2035 from the *filiformis* graptolite Biozone in the lower part of the Yingpan Formation. The overlying samples AFI2037, AFI2038, and AFI2040 are from the *eobifidus* graptolite Biozone, and AFI2042 from the *deflexus* graptolite Biozone. The samples AFI2045, AFI2048, and

AFI2049 come from the *suecicus* graptolite Biozone, while AFI2050, AFI2053, and AFI2055 are from the *hirundo* graptolite Biozone in the upper part of the Yingpan Formation.

All samples were treated in the Palynological Laboratory of the Nanjing Institute of Geology and Palaeontology following standard palynological techniques. The slides and residues are housed in the collections of the Nanjing Institute of Geology and Palaeontology, Chinese Academy of Sciences, Nanjing, China.



Figure 1. Location of the Houping section, South of Chengkou county, Chongqing city.

RESULTS

In the palynomorph assemblages from the Dacao and Yingpan formations of the Houping section, a total of 41 species have been identified, attributed to 25 genera (Table 1).

The acritarch assemblage analysed from the *approximatus* graptolite Biozone in three samples, containing 25 species attributed to 16 genera, is dominated by the genera *Polygonium* (48.0-56.4%), *Rhopaliophora* (8.5-30.1%), *Leiosphaeridia* (12.9-20.1%), and *Dactylofusa* (0-19.4%).

An acritarch assemblage consisting of 17 species attributed to 10 genera was recorded from one sample of the *filiformis* graptolite Biozone. This assemblage is dominated by *Polygonium* (27.2%), *Leiosphaeridia* (17.9%), and *Micrhystridium* (13.5%).

From the *eobifidus* graptolite Biozone, 25 species attributed to 17 genera are recorded in the three samples in an assemblage that is dominated by *Leiosphaeridia* (25.2-33.6%), *Polygonium* (16.8-30.8%), and *Cymatiogalea* (10.3-14.5%).

Within the *deflexus* graptolite Biozone, the acritarch assemblage contains 14 species assigned to 12 genera, recorded from a single sample, and is dominated by *Polygonium* (40.5%), *Leiosphaeridia* (19.3%), and *Cymatiogalea* (13.8%).

The acritarch assemblage analysed from three samples within the *suecicus* graptolite Biozone is composed of 20 species attributed to 16 genera, being dominated by *Polygonium* (19.7-32.3%), *Leiosphaeridia* (12.7-21.7%), *Cymatiogalea* (8.1-18.3%) and *Peteinosphaeridium* (5.9-18.3%).

Within the *hirundo* graptolite Biozone, the acritarch assemblage consisting of 25 species attributed to

19 genera from three samples, is dominated by *Polygonium* (35.4-44.0%), *Rhopaliophora* (2.4-20.8%), and *Cymatiogalea* (0.7-17.5%).

acritarchs/Samples(AFI)	2025	2030	2033	2035	2037	2038	2040	2042	2045	2048	2049	2050	2053	2055
<i>Acanthodiacrodium burmanniae</i>	c												r	
<i>Acanthodiacrodium tassellii</i>	r												r	
<i>Ampullula composta</i>														r
<i>Ampullula crassula</i>					r			r						
<i>Ampullula erchunensis</i>					c	r	c	r	c		c	c		r
<i>Arbusculidium filamentosum</i>						c	c	r	c				c	
<i>Athabascaella playfordi</i>								r		r				
<i>Aureotesta clathrata</i> var. <i>simplex</i>	r					c			r		r			r
<i>Baltisphaeridium microspinosum</i>				c	c	r					r			
<i>Baltisphaeridium</i> sp.	r			c	c	c	r		c	c	a	a	r	c
<i>Coryphidium bohemicum</i>				a	r	a	c	c	a	r	a	a	a	
<i>Cristallinium dentatum</i>	c													
<i>Cymatiogalea granulata</i>	c	c	c		a	a	c	a	a	c	a	c	a	a
<i>Dactylofusa velifera</i> var. <i>brevis</i>		a	c			c								
<i>Ferromiact. pellita</i>	a		c											c
<i>Michrystidium acum brevispinosum</i>			r	c										
<i>Michrystidium accuminosum</i>	r		c	c	a	c	c	c	c				c	a
<i>Michrystidium henry</i>	a			c	r	r	r				c			r
<i>Multiplicisphaeridium</i> cf. <i>irregulare</i>				r		c	r						r	c
<i>Multiplicisphaeridium</i> sp.				r										c
<i>Pachysphaeridium rhabdocladum</i>	c					r								
<i>Petalofleridium bulliferum</i>	r										r			r
<i>Petalofleridium florgerum</i>	r										r			r
<i>Peteinosphaeridium armatum</i>					c	a	c	c	a	c	c	a	c	c
<i>Peteinosphaeridium tenuifolium</i>				r			c						r	c
<i>Polygonium gracile</i>	a	c	a	a	a	a	a	a	a	c	a	a	a	a
<i>Polygonium</i> sp.	c	a	r	c	c	c	c							c
<i>Rhopaliophora membrana</i>					r		r							
<i>Rhopaliophora palmata</i>	c	c	a	a	a	c	c	c	c	c	a	a	c	c
<i>Rhopaliophora pilata</i>	a		c	c		c		c	c			c	c	r
<i>Sacculidium inornatum</i>				r										r
<i>Sacculidium macropylum</i>				r										
<i>Stelliferidium striatulum</i>	r			c	c									c
<i>Striatotheca picipalis parva</i>			c			c		a	a			r	a	c
<i>Tongzia meitana</i>											r			
<i>Veryhachium ?estrellitae</i>	r													
<i>Veryhachium lairdii</i>	r						r							
<i>Veryhachium symmetricum</i>	c													
<i>Veryhachium trispinosum</i>	c						r	c	c		c			
<i>Leiosphaeridia</i> spp.	a	c	c	a	a	a	c	a	a	c	a	a	c	a
<i>Cymatiosphaera</i> sp.													r	

Table 1. Distribution and relative abundances of acritarchs and prasinophytes in the studied samples from the Houping section. All specimens present on the two slides for each processed sample were counted to determine relative abundances of species for each sample. A = Abundant (>50 specimens), C = Common (10–50 specimens), and R = Rare (<10 specimens).

STRATIGRAPHICAL IMPLICATIONS

Ordovician biostratigraphical divisions are mostly based on graptolite, conodont and latterly chitinozoan studies. In recent years, several easily recognizable taxa were selection for the recognition of

some Ordovician stage boundaries (e.g. Servais and Molyneux, 1997; Servais and Mette, 2000; Li et al., 2002b, 2003; Molyneux et al., 2007; Li et al., 2010).

Based on the material from the Dacao and Yingpan formations in the Houping section, Chengkou, Chongqing city, we are able to select several acritarch taxa with high biostratigraphical potential.

Aureotesta, *Petaloferidium*, *Striatotheca*, and the *Veryhachium lairdii* group-*V. trispinosum* group first appear in the *approximatus* graptolite Biozone, the first graptolite Biozone of the Floian in the Houping section. *Aureotesta* first appears in sample AFI2025 of the *approximatus* graptolite Biozone in the Dacao Formation from Houping section, Chongqing city, which indicates that the FAD is also in the lower part of the Floian in South China. The FAD of *Petaloferidium bulliferum* and *P. florigerum* probably occur in the *approximatus* graptolite Biozone, as these species first appear in sample AFI2025 from the Dacao Formation in the Houping section, Chengkou, Chongqing city. *Striatotheca principalis parva* first appears in the sample AFI2033 in the uppermost *approximatus* graptolite Biozone in the Houping section, Chengkou Chongqing city. The previous studies inferred that *Veryhachium lairdii* group-*V. trispinosum* group first appear in the *approximatus* graptolite Biozone in South China (Xu, 2001). In this study the *V. lairdii* group is present in the samples AFI2025 and AFI2040, while the *V. trispinosum* group appears in the samples AFI2025, AFI2040, AFI2042, AFI2045, and AFI2049. The FAD of both genera are thus in the *approximatus* graptolite Biozone in the Dacao Formation of the Houping section, Chengkou.

Ampullula, *Coryphium bohemicum*, *Sacculidium* first appear slightly higher in the *filiformis* graptolite Biozone, while *Arbusculidium filamentosum* and *Tongzia* first appears in the *eobifidus* graptolite Biozone in South China. Yan et al. (2010) revised *Ampullula* and discussed its biostratigraphical implications. The FAD of *Ampullula* is at the base of the *eobifidus* graptolite Biozone in South China. *A. crassula* and *A. erchunensis* appear in the sample AFI2037 at the base of the *eobifidus* graptolite Biozone in the Yingpan Formation of the Houping section, Chengkou, Chongqing city while *A. composita* is present in the sample AFI2055 in the *hirundo* graptolite Biozone in the Yingpan Formation. The genus *Coryphidium* has been reviewed taxonomically and its biostratigraphical implication has been discussed in detail (Servais et al., 2008). *Coryphidium bohemicum* is the most common species present in South China. It first appears in sample AFI2035 at the base of the *filiformis* graptolite Biozone in the Houping section, Chengkou, Chongqing city. *Sacculidium* appears in the sample AFI2035 and AFI2055 in the *filiformis* graptolite Biozone in the Yingpan Formation of the Houping section, Chengkou, Chongqing city. The present study shows that *A. filamentosum* first appears in sample AFI2038 of the *eobifidus* graptolite Biozone from the Houping section, Chengkou, Chongqing city as well as in the sample AFI1010 from the Honghuayuan section, Tongzi, Guizhou province. In this study, *Tongzia* appears in sample AFI2049 of the *suecicus* graptolite Biozone of the Yingpan Formation of the Houping section which is present later than that of the Huanghuachang section.

As a result, the FAD of the acritarch taxa mentioned above have potential for recognition of the Floian, and several other acritarch taxa have been selected by Li et al. (2010) for recognition the Middle Ordovician in South China. Based on these progresses on acritarch biostratigraphical, we can push forward to establish the acritarch biostratigraphical assemblage zones.

PALAEOENVIRONMENTAL IMPLICATIONS

During the Ordovician, South China can be divided from northwest to southeast into three major regions, the Yangtze, Jiangnan and Zhujiang regions, representing platform, slope, and basin depositional

environments, respectively (Zhang et al., 2010). Several depositional belts are recognised, which are approximately parallel to one another.

We selected acritarch assemblages from six sections in the *deflexus-suecicus* graptolite Biozones to analyse the possible environmental implications. Three groupings are recognised with the cluster and principle component analyses. The Honghuayuan section, Tongzi, the Guanyinqiao section, Qijiang and Houping section, Chengkou were located in the inner-shelf mud-carbonate belt which are dominated by the genera *Polygonium*, *Leiosphaeridia*, and *Cymatogalea*. The assemblages from the Honghuayuan and Guanyinqiao sections are much more similar because they are located nearby. The Houping section in Chengkou should be closer to the continent and the relative abundance of *Stelliferidium* is much higher in the Honghuayuan, Tongzi and Guanyinqiao sections, Qijiang. Another grouping consists of the two sections from the Yichang area which were attributed to an outer-shelf carbonate-mud belt environment. The acritarch assemblages from these two sections are dominated by *Baltisphaeridium*, *Peteinosphaeridium*, and *Rhopaliophora*, which show a typical offshore acritarch assemblage. The acritarch assemblage from the Huangnitang section in Changshan is dominated by *Leiosphaeridia* and *Baltisphaeridium* with a fairly low diversity implying that the Huangnitang section was probably located on the slope.

The acritarch assemblage distribution models in South China studied both Li et al. (2004) and in this study are therefore similar to those of Jacobson (1979) and Dorning (1981). Comparing with the previous studies (Jacobson, 1979; Dorning, 1981; Li et al., 2004), it is thus possible to use the cluster and principle component analyses for distinguishing acritarch assemblage groups.

CONCLUSION

A diverse acritarch and prasinophyte assemblage consisting of 41 species attributed to 25 genera is described from the Dacao and Yingpan formations of the Chengkou section, Chongqing. The palynomorph assemblages are dominated by the genera *Polygonium*, *Leiosphaeridia*, and *Rhopaliophora*.

Aureotesta, *Petaloferidium*, *Striatotheca*, and the *Veryhachium lairdii* group-*V. trispinosum* group first appear in the *approximatus* graptolite Biozone, the first graptolite Biozone of the Floian in South China. *Ampullula*, *Coryphium bohemicum*, *Sacculidium* first appear slightly higher in the *filiformis* graptolite Biozone, while *Arbusculidium filamentosum* and *Tongzia* first appears in the *eobifidus* graptolite Biozone in South China. These selected acritarch taxa may thus be useful for stratigraphical correlations in South China and at a global level.

Three distinguished associations or groupings can be recognised by cluster and principle component analyses based on data from six Floian-Dapingian sections in South China. The association recovered from the Honghuayuan section, Tongzi, the Guanyinqiao section, Qijiang, and Houping section, Chengkou, show similar acritarch assemblages. These three sections were located in the inner-shelf mud-carbonate belt during the Early-Middle Ordovician in South China.

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BIOSTRATIGRAPHY AND PALEOENVIRONMENTS OF THE SANTA ROSITA FORMATION (LATE FURONGIAN–TREMADOCIAN), CORDILLERA ORIENTAL OF JUJUY, ARGENTINA

F.J. Zeballo¹, G.L. Albanesi^{1,2} and G. Ortega^{1,2}

¹ Museo de Paleontología, Facultad de Ciencias Exactas, Físicas y Naturales, Universidad Nacional de Córdoba, Casilla de Correo 1598, 5000 Córdoba, Argentina. fzeballo@efn.uncor.edu, galbanes@com.uncor.edu, gcortega@arnet.com.ar

² CONICET.

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INTRODUCTION

A comprehensive stratigraphic scheme from the Cordillera Oriental of northwestern Argentina is necessary to accomplish a synthesis on the current amount of nominated units, as well as to establish a reliable scheme for future geological research. An attempt to correlate the different sequences of the Santa Rosita Formation was carried out by Buatois et al. (2006). Accordingly, it is necessary to achieve a precise dating of the sequence boundaries and tectono-eustatic events over the involved interval, particularly related to the architecture of different sedimentary facies.

This contribution provides an integrated biostratigraphic scheme based on the three main index fossils of the Lower Paleozoic (*i.e.*, conodonts, graptolites and trilobites) for the rock series exposed on the western flank of the sierra de Tilcara, and other classical localities (*e.g.*, Purmamarca, Parcha) from the Argentine Cordillera Oriental of Jujuy and Salta provinces (Fig. 1).

BIOSTRATIGRAPHY

Cordylodus intermedius Zone

At the study area, the *Cordylodus intermedius* Zone is located in the lower half of the first transgressive-regressive cycle or lower interval of the Alfarcito Member, cropping out in the Salto Alto, El Arenal and Tres Ciénagas creeks. Conodonts from the Salto Alto section were recovered from a coquina level that is intercalated between sandstones at the basal interval of the Alfarcito Member and from five calcarenite levels from the lower part of the same unit. Determined conodont species include: *Cordylodus caboti* Bagnoli, Barnes and Stevens, *C. intermedius* Furnish, *C. proavus* Müller, *C. viruanus* Viira and Sergeyeva, *C. cf. andresi* Viira and Sergeyeva, *C. cf. tortus* Barnes, *Cordylodus* n. sp., *Drepanodus* sp.,

Hirsutodontus simplex (Druce and Jones), *H. galerus* Tolmacheva and Abaimova, *Variabiloconus datsonensis* (Druce and Jones), *V. bicuspatus* (Druce and Jones), *Semiacontiodus* sp., *Teridontus nakamurai* (Nogami), *T. gallicus* Serpagli, Ferretti, Nicoll and Serventi, and a high frequency of proto and paraconodonts is apparent, such as *Albiconus postcostatus* Miller, *Phakelodus elongatus* (Zhang), *P. tenuis* (Müller), *Problematoconites perforatus* Müller, ? *Prooneotodus* 'mitriformis' Dubinina, *Prosagittodontus* sp. and *Westergaardodina polymorpha* Müller and Hinz (Zeballo and Albanesi, 2009; Zeballo et al., 2009).

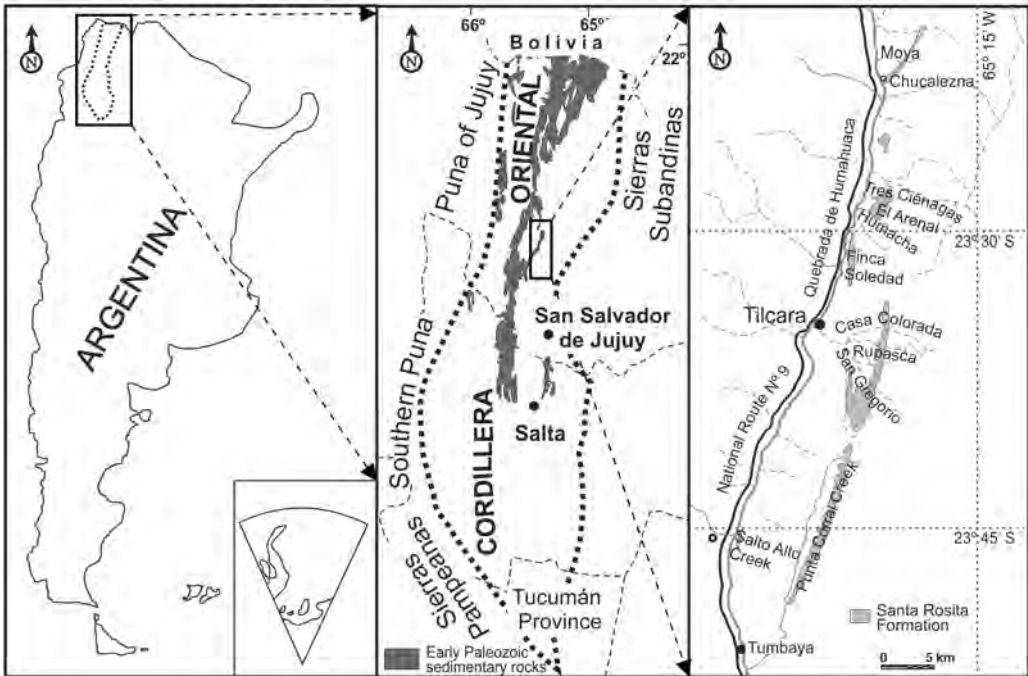


Figure 1. Location map of the study area in the Cordillera Oriental of northwestern Argentina showing Cambro-Ordovician outcrops in the surveyed sections.

The collected conodont fauna corresponds to the *Cordylodus intermedius* Zone, whose chronostratigraphic position occurs in the Furongian Stage 10, i.e., upper Cambrian. Following documented records of species (e.g., Druce and Jones, 1971; Miller et al., 2006), the nominal taxon of the *Hirsutodontus simplex* Subzone and associated taxa such as *Albiconus postcostatus*, *Variabiloconus datsonensis*, *V. bicuspatus* and *Westergaardodina polymorpha* do not cross over the Cambrian-Ordovician boundary. Incidentally, we interpret that the taxa *Cordylodus* cf. *andresi* and *Hirsutodontus galerus* are derived forms of *C. andresi* and *H. hirsutus*, respectively, whose ranges do not surpass the *C. proavus* and *C. intermedius* zones, respectively (Miller et al., 2006; Tolmacheva and Abaimova, 2009). The homotaxial relationships constrain the reported conodont association to the lower part of the *C. intermedius* Zone, in the *Hirsutodontus simplex* Subzone.

Trilobites from the *Jujuyaspis keideli keideli* Zone were described from the same stratigraphic levels, where the most abundant forms include *Angelina hyeronimi* (Kayser), *Bienvilla* sp. and *Parabolinella* sp. together with the eponymous species (Zeballo and Albanesi, 2009) (Fig. 2).

SYSTEM	SERIES	STAGE	BIOZONES			GROUP	CORDILLERA ORIENTAL		TRANSGRESSIVE- REGRESSIVE EVENTS				
			GRAPTOLITES	CONODONTS	TRILOBITES		WESTERN	EASTERN	T	R			
							Formation	Formation			Member		
ORDOVICIAN	LOWER	TREMADOCIAN	phyllograptoides	Acodus deltatus- Paroistodus proteus	Thysanopyge	Santa Victoria	Parcha	San Bernardo					
			H. copiosus				TUMBAYA	Áspero- Humacha					
			Araneograptus murrayi								Saladillo	Floresta- Rupasca	
			K. supremus	Devendeus	Cardonal		Alfarcito	San José La Pedrera					
			Aorograptus victoriae								Cordylodus angulatus	K. meridionalis	Kainella andina
			Bryograptus	Cordylodus proavus	Parabolina (N.) frequens argentina		Lampazar	Casa Colorada					
			Adelograptus								C. lindstromi	Jujuyaspis keideli	Padric- Sococha
			"R. f. anglica"	C. intermedius	hiatus		IRUYA	Chalhualmayoc					
			A. matanensis								Unnamed interval	Iapetognathus	C. lindstromi
			Unnamed interval	Iapetognathus	C. lindstromi		Jujuyaspis keideli	hiatus					
Unnamed interval	Iapetognathus	C. lindstromi	Jujuyaspis keideli			hiatus			IRUYA	Chalhualmayoc			

Figure 2. Conodont, graptolite and trilobite biostratigraphic scheme of the Furongian – Tremadocian related to lithostratigraphic units and the sea level curve, proposed for the study area. LREE: Lange Ranch Eustatic Event, BHL: Basal House Lowstand, ARE: *Acerocare* Regressive Event, BMEE: Black Mountain Eustatic Event, PRE: *Peltocare* Regressive Event, CRE: *Ceratopyge* Regressive Event (modified after Albanesi et al., 2008; Miller et al., 2006; Waisfeld and Vaccari, 2008; Vaccari et al., 2010).

Cordylodus lindstromi sensu lato Zone

The *Cordylodus lindstromi* sensu lato Zone is poorly represented in the study area, being localized only at Punta Corral creek, from the lowest coquina of the Alfarcito Member (Zeballos et al., 2009). This level corresponds to the upper half of the first cycle (lower interval) in the Alfarcito Member, just above the basal sandstones and siltstones. The conodont association includes *C. prolindstromi* Nicoll, *Striatodontus* sp., *Teridontus gallicus*, *Variabiloconus datsonensis* and *Phakelodus elongatus*. This record represents the upmost approximation to the Cambrian-Ordovician boundary in the Quebrada de Humahuaca area so far. The boundary should be expected in the upper strata of the lower interval from the Alfarcito Member; however, the erosive regressive surface reported for this part of the sequence is precluding a complete stratigraphic record (Zeballos, 2010).

Cordylodus angulatus Zone

The *Cordylodus angulatus* Zone is the best represented in the study area regarding number of specimens and faunal diversity. It is recorded in the Moya, Chucalezna, Tres Ciénagas, El Arenal, Casa Colorada, Rupasca and San Gregorio sections, and out of the study area in the Trampeadero creek, at Pascha area. Fossiliferous rocks were sampled from top levels of the lower interval, throughout the middle-upper interval of the Alfarcito Member, and the lower Rupasca Member.

Recorded conodonts belong to the species *Acanthodus* n. sp. A, *Cordylodus angulatus* Pander, *C. caseyi* Druce and Jones, *C. intermedius*, *C. cf. prion* Lindström, *Cordylodus* sp., *Drepanoistodus alfarcitensis* Zeballo, Albanesi and Ortega, *D. chucaleznsis* Albanesi and Aceñolaza, *Iapetognathus* sp., *Kallidontus* n. sp., *Paltodus* sp., *Problematoconites perforatus*, *Rossodus manitouensis* Repetski and Ethington, *Rossodus* sp., *Semiacontiodus striatus* Zeballo, Albanesi and Ortega, *S. minutus* Zeballo, Albanesi and Ortega, *Striatodontus* sp., *Teridontus gallicus*, *Tilcarodus humahuacensis* (Albanesi and Aceñolaza), *Utahconus* n. sp. A, *Utahconus* sp., *Variabiloconus cf. datsonensis*, *Variabiloconus* n. sp., together with *Coelocerodontus* sp., *Phakelodus elongatus* and *P. tenuis*. The more complex association is found in the upper part of the biozone, which is located in the third stratigraphic cycle of the Alfarcito Member and the lowest Rupasca Member. The El Arenal creek is one of the sections that provided the most complete conodont collection, including most of new species holotypes.

A faunal assemblage related to this biozone is documented in the Trampeadero creek, Pascha area, from the Devendeus Formation (Zeballo and Albanesi, 2007). It indicates a youngest age corresponding to the uppermost part of the *Cordylodus angulatus* Zone, being this biozone restricted to the lower Tremadocian, as the subsequent unit that follows the Cambrian-Ordovician boundary interval.

Previous studies in the Alfarcito area by Zeballo et al. (2005a, 2005b) and Zeballo and Tortello (2005) identify the *C. angulatus* Zone in the Alfarcito Member, corresponding to the *Kainella meridionalis* Zone. The taxonomic assignment of those faunas is re-interpreted according to the new material recorded in the study area, and the new definition of trilobite biozones by Waisfeld and Vaccari (2008) and Vaccari et al. (2010). Moreover, the graptolite *Adelograptus* n. sp. (Zeballo et al., 2009; Zeballo, 2010) that characterizes the *Adelograptus* Zone is constrained to the upper *C. angulatus* Zone.

Paltodus deltiifer Zone

The lower boundary of this biozone in the study area is located between 8 and 10 m above the base of the Rupasca Member, in the Chucalezna, El Arenal, and San Gregorio sections (Albanesi and Aceñolaza, 2005; Zeballo et al., 2005a; Zeballo, 2010). The entrance of *Paltodus deltiifer* in the Casa Colorada creek section is missing because of virtually barren strata in the critical interval (Zeballo et al., 2005a). This bias is assuming the base of the biozone to upper levels in the latter section. The biozone is also documented in the Coquena creek, where the upper interval is controlled by conodont appearances of the overlying biozone (Zeballo et al., 2008, 2009).

Figure 3. Representative conodonts and graptolites recovered from the eastern flank of the Quebrada de Humahuaca. 1, *Cordylodus cf. andresi* Viira and Sergeyeva (CORD-MP 12554/1); 2, *Hirsutodontus simplex* (Druce and Jones) (CORD-MP 12546/3); 3, *Hirsutodontus galerus* Tolmacheva and Abaimova (CORD-MP 12566/1); 4, *Cordylodus proavus* Müller (CORD-MP 12536/1); 5, *Albiconus postcostatus* Miller (CORD-MP 12550/1); 6, *Westergaardodina polymorpha* Müller and Hinz (CORD-MP 12548/2); 7, *Cordylodus prolindstromi* Nicoll (CORD-MP 16558/1); 8, *C. cf. prion* Lindström (CORD-MP 16564/1); 9, *Drepanoistodus alfarcitensis* Zeballo, Albanesi and Ortega (CORD-MP 16477/1); 10, *Cordylodus intermedius* Furnish (CORD-MP 16475/1); 11, *Cordylodus angulatus* Pander (CORD-MP 16474/2); 12, *Paltodus cf. subaequalis* Pander (CORD-MP 11295/1); 13, *Drepanoistodus chucaleznsis* Albanesi and Aceñolaza (CORD-MP 16619/1); 14, *Paltodus deltiifer pristinus* (Viira) (CORD-MP 16620/2); 15, *Drepanoistodus cf. concavus* (Branson and Mehl) (CORD-MP 16651/1); 16, *Cornuodus* sp. (CORD-MP 16657/1); 17, *Paltodus deltiifer deltiifer* (Lindström) (CORD-MP 11356/1); 18, *Drepanoistodus nowlani* Ji and Barnes (CORD-MP 11313/1); 19, 20, *Rhabdinopora flabelliformis flabelliformis* (Eichwald) (CORD-PZ 30808, CORD-PZ 30810); 21, *Aorograptus victoriae* (T.S. Hall) (CORD-PZ 32001); 22, '*Adelograptus*' cf. *altus* Williams and Stevens (CORD-PZ 31727); 23, *Ancoragraptus bulmani* (Spjeldnaes). 1-6: *Cordylodus intermedius* Zone, 7: *C. lindstromi* sensu lato Zone, 8-11: *C. angulatus* Zone, 12-18: *Paltodus deltiifer* Zone, 19-20: *Adelograptus* Zone, 21-23: *Aorograptus victoriae* Zone. 1-18, scale bar: 0.1 mm, 19-23, scale bar: 1 mm.

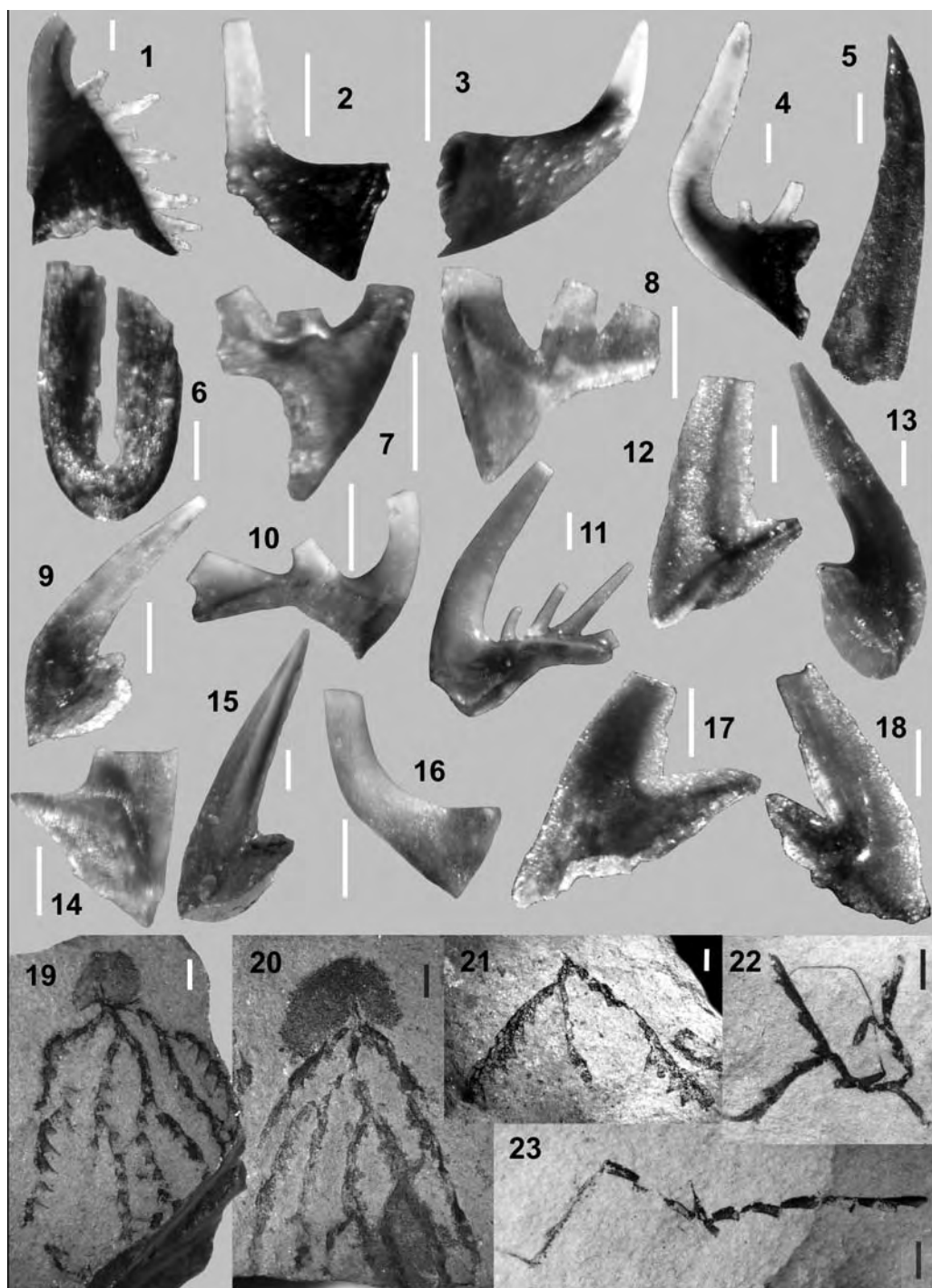


Figure 3.

The conodont fauna from the lower *Paltodus deltifer pristinus* Subzone consists of the species *Coelocerodontus* sp., *Cordylodus angulatus*, *Drepanodus arcuatus* Pander, *Drepanoistodus chucaleznsensis*, *D. cf. concavus* (Branson and Mehl), *Filodontus* sp. A (Ji and Barnes), *Kallidontus* n. sp., *Paltodus deltifer pristinus* (Viira), *P. deltifer* n. ssp., *P. cf. subaequalis* (Pander), *Teridontus gallicus*, *Tilcarodus humahuacensis*, *Utahconus* n. sp. A, *Utahconus* n. sp. B, *Variabiloconus* n. sp. including *Furnishina?* sp., *Phakelodus elongatus* and *P. tenuis* as earlier forms.

The upper *Paltodus deltifer deltifer* Subzone spans the top level of the Rupasca Member (previously assigned to the base of the Humacha Member by Zeballo et al., 2008) and the Humacha Member, both units exposed in the Humacha creek. A similar fauna is reported from the top strata of the Lower Member of the Coquena Formation. The conodont species that compose this fauna are: *Acanthodus* n. sp. A, *Acanthodus* n. sp. B, *Acodus* n. sp., *Drepanoistodus chucaleznsensis*, *D. costatus* (Abaimova), *D. nowlani* Ji and Barnes, *D. cf. concavus*, *Kallidontus* n. sp., *Paltodus deltifer deltifer* (Lindström), *P. deltifer* n. ssp., *P. cf. subaequalis*, *Phakelodus elongatus*, *P. tenuis*, *Utahconus* n. sp. B, *Utahconus* n. sp. C and *Variabiloconus variabilis* (Lindström). The taxon *Acodus* n. sp., previously considered as *A. deltatus* by Zeballo et al. (2008), is now interpreted as an ancestral form of the lineage appearing in the uppermost interval of the *Paltodus deltifer* Zone. *Acodus* n. sp. would be phylogenetically linked to younger forms that appear in other sections of the Cordillera Oriental, associated to conodonts of the *Paroistodus proteus* – *Acodus deltatus* Zone and to the graptolite *Araneograptus murrayi*, which determine a younger age for the bearer levels (Albanesi et al., in press).

The *P. deltifer pristinus* and *P. deltifer deltifer* subzones are correlative with the trilobite zones *Bienvillia tetragonalis* and *Notopeltis orthometopa*, respectively. In association with these fauna, the record of '*Adelograptus*' cf. *altus* Williams and Stevens, *Ancoragraptus bulmani* (Spjeldnaes) and *Aorograptus victoriae* (T.S. Hall), represent the *Aorograptus victoriae* Zone (Zeballo et al., 2008, 2009) (Fig. 3).

PALEOENVIRONMENTS

The conodont faunas can be merged in biofacies that typifies diverse paleoenvironments. The *Acanthodus-Utahconus* biofacies characterizes shallow water environments (e.g., Ji and Barnes, 1994) and is recurrent through different levels of the Santa Rosita Formation. Alternatively, the *Tilcarodus-Drepanoistodus* biofacies clearly represents deeper water environments, while the *Variabiloconus-Teridontus* biofacies is overlapping all sampled lithofacies.

The conodont associations show affinities with Baltoscandinavian faunas, including index species (*Paltodus deltifer*), although warm water taxa are recorded as well (e.g., *Hirsutodontus*). This conodont fauna integrates a compositional mixture that incorporates endemic genera of the Gondwanan margin, such as *Tilcarodus*. The conodont assemblages are diagnostic of a new paleobiogeographic unit; i.e., the Southwestern Gondwanan Province, composing the Cold Domain of the Shallow Water Realm (cf. Albanesi et al., 2007).

The graptolites *Aorograptus victoriae* and *Ancoragraptus bulmani* are documented from proximal off-shore facies in the study area, though expanding its paleoenvironmental record that referred them to outer-platform to slope environments.

CONCLUSIONS

The biostratigraphic analysis demonstrates the homotaxial succession of conodont zones, linked to particular graptolite and trilobite units. The correlation of the *Cordylodus intermedius* Zone (*Hirsutodontus simplex* Subzone) with the *Jujuyaspis keideli keideli* Zone indicates that at least the lower interval of the latter unit is late Cambrian (late Furongian) in age, occupying the lower part of the Alfarcito Member (Zeballo and Albanesi, 2009).

The *Cordylodus angulatus* Zone is identified from the middle-upper tracts of the Alfarcito Member up to the base of the Rupasca Member, being related to the *Adelograptus* and *Kainella* spp. zones. The *C. angulatus* Zone is also well-represented in the Devendeus Formation, exposed to the southwest of the study area, in the Cordillera Oriental of Salta Province.

The subsequent *Paltodus deltiifer pristinus* and *P. deltiifer deltiifer* subzones are determined within the Rupasca and Humacha members, being partly correlated with the *Bienvillia tetragonalis* and *Notopeltis orthometopa* zones, respectively. Graptolites of the *Aorograptus victroriae* Zone are associated to this conodont fauna.

Particular species, such as *Hirsutodontus galerus*, *H. simplex*, *Variabiloconus datsonensis* and *V. bicuspatus*, previously reported in low paleolatitude regions (Australia, Laurentia, Siberia, North China) are documented in the study area. These records, together with the trilobite *Onychopyge* and the gasteropod *Peelerophon oehlerti* in the Cordillera Oriental (Benedetto, 2003; Benedetto et al., 2009) verify the occurrence of a perigondwanic corridor that would have favored the connection between faunas from Australia and New Zealand with those from other distant tropical regions through the northwestern Argentine basins.

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ON THE MACROEVOLUTION OF *EOSPIRIFER* SCHUCHERT, 1913 (SPIRIFERIDA, BRACHIOPODA)

R. Zhan, Y. Liang and L. Meng

State Key Laboratory of Palaeobiology and Stratigraphy, Nanjing Institute of Geology and Palaeontology, Chinese Academy of Sciences, Nanjing 210008, China. rbzhan@nigpas.ac.cn, liangyan@nigpas.ac.cn

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INTRODUCTION

Spiriferida, one of the major groups within the Brachiopoda, experienced its origination and major development during the Palaeozoic Era (Rong et al., 1994). It usually constitutes the dominant component of the brachiopod fauna between Silurian and Permian periods when the Paleozoic Evolutionary Fauna flourished (Sepkoski, 1995). Eospiriferines, the stem group of Spiriferida, were taking the leading position of the entire order during the Silurian and Early Devonian. *Eospirifer*, the root and the earliest known genus of this major group, has a similar macroevolutionary pattern to eospiriferines and the Spiriferida. This paper is trying to briefly investigate the macroevolutionary route of *Eospirifer*, such as its origination, radiation (including its palaeogeographic dispersal and global distribution) and extinction, in order to reveal its implication to the macroevolution of the entire group and phylum.

TYPE SPECIES AND THE EARLIEST KNOWN SPECIES OF *EOSPIRIFER*

Eospirifer was named by Schuchert in 1913 on the basis of its type species *Spirifer radiatus* Sowerby, 1835 from the Wenlock Limestone of Dudley, English Midlands. St. Joseph (1935) was the first person who described in detail the interiors of *Eospirifer radiatus*, particularly its crural structure and spiralia. Rong and Zhan (1996) sectioned a few topotype specimens of *E. radiatus* and reconstructed its crura and spiralia while they systematically studied a series of Late Ordovician to Silurian eospiriferine taxa around the world. And, for the first time, the spiralia and early evolution of eospiriferines had been convincingly discussed.

The earliest known species of *Eospirifer*, *E. praecursor*, was discovered and systematically described by Rong and his colleagues in 1994 from the upper Changwu Formation (late Katian, Late Ordovician, Zhan and Cocks, 1998) at Pengli of Hejiashan, Jiangshan County, western Zhejiang Province, East China. It was

collected from the greenish yellow (weathered color) mudstone and all the specimens are preserved as external and internal moulds. So, nothing was known about its spiralia when it was named. Fortunately, just one year after its first discovery in 1991, thousands of conjoined valves (loose specimens) of *Eospirifer praecursor* were found and collected from the Xiazhen Formation (corresponding rocks of the Changwu Formation, Zhan and Fu, 1994) at Zhuzhai, Yushan County, northeastern Jiangxi Province, about 50 km southwest of the type locality of *E. praecursor*. Serial sections were made for more than 20 individuals (normally about 3-5 mm long/wide) of *E. praecursor*, and the spiralia were found and reconstructed (Rong and Zhan, 1995, 1996).

Compared with the oldest species *Eospirifer praecursor*, the type species *E. radiatus* is much younger (Wenlock age). Morphologically, it is much bigger in shell size (normally 20-30 mm wide), and has much stronger radial costellae on the entire shell surface (Fig. 1). Its comb-like, striated cardinal process is well-developed in almost all sectioned specimens, but similar structure never occurs in *E. praecursor*. Its spiralia has much more whorls (usually around 10 whorls) than that of *E. praecursor* which is normally 3-4 whorls in adults (Rong and Zhan, 1996).

TEMPORAL AND SPATIAL OCCURRENCES OF *EOSPIRIFER*

Up to now, *Eospirifer* has been reported from the Late Ordovician to Mid Devonian rocks of more than 25 palaeoplates or terranes. On some of these blocks, there are many localities and horizons with occurrences of different species of *Eospirifer*, e.g., Laurentia, South China and Avalonia.

Late Katian

Eospirifer originated at a near shore shallow water benthic regime represented by *Eospirifer praecursor* Rong, Zhan and Han, 1994. On the narrow Zhe-Gan Platform and the upper part of the Zhexi Slope, eastern South China palaeoplate (Rong and Chen, 1987), it has a wide ecological distribution and various types of substrates, which may indicate that those evolutionary novelties enable it a strong potential in adapting various benthic environments. The morphological novelties holden by *E. praecursor* include: 1) a well-developed interarea; 2) a wider and straight hinge-line; 3) a well-defined dorsal fold and ventral sulcus; 4) fine radial microsculpture; 5) spiralia directed ventro-laterally; and 6) a pair of small jugal processes rather than jugum, amongst which the development of spiralia and jugal processes are the most important.

Hirnantian

Affected by the first episode of the end Ordovician mass extinction, *Eospirifer* disappeared from South China and was not found anywhere in the world for about two graptolitic biozones. At Honghuayuan of Tongzi County, northern Guizhou Province, South China, some specimens of *Eospirifer* sp. (Chen et al., 2000; Rong et al., 2002) were collected from the yellow mudstone of the upper Kuanyinchiao Formation (the middle part of the *Undulograptus persculptus* Biozone, late Hirnantian, latest Ordovician). All specimens found are external and internal moulds, and about the same shell sizes as those found in eastern South China of late Katian age. They shall several common morphological features such as interarea, straight hinge-line, fold and sulcus and fine radial ornamentation. But nothing is known about its spiralia which leaves its specific name unidentifiable.

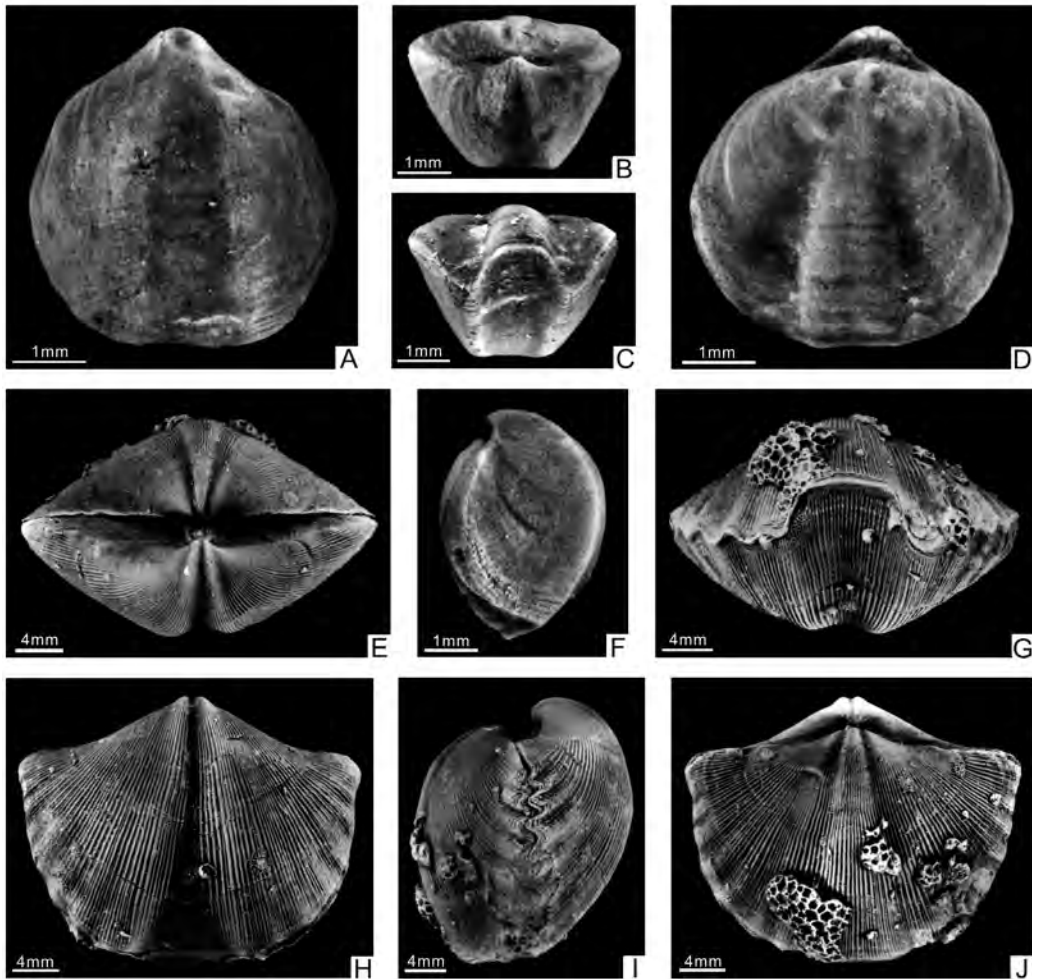


Figure 1. A–D, F, *Eospirifer praecursor* Rong, Zhan and Han, 1994, NIGP 124756, ventral, posterior, anterior, dorsal and lateral views of a conjoined individual from the Xiazhen Formation (late Katian, Late Ordovician) at Zhuzhai, Yushan County, northeastern Jiangxi Province, East China. E, G–J, *Eospirifer radiatus* (Sowerby, 1834), NIGP 124759, posterior, anterior, ventral, lateral and dorsal views of a conjoined individual from the Mulde Beds (middle-upper Wenlock, Silurian), Gotland, Sweden.

Rhuddanian

Possibly benefitted from the evolutionary novelties it holds, *Eospirifer* enjoyed a major development immediately after the second episode of the end Ordovician mass extinction. Besides its wider distribution in South China, it rapidly expanded to 8 other palaeoplates and terranes during the Rhuddanian, such as Tasmania Australia, Chinghiz Kazakhstan, many localities of Laurentia, Oslo Norway, northern Greenland, Myanmar Sibumasu, Gieben Germany, and T'ien-Shan (see Appendix for details). *E. radiatus* found from the Aroostook Limestone (Rhuddanian) of Aroostook County, northern Maine (Twenhofel, 1941) might represent the earliest known occurrence of this species.

Aeronian

The continuous growing species diversity indicates that *Eospirifer* of this interval continued its major development although no big changes happened to its palaeogeographic distribution except two more palaeoplates with its new occurrences: North Africa and North China. Its distribution on the Upper Yangtze Platform (South China) became even broader from south to north and west to east covering a vast area of thousands of square kilometers. Several new species originated in southwestern South China. Besides, *Eospirifer* was also reported from Tasmania Australia, Chinghiz, Tuva, many localities of Laurentia, Morocco Africa, Ningxia North China, Northern Greenland, Myanmar Sibumasu, Gieben Germany, and T'ien-Shan (see Appendix).

Telychian

On species level, the diversity of *Eospirifer* got its acme in this interval by having 24 species reported from more than 12 palaeoplates or terranes. The major contribution to this diversity acme comes from Laurentia where different species of *Eospirifer* have been reported from many localities of the United States and Canada, such as Tennessee, Newfoundland, Maine, New York, Pennsylvania, Indiana, Maryland, Kentucky, Anticosti Island, Yukon, Nova Scotia, Massachusetts, Oklahoma, Arkansas, New Brunswick, and Chihuahua Mexico. There are also some new occurrences reported from northern Europe, e.g. Estonia and Gotland Sweden. Continuous from the previous interval, *Eospirifer* was also found in Tasmania Australia, central Kazakhstan and Tuva, Morocco North Africa, Northern Greenland, Podolia Ukraine, Myanmar Sibumasu, South China, Gieben Germany of Southern Europe, and T'ien-Shan (see Appendix).

Sheinwoodian

After its macroevolutionary climax in Telychian, *Eospirifer* experienced a gradual decrease starting from this interval. Although its species diversity has no big change compared with that of previous interval, it disappeared from a few palaeoplates or terranes, such as Australia, Northern Greenland, and Southern Europe. Its occurrences in Northern Europe are also becoming fewer with only one documentation from the Lower Visby Marl (Sheinwoodian) of Gotland Sweden. The most outstanding character of this interval is that *Eospirifer* was extremely flourishing in Laurentia with occurrences at many places, such as Arkansas, Tennessee, Oklahoma, Maine, New York, Pennsylvania, Indiana, Maryland, Wisconsin, Ohio, Gaspé, Anticosti Island, Newfoundland, Ontario, Yukon, Massachusetts, and New Hampshire (see Appendix). Another event that should be mentioned here is that *Eospirifer* successfully expanded to Bohemia in this interval with the occurrence of *E. pollens* (Barrande, 1848) from the upper Motol Formation (Sheinwoodian) in the area between Hills Kolo and Branzovy, Kozolupy, Bohemia, Czech Republic (Havliček, 1980).

Homerian

Although keeping similar species diversity to that of former interval, *Eospirifer* of Homerian age had a slightly wider palaeogeographic distribution with new occurrences in Avalonia (many localities of UK) and Tarim (Rong and Chen, 2003). Again, concerning the macroevolution of *Eospirifer*, its most flourished area is still in Laurentia where it was reported from almost all states marginal to the continent (see Appendix). Its distribution in South China was becoming slightly smaller confined to northeastern Upper Yangtze Platform, such as Hubei Province (Zeng, 1977).

Gorstian

Further decrease in species diversity, *Eospirifer* in this interval had an even wider palaeogeographic distribution by having been documented in 13 palaeoplates or terranes. Although it was again most flourished in Laurentia, the number of occurrences became much fewer than before concentrating mainly in the eastern margin of the continent (see Appendix). It disappeared from Tarim but was back to Australia after two intervals represented by *E. eastoni* Gill, 1949 from the Dargile Formation (Gorstian) of Melbourne Trough, Victoria. The occurrence of *E. ferganensis* Nikiforova, 1937 from the *Marginalis* and *Isfara* Beds of Turkestan is questionable because this species was used by Boucot (1962) as the type species to establish a new genus *Nikiforovaena*. Some specimens from the *Striatus* Beds of Urals (Russia) were also identified as *E. cf. radiatus* by Khodalevitch (1939).

Ludfordian

Having similar species diversity and palaeogeographic distribution as that of previous interval, *Eospirifer* of Ludfordian age experienced no major macroevolutionary changes except for its occurrence in the Fukuji Formation (Ludfordian) of central Japan (Ohno, 1977).

Pridoli

Coming to the end of Silurian, *Eospirifer* experienced an apparent decrease of species diversity and a shrink of palaeogeographic distribution. It became extinct in Avalonia, and provisionally disappeared in Bohemia, Turkestan and Urals (Russia). The number of its occurrences in Laurentia also became much fewer with only a few specimens found in New Hampshire, Chihuahua Mexico and Canadian Cordillera (see Appendix).

Lochkovian

There are four blocks which were reoccupied by *Eospirifer* in Lochkovian: Germany, Bohemia, Turkestan and Urals. Southeastern Alaska records the occurrence of *Eospirifer* for the first time (see Appendix). And the number of occurrences in Australia becomes more than before, such as Victoria, New South Wales, and New Zealand. So, the species diversity of this interval is much higher than that of former one, and the palaeogeographic distribution is also slightly wider, representing the second macroevolutionary climax of *Eospirifer* although it was much smaller than the first one (Telychian) in scale. One thing should be mentioned here is that *Eospirifer* became extinct in South China starting from this interval.

Pragian

The species diversity of this interval is lower than that of former one. Major changes are: 1) it disappeared from Podolia, Southern Europe and Turkestan; 2) new documentations were reported from the Altai Mountains and Xainza Tibet. Its occurrence in North Africa moved from Morocco to Algeria. And its distribution in Australia remained the same as ever: Victoria, New South Wales, and New Zealand.

Emsian

Compared with the species diversity and palaeogeographic distribution of *Eospirifer* in Pragian, *Eospirifer* of this interval experienced a moderate development by having a slightly wider distribution and higher species diversity. There are some new documentations from Asia Minor, Brittany, Harz Mountains and Carnic Alps (see Appendix). Besides, all the localities yielding *Eospirifer* species in previous interval have the occurrences continually.

Eifelian

Only four reliable species have been reported from three blocks of this interval: Novaya Zemlya, Turkestan, and Urals.

There are no *Eospirifer* species documented from the younger rocks anywhere in the world.

DISCUSSION: MACROEVOLUTION OF *EOSPIRIFER*

Eospirifer originated during the last climax of the great Ordovician biodiversification event (GOBE), i.e. the Ordovician radiation in a near shore shallow water benthic regime on the Zhe-Gan Platform in late Katian (Rong and Zhan, 1996), and experienced the end Ordovician mass extinction almost immediately after its origination (Rong and Zhan, 2004). It not only survived the crisis but also flourished in the following macroevolutionary intervals in Silurian (Rong et al., 2003). This is probably explained by its occupation of a series of morphological novelties. During its macroevolutionary process from Late Ordovician to early Mid Devonian, *Eospirifer* experienced twice dramatic change in species diversity and palaeogeographic distribution: the Hirnantian/Rhuddanian boundary (i.e. O/S boundary) and the Emsian/Eifelian boundary (i.e. Early/Mid Devonian boundary). The rapid increase of species diversity and expansion of palaeogeographic distribution in the earliest Silurian were accompanied by high origination rate and low extinction rate, and vice versa by the end of Early Devonian. But during the long interval of Silurian and Early Devonian, the species diversity and palaeogeographic distribution seem not to have close relationship with its specific origination and extinction rates (Table 1; Fig. 1). Throughout the macroevolutionary history, *Eospirifer* flourished in Silurian, particularly from Aeronian to Homerian. Although there was a small scale increase of species diversity in the earliest Devonian, the general decreasing trend did not change and it survived for only three more intervals before its extinction in early Mid Devonian. It originated in South China and got widespread on the Upper Yangtze Platform later on, but it was most flourished in those marginal areas of Laurentia where it has the most speices (most diversified), and the widest geographic and the longest stratigraphic distributions. It became extinct in Novaya Zemlya, Turkestan and Urals, regions with comparatively higher palaeolatitudes, which might indicate a new ecological experimentation conducted by *Eospirifer* before its extinction but eventually failed.

Morphologically, *Eospirifer* experienced two different kinds of macroevolution from Late Ordovician through Silurian to early Mid Devonian. On the one hand, it evolves many different species from *E. praecursor* in different palaeoplates or terranes, or at different localities of a single block. All those documented species of *Eospirifer* have different palaeographical or stratigraphical or ecological affiliations, and are also slightly different in their external and internal morphologies. There are some general trends of evolution within *Eospirifer*. 1) Shell size is becoming larger and larger. All Late Ordovician representatives are generally smaller than 5 mm, but most of the Silurian and Devonian species are larger than 10 mm, and some even larger than 20 mm or 30 mm. 2) Shell fine radial ornamentation is becoming stronger and stronger, while those Ordovician species look like smooth-shelled (Fig. 1, figs A-D). 3) The number of whorls of the spiralia is becoming larger with the growing of shell sizes. 4) The angle between the jugal process is becoming smaller and smaller. It is about 115° at *E. praecursor* (late Katian), about 70° at *E. sinensis* (Rhuddanian), about 42° at *E. cf. radiatus* (Telychian) and about 30° at *E. radiatus* (Wenlock) (Rong and Zhan, 1996). On the other hand, as a root genus of the eospiriferines, *Eospirifer* evolves into many different genera making the group become larger and larger, and an important group in those Silurian and Early

Chronostratigraphy			No. of species	No. of sustained species	No. of palaeoplates or terranes	Rate of origination (%)	Rate of extinction (%)
Devonian	Middle	Eifelian	4	2	3	50	100
	Lower	Emsian	15	11	13	26.7	80
		Pragian	14	11	10	21.4	21.4
		Lochkovian	17	5	11	70.6	35.3
Silurian	Pridoli		10	9	9	10	50
	Ludlow	Ludfordian	16	14	13	12.5	43.8
		Gorstian	17	10	13	41.2	17.6
	Wenlock	Homerian	22	20	11	9.1	54.5
		Sheinwoodian	23	13	10	43.5	17.4
	Llandovery	Telychian	24	12	13	50	45.8
		Aeronian	21	12	12	42.9	42.9
		Rhuddanian	14	0	9	100	35.7
		Hirnantian	1	0	1	100	100
Ordovician	Upper	Katian	1	0	1	100	100

Table 1. Species diversity, palaeogeographic distribution, and other statistics of *Eospirifer* species at each interval during its macroevolutionary history from Late Ordovician to early Mid Devonian. The "sustained species" means the species extended from the former interval. The originating species means the species newly occurred in the current interval, and the extinct species refers to the species of this interval that does not extend upward to its younger interval.

Devonian brachiopod faunas. For example, those Early Silurian genera, *Striispirifer*, *Janius*, *Nikiforovaena*, *Cyrtia* and *Yingwuspirifer* are all having *Eospirifer* as their ancestor. Both kinds of macroevolutions are proving that those morphological characters obtained with the origination of *E. praecursor* are evolutionary novelties with great potential of development, taking an important role in the brachiopod faunas from Silurian to the end of Permian.

CONCLUSIONS

Being the oldest known species of the genus *Eospirifer* and the entire group of eospiriferines, *E. praecursor* holds a series of morphological innovations including both external and internal characters. Just because of these evolutionary novelties, it enjoyed a large scope of ecological distribution immediately after its origination, all well-developed from the Zhe-Gan Platform to the upper Zhexi Slope, but most flourished in the near shore, shallow water benthic regime on the Zhe-Gan Platform. It evolves into many different species of *Eospirifer* itself and many species of other eospiriferine genera within South China and many other palaeoplates or terranes in Silurian and Early Devonian. It experienced several morphological changes with the ever growing species diversity and wider palaeogeographic distribution. Its macroevolutionary climax reached in Telychian (late Llandovery) and lasted for two more intervals, i.e. the Sheinwoodian and the Homerian (Wenlock), and then decreased gradually to its extinction in early Mid Devonian punctuated by a small increasing in early Devonian (Lochkovian). During its macroevolutionary history, there was one sharp increase and decrease respectively of its species diversity and palaeogeographic distribution. The former was at the beginning of Silurian (early Rhuddanian) with a very high species origination rate, and the latter was at early Mid Devonian with a very high extinction rate. But, during its long evolutionary history from Early Silurian to Early Devonian, the species diversity and palaeogeographic distribution of *Eospirifer* seem not to have close relationship with the rates of origination and extinction at each interval. Up to now, *Eospirifer* has been documented from almost all major continents (palaeoplates or terranes) during the interval from Late Ordovician to Mid Devonian, except for

Antarctica and South America, but North America (Laurentia) is proven to have the most species, the widest distribution, the longest stratigraphical range, and the most abundant individuals of *Eospirifer*.

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Appendix: List of *Eospirifer* occurrences

Late Katian

Eospirifer praecursor Rong, Zhan and Han, 1994; Changwu Formation and Xiazhen Formation; Zhejiang-Jiangxi border region (JCY area), East China (South China palaeoplate).

Hirnantian

Eospirifer sp.; Chen et al., 2000, Rong et al., 2002; upper part of the Kuanyinchiao Formation (middle part of the *Normalograptus persculptus* Biozone); Honghuayuan, Tongzi, Guizhou, southwest China.

Rhuddanian

Eospirifer cinghizicus Borisiak, 1955; the transitional beds between the Llandovery and Wenlock Series; Chinghiz, Kazakhstan. *Eospirifer fusus* Borisiak, 1955; the *Pentamerus* Beds; central Kazakhstan.

Eospirifer inchoans (Barrande), Kegel, 1953; *U. marginalis* Beds; Gieben, Germany.

Eospirifer kassini Borisiak, 1955; rocks of Llandovery and early Wenlock age; Kazakhstan.

Eospirifer marklini Kiaer, 1908; lower Llandovery; Oslo region, Norway. But most of the specimens Kiaer studied are questionable in their identification or their ages (Boucot, 1962).

Eospirifer quinqueplicatus Poulsen, 1934; the Cape Schuchert Formation; St. George Fiord, Northern Greenland.

Eospirifer sinensis Rong, Xu and Yang, 1974; the base Xiangshuyuan Formation (late Rhuddanian-early Aeronian); Leijiatun, Shiqian County, northeastern Guizhou, southwest China.

Eospirifer tasmaniensis Sheehan and Baillie, 1981; Arndell Sandstone; Locality F5, Range Road Section, Tasmania, Australia.

Eospirifer radiatus (Sowerby, 1835); Twenhofel, 1941; the Aroostook Limestone; Aroostook County, northern Maine, USA.

Eospirifer cf. *radiatus* (J. de C. Sowerby, 1835)

Reed 1906; Namhsim ss.; Myanmar.

Eospirifer sp., Sheehan and Baillie, 1981; Arndell Sandstone, Locality F2, Westfield Quarry, Tasmania, Australia.

Eospirifer sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.

Eospirifer sp., Sheehan, 1975; the Solis Limestone (Quadalupe section); Chihuahua, Mexico.

Eospirifer sp.; Rong et al., 1994; the Wulipo Formation; Wulipo, Yanjiazhai, Meitan, northeastern Guizhou, SW China.

Aeronian

Eospirifer cinghizicus Borisiak, 1955; the transitional beds between the Llandovery and Wenlock Series; Chinghiz, Kazakhstan.

Eospirifer fusus Borisiak, 1955; the *Pentamerus* Beds; central Kazakhstan.

Eospirifer inchoans (Barrande), Kegel, 1953; *U. marginalis* Beds; Gieben, Germany.

Eospirifer kassini Borisiak, 1955; rocks of Llandovery and early Wenlock age; Kazakhstan.

Eospirifer minutes Rong and Yang, 1978; the middle Xiangshuyuan Formation (early Aeronian); Yingwuxi, Sinan County, northeastern Guizhou, southwest China.

Eospirifer? *plicatus* Xian and Jiang, 1978; the middle lower Xiangshuyuan Formation (early Aeronian); Leijiatun, Shiqian County, northeastern Guizhou, southwest China.

Eospirifer quinqueplicatus Poulsen, 1934; the Cape Schuchert Formation; St. George Fiord, Northern Greenland.

Eospirifer radiatus (Sowerby, 1835)

Boucot, 1962; many places in the United States.

Gigout, 1951; Morocco, northern Africa.

Eospirifer sinanensis Jiang in Xian and Jiang, 1978; the upper lower Xiangshuyuan Formation (early Aeronian); Yingwuxi, Sinan County, northeastern Guizhou, southwest China.

Eospirifer sinensis Rong, Xu and Yang, 1974; the base Xiangshuyuan Formation (late Rhuddanian-early Aeronian); Leijiatun, Shiqian County, northeastern Guizhou, southwest China.

Eospirifer sinensis dasifiliformis Fu, 1982; lower Zhaohuajing Formation; Zhaohuajing, Tongxin, Ningxia, North China.

Eospirifer tasmaniensis Sheehan and Baillie, 1981; Arndell Sandstone; Locality F5, Range Road Section, Tasmania, Australia.

Eospirifer transversalis Rong and Yang, 1981; the middle Xiangshuyuan Formation (early Aeronian); Donghuaxi, Sinan County, northeastern Guizhou.

Eospirifer tuvaensis Chernyshev, 1937; the Kyzylchirinsk Bed; Kyzyl-Chiraa, Tuva (Aeronian age according to Rong et al., 1994).

Eospirifer cf. *radiatus* (J. de C. Sowerby, 1835)

Reed 1906; Namhsim ss.; Myanmar.

Kul'kov et al., 1985; Kyzylchirinsk Bed; Tuva, Kazakhstan.

Eospirifer sp., Barnes, Boucot, Cloud and Palmer, 1966; Starcke Limestone; Llano uplift, central Texas (Wenlock age according to Boucot).

Eospirifer sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.

Eospirifer sp., Sheehan and Baillie, 1981; Arndell Sandstone, Locality F2, Westfield Quarry, Tasmania, Australia.

Eospirifer sp., Sheehan, 1975; the Solis Limestone (Quadalupe section); Chihuahua, Mexico.

Eospirifer sp.; Rong and Yang, 1981; the middle Xiangshuyuan Formation; Leijiatun, Shiqian, northeastern Guizhou, SW China.

Eospirifer sp.; Zeng et al., 1993; Zhangwan Formation; Shiyanhe, Xichuan, Henan, central China.

Telychian

Eospirifer consobrinus Poulsen, 1943; the Offley Island Formation; north coast of Offley Island, Cape Godfred Hansen, northern Greenland.

Eospirifer foggi Foerste, 1935; Lobelville Formation; Tennessee, USA.

Eospirifer globosus; Davidson, 1866-1867; British Silurian Brachiopoda.

Hede, 1921; Lower Visby Marl; Gotland, Sweden.

Eospirifer inchoans (Barrande), Kegel, 1953; *U. marginalis* Beds; Gieben, Germany.

Eospirifer kassini Borisiak, 1955; rocks of Llandovery and early Wenlock age; Kazakhstan.

Eospirifer marklini (de Verneuil, 1848); Bassett and Cocks, 1973; the Lower Visby Beds; Gotland, Sweden.

Eospirifer minutes Rong and Yang, 1978; the middle Xiangshuyuan Formation (early Aeronian); Yingwuxi, Sinan County, northeastern Guizhou, southwest China.

Eospirifer profusus Rubel, 1970; the Adavere Stage; Estonia (Bassett and Cocks 1973 treated it as a junior synonym of *Eospirifer marklini*).

Eospirifer quinqueplicatus Poulsen, 1934; the Cape Schuchert Formation; St. George Fiord, Northern Greenland.

Eospirifer radiatus (J. de C. Sowerby, 1835)

Boucot, 1962; many places in the USA.

Shrock and Twenhofel, 1938; Pike Arm Formation; northern Newfoundland, Canada.

Poulsen, 1934; northern Greenland.

Beecher and Dodge, 1892; Ames Knob Formation; Coastal Maine, USA.

Gillette, 1947; Williamson Shale; New York, USA.

Lesley, 1890; Clinton Shale; Pennsylvania, USA.

Tillman, 1961; Osgood Formation; Indiana, USA.

Prouty and Swartz, 1923; Rose Hill Formation; Maryland, USA.

Foerste, 1909; West Union Bed; Kentucky, USA.

Twenhofel, 1928; Jupiter Formation; Anticosti Island, Canada.

Kindle in Cairnes, 1914; Unnamed Beds; Yukon, Canada.

Nikiforova, 1954; Kitaygorod Formation; Podolia, Ukraine.

Borisiak, 1955; Kazakhstan.

Gigout, 1951; Morocco, northern Africa.

Eospirifer songkanensis Wu; Rong and Yang, 1978; upper Shiniulan Formation and Leijiatun Formation; northeastern Guizhou, South China.

Eospirifer stonehousensis McLearn, 1924; Maehl, 1961; French River Formation; Nova Scotia, Canada.

Eospirifer subradiatus Wang, 1956; from the fine grained yellowish green sandstone near the first fault, east of the Changning county town, Sichuan Province, southwest China. According to Wang's description, the fossils of this species should be collected from the Xiushan Formation which is of Telychian age.

Eospirifer tasmaniensis Sheehan and Baillie, 1981; Arndell Sandstone; Locality F5, Range Road Section, Tasmania, Australia.

Eospirifer tuvaensis Chernyshev, 1937; the Kyzylchirinskie Beds; Kyzyl-Chiraa, Tuva (Aeronian age according to Rong et al., 1994).

Eospirifer cf. *radiatus*, Reed 1906; Namhsim ss.; Myanmar.

Boucot et al., 1958; quartzite of Bernardston Formation; Massachusetts, USA.

Eospirifer sp., Amsden 1957; Clarita Member of the Chimneyhill Formation; Oklahoma, USA.

Eospirifer sp., Amsden 1957; St. Clair Formation; Arkansas, USA.

Eospirifer sp., Boucot, 1962; vicinity of Ciudad Victoria, State of Tamaulipas, Mexico.

Eospirifer sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.

Eospirifer sp., Boucot et al., 1966; Long Reach Formation; south New Brunswick, Canada.

Eospirifer sp., Barnes, Boucot, Cloud and Palmer, 1966; Starcke Limestone; Llano uplift, central Texas.

Eospirifer sp., Sheehan and Baillie, 1981; Arndell Sandstone, Locality F2, Westfield Quarry, Tasmania, Australia.

Eospirifer sp., Sheehan, 1975; the Solis Limestone (Quadalupe section); Chihuahua, Mexico.

Sheinwoodian (lower Wenlock)

- Eospirifer (Acutilineolus) acutolineatus* Amsden, 1968; Fitzhugh Member, Clarita Formation; Arkansas, USA
Eospirifer acutolineatus acutolineatus Amsden, 1968; St. Clair Formation; Batesville District, Arkansas, USA.
Eospirifer dilectus Rong and Yang, 1978; lower Xiushan Formation; northeastern Guizhou.
Eospirifer foggi Foerste, 1935; Lobelville Formation; Tennessee, USA.
Eospirifer globosus; Hede, 1921; Lower Visby Marl; Gotland, Sweden.
Eospirifer (Acutilineolus) inferatus Amsden, 1978; Marble City Member, Quarry Mountain Formation; eastern Oklahoma, USA.
Eospirifer kassini Borisiak, 1955; rocks of Llandovery and early Wenlock age; Kazakhstan.
Eospirifer pentagonus Amsden, 1968; Fitzhugh Member, Clarita Formation; Oklahoma, USA.
Eospirifer pollens (Barrande, 1848); Havlíček, 1980; upper Motol Formation; area between Hills Kolo and Branžovy, Kozolupy, Bohemia, Czech Republic.
Eospirifer praeseicans Havlíček, 1959; Havlíček, 1980; Motol Formation; area between Lužce and Lodnice, Bohemia, Czech Republic.
Eospirifer radiatus (J. de C. Sowerby, 1835)
 Twenhofel, 1941; the Aroostook Limestone; Aroostook County, northern Maine, USA.
 Beecher and Dodge, 1892; Ames Knob Formation; Coastal Maine, USA.
 Gillette, 1947; Rochester Shale, Herkimer ss.; New York, USA.
 Lesley, 1890; Clinton Shale; Pennsylvania, USA.
 Nettleroth, 1889; Louisville ls.; Indiana, USA.
 Prouty and Swartz, 1923; Rochester Formation; Maryland, USA.
 Hall and Clarke, 1893; Racine Dolomite; Wisconsin, USA.
 Foerste, 1935; Massie Clay; Ohio, USA.
 Northrop, 1939; La Vieille Formation; Gaspé, Canada.
 Twenhofel, 1928; Chicotte Formation; Anticosti Island, Canada.
 Shrock and Twenhofel, 1938; Pike Arm Formation; northern Newfoundland, Canada.
 Bolton, 1957; 'Irondequoit' Formation, Rochester Formation, and Ancaster chert of Goat Island Member of Lockport Formation; southwestern Ontario, Canada.
 Kindle in Cairnes, 1914; Unnamed Beds; Yukon, Canada.
 Hede, 1927; Hogklint ls., Slite Group, Halla ls., Mulde Marl; Gotland, Sweden.
 Havlíček, 1980; tuffaceous limestone, Motol Formation; Tetín and "V Kozle" near Beroun, Bohemia, Czech Republic.
 Zeng, 1977; Shamao Group; western Hubei, South China.
 Rong and Chen, 2003; Shamao Formation; southern Hubei, central South China.
 Rong and Chen, 2003; Shaerbuer Formation; Xinjiang, NW China.
 Nikiforova, 1954; Borshchov Formation; Podolia, Ukraine.
 Gigout, 1951; Morocco, northern Africa.
Eospirifer radiatus globosus (Salter, 1848); Bassett and Cocks, 1974; Slite Beds, Wenlock Limestone; Dudley, Gotland, Sweden.
Eospirifer subradiatus (Wang); Zeng, 1977; Shamao Group; Xianfeng County, western Hubei, South China.
Eospirifer togatus Barrande, 1848; Vascutanu, 1931; Schistes Marneaux a Strophomenides; Podolia, Ukraine.
Eospirifer tuvaensis Chernyshev, 1937; the Kyzylchirinsk Bed; Kyzyl-Chiraa, Tuva (Aeronian age according to Rong et al., 1994).
Eospirifer xianfengensis Zeng, 1977; Shamao Group; Datianba, Xianfeng County, western Hubei, South China.
Eospirifer cf. *radiatus*, Reed 1906; Namhsim ss.; Myanmar.
 Boucot et al., 1958; quartzite of Bernardston Formation; Massachusetts, USA.
 Boucot and Thompson, 1963; Clough Formation; westcentral New Hampshire, USA.
 Twenhofel, 1941; Ashland Limestone; Aroostook County, northern Maine, USA.
Eospirifer sp., Amsden 1957; Clarita Member of the Chimneyhill Formation; Oklahoma, USA.
Eospirifer sp., Amsden 1957; St. Clair Formation; Arkansas, USA.
Eospirifer sp., Bolton, 1957; De Cew Formation; southwestern Ontario, Canada.
Eospirifer sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.
Eospirifer sp., Boucot, 1962; vicinity of Ciudad Victoria, State of Tamaulipas, Mexico.
Eospirifer sp., Boucot et al., 1966; Long Reach Formation; coastal New Brunswick, Canada.
Eospirifer sp., Sheehan, 1975; the Solis Limestone (Quadalupe section); Chihuahua, Mexico.

Homerian (upper Wenlock)

- Eospirifer acutolineatus acutolineatus* Amsden, 1968; Fitzhugh Member, Clarita Formation; Oklahoma, USA
Eospirifer acutolineatus pentagonus Amsden, 1968; St. Clair Limestone Formation; Batesville District, Arkansas, USA.

- Eospirifer devonians* Havlíček, 1959; Kopanina Beds; Czech Republic.
- Eospirifer foggi* Foerste, 1935; Lobelville Formation; Tennessee, USA.
- Eospirifer globosus* (Salter in Phillips and Salter, 1848)
- Phillips and Salter, 1848; Much Wenlock Limestone Formation; Dudley, West Midlands, England.
- Hede, 1921; Lower Visby Marl; Gotland, Sweden.
- Eospirifer (Acutilineolus) inferatus* Amsden, 1978; Marble City Member, Quarry Mountain Formation; eastern Oklahoma, USA.
- Eospirifer pollens* (Barrande, 1848); Havlíček, 1980; upper Motol Formation; area between Hills Kolo and Bránžovy, Kozolupy, Bohemia, Czech Republic.
- Eospirifer praesecans* Havlíček, 1959; Havlíček, 1980; Motol Formation; area between Lužce and Lodnice, Bohemia, Czech Republic.
- Eospirifer radiatus* (J. de C. Sowerby, 1835)
- Twenhofel, 1941; the Aroostook Limestone; Aroostook County, northern Maine, USA.
- Beecher and Dodge, 1892; Ames Knob Formation; Coastal Maine, USA.
- Gillette, 1947; Rochester Shale, Herkimer ss.; New York, USA.
- Lesley, 1890; Clinton Shale; Pennsylvania, USA.
- Nettleroth, 1889; Louisville ls.; Indiana, USA.
- Prouty and Swartz, 1923; Rochester Formation; Maryland, USA.
- Foerste, 1935; Massie Clay; Ohio, USA.
- Hall and Clarke, 1893; Racine Dolomite; Wisconsin, USA.
- Amsden, 1978; Marble City Member, Quarry Mountain Formation; eastern Oklahoma, USA. (The author identified his specimens as this species with a question mark.)
- Northrop, 1939; La Vieille Formation; Gaspé, Canada.
- Shrock and Twenhofel, 1938; Pike Arm Formation; northern Newfoundland, Canada.
- Twenhofel, 1928; Chicotte Formation; Anticosti Island, Canada.
- Bolton, 1957; 'Irondequoit' Formation, Rochester Formation, and Ancaster chert of Goat Island Member of Lockport Formation; southwestern Ontario, Canada.
- Kindle in Cairnes, 1914; Unnamed Beds; Yukon, Canada.
- Boucot, 1962; Cocks, 2008; upper Llandovery to lower Ludlow; many places in UK.
- Shergold and Bassett, 1970; Wenlock Limestone; Wenlock Edge, Shropshire, England.
- Havlíček, 1980; tuffaceous limestone, Motol Formation; Tetín and "V Kozle" near Beroun, Bohemia, Czech Republic.
- Bassett and Cocks, 1974; Much Wenlock Limestone; Dudley, West Midlands, Gotland, Sweden.
- Hede, 1927; Hogklint ls., Slite Group, Halla ls., Mulde Marl; Gotland, Sweden.
- Rong and Chen, 2003; Shaerbuer Formation; Xinjiang, NW China.
- Rong and Chen, 2003; Shamao Formation; southern Hubei, central South China.
- Nikiforova, 1954; Borshchov Formation; Podolia, Ukraine.
- Gigout, 1951; Morocco, northern Africa.
- Eospirifer radiatus globosus* (Salter, 1848); Bassett and Cocks, 1974; Slite Beds, Wenlock Limestone; Dudley, Gotland, Sweden.
- Eospirifer togatus* Barrande, 1848; Vascautanu, 1931; Schistes Marneux a Strophomenides; Podolia, Ukraine.
- Eospirifer tuvaensis* Chernyshev, 1937; the Kyzylchirinskie Beds; Kyzyl-Chiraa, Tuva (Aeronian age according to Rong et al., 1994).
- Eospirifer xianfengensis* Zeng, 1977; Shamao Group; Datianba, Xianfeng County, western Hubei, South China.
- Eospirifer* cf. *eudora* Northrop, 1939; Gascons Formation; Gaspé, Canada.
- Eospirifer* cf. *radiatus*, Reed 1906; Namhsim ss.; Myanmar.
- Boucot et al., 1958; quartzite of Bernardston Formation; Massachusetts, USA.
- Boucot and Thompson, 1963; Clough Formation; westcentral New Hampshire, USA.
- Twenhofel, 1941; Ashland Limestone; Aroostook County, northern Maine, USA.
- Eospirifer* sp., Amsden 1957; Clarita Member of the Chimneyhill Formation; Oklahoma, USA.
- Eospirifer* sp., Amsden 1957; St. Clair Formation; Arkansas, USA.
- Eospirifer* sp., Bolton, 1957; De Cew Formation; southwestern Ontario, Canada.
- Eospirifer* sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.
- Eospirifer* sp., Boucot, 1962; vicinity of Ciudad Victoria, State of Tamaulipas, Mexico.
- Eospirifer* sp., Boucot et al., 1966; Long Reach Formation; coastal New Brunswick, Canada.
- Eospirifer* sp., Sheehan, 1975; the Solis Limestone (Quadalupe section); Chihuahua, Mexico.

Gorstian (lower Ludlow)

Eospirifer contortus Havlíček, 1959; Ludlow; Czech Republic.

Eospirifer devonicans Havlíček, 1959; Kopanina Beds; Barrande area, Bohemia, Czech Republic.
Eospirifer eastoni Gill, 1949; Garratt, 1983; Dargile Formation; Melbourne Trough, Victoria, Australia.
Eospirifer ferganensis Nikiforova, 1937; Marginalis and Isfara Beds; Turkestan. Boucot (1962) used this species as the type species to establish a new genus *Nikiforovaena*.
Eospirifer praeseecans Havlíček, 1959; Havlíček, 1980; Motol Formation; area between Lužce and Lodnice, Bohemia, Czech Republic.
Eospirifer radiatus (Sowerby, 1835);
 Northrop, 1939; La Vieille Formation; Gaspé, Canada.
 Bolton, 1957; 'Irondequoit' Formation, Rochester Formation, and Ancaster chert of Goat Island Member of Lockport Formation; southwestern Ontario, Canada.
 Kindle in Cairnes, 1914; Unnamed Beds; Yukon, Canada.
 Boucot, 1962; Watkins, 1981; Cocks, 2008; upper Llandovery to lower Ludlow; many places in UK.
 Shergold and Bassett, 1970; Lower and basal Middle Elton Beds; Wenlock Edge, Shropshire, England.
 Havlíček, 1980; tuffaceous limestone, Motol Formation; Tetín and "V Kozle" near Beroun, Bohemia, Czech Republic.
 Bassett and Cocks, 1974; the Upper Visby Beds to the Klinteberg Beds, Much Wenlock Limestone; Dudley, West Midlands, Gotland, Sweden.
 Nikiforova, 1954; Borshchov Formation; Podolia, Ukraine.
 Gigout, 1951; Morocco, northern Africa.
 Rong and Chen, 2003; Shamao Formation; southern Hubei, central South China.
Eospirifer tingi Grabau, 1926; Tsin, 1956; the Gaozhaitian Formation; Wudang, Guiyang, Guizhou, South China.
Eospirifer togatus Barrande, 1848; Nikiforova, 1937; Marginalis Beds; Turkestan.
Eospirifer tuvaensis Chernyshev, 1937; the Kyzylchirinskies Beds; Kyzyl-Chiraa, Tuva (Aeronian age according to Rong et al., 1994).
Eospirifer uniplicatus Tsin, 1956; Gaozhaitian Formation; Wudang, Guiyang, Guizhou, South China.
Eospirifer cf. *radiatus*, Reed 1906; Namhsim ss.; Myanmar.
 Khodalevitch, 1939; *Striatius* Beds; Urals, Russia.
 Boucot et al., 1958; quartzite of Bernardston Formation; Massachusetts, USA.
 Boucot and Thompson, 1963; Clough Formation; westcentral New Hampshire, USA.
Eospirifer cf. *tenuis* (Barrande); Havlíček, 1959; Kopanina Beds; Barrande area, Bohemia, Czech Republic.
Eospirifer sp., Amsden 1957; St. Clair Formation; Arkansas, USA.
Eospirifer sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.
Eospirifer sp., Boucot, 1962; vicinity of Ciudad Victoria, State of Tamaulipas, Mexico.
Eospirifer sp., Sheehan, 1975; the Solis Limestone (Quadalupe section); Chihuahua, Mexico.
Eospirifer sp.; Lenz, 1977; the Road River Formation; Canadian Cordillera, Canada.

Ludfordian (upper Ludlow)

Eospirifer contortus Havlíček, 1959; Ludlow; Czech Republic.
Eospirifer devonicans Havlíček, 1959; Kopanina Beds; Barrande area, Bohemia, Czech Republic.
Eospirifer eastoni Gill, 1949; Garratt, 1983; Dargile Formation; Melbourne Trough, Victoria, Australia.
Eospirifer ferganensis Nikiforova, 1937; Marginalis and Isfara Beds; Turkestan. Boucot (1962) used this species as the type species to establish a new genus *Nikiforovaena*.
Eospirifer plicatellus var. *interlineatus* Lindström; Walmsley, 1958; the Lower Llangibby Beds; Usk inlier, Monmouthshire, England.
Eospirifer radiatus (Sowerby, 1835);
 Northrop, 1939; La Vieille Formation; Gaspé, Canada.
 Bolton, 1957; 'Irondequoit' Formation, Rochester Formation, and Ancaster chert of Goat Island Member of Lockport Formation; southwestern Ontario, Canada.
 Kindle in Cairnes, 1914; Unnamed Beds; Yukon, Canada.
 Boucot, 1962; Cocks, 2008; upper Bringewood Beds; many places in UK.
 Nikiforova, 1954; Borshchov Formation; Podolia, Ukraine.
 Gigout, 1951; Morocco, northern Africa.
Eospirifer tingi Grabau, 1926; Grabau, 1931; the Miaokao Formation; Qujing, Yunnan, Southwest China.
Eospirifer togatus Barrande, 1848; Nikiforova, 1937; Marginalis Beds; Turkestan.
Eospirifer tuvaensis Chernyshev, 1937; the Kyzylchirinskies Beds; Kyzyl-Chiraa, Tuva (Aeronian age according to Rong et al., 1994).
Eospirifer uniplicatus Tsin, 1956; Gaozhaitian Formation; Wudang, Guiyang, Guizhou, South China.

Eospirifer variplicatus Ohno, 1977; the Fukuji Formation; Hida Massif, Central Japan.
Eospirifer cf. *radiatus*, Reed 1906; Namhsim ss.; Myanmar.
 Khodalevitch, 1939; *Striatus* Beds; Urals, Russia.
 Boucot et al., 1958; quartzite of Bernardston Formation; Massachusetts, USA.
 Boucot and Thompson, 1963; Clough Formation; westcentral New Hampshire, USA.
Eospirifer sp., Amsden 1957; St. Clair Formation; Arkansas, USA.
Eospirifer sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.
Eospirifer sp., Boucot, 1962; vicinity of Ciudad Victoria, State of Tamaulipas, Mexico.
Eospirifer sp.; Lenz, 1977; the Road River Formation; Canadian Cordillera, Canada.

Pridoli

Eospirifer eastoni Gill, 1949; Garratt, 1983; Dargile Formation; Melbourne Trough, Victoria, Australia.
Eospirifer radiatus (Sowerby, 1835);
 Nikiforova, 1954; Borshchov Formation; Podolia, Ukraine.
 Gigout, 1951; Morocco, northern Africa.
Eospirifer tingi Grabau, 1926; Grabau, 1931; the Miaokao Formation; Qujing, Yunnan, Southwest China.
Eospirifer tuvaensis Chernyshev, 1937; the Kyzylchirinskie Beds; Kyzyl-Chiraa, Tuva (Aeronian age according to Rong et al., 1994).
Eospirifer uniplicatus Tsin, 1956; Gaozhaitian Formation; Wudang, Guiyang, Guizhou, South China.
Eospirifer variplicatus Ohno, 1977; the Fukuji Formation; Hida Massif, Central Japan.
Eospirifer cf. *radiatus*, Reed 1906; Namhsim ss.; Myanmar.
 Boucot and Thompson, 1963; Clough Formation; westcentral New Hampshire, USA.
Eospirifer sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.
Eospirifer sp., Sheehan, 1975; the Solis Limestone (Quadalupe section); Chihuahua, Mexico.
Eospirifer sp.; Lenz, 1977; the Road River Formation; Canadian Cordillera, Canada.

Lochkovian (Early Devonian)

Eospirifer admirabilis Nikiforova, 1937; Ged. (Lochkovian); Turkestan. Boucot (1962) named two new genera *Macroleura* and *Nikiforovaena*, and revised the specimens of this species into these two genera respectively.
Eospirifer balchaaschensis Nikiforova, 1937; the Marginalis Beds (early Lochkovian); Turkestan and western Balkhash Land (Boucot 1962 reviewed this species as *Macroleura balchaaschensis* (Nikiforova, 1937)).
Eospirifer eastoni Gill, 1949; Garratt, 1983; Dargile Formation; Melbourne Trough, Victoria, Australia.
Eospirifer ignobilis Khodalevitch, 1939; Upper Marginalis Beds; Urals, Russia.
Eospirifer inchoans (Barrande), Kegel, 1953; *U. marginalis* Beds, Eichelstueckschacht Formation; Giessen, north of Frankfurt am Main, Hessen, Germany. According to Bahlburg's (1985) revision, this horizon is of late Lochkovian age.
Eospirifer parahentius Gill, 1950; Savage, 1974; Maradana Shale; Cowra Trough, New South Wales, Australia.
Eospirifer radiatus (Sowerby, 1835)
 Nikiforova, 1954; Borshchov Formation; Podolia, Ukraine.
Eospirifer secans (Barrande, 1848); Shirley, 1938; Baton River Beds; Baton River, New Zealand.
Eospirifer secans var. *rarus* Khodalevitch, 1951; Lower Devonian; Urals, Russia. Boucot (1962) named a new genus *Havlicekia*, and put this species into this genus.
Eospirifer subviator Khodalevitch, 1951; Lower Devonian to Eifelian; Urals, Russia. Boucot (1962) revised this species to *Strispirifer*.
Eospirifer tenuis (Barrande, 1879); Walmsley et al., 1974; Lochkov Formation; Reporyje and Lode ice, Bohemia, Czech Republic.
Eospirifer togatus Barrande, 1848; Vascautanu, 1931; Schiste d'Onut; Podolia, Ukraine.
Eospirifer variplicatus Ohno, 1977; the Fukuji Formation; Hida Massif, Central.
Eospirifer cf. *togatus* (Barrande, 1879); Termier, 1936; Morocco, North Africa.
 Xu, 1987; Da'erdong Formation; Xainza, northern Tibet, China.
Eospirifer sp.; Kirk and Amsden, 1951; unnamed limestone; northeastern end of Heceta Island, southeastern Alaska, USA.
Eospirifer sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.
Eospirifer sp., Sheehan, 1975; the Solis Limestone (Quadalupe section); Chihuahua, Mexico.

Pragian

Eospirifer bascuscanicus Rzonitskaya, 1952; Pesterevo Beds (late Pragian); Kuznetsk Basin, Kazakhstan. Boucot (1962) named a new genus *Macroleura*, and put this species into this genus.

Eospirifer eastoni Gill, 1949; Garratt, 1983; Dargile Formation; Melbourne Trough, Victoria, Australia.
Eospirifer parahentius Gill, 1950; Savage, 1974; Maradana Shale; Cowra Trough, New South Wales, Australia.
Eospirifer pseudotogatus Khalfin, 1948; *Pseudotogatus* Horizon; Altai Mountains, Kazakhstan.
Eospirifer secans (Barrande, 1848); Shirley, 1938; Baton River Beds; Baton River, New Zealand.
Eospirifer secans var. *rarus* Khodalevitch, 1951; Lower Devonian; Urals, Russia. Boucot (1962) named a new genus *Havlicekia*, and put this species into this genus.
Eospirifer subviator Khodalevitch, 1951; Lower Devonian to Eifelian; Urals, Russia. Boucot (1962) revised this species to *Striispirifer*.
Eospirifer togatus; Barrande, 1879;
Konieprus (f2) and Mnienian (f2); Czech Republic (Boucot, 1962 thought it is of early Emsian age).
Shirley, 1938; Baton River Beds; Baton River, New Zealand.
LeMaitre, 1952; Gisement du Kilometre 30, Algeria, North Africa.
Eospirifer variplicatus Ohno, 1977; the Fukuji Formation; Hida Massif, Central Japan.
Eospirifer cf. *togatus* (Barrande, 1879); Termier, 1936; Morocco, North Africa.
Xu, 1987; Da'erdong Formation; Xainza, northern Tibet, China.
Eospirifer sp.; Gill, 1942; the Yeringian Series; Victoria, Australia. The author thought the specimens of this species could be conspecific as *E. togatus* named in New Zealand.
Eospirifer sp.; Kirk and Amsden, 1951; unnamed limestone; northeastern end of Heceta Island, southeastern Alaska, USA.
Eospirifer sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.
Eospirifer sp., Sheehan, 1975; the Solis Limestone (Quadalupe section); Chihuahua, Mexico.

Emsian

Eospirifer davousti; Barrois, 1888; Calcaire d'Erbray; Brittany.
Eospirifer eastoni Gill, 1949; Sandy's Creek Beds; Sandy's Creek, Gippsland, Victoria, Australia. (Gill 1949 discussed the age of *E. eastoni*, and thought it as late Early Devonian in age.)
Garratt, 1983; Dargile Formation; Melbourne Trough, Victoria, Australia.
Eospirifer parahentius Gill, 1950; Savage, 1974; Maradana Shale; Cowra Trough, New South Wales, Australia.
Eospirifer secans (Barrande, 1848);
Shirley, 1938; Baton River Beds; Baton River, New Zealand.
Rzonsnitskaya, 1952; Baskukan Beds; Kuznetsk Basin, Kazakhstan.
Eospirifer secans var. *rarus* Khodalevitch, 1951; Lower Devonian; Urals, Russia. Boucot (1962) named a new genus *Havlicekia*, and put this species into this genus.
Eospirifer subviator Khodalevitch, 1951; Lower Devonian to Eifelian; Urals, Russia. Boucot (1962) revised this species to *Striispirifer*.
Eospirifer togatoides; Paeckelmann, 1925; Pendik Schichten; Bosphorus Region, Asia Minor.
Eospirifer togatus; Barrande, 1879;
Kayser, 1878; Kalk des Joachimskopfes; Harz Mountains.
Gortani, 1915; Capolago; Carnic Alps.
Shirley, 1938; Baton River Beds; Baton River, New Zealand.
Eospirifer togatus insidiosus; Havlíček, 1959; Koneprusy ls.; Bohemia, Czech Republic.
Eospirifer togatus var. *subsiniuata*; Kayser, 1878; Kalk des Schneckenberges und Badeholzes bei Magdesprung; Harz Mountains.
Eospirifer variplicatus Ohno, 1977; the Fukuji Formation; Hida Massif, Central Japan.
Eospirifer cf. *togatus* (Barrande, 1879); Termier, 1936; Morocco, North Africa.
Xu, 1987; Da'erdong Formation; Xainza, northern Tibet.
Eospirifer sp.; Kirk and Amsden, 1951; unnamed limestone; northeastern end of Heceta Island, southeastern Alaska, USA.
Eospirifer sp., Boucot, 1962; locality 14 of Arpishmebulag Series; T'ien-Shan.
Eospirifer sp., Sheehan, 1975; the Solis Limestone (Quadalupe section); Chihuahua, Mexico.

Eifelian

Eospirifer davousti; Nalivkin, 1930; highest Eifelian; Turkestan.
Eospirifer pseudoindifferens; Nalivkin, 1930; highest Eifelian; Turkestan.
Eospirifer subviator Khodalevitch, 1951; Lower Devonian to Eifelian; Urals, Russia. Boucot (1962) revised this species to *Striispirifer*.
Eospirifer vetuloides Nalivkin, 1960; upper Eifelian; Novaya Zemlya, Russia. Boucot (1962) revised this species into the genus *Janius* Havlíček, 1957.

LATE DARRIWILIAN TO EARLY SANDBIAN GRAPTOLITE BIOSTRATIGRAPHY IN WESTERN ZHEJIANG AND EASTERN JIANGXI PROVINCES, SE CHINA

Y.D. Zhang, Y.Y. Song and J. Zhang

State Laboratory of Palaeobiology and Stratigraphy, Nanjing Institute of Geology and Palaeontology, Nanjing 210008, China.
ydzhang@nigpas.ac.cn, duyuan1986@163.com, zhangju.0331@163.com

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INTRODUCTION

The Ordovician System in western Zhejiang and eastern Jiangxi provinces, SE China has been well-known for its well-preserved, three-dimensional pyritic graptolites, and the continuous graptolite successions of Tremadocian to Sandbian (Zhang et al., 2007a). However, the graptolites of the late Darriwilian to early Sandbian interval has been relatively poorly known compared to some other intervals. In the Baijiawu section, Yushan, Jiangxi Province (Fig. 1), the concerned interval has been recognized as including in ascending order the *Pterograptus elegans*, the *Didymograptus jiangxiensis*, the *Glossograptus hincksii*, the *Nemagraptus gracilis* and the *Dicranograptus sinensis* zones (Xiao et al., 1991). Unfortunately, most of these graptolite zones are poorly defined, partially due to the poor exposure of the interval. For example, the *D. jiangxiensis* Zone was defined with its base at the first appearance of the eponymous species, but the boundary lies above the last appearance of *Pterograptus elegans*, which is the index species of the underlying biozone, leaving an interval in between without the occurrences of either index species. In Wuning area, northern Jiangxi Province, the *Didymograptus jiangxiensis* Zone was defined at the last appearance of the *Pterograptus*, rather than the FAD of the eponymous species (Ni, 1991). For the topmost part of the Darriwilian in Yushan, the *Glossograptus hincksii* Zone was named with its base at the FAD of the eponymous species or *Hustedograptus teretiusculus*. But in Wuning area, *G. hincksii* was found mostly from the underlying *D. jiangxiensis* Zone, and instead the *Glyptograptus teretiusculus siccatus* Zone was suggested, whose base was poorly defined at the last appearance of *Didymograptus jiangxiensis*.

In western Zhejiang Province, the graptolite zones of late Darriwilian are also poorly defined, partially due to the poor exposure of the interval in several classic sections. In the Huangnitang section (GSSP for the base of the Darriwilian Stage), the base of the Sandbian was successfully and precisely identified and a diverse graptolite fauna of the *Nemagraptus gracilis* Zone was recorded (Chen et al., 2006). However, the *Pterograptus elegans* Zone was not well represented with the absence of the eponymous species and other diagnostic species due to the disturbance of Mesozoic volcanic dikes. Overlying this interval the "*Hustedograptus teretiusculus* Zone" was adopted provisionally due to the poor recovery of

stratigraphically diagnostic graptolite species, and that the FAD of the *H. teretiusculus* was coincident with the base of the supposed *P. elegans* Zone (Chen et al., 2006). In the Hengtang section, the late Darriwilian interval is completely covered by local farmers' house and thick Quarternary deposits, while in the Fengzu section of Jiangshan, the interval is truncated by a fault (Zhang et al., 2007a).

Herein we report two new sections excavated and discovered in recent years, the Hengdu Quarry section and the Liujia section in western Zhejiang Province (Fig. 1), with well developed and exposed late Darriwilian to Sandbian outcrops. Both sections include lithologically in ascending order the Yinchufu (top part), the Ningkuo, the Hulo, and the Yenwashan formations. The late Darriwilian to Sandbian interval corresponds to the upper part of the Hulo Formation which underlies conformably the Yenwashan Formation. A preliminary biostratigraphic study of the two sections is briefly reported herein.

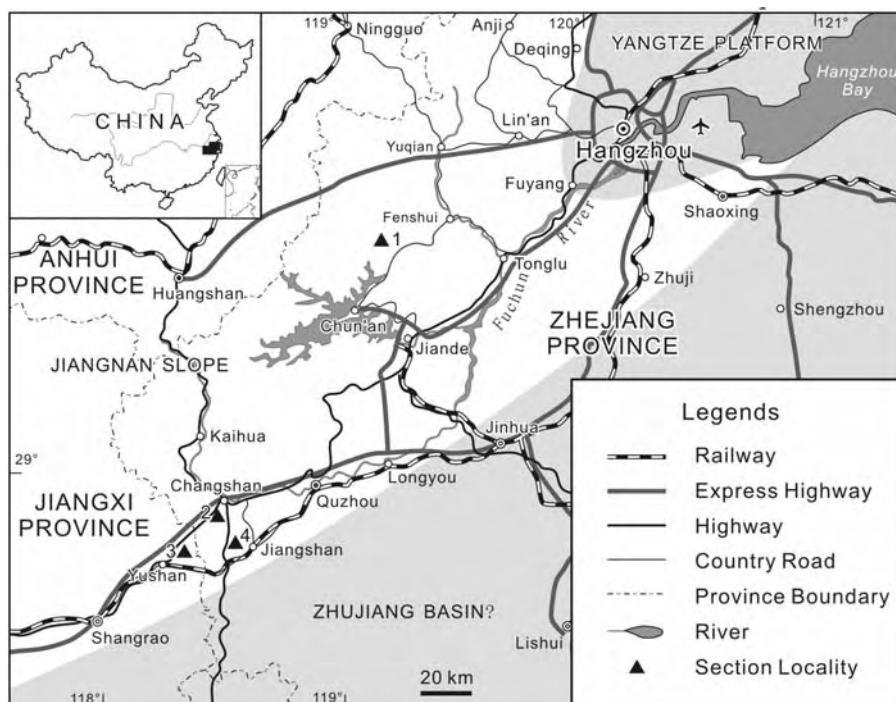


Figure 1. Location of the studied and discussed sections in western Zhejiang and eastern Jiangxi provinces, SE China. 1, Liujia section, Tonglu; 2, Huangnitang section, Changshan; 3, Baijiawu section, Yushan, Jiangxi Province; 4, Hengdu Quarry section, Jiangshan.

GEOLOGICAL SETTING OF THE SECTIONS

The western Zhejiang is widely known as a part of the Jiangnan Slope, where the Ordovician rocks are dominated by black shale, mudstone, chert and some intercalated nodular or bedded limestones, and are abundant in graptolites, acritarchs, chitinozoans and some other planktons, but short of benthos, and is regarded as belonging to slope facies. To the northwest of the Jiangnan Slope locates the Yangtze Platform which is typified by the prevailing development of carbonates with some mudstones, whereas in the

southeast the facies is truncated by the magnificent Jiangshan-Shaoxing Fault. Beyond the fault is a vast poorly known region, where may be presumably a part of the Zhujiang Basin.

(1) Hengdu Quarry section. The section is located 1 km northeast of the Hengdu Town, Jiangshan. It has been excavated by local farmers since 2007, to take the limestones and the black shale for cement production. The quarry exposes the upper part of the Hulo Formation, which is characterized by black shale, chert intercalated with some layers of siliceous limestone, and the basal Yenwashan Formation typified by grey nodular limestone (Fig. 2). During 2008-2010 when the Quarry was being mined, we visited the quarry for three times, and collected systematically the fossils from the black shale and chert, and organic carbon isotope samples from the siliceous limestone layers. The fossils are dominated by abundant graptolites, including commonly *Nemagraptus gracilis*, *Dicellograptus sextans*, *Dicell. vagus*, *Pseudoclimacograptus scharenbergi*, *Hustedograptus teretiusculus* and *Pterograptus elegans* etc., together with some inarticulate brachiopods (common within the topmost 2 meters of the Hulo Formation), No trilobites have been found. Most of the graptolites are pyritic in three-dimensions, while some are flattened.

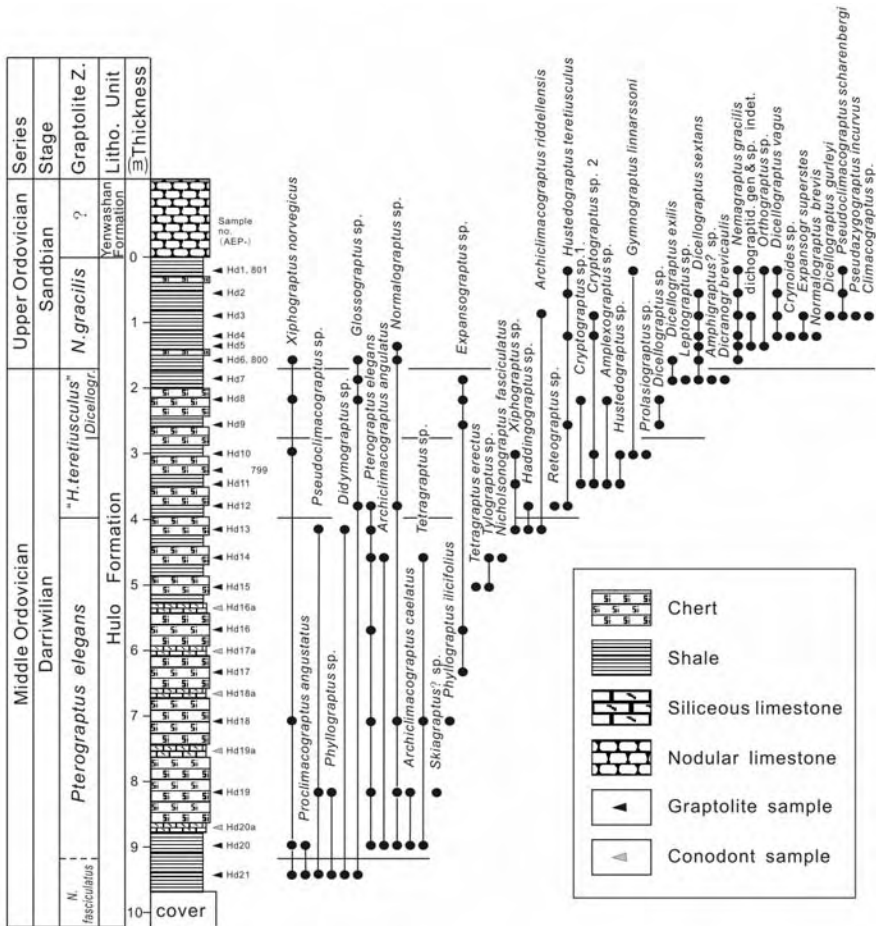


Figure 2. Graptolite range chart of the uppermost Hulo Formation in the Hengdu section, Jiangshan, western Zhejiang Province.

(2) Liujia section. The section is located near the Liujia Village, ca. 40 km to the northwest of the Tonglu county town, western Zhejiang. Along this road-cut section, a nearly complete Ordovician sequence is continuously well exposed, including the Yinchufu (top part), Ningkuo, Hulo, Yenwashan, Huangnekang and Changwu (basal part) formations. The top part of the Hulo Formation yields abundant graptolites including *Nemagraptus gracilis*, *Dicellograptus vagus*, *Haddingograptus scharenbergi*, *Hustedograptus teretiusculus*, and *Archiclimacograptus riddellensis*, among some others. The FAD of *N. gracilis* lies at the horizon half meter below the top of the formation, being well comparable to that in the Huangnitang section, where the FAD of the species is one meter below the top of the Hulo Formation. Significantly, the first appearance of *Dicellograptus vagus* occurs at 8 meters below the top, in association with species of *Dicranograptus*, *Orthograptus* and *Haddingograptus*.

GRAPTOLITE BIOSTRATIGRAPHY AND INTERNATIONAL CORRELATION

Based on the graptolite successions from the upper part of the Hulo Formation in the Hengdu and Liujia sections, four graptolites zones are recognized, which are discussed as below (Fig. 2).

Nicholsonograptus fasciculatus Zone

The biozone is weakly identified herein for a short interval below the FAD of *Pterograptus elegans*. Although the eponymous species has not yet been confirmed, it is clear the interval (HD21) is readily absent of *P. elegans*, and is most likely to correspond to the top part of the *Nicholsonograptus fasciculatus* Zone, compared to the nearby Huangnitang and the Baijiawu sections (see Xiao et al., 1991; Chen et al., 2006). As the rocks below the sample number HD21 are yet to uncovered, it seems probable that the occurred species within this interval extend downwards into lower horizons.

Pterograptus elegans Zone

The biozone is identified with its base at the FAD of the *Pterograptus elegans* (Fig. 3A, 3E), which itself is rather common within the zone and ranges through up into the lower part of the succeeding graptolite zone. At the base, *Archiclimacograptus angulatus*, *A. caelatus* and *Normalograptus* sp. also make their first appearances. Some species extend up from the preceding graptolite zone, such as *Xiphograptus norvegicus*, *Pseudoclimacograptus* sp., *Glossograptus* sp., and *Didymograptus* sp. In the middle part of the zone, there are some long-ranging species like *Phyllograptus ilicifolius*, *Tetragraptus* sp. and *Expansograptus* sp. Significantly, *Archiclimacograptus riddellensis* and *Haddingograptus* sp. make their first appearances in the top part, exhibiting correlation potential to Australasia (VandenBerg and Cooper, 1992). The single or rare occurrences of *Tylograptus* sp. and *Nicholsonograptus fasciculatus* in the upper part of this zone are somewhat surprising, and it is quite likely that they range up from much lower horizons.

The graptolite succession of the *Pterograptus elegans* Zone has been relatively poorly understood in eastern Jiangxi and western Zhejiang provinces, although the eponymous species itself has been recorded in many localities, and the zone has long been suggested and adopted for the specific interval across the entire area. In the Huangnitang section, this interval was disturbed by Mesozoic volcanic dikes, and contains poorly preserved graptolite specimens. In the Baijiawu section, Yushan, the recognized *P. elegans*

Zone contains a rather diverse graptolite fauna as reported by Xiao et al. (1991), but the sampling was in relatively low density and many species of the fauna need to be taxonomically restudied. At some other localities, this zone was also recognized, such as in the Huiyingting section, Yushan (Chen and Han, 1964), the Poponong section, Longyou (Ge, 1962), the Laohuwu section, western Jiangshan (identified as *Pterograptus* sp. by Han, 1966), and the Banqiao section, Lin'an (Zhang et al., 2010), based largely on single or rare occurrences of the eponymous species.

The *P. elegans* Zone, as recognized herein at Hengdu, can be well correlated to the same zone in southern Anhui (Li, 1983), southern Jiangxi (Li et al., 2000), Scandinavia (Maletz, 1995, 1997), the *Didymograptus murchisoni* Zone in Britain (Fortey et al., 2000), Tarim (Bergström et al., 1999), and the lower part of the *Hustedograptus teretiusculus* Zone in North America (Maletz and Mitchell, 1995)

"*Hustedograptus teretiusculus* Zone"

The biozone herein is provisionally defined with its base at the FAD of the eponymous species, which coincides with the occurrence of *Reteograptus* sp.. Right below the basal boundary, *Archiclimacograptus riddellensis* makes its first appearance, and slightly above the boundary there are some indeterminate, biostratigraphically undiagnostic species. In the middle of the biozone, *Gymnograptus linnarssoni* and *Prolasiograptus* sp. make their first appearances significantly.

In the upper part of the biozone, *Dicellograptus* sp. first occurs, and slightly higher up in the top part more species of *Dicellograptus*, including *D. sextans*, *D. exilis*, together with *Leptograptus* sp., *Dicranograptus brevicaulis* first occur, in association of *Glossograptus* sp. and *Expansograptus* sp. The graptolite assemblage in the upper part of the biozone is significantly different from the lower part, as typified by the occurrence of abundant and diverse early dicellograptids, based on which an upper *Dicellograptus* Subzone is tentatively suggested in the present paper. The specimens of *Dicellograptus* sp. in the basal part of this subzone are poorly preserved and cannot be identified to any specific species. Further collections of well-preserved specimens may help identify the presumably earliest species of dicellograptids and solve the problem.

The "*Hustedograptus teretiusculus* Zone" has been adopted for the corresponding interval in many other regions including Britain, Scandinavia, North America, etc. but latest study reveals that this biozone is problematic as its base is not defined at the first appearance of the eponymous species which itself is also long-ranging (Maletz, 1997). In Britain, where the *H. teretiusculus* Zone was originally proposed by Miss Elles, the basal boundary has been traditionally placed at the last appearance of *Didymograptus murchisoni*, corresponding to the extinction of pendent didymograptids, and the FAD of the eponymous species lies actually in the middle of the biozone (Hughes, 1989; Zalasiewicz et al., 2009). In the Oslo region, the base of the *H. teretiusculus* Zone was chosen at the first appearance of *Dicellograptus vagus*, together with either *Gymnograptus linnarssoni*, *Orthograptus calcaratus* cf. *acutus*, or *Reteograptus* sp., and the biozone is characterized by the common occurrences, rather than the first appearance, of the eponymous species, together with *Dicellograptus vagus*, *Dicranograptus irregularis*, *Glossograptus hincksii*, *Gymnograptus linnarssoni*, *Orthograptus propinquus* and *O. calcaratus* cf. *acutus* (see Berry, 1964, p.77-78). The FAD of the *H. teretiusculus* itself is much lower at approximately the base of *Pterograptus elegans* Zone (Maletz, 1997). Recently, Maletz et al. (2007) suggested a replacement of the poorly-defined *H. teretiusculus* Zone with the proposed *Dicellograptus vagus* Zone in Scandinavia. In Alabama, USA, the basal Athens Shale yields a graptolite fauna corresponding to the topmost part of *H. cf. teretiusculus* Zone, which include *Dicellograptus geniculatus*, *Archiclimacograptus angulatus*, *Normalograptus euglyphus*,

Cryptograptus marcidus, *Glossograptus ciliatus*, *Reteograptus geinitzianus*, *Haddingograptus* cf. *eurystoma*, and *Lasiograptus* sp. (Finney, 1977, 1984). In the Marathon region of West Texas, the *H. cf. teretiusculus* Zone includes *Archiclimacograptus riddellensis*, *Pseudoclimacograptus confertus*, *Phyllograptus nobilis*, and the eponymous species (Berry, 1960).

Based on the included graptolite assemblage, the “*H. teretiusculus* Zone” herein is somewhat similar to the definition of the same zone in Britain, but with the basal boundary slightly higher at the FAD of eponymous species, rather than the LAD of the preceding *D. murchisoni*. The *Dicellograptus* Subzone is approximately equivalent to the *H. teretiusculus* Zone in its original sense in Scandinavia (i.e. the *Dicellograptus vagus* Zone by Maletz et al., 2007), and well correlated to the *Gymnograptus linnarssoni* Zone in Yangtze Region of South China (see Zhang et al., 2007b; Chen et al., 2010). In the lower interval of the “*H. teretiusculus* Zone”, no diagnostic species are available to name a lower subzone, which will requires further collections. This interval is roughly equivalent to the *Pseudamplexograptus distichus* Zone in Scandinavia (Maletz, 1997; Maletz et al., 2007).

Nemagraptus gracilis Zone

This biozone is well defined with its base at the FAD of *Nemagraptus gracilis*, and contains a rather diverse fauna (Figs. 2, 3). Slightly above the basal boundary, *Dicellograptus vagus* (Fig. 3G), *D. gurleyi* (Fig. 3K), *Orthograptus* sp., *Crynoides* sp., *Expansograptus superstes* (Fig. 3I), *Normalograptus brevis* (Fig. 3L), *Pseudazygograptus incurvus* (Fig. 3H), *Pseudoclimacograptus scharenbergi* (Fig. 3M) and *Climacograptus* sp. make their first appearances, together with some species extending up from the underlying “*H. teretiusculus* Zone”, including *Dicellograptus sextans*, *Xiphograptus norvegicus*, *Archiclimacograptus riddellensis*, *Gymnograptus linnarssoni*, and *Hustedograptus teretiusculus*. At Fågelsång, Sweden, *D. vagus* first appears at a horizon significantly lower than the boundary (see Bergström et al., 2000), and the species was recently adopted to name a biozone as replacement of the previous *H. teretiusculus* Zone (Maletz et al., 2007). In the Liujia section, *D. vagus* does first appear significantly lower than that of *N. gracilis*, implying that the species may also range downwards into the “*H. teretiusculus* Zone” in western Zhejiang. The occurrences of *E. superstes* within the *N. gracilis* Zone at Hengdu, and Huangnitang (Chen et al., 2006), indicate that the expansograptids range up into early Sandbian, slightly younger than in southern Sweden.

The graptolite assemblage of the *N. gracilis* Zone in Hengdu Quarry is very similar to that in the adjacent Huangnitang section and the Dawangou section in Tarim, suggesting that they are well correlated to each other. No graptolites diagnostic of age younger than the *N. gracilis* Zone are recorded in the Hengdu and Liujia sections, indicate that top of the Hulo Formation in western Zhejiang is no younger than early Sandbian.

Figure 3. Some biostratigraphically significant graptolites of late Darriwilian to early Sandbian from the Hengdu section, Jiangshan, Zhejiang Province. A, E, *Pterograptus elegans* Holm, 1881: A, NIGP153681 (AEP-HD-13-20a); E, NIGP153682 (AEP-HD-13-19b). B, *Archiclimacograptus riddellensis* (Harris, 1924), NIGP153683 (AEP-HD3-45a-7). C, *Archiclimacograptus caelatus* (Lapworth, 1875), NIGP153684 (AEP-HD3-4-1). D, *Nemagraptus gracilis* (J. Hall, 1847), Hulo Fm. at Hengdu, Jiangshan, Zhejiang, NIGP150352 (AEP800). F, *Dicellograptus sextans* (J. Hall, 1843), NIGP153685 (AEP-HD3-39-1). G, *Dicellograptus vagus* Hadding, 1913, NIGP153686 (AEP-HD4-8b-2). H, *Pseudazygograptus incurvus* (Ekström), NIGP153687 (AEP-HD3-25a-2). I, *Expansograptus superstes* (Lapworth, 1876), NIGP153688 (AEP-HD4-4). J, *Gymnograptus linnarssoni* Moberg, 1896, NIGP153689 (AEP-HD1-34a). K, *Dicellograptus gurleyi* Ruedemann, 1908, NIGP153690 (AEP-HD3-32-1). L, *Normalograptus brevis* (Elles and Wood, 1906), NIGP153691 (AEP-HD4-10). M, *Pseudoclimacograptus scharenbergi* (Lapworth, 1876), NIGP153692 (AEP-HD1-25b-2).

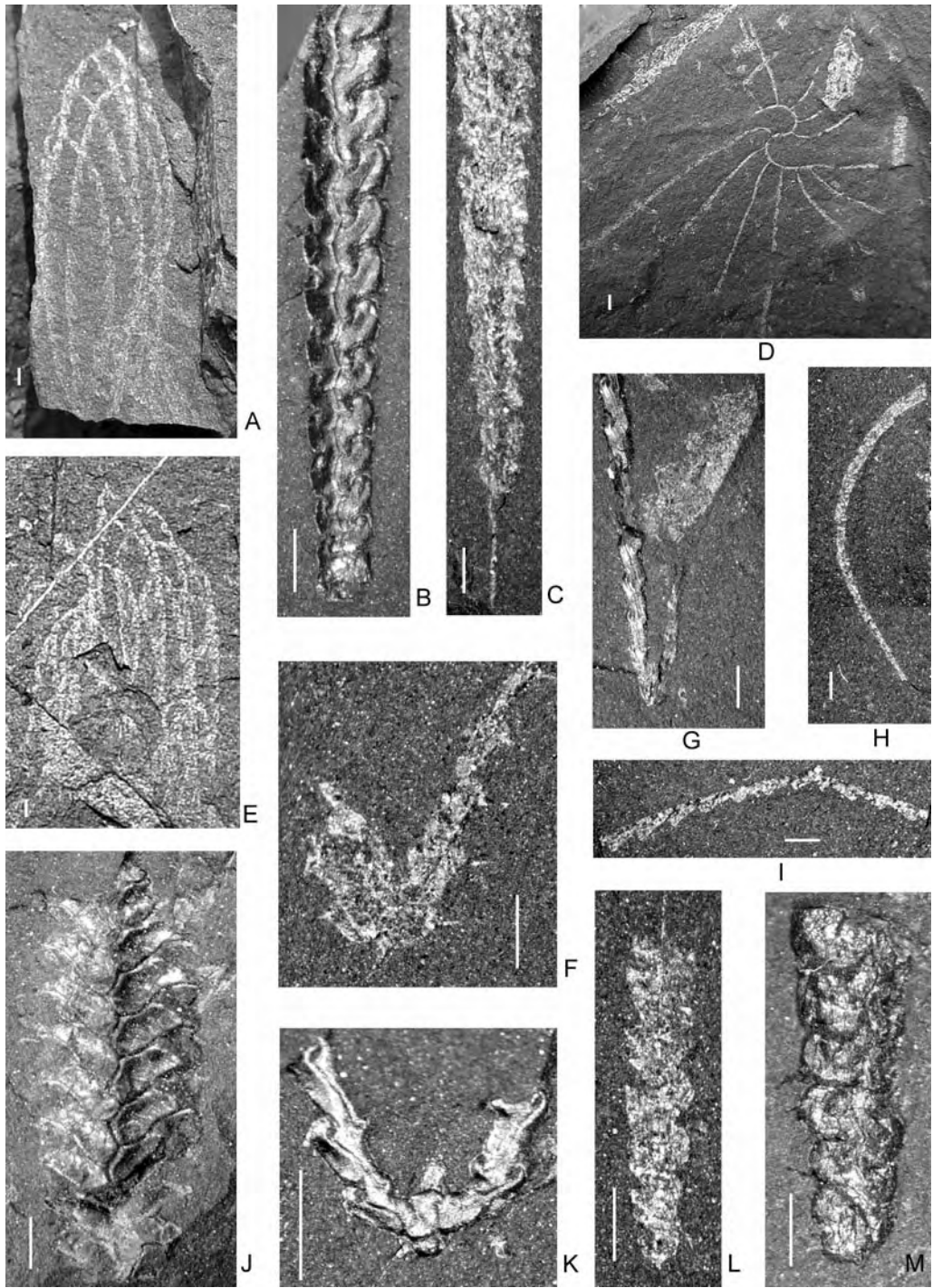


Figure 3

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DETRITAL SOURCE ANALYSES OF LATE ORDOVICIAN (HIRNANTIAN?) TO SILURIAN DEPOSITS OF NORTHWESTERN AND EASTERN ARGENTINA AND CONSTRAINTS FOR PALAEOTECTONIC EVOLUTION

U. Zimmermann

Universitetet i Stavanger, Institutt for petroleumsteknologi, Ullandhaug, 4036 Stavanger, Norway.
udo.zimmermann@uis.no

Keywords: Palaeogeography, provenance, Upper Ordovician, northwest Argentina.

INTRODUCTION

Late Ordovician to Early Silurian successions in northwest and central Argentina can be interpreted as overlap sequences regarding the Ordovician orogenic processes as a result of active continental margin settings. This work presents a provenance study of several latest Ordovician to Early Silurian successions in northwestern and central Argentina to reveal a better understanding of the Ordovician palaeogeography. The sediments were deposited in small-scale depositional centres (Fig. 1), close to Pucará in the Santa Victoria Range, at Zapla in the Sierras Subandinas (Zapla and Lipeón Formations), in the eastern Puna at Los Colorados (Zapla, Lipeón and Arroyo de los Colorados Formations), the western Puna at Salar del Rincón (Salar del Rincón and Lipeón Formations) and in the Precordillera region (Los Espejos and Don Braulio Formations).

The palaeogeography in northwestern Argentina at the boundary of the Ordovician and Silurian is interpreted as having been dominated by the exhumation of the Ordovician arc and older siliciclastic successions of Cambrian to Ediacaran age (Mesón Group and Puncoviscana Formation) during the development of the so-called Puna-arch (e.g. Mon and Salfity, 1995). In central Argentina, the Precordillera terrane records a partly very different supracrustal Early Palaeozoic lithostratigraphy compared to northwest Argentina (e.g. Kay et al., 1984; Fernandez Noia et al., 1990; Keller, 1999). However, the recycled basement material seems to be similar to the basement in northwest Argentina. The source rocks are possibly situated below the supracrustal rocks of the Precordillera as detrital zircon populations, Sm-Nd and Pb-Pb and also geochemical whole-rock data have shown (Bock et al., 2000; Zimmermann and Bahlburg, 2003; Gleason et al., 2007; Bahlburg et al., 2009; Abre, 2009; Rapela et al., 2010).

SEDIMENTOLOGY AND PETROGRAPHY

In northwest Argentina the lower part of the Los Colorados section has been sampled (Fig. 1a). The rocks are mainly composed of originally sub-angular quartz, coated by silica-rich rims, which mimic well-

rounded grains. Subordinated are muscovite, weathered feldspar and sedimentary lithoclasts. Prominent precipitation of iron-oxides cannot be observed. Few cross-bedding measurements show a transport direction towards the modern east. The succession includes a diamictite (Astini et al., 2004).

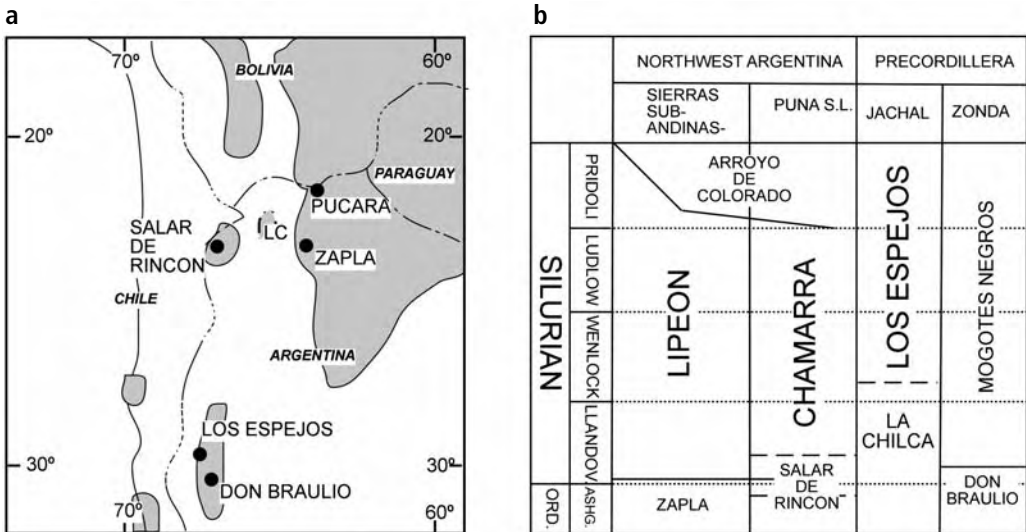


Figure 1. a, Outcrops location on a sketch of the NW Argentina. Grey shaded areas are proposed deposition centres during the Silurian (after França et al., 1995). (LC= Los Colorados). b, Stratigraphic table of Ordovician to Silurian formation here in discussion (after Aceñolaza et al., 1999; Moya and Monteros, 1999).

The Salar del Rincón Formation (Fig. 1a) is composed of quartz-rich arenites with a rich brachiopode fauna, and a 2-3 m thick layer enriched in iron-oxides and thin beds of conglomerates. However, a distinct glacial unit (Zapla Formation) is here not preserved. The Salar del Rincón Formation is mainly composed of well-rounded and well-sorted undulose quartz grains, again as a result of diagenetic processes, as grains in CL light appear to be sub-angular. Feldspar, sedimentary and metamorphic lithoclasts are rare. The matrix content is low (0% to 3%). The fauna and sedimentology points to a very shallow depositional setting in a low tidal environment (Isaacson et al., 1976) causing the sorting of labile fragments and crystals. Palaeocurrents are interpreted from flute marks ($n=2$) and cross-bedding ($n=4$) and suggests a transport from east to west.

At Zapla (Fig. 1a), close to the Mine 9th of October, the Lipeón Formation was sampled. The rocks are characterised by a high concentrations of mica and iron-rich minerals, and are described in detail by Boso and Monaldi (1999). Most common minerals are well-rounded and well-sorted quartz and mica. Other grains such as feldspar and lithoclasts are extremely rare, pointing to a facies environment where recycling and sorting were dominant. Boso and Monaldi (1999) interpret a shallow marine shelf environment in a sub-tropical to tropical climate. However, the mineralization is not precisely dated but proposed to be early diagenetic. Palaeocurrents are deduced from only few cross-bedding measurements and show a sediment transport from west to east.

At Pucará (Fig. 1a), 8 km to the north of Santa Victoria, diamictites (Zapla Formation) are concordantly overlain by shales and iron-rich rocks (Lipeón Formation). The outcrop has been described in detail by Turner (1964). The proposed to be glacial-marine deposits are dominated by well-rounded quartz, few feldspar,

abundant mica and clay minerals. Pebbles of granitic and arenitic origin can be found with clast sizes between 2 and 15 cm. Reworking of shaley and silty fragments from the underlying Lower Ordovician deposits (Santa Rosita and Acoite Formations) are obvious. The rocks of the Lipeón Formation are composed of quartz-rich yellow to orange sandstones intercalated with iron-rich layers. The sandstones dominated by angular to rounded quartz and mica and a matrix content of less than 3%. Palaeocurrent indicators determine the source for the sediment in the southeast.

In the Precordillera, the rocks of the Los Espejos Formation indicate platform deposits composed of mainly fine-grained sediments deposited as turbidity currents (Baldis and Peralta, 1999). The main component of the rocks is fine to medium-grained well-sorted sub-angular quartz, while only larger grains are well-rounded. Feldspar and lithoclasts are rare, and the matrix (<10 %) rich in mica, iron-oxides and clay minerals. Paleocurrents point to a source for the sediment in the east. For comparisons and complement the Don Braulio Formation deposited in the Precordillera is shown, which contains a thick bed of diamictites, interpreted to be related to the Hirnantian glacial events (Buggisch and Astini, 1993). The main components are rarely undulose sub-angular fine to medium grained quartz grains, feldspar and lithoclasts are scarce in an iron-oxide rich matrix (Abre et al., 2005; Abre, 2009).

PROVENANCE

Zr/Ti and Nb/Y ratios (Fig. 2a) demonstrate a mainly rhyolitic to dacitic composition for all rocks from northwest Argentina besides the Lipeón Formation with significant lower Zr/Ti ratios. The rocks of the Precordillera are different and characterized by the lowest Zr/Ti ratios (Don Braulio Formation) and slightly higher Nb/Y ratios with a larger scatter in Zr/Ti in the Los Espejos Formation despite its fine-grained

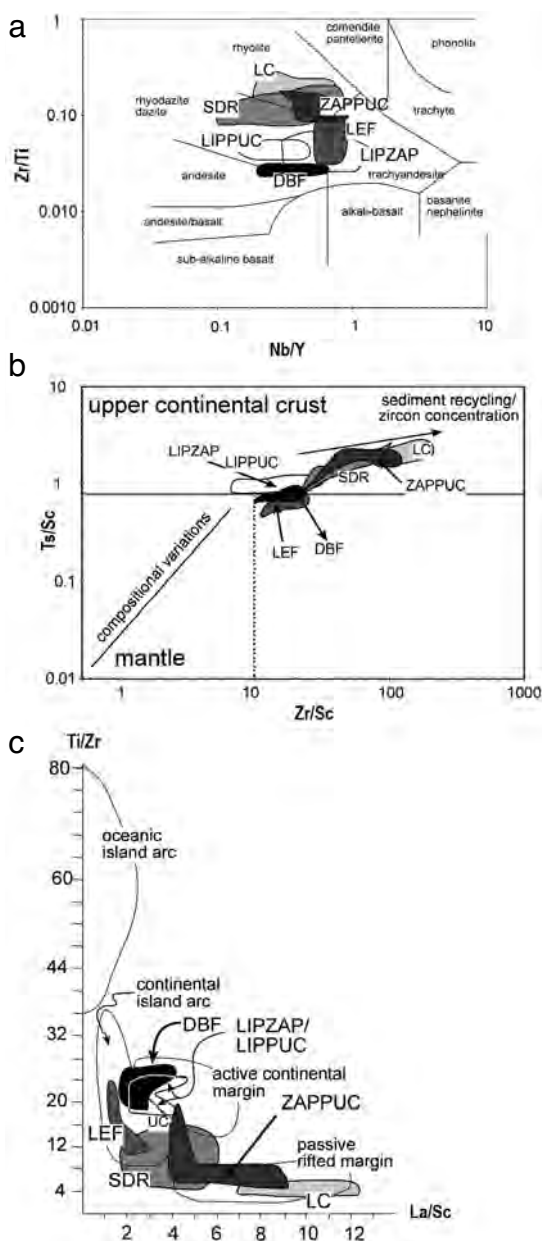


Figure 2. a, Composition of Silurian strata after Winchester and Floyd (1977). b, Plot of Th/Sc vs. Zr/Sc (after McLennan et al., 1990). c, Provenance plot after Bhatia and Crook (1986) using Ti/Zr vs. La/Sc ratios. DBF = Don Braulio Fm; LEF = Los Espejos Fm; LC = "Los Colorados"; SDR = Salar del Rincón Fm; LIPZAP = Lipeón Fm, Zapla; LIPPUC = Lipeón Fm, Pucará; ZAPPUC = Zapla Fm, Pucará.

Late Ordovician to Silurian

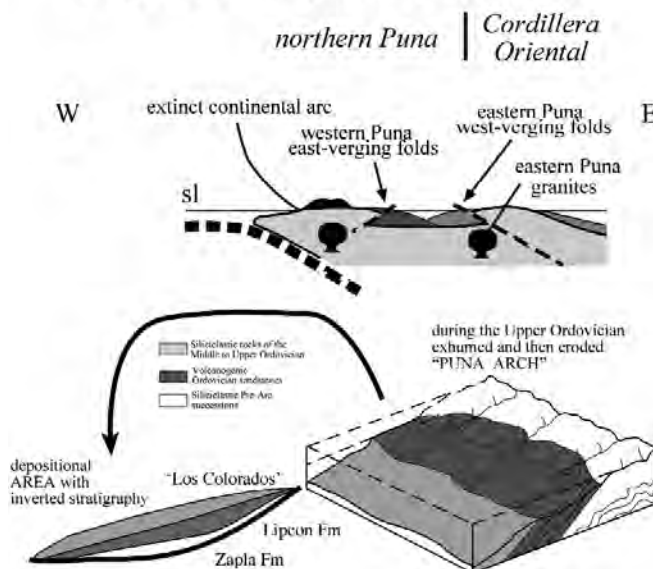


Figure 3. Source model for Late Ordovician to Silurian successions in northwest Argentina.

character. Th/Sc versus Zr/Sc ratios (Fig. 2b) identify the effective recycling of the sediments in the rocks from the northwest of Argentina, again with the exception of the Lipeón Formation with Zr/Sc and Th/Sc ratios similar to unrecycled upper continental crust (UCC). The Don Braulio and Los Espejos Formations are different showing the lowest Th/Sc ratios with UCC comparable Zr/Sc ratios. Assuming the sedimentary debris is related to Ordovician geological processes then the detritus in the Precordilleran formations and the Lipeón Formation might point to an arc source (Fig. 2c). All other deposits are devoid of significant intermediate or mafic detritus including volcanic lithoclasts, as shown above.

DISCUSSION

In northwest Argentina a continental arc was prominent and is well established (Zimmermann and Bahlburg, 2003). Tremadocian and Arenigian successions in northwest Argentina record, in contrast, significant 'Famatinian' magmatic zircons (e.g. Baldo et al., 2003; Zimmermann et al., 2010). Other interpretations of the evolution of the Precordillera argue for a large-scale strike-slip movement along Gondwana and the scarcity of 'Famatinian' detrital zircons is either related to sedimentary processes or this specific magmatic arc was not active or not even present in this area of Argentina. The demonstrated geochemistry of the rocks point to the influence of mafic and intermediate rocks in the detritus for the successions in the Precordillera and the Lipeón Formation of northwest Argentina. Although the age of this detrital component is not secured, palaeocurrents point to a source towards the west of the outcrop and might indicate the Early Ordovician arc as a source. High Sc and low Ta and Nb concentrations can support this interpretation. The absence of this intermediate to mafic component in all other three exposures in northwest Argentina can be related to the diametral directed palaeocurrents and to the interpretation that

most of the Late Ordovician to early Silurian basins were smaller basins without representing a large depositional area adjacent to the extinct magmatic arc. Alternatively, the oldest formations in this study, Zapla and Salar de Rincón, received material from pre-arc successions, possibly the Ediacaran to Lower Cambrian Puncoviscana Formation, the Cambrian Mesón Group and siliclastic Tremadocian successions. After the erosion of the arc deposits, Middle to Upper Ordovician rocks have been reworked in younger Silurian deposits ('Los Colorados').

The southern prolongation of the Puna continental arc extended definitely towards the Famatina area (e.g. Clemens and Miller, 1996). The proposed collision of the Precordillera terrane during the Early Palaeozoic (e.g. Astini et al., 1995) requires a subduction zone with a possible magmatic arc. Remnants of this arc as detritus in clastic successions of the Precordillera are so far not published, especially in regard of detrital zircons with ages typical for this magmatic event (Gleason et al., 2007; Abre et al., 2011). The here discussed rocks from the Precordillera are different, as they carry high Ti, Ta and Nb concentrations, exceeding significantly UCC values, not typical for a volcanic arc source. The intermediate to mafic component can, therefore, as well explained as derived from the existing mafic basement rocks (Rapela et al., 2010).

CONCLUSIONS

The petrography and geochemistry of Upper Ordovician to Lower (Middle) Silurian rocks in northwestern and central Argentina could reveal significant differences in their compositions and source areas which fed these sediments. The deposits in northwest Argentina seemed to be disconnected from each other and deposited in smaller restricted basins, controlled in their composition by their surrounding rocks. Only one formation shows direct influence of arc detritus, most probably derived from the Lower Ordovician Puna-Famatina continental arc. In other exposures the Upper Ordovician to Lower Silurian rocks received their material from either pre-arc successions or post-arc rocks devoid of major volcanoclastic detritus. Although most of the rocks show geochemical proxies pointing to high-grade recycling in some exposures quartz grains have been originally angular, but diagenetic processes let the grains appear to be rounded in common light microscope analysis. The absence of feldspar might be related to climatic constraints but still is remarkable in the proposed to be glacial deposits. The low abundance of matrix might point to a strong reworking in a local area, as during diagenetic processes matrix should develop from deposited fragile lithoclasts, which is not the case. The relatively absence of obvious metamorphic basement material in form of lithoclasts, besides possible metamorphic quartz, points to mainly supracrustal rocks as sources (Fig. 3). In the rocks of the Precordillera the provenance interpretation is more complex. Although intermediate to mafic sources play a role in the composition of the rocks, they cannot be identified as being related to an arc terrane according to the geochemistry and petrography. This trend might be related to a different palaeotectonic evolution in central Argentina compared to the northwestern part in regard of the Early Ordovician to Early and Middle Silurian.

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FROM FORE-ARC TO FORELAND: A CROSS-SECTION OF THE ORDOVICIAN IN THE CENTRAL ANDES

U. Zimmermann

Universitet i Stavanger, Institutt for petroleumsteknologi, Ullandhaug, 4036 Stavanger, Norway.
udo.zimmermann@uis.no

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INTRODUCTION

The tectonic evolution of the proto-Andean margin of western Gondwana has been commonly seen in terms of terrane accretion processes, requiring the existence of early Palaeozoic terrane boundaries and associated sutures. However, recent studies in supracrustal succession revealed a different scenario for the northwestern of Argentina, which are here compiled and new data added. The Ordovician of the northwestern of Argentina (Fig. 1) is dominated by the evolution of an active continental arc with it associated basins and magmatic record. The occurrence of intrusive mafic and ultra-mafic rocks has been often and repeatedly interpreted as representatives of a suture and as relicts of an oceanic basin (e.g. Allmendinger et al., 1983; Ramos et al., 1986; Ramos, 2008). However, there are no data, which can support such an assumption. Field geology relations, besides geochemical and isotope geochemical analyses also refuse this hypothesis. Associated sedimentary rocks are devoid of any influence of detritus of rocks related to oceanic crustal, regarding their mineralogy, geochemistry or isotope geochemistry (Bock et al., 2000; Zimmermann and Bahlburg, 2003).

SEDIMENTOLOGY

The Ordovician sedimentary rocks reflect the evolution from a shelf area to the establishment of a continental arc and the syn- and post-arc developed retro-arc and foreland basins (Fig. 2a-d).

In the west of the Ordovician basin, located in northern Chile, fore-arc and intra-arc successions have been deposited with high amounts of volcanoclastic detritus associated with a bi-modal volcanic succession of tholeiitic basalts, andesites and rhyolites with an age of c. 478 Ma (Zimmermann et al., 2010). The sedimentary rocks can be interpreted as turbidity currents and deep-marine shales. Section up, the influence of volcanic debris decreases and the rocks are richer in siliciclastics with the occurrence of brachiopods (Benedetto et al., 2008). Further east, in the Argentinean Puna (Fig. 1), the Ordovician represents during the

Tremadoc and Arenig a retro-arc basin environment (Zimmermann and Bahlburg, 2003) and a shallow marine basin margin environment in the Cordillera Oriental (Moya, 1999). The retro-arc basin deposits are characterized by associations of highly volcani-clastic rocks, andesites and rhyolites, represented by lavas, tuffs and pyroclastic flows (Moya et al., 1993; Zimmermann and Bahlburg, 2003). The sediments here were deposited in very different water depth, partly close to volcanic centres, partly as turbidites, conglomerates and fine-grained siltstones, as well as immature coarse-grained wackes carrying various different faunas like graptolites and brachiopodes. The former can date the earliest recorded occurrence of arc related volcanism into the earliest Tremadoc in one specific area (Salar de Rincón; Moya et al., 1993), while further south lava flows related to the continental arc are younger than Lancefield 2. This facies complex is named 'Puna Volcanic Complex'.

During the late Arenig, Llanvirn and Llandeilo the retro-arc basin widens and a foreland basin developed, which is especially in its fully dimension exposed in the northern Puna (Bahlburg, 1990; Bahlburg and Furlong, 1996), while in the southern Puna and northern Chile those deposits are scarce (Zimmermann et al., 2002). The successions are dominated by large turbidite systems and channel systems and combining the Lower and Upper Turbidite System (Bahlburg, 1990. 1998) to the 'Puna



Figure 1. Geological map of the Puna after Zimmermann and Bahlburg (2003).

Complex' (Bahlburg, 1990; Bahlburg and Furlong, 1996), while in the southern Puna and northern Chile those deposits are scarce (Zimmermann et al., 2002). The successions are dominated by large turbidite systems and channel systems and combining the Lower and Upper Turbidite System (Bahlburg, 1990. 1998) to the 'Puna

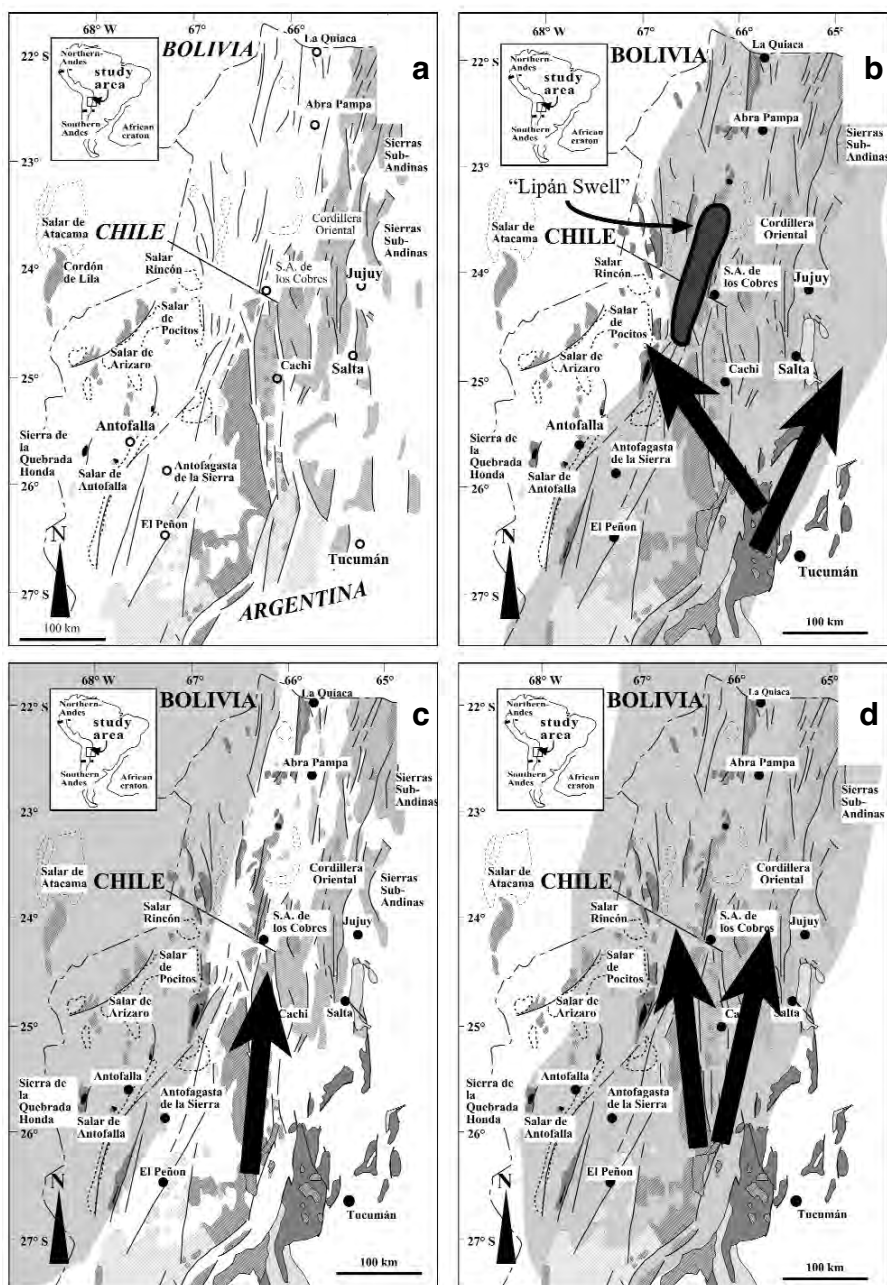


Figure 2. Geological map from Fig. 1 (see Fig. 1 for references). a, Outline of the basin area before the sedimentation of the Ordovician. Note that Puncoviscana Formation equivalents are exposed to the west of the Salar de Antofalla and in the southern Puna. b, Extension of the Puna Shelf Complex (+/- Tremadoc) (shaded region) with the well established palaeohigh at Lipán (Moya, 1999). c, Extension of the Puna Volcanic Complex (+/- Arenig) (shaded region). d, Distribution of the Puna Turbidite complex (+/- Llanvirn) (shaded region). Black arrows point to the main palaeocurrent directions for each facies association.

Turbidite Complex'. These successions are associated with intrusive magmatism of mainly felsic nature, which, however, occurs only in one specific region (Fig. 1).

The evolution of the continental arc did not affect the entire region at the same time, hence was diachronic. Therefore, distal deposits are only partly affected by volcanic detritus and represented by reworked arenites in the central Puna and further east in the Cordillera Oriental. In the latter, the Ordovician deposits were mostly fed by underlying Cambrian to Ediacaran sedimentary rocks and basement successions. Those deposited in the Puna region are compiled as 'Puna Shelf Complex' and equivalent, in age and facies, with shallow-marine deposits recorded in the Cordillera Oriental during the late Arenig and Llanvirn until the Caradoc (Bahlburg, 1990; Moya, 1999). Although distal to the active arc, some of the successions are affected by arc related tuff layers.

GEOCHEMISTRY AND ISOTOPE GEOCHEMISTRY

The same basin evolution can be identified in the whole-rock geochemistry of the Ordovician successions (Fig. 3a). The west-east trend from fore-arc to basin margin deposits is beyond doubt. This can be substantiated by the geochemical signature of the associated intrusive and extrusive rocks (Fig. 3b,c). Intrusion ages of arc related plutons support the continental arc setting during the Arenig (Kleine et al., 2004; Poma et al., 2004). Arc related tuff layers could be observed in the Cordillera Oriental at Salta in the San Bernardo Formation (Fig. 3a). Sm-Nd isotope analysis and Pb-Pb data on whole rock samples reflect typical signatures for the basement in northwestern Argentina with T_{DM} around 1.5 Ga but as old as 2.0 Ga, which points to the existence of Palaeoproterozoic basement, so far undiscovered. Only in few samples of arc related lavas, a subordinated juvenile component could be identified. This supports the interpretation of a

tectonic setting dominated by a relatively evolved crust affected only locally by juvenile magmas. The active margin setting during the Ordovician recycled mainly underlying crust of Gondwana.

DISCUSSION

Decades ago the area was pin-pointed as being dominated by collisional tectonics during the Ordovician with the accretion of several micro-terranes and the evolution of two subduction zones (e.g. Ramos et al., 1986, Ramos, 2008 and references therein). This was mainly based on very few geochemical analyses of mafic and ultramafic rocks of unknown age (Coira et al., 1982; Allmendinger et al., 1983; Zappettini et al., 1994). After thorough mapping and extensive fieldwork the abundance of ultramafic could be limited to few small bodies in mainly gabbroic rocks (Zimmermann, 1999). Moreover, the mafic successions are by far less homogeneous than interpreted and might be related to different tectonic events, as they are until today not dated. A larger sample set of the mafic succession demonstrated that the rocks are not related to MORB or an oceanic environment (Zimmermann et al., 1999). Sedimentological studies on both sides of the suggested "suture" could not reveal any significant difference in composition or facies or identify any juvenile detritus (Zimmermann and Bahlburg, 2003; Zimmermann et al., 2010). Moreover, during the Ordovician the northwest of Argentina has not been affected by several volcanic belts ('fajas') as proposed (e.g. Coira et al., 1982), but records a well understood basin evolution from a compressive subduction zone setting (Tremadoc to Arenig) towards extensional tectonics in post-arc times (Llanvirn and

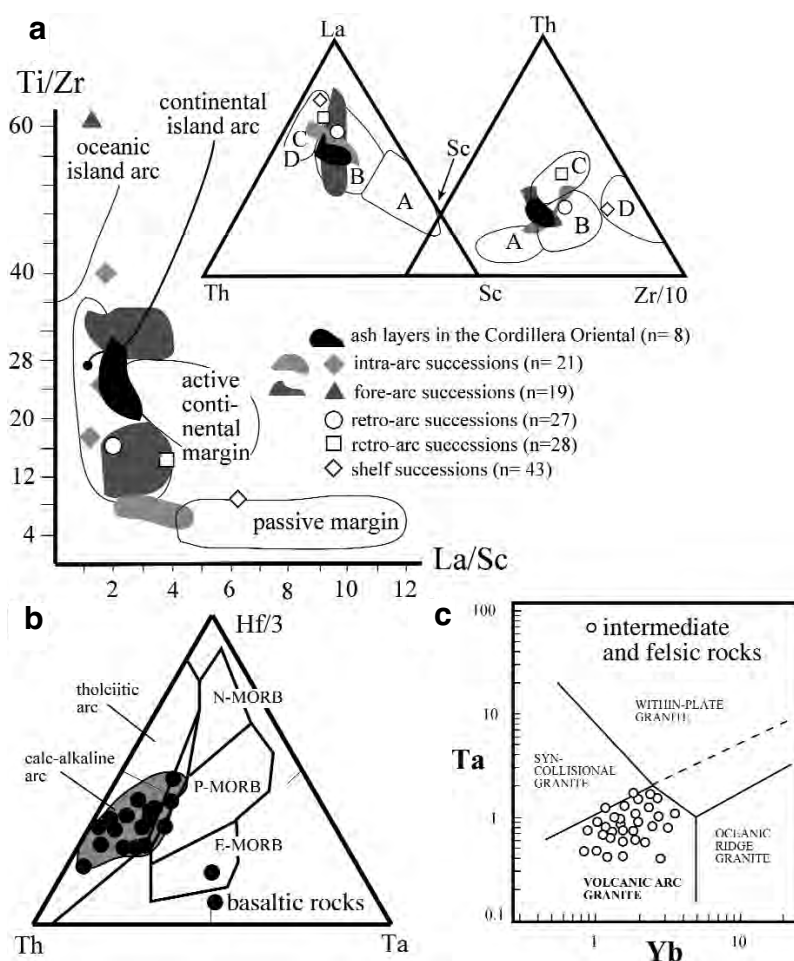


Figure 3. a, La/Sc versus Ti/Zr ratios. Data for age equivalent successions of the Argentinean Puna and northern Chile. Inlet: Ternary provenance diagrams. (Diagrams after Bhatia and Crook, 1986). A = oceanic arc; B = continental arc; C = active continental margin; D = rifted/passive margin. Average values from Bahlburg (1990, 1998), Zimmermann and Bahlburg (2003), Zimmermann et al. (2010) and new data. b, Hf/3-Th-Ta relations after Wood (1980) to define the type of basaltic rock. Most of the samples are characterised by increase in Th and depletion in Hf and Ta. c, Felsic magmatic rocks discriminated regarding their tectonic setting using Ta-Nb-Yb-Rb concentrations (after Pearce et al., 1984), pointing to a continental arc related tectonic setting.

younger), which in turn amounts for magmatic activity (Bahlburg, 1990) and caused the variety of magmatic rocks. Hence, the terrane model has to be abandoned.

Palaeomagnetic data have been often used to demonstrate a rotation of the Puna region to support the terrane hypothesis. This rotation argues for a counter-clockwise movement and describes a basin opening from north to south (Forsythe et al. 1993; Conti et al., 1996). These interpretations are diametral to several hundred palaeocurrent measurements in Ordovician successions in northern Chile and the Puna region, where detritus was transported mainly from the arc into the basin towards the east and from south to north. Unpublished paleomagnetic data from Ordovician rocks of the Puna showed for more than 25

sites that Permian overprints disturbed the original magnetisation. Recently, a variety of rotational movements were described from northwestern Argentina and assigned to the Ordovician but interpreted as a result of escape tectonics (Spagnuolo et al., 2010). This interpretation does not support the terrane model or the hypothesis of the existence of 'back-arc' basins (Coira et al., 2009) - in contrast to 'retro-arc' basins - as the region has not been affected by the evolution of oceanic crust during the Ordovician.

The peak of the volcanism in northwest Argentina can be defined during the Middle Arenig (García et al., 1962; Breitzkreuz, 1986; Bahlburg, 1990; Zimmermann and Bahlburg, 2003). This seems to be diachronic to arc related peak volcanism in southern Peru (Middle Ordovician; Bahlburg et al., 2006) and the Sierra Famatina (Llanvirn; Clemens and Miller, 1996; Mángano and Buatois, 1996) - a situation comparable to the modern one in the Andes.

CONCLUSIONS

In the northwestern of Argentina the Ordovician successions reflect a basin evolution from the development of a continental arc to a retro-arc foreland basin. There are no data available, which could point to terrane accretions, terrane definitions or multiple synchronic magmatic belts related to the opening of wide basins floored by oceanic crust. The entire record of supra-crustal rocks reveals an exemplar for a Palaeozoic succession in which the change of the tectonic regimes can be observed in mineralogical, sedimentological, geochemical and isotope geochemical data and interpreted according to actualistic models. The western margin of Gondwana has been situated in northern Chile, and the eastern margin of the developed foreland basin associated to the active continental margin located in the Cordillera Oriental. Absence of any sutures across strike is consistent with an evolving continental margin arc constructed on attenuated crust of the proto-Andean margin and, therefore, the concept of an Antofalla Terrane is unnecessary for the Ordovician.

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The Early to Middle Paleozoic Revolution

Bridging the Gap between the Great Ordovician Biodiversification Event and the Devonian Terrestrial Revolution

International Geoscience Programme (IGCP) Project 591

The Early Ordovician to Early Devonian interval contains several of the most significant paleoclimate and paleobiological events in Earth history including paleobiodiversity events and/or perturbations to the global carbon cycle associated with the Great Ordovician Biodiversification Event (GOBE), near the base of the Katian, Ordovician-Silurian boundary, Llandovery-Wenlock boundary, middle Homerian, middle Ludfordian, and Silurian-Devonian boundary, among others. This interval of Earth history also contains the acme and amelioration of the Early Paleozoic Ice Age, which provides an important historical analogue for researchers of modern climate change. Additionally, this interval contains the roots of the invasion of life onto land. The Earth did not go quietly into the Middle Paleozoic and the primary research objective of IGCP 591 – ‘*The Early to Middle Paleozoic Revolution*’ is to investigate this dynamic and important interval in the history and evolution of life and our planet.

IGCP 591 is designed to allow the Early to Middle Paleozoic global community an opportunity to build on the momentum gained by the highly successful IGCP projects 410 and 503 by providing a regular venue in which to continue their research and dialogue so effectively begun during those projects. We are pleased to announce the commencement of this project with the 2011 Field Meetings of the International Subcommissions on Ordovician and Silurian Stratigraphy in Madrid, Spain, and Ludlow, England, respectively. Scheduled to run 2011-2015, annual meetings have been scheduled in the following locations.

2012: Cincinnati, Ohio

2013: Lund, Sweden

2014: Vilnius, Lithuania

2015: Lille, France

A host of other IGCP 591 related field trips and symposia have already been scheduled, but we would love to hear of anyone interested in hosting further activities. As with all IGCP projects, a small amount of funds are made available each year to help researchers from developing countries, students, and early career researchers attend project meetings. We look forward to IGCP 591 getting underway, and thank everyone in the community who emailed their support for the project.

IGCP 591 – *The Early to Middle Paleozoic Revolution*

Bradley D. Cramer (USA)

Thijs R.A. Vandenbroucke (France)

Renbin Zhan (China)

Michael J. Melchin (Canada)

Živilė Žigaitė (Lithuania)

Kathleen Histon (Italy)

Guillermo L. Albanesi (Argentina)

Mikael Calner (Sweden)

AUTHORS' INDEX

Abati, J.	43	Chemale Jr., F.	29
Abre, P.	29	Cheng, J.	101
Aceñolaza, G.F.	483	Cingolani, C.	29, 103
Aceñolaza, F.G.	35	Cirés, J.	95
Adachi, N.	309	Colmenar, J.	189
Adrain, J.M.	41	Cooper, R.A.	421, 499
Alfaro, M.B.	103	Copper, P.	109
Ahlberg, P.	327	Couto, H.	113
Ainsaar, L.	143, 353		
Albanesi, G.L.	409, 611, 625	Dai, X.	309
Aleinikoff, J.	95	David, M.	65
Aleksandrov, P.A.	589	Degtyarev, K.E.	589
Álvaro, J.J.	169	Delabroye, A.	119
Andonaegui, P.	43	Dias da Silva, I.	121
Arenas, R.	43	Díaz-Martínez, E.	127
Armstrong, H.A.	607	Dietsch, C.	95
		Donovan, S.K.	215
Bagnoli, G.	587	Dronov, A.V.	135, 143, 279
Barba, P.	49, 121	Du, P.	581
Barnes, C.R.	611		
Barrick, J.E.	295	Egenhoff, S.	333
Bartošová, J.	277	Ehsani, M.H.	169
Benedetto, J.L.	55	Einasto, R.	143
Bergström, S.M.	179, 295, 301, 559	Epard, J.L.	75
Bertero, V.	55	Ezaki, Y.	309
Bidone, A.R.	103		
Boyce, W.D.	339	Fatka, O.	65, 151
Brock, G.A.	253	Fernandes, P.	313
Bruton, D.L.	61	Fernández-Suárez, J.	43
Budil, P.	65, 151	Ferretti, A.	515
Bukolova, E.V.	69, 547, 553	Finnegan, S.	155
Bussy, F.	75	Finney, S.C.	161
		Fischer, W.W.	155
Cairncross, B.	29	Fryda, J.	277
Carlorosi, J.	83	Fuenlabrada, J.M.	43
Carrera, M.G.	55, 89		
Carreras, J.	95	Gaggero, L.	515
Casas, J.M.	95, 441	Galeano Inchausti, J.C.	103
Castiñeiras, P.	95, 391	Gámez Vintaned, J.A.	163
Challands, T.J.	607	García-Bellido, D.C.	483

Genge, M.J.	611	Kraft, P.	275, 277
Ghavidel-syooki, M.	169	Kushlina, V.B.	279
Ghienne, J.-F.	13	Legrand, P.	287
Ghobadi Pour, M.	169, 171, 413	Leslie, S.A.	295, 301, 537
Goggin, K.E.	221	Li, J.	617, 619
Goldman, D.	179, 199, 559	Liang, Y.	633
Gómez, J.	199	Liao, H.	309
Gonta, T.V.	135	Liesa, M.	95
González-Clavijo, E.	49, 121	Liu, J.	215, 309
González Cuadra, P.	43	Lopes, G.	313
González Menéndez, L.	473	Lourenço, A.	113
Grahn, C.Y.	127		
Gruenwald, R.	347	Maletz, J.	327, 333, 537, 597
Gutiérrez-Marco, J.C.	189, 199, 371, 463, 483, 505	Martín Algarra, A.	505
		Martínez, J.F.	95
Halpern, K.	55, 89	Martma, T.	267
Hammarlund, E.	515	McCobb, L.M.E.	171, 339
Hansen, J.W.	207	McDougall, N.D.	347
Harper, D.A.T.	3, 207, 215, 253, 455, 515	Meidla, T.	143, 353
Haynes, J.T.	221, 229	Melchin, M.J.	371
Heredia, S.	83, 237	Melson, W.G.	229
Hints, O.	243	Meng, L.	633
Histon, K.	515	Mergl, M.	359, 367
Holmden, C.	371	Mestre, A.	237
Howard, A.	301	Mitchell, C.E.	371
Hroch, T.	275, 277	Monaldi, C.R.	409
Huff, W.D.	135	Monteros, J.A.	379
		Mosley, J.	295
Ingham, J.K.	605	Moya, M.C.	83, 379
Iriondo, A.	95	Munnecke, A.	101, 119
Isaacson, P.E.	127		
Isozaki, Y.	251	Navas-Parejo, P.	505
		Navidad, M.	95, 391
Jakobsen, K.G.	253	Nestor, H.	109
Jiménez-Sánchez, A.	259	Nielsen, A.T.	207, 253, 399, 455, 605
		Nölvak, J.	243, 607
Kaljo, D.	143, 267		
Kanygin, A.V.	135	Obut, O.T.	403, 547, 553
Knight, I.	339	Oliveira, J.T.	609
Kolář, P.	65	Ortega, G.	409, 611, 625
Koptíková, L.	277		
Kotov, A.B.	589	Paluveer, L.	243
		Pantle, C.	179, 559

Paris, F.	607, 609
Paz, A.	493
Percival, I.G.	171, 413, 421
Pereira, Z.	313, 429
Péronnet, V.	75
Peters, S.	155
Piçarra, J.M.	313, 429, 483
Popov, L.	169, 171, 413, 589
Priewalder, H.	515
Puddu, C.	441
Rajchl, M.	275
Rak, Š.	151
Rasmussen, C.M.Ø.	447
Rasmussen, J.A.	455
Reche, J.	95
Repetski, J.E.	301
Reyes-Abril, J.	199, 463
Rocha, D.	493
Rodríguez-Cañero, R.	505
Rodríguez Sastre, M.A.	473
Ryazantsev, A.V.	589
Sá, A.A.	483, 493
Saadre, T.	143
Sabbe, K.	607
Sadler, P.M.	499
Saltzman, M.R.	301
Sánchez, T.M.	55
Sánchez Martínez, S.	43
Sarmiento, G.N.	83, 505
Schmitz, U.	163
Schönlaub, H.P.	515
Sell, B.K.	527, 537
Semenova, A.M.	403
Sennikov, N.V.	547, 553
Sequeira, A.J.D.	313
Servais, T.	119, 607, 617, 619
Seward, A.M.	301
Sheets, H.D.	179, 559
Simes, J.E.	421
Song, Y.Y.	649
Spötl, C.	515
Stampfli, G.M.	567

Steinová, M.	575
Stemmerik, L.	215
Stock, C.	109
Štorch, P.	371, 515
Stouge, S.	215, 581, 587
Suyarkova, A.	169
Tammekänd, M.	243
Terfelt, F.	587
Tolmacheva, T.Yu.	553, 589
Toro, B.A.	597
Tortello, F.	103
Truuver, K.	353
Ugidos, J.M.	49, 121
Ulianov, A.	75
Uriz, N.J.	103
Valladares, M.I.	49, 121
Van Roy, P.	21
Vandenbroucke, T.R.A.	119, 605, 607
Vavrdová, M.	127
Vaz, N.	313, 483, 609
Vecoli, M.	119
Verniers, J.	607
Villas, E.	463
Voldman, G.G.	611
von Raumer, J.	75, 567
Wilhelm, C.	567
Williams, M.	607
Wright, A.J.	421
Yan, K.	617, 619
Zalasiewicz, J.A.	607
Zeballo, F.J.	625
Zhan, R.	215, 309, 413, 633
Zhang, J.	597, 649
Zhang, Y.D.	101, 597, 649
Zhao, Z.	581
Zhen, Y.Y.	421
Zhou, C.	101
Zimmermann, U.	659, 667



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