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Chapter 8

From Greenhouse to Icehouse – The Eocene/Oligocene in Antarctica

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ABSTRACT

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The change from a warm, ice-free greenhouse world to the glacial Antarctic icehouse occurred during the latest Eocene–earliest Oligocene. Prior to this, during the Early–Middle Eocene, Antarctica experienced warm climates, at least on the margins of the continent where geological evidence is present. Climates appear to have been warm and wet, the seas were warm and plants flourished in a frost-free environment, although there is some suggestion of valley glaciers on King George Island. Climate signals in the geological record show that the climate then cooled but not enough to allow the existence of significant ice until the latest Eocene.

Glacial deposits on Seymour Island indicate that ice was present there at Eocene/Oligocene boundary times. Further south in the Ross Sea region, ice-rafted clasts in drill cores and deposits of tidewater glacier origin in Prydz Bay confirm the presence of ice at the continental shelf by the earliest Oligocene. This matches the major Oi-1 oxygen isotope event in the marine record. On land, *vegetation was able to persist but the warmth-loving plants of the Eocene were replaced by shrubby vegetation with the southern beech* Nothofagus*, mosses and ferns, which survived in tundra-like conditions. Throughout the Oligocene, glaciation waxed and waned until a major glacial phase in the Miocene.*

Coupled climate–ice sheet modelling indicates that changing levels of atmospheric CO_2 controlled Antarctica's climate. Factors such as mountain uplift, vegetation changes and orbital forcing all played a part in cooling the polar climate, but only when CO_2 levels reached critical thresholds was Antarctica tipped into its icy glacial world.

8.1. Introduction

One of the most intriguing challenges in Antarctic Earth history is to understand the fundamental climate change from the past greenhouse world with no major polar ice caps to our present icehouse that is dominated by the vast ice sheets on the Antarctic continent. This change across a major climate threshold holds many clues that will help us understand the potential changes our world may undergo in future.

Geological evidence from rocks and fossils from the Antarctic continent and from marine oxygen isotopes that record changes in temperature and water masses indicate that ice sheets built up on Antarctica from about Eocene/Oligocene (E/O) boundary times, approximately 34 million years ago. This is named the Oi-1 event in the marine oxygen isotope record and is represented by the appearance of several glacial deposits in the rock record. However, the actual pattern of climate cooling and the causes of glaciation are far from understood, and there are hints of the presence of ice in the Late Eocene. Indeed, it is even possible that ice existed on Antarctica even during the Cretaceous (Miller et al., 2005).

This chapter reviews our current understanding of the greenhouse–icehouse transition in Antarctica. It covers the interval of climate change from the warm greenhouse climates of the Early Eocene through to the first appearance of ice and the establishment of glacial conditions during the Oligocene. Several lines of evidence are presented for climate and environmental change: the sedimentary rock record on the continent provides clues to the nature of the cooling climate during the Eocene; latest Eocene and Oligocene sediments recovered from marginal basins in drill cores contain the earliest undisputed glacial deposits; fossil plants and palynomorphs from both the continent and marginal basins have yielded information about cooling climates in the terrestrial realm; marine microfossils hold clues to ocean circulation and the significance of ocean gateways; the marine isotope record of the open oceans tells us about changes in deep ocean temperatures as the result of climate cooling and ice growth; and finally, the cause of cooling and the birth of the icehouse world is explored through computer modelling. The final section summarizes our current understanding of the greenhouse–icehouse transition in Antarctica.

8.2. Climate Signals from the Sedimentary Record

Sedimentary strata of Eocene and Oligocene age are exposed in the Antarctic Peninsula region and provide an important record of environmental conditions on land and in shallow marine settings during this period. In addition, there are intriguing hints about environments at higher latitudes, extracted from erratic boulders composed of Eocene and Oligocene fossiliferous sediments in the Ross Sea region, derived from sub-glacial outcrops. A summary of the important outcrops and their environmental signal is given below.

Sedimentary rocks of Palaeogene age are exposed around the northern part of the Antarctic Peninsula, on the South Shetland Islands and on Seymour Island (Fig. 8.1). The sediments were deposited in very different tectonic settings and environments – the South Shetland sequence is of terrestrial volcanic and sedimentary deposits that represent an outer-arc (Birkenmajer, 1995a) or fore-arc (Elliot, 1988) succession; the sequence on Seymour Island consists of marine clastics deposited in a back-arc basin, the uppermost beds of a regressive megasequence (Hathway, 2000). Both contain evidence of Palaeogene cooling and the first appearance of ice but in different settings. AU :11

8.2.1. Antarctic Peninsula Region

The Palaeogene back-arc deposits exposed on Seymour Island and Cockburn Island comprise more than 1,000 m of shallow marine to coastal fossiliferous clastic sedimentary rocks mainly of Palaeocene and Eocene age (Elliot, 1988; Sadler, 1988; Marenssi et al., 1998). The Maastrichtian–Danian sequence forms a simple ~N–S homocline dipping gently to the east while the Late Palaeocene and the Eocene fill incised valleys trending NW–SE. The Early Eocene to latest Late Eocene La Meseta Formation (Elliot and Trautman,

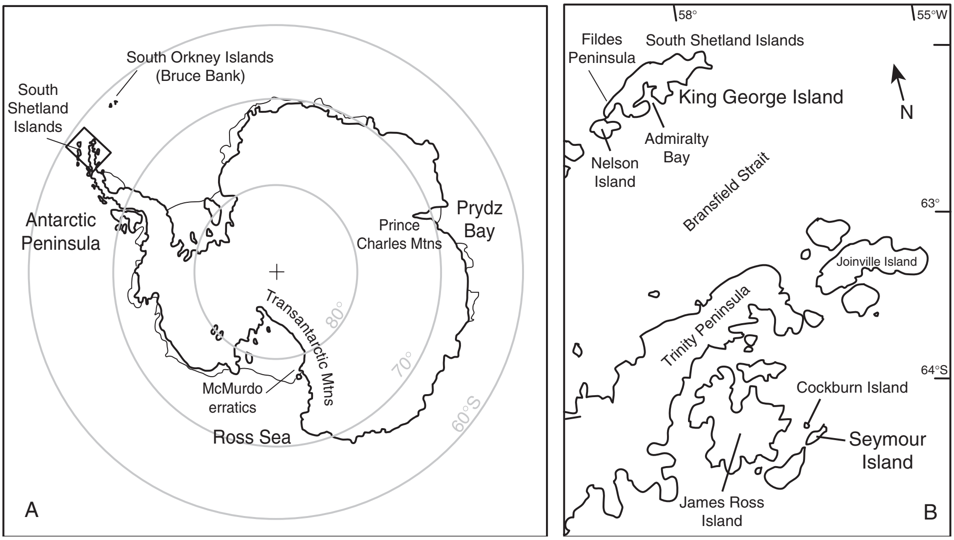

Figure 8.1: Map of Antarctica Continent (A) and the Antarctica Peninsula Region (B) Showing some of the Locations mentioned in the Text.

1982) is an unconformity-bounded unit (La Meseta Alloformation of Marenssi et al., 1996, 1998) of maximum composite thickness of 720 m, which fills a 7 km wide valley cut down into older strata after the regional uplift and tilting of the Palaeocene and Cretaceous older beds.

The La Meseta Formation comprises mostly poorly consolidated siliciclastic fine-grained sediments deposited in deltaic, estuarine and shallow marine environments as a part of a tectonically controlled incised valley system (Marenssi, 1995; Marenssi et al., 2002). The richly fossiliferous Eocene sediments have yielded the only fossils of land mammals in the whole Antarctic continent (Reguero et al., 2002) along with fossil wood, fossil leaves, a rare flower, plus marine vertebrates (including giant penguins) and invertebrates (Stilwell and Zinsmeister, 1992; Gandolfo et al., 1998a,b,c; AU :12
Francis et al., 2004a,b; Tambussi, 2006). AU :13

Overall, evidence from fossils, sediments and geochemistry from Seymour Island indicates generally warm and ice-free conditions during the earliest part of the Eocene but followed by gradual cooling. Dingle et al. (1998), based on chemical analysis (chemical index of alteration, CIA) and clay mineralogy, recorded a climatic deterioration from warm, non-seasonally wet conditions during the early Middle Eocene (smectite-dominated clay assemblage and CIA values <0.7 to >0.6) to a latest Eocene cold, frost-prone and relatively dry regime (illite-dominated clay association and CIA values <0.6). Gazdzicki et al. (1992) showed a 6 ppmil decrease in δC13 values in biogenic carbonates and Tatur et al. (2006) recorded an increase in AU :14 the Cd/Ca ratio in bivalve shells from the upper part of the La Meseta Formation. They interpreted these results as a change from stratified to vigorously mixed oceanic conditions related to the cryosphere development in the Southern Ocean by the end of the Eocene. Unpublished stable isotope (δO18 and δC13) measurements (Feldmann and Marenssi) obtained from molluscan shells of the La Meseta Formation show a period of warmer seawater temperatures during 51–47 Ma and a drop of 1.5°C during 35–34 Ma. This information is consistent with ice-rafted debris reported for the upper part of the La Meseta Formation (Doktor et al., 1988) and the palaeoclimatic evidence of a severe climatic deterioration towards the end of the Eocene. It is possible that by the end of the Eocene, limited ice, perhaps AU :15 as valley glaciers, was already present in the area.

By the end of the Eocene, it is possible that an ice sheet extended over much of the peninsula. Ivany et al. (2006) reported 5–6 m thick glacial deposits that conformably overlie the typical marine sandstones of the La Meseta Formation but are beneath the glacial diamictites of the younger Weddell Sea Formation. Based on dinocyst stratigraphy and strontium isotopes, these authors suggested an age of 33.57–34.78 Ma for these glaciomarine deposits (supported by marine Sr dates from Dingle and Lavelle, 1998; Dingle et al., 1998; Dutton et al., 2002), at or very close to the Eocene–Oligocene boundary. There is therefore a small but intriguing window into the early stages of the icehouse world in the James Ross Basin.

8.2.2. King George (25 de Mayo) Island, South Shetland Islands

The South Shetland Islands contain a different story of early glacial events within a fore-arc/outer-arc terrestrial setting.

King George Island and the neighbouring Nelson Island consist of several tectonic blocks bounded by two systems of strike–slip faults of Tertiary (54–21 Ma) age (Birkenmajer, 1989). Thus, considerable differences in stratigraphic succession, age and character of the rocks occur between particular blocks. The stratigraphic sequence includes mainly Late Cretaceous to Early Miocene island-arc extrusive and intrusive rocks comprising mainly terrestrial lavas, pyroclastic and volcaniclastic sediments often with terrestrial plant fossils. Hypabyssal dykes and plutons intrude the latter. Fossiliferous marine and glaciomarine sediments are also represented that provide clues to palaeoclimates.

Several sequences of glacial tillites crop out within these complicated sequences, representing glacial and interglacial events. Reports of supposed

Eocene-age tillites at Magda Nunatak (Birkenmajer, 1980a,b), named the Krakow Glaciation and dated at 49 Ma, have been disproved by Sr dating (Dingle and Lavelle, 1998). However, Birkenmajer et al. (2005) describe valley-type tillites between lava sequences in the Eocene–Oligocene Point Thomas Formation in Admiralty Bay. K–Ar dating gives ages of 41–45 Ma for the lavas below the tillites and 45–28 Ma in the lavas above. Birkenmajer et al. thus propose that a glacial period occurred at 45–41 Ma during the Middle Eocene, being the oldest record of alpine glaciers in West Antarctica.

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A clear record of glacial activity is present as diamictites and ice-rafted deposits within the Polonez Cove Formation, of mid-Oligocene age (26–30 Ma) (Troedson and Smellie, 2002). This is called the Krakowiak Glacial Member. At its maximum extent, ice was grounded on a shallow marine shelf. Interestingly, exotic clasts within this sequence may represent ice-rafted debris that was derived from as far away as the Transantarctic Mountains, suggesting marine-based glaciation in the Weddell Sea region. Non-glacial sediments that overlie the Polonez Cove Formation signal an interglacial period in the Late Oligocene (26–24.5 Ma), before another glacial phase in the Miocene. The only other terrestrial evidence for Oligocene ice, possibly representing local alpine glaciation, is from Mount Petras in Marie Byrd Land, West Antarctica, where deposits indicate volcanic eruptions beneath ice (Wilch and McIntosh, 2000).

8.2.3. Eocene Environments in the Ross Sea Region

There is an intriguing record of Eocene pre-glacial environments in the Ross Sea region, around McMurdo Sound. Although the location of the Eocene outcrop is presumably under the ice sheet, there are no major exposures. A glimpse of Eocene environments is, however, provided by several hundred erratic boulders and cobbles recovered from coastal moraines around the shores of Mount Discovery, Brown Peninsula and Minna Bluff (Fig. 8.1).

These fossiliferous glacial erratics (called the McMurdo Erratics), recovered from moraine around the northwest coast of Mount Discovery and Minna Bluff in southern McMurdo Sound, provide a window into the environment that may have existed along the coast of the gradually rising Transantarctic Mountains during the Eocene. The erratics are most likely derived from sub-glacial basins, such as Discovery Deep, that lie along the coast of the Transantarctic Mountains or basement highs situated to the east of the discovery accommodation zone (Wilson, 1999; Wilson et al., 2006). The erratics were distributed into their distinctive pattern of terminal and lateral retreat moraines during relatively recent advance and retreat of grounded ice into southern McMurdo Sound (Wilson, 2000). Subsequent basal adfreezing and surface ablation has transported the erratics to the surface of the McMurdo Ice Shelf. Although currently out of their original stratigraphic position, this suite of erratics provides us with a mechanism to obtain geologic data that are otherwise buried beneath the Antarctic ice sheets and fringing ice shelves.

The McMurdo Erratics comprise a range of lithotypes and ages. Eocene rocks contain a rich suite of fossil floras and faunas including marine and terrestrial palynomorphs, diatoms, ebridians, marine vertebrates and invertebrates, terrestrial plant remains and a bird humerus. Biostratigraphic data from dinoflagellate cyst, ebridian and mollusc assemblages recovered from many of the erratics indicate that the majority of fossiliferous rocks range from Middle to Late Eocene, ~43–34 Ma. Several hundred samples collected between 1993 and 1996 (Stilwell and Feldmann, 2000) include Oligocene, Miocene and Pliocene samples. Dinoflagellate cyst assemblages in post-Eocene erratics comprise few taxa (typically <5 species). In contrast, diatom assemblages in Oligocene and Miocene erratics are relatively rich. This general paucity of dinocyst species in late Palaeogene and Neogene sequences is observed in several other sites from the southern high latitudes (Wilson, 1989; Mao and Mohr, 1995) and may reflect geographical and thermal isolation of Antarctica (McMinn, 1995; Williams et al., 2004).

The majority of the Eocene erratics record a suite of lithofacies that were deposited in coastal–terrestrial to inner shelf marine environments (Levy and Harwood, 2000a,b). These sediments were probably deposited within fan deltas that formed along the rugged coastline of the rapidly rising Transantarctic Mountains. Abundant macroinvertebrate faunas, including bivalves, gastropods, scaphopods, cirripeds, bryzoans, decapods and brachiopods, indicate that many of these sediments were deposited in a spectrum of predominantly shallow marine environments. The presence of terrestrial plant material and palynomorphs also suggests that the majority of the rocks were formed in nearshore environments. However, the occurrence of open marine dinocyst species and the absence of benthic diatom taxa in many of the fine-grained lithofacies indicate that outer shelf open marine environments were also present in the source region. The Eocene erratics contain no direct or unequivocal sedimentological evidence for the presence of ice close to the basins in which the sediments were originally deposited. It is notable that erratics composed of diamictites recovered from the coastal moraines are all Oligocene and younger in age.

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Although rare, fossil leaves, wood and pollen recovered from several erratics provide a glimpse of the Eocene climate for the region. One erratic contains wood and leaves from *Araucaria* and *Nothofagus trees*, which suggests that the climate was not extreme. Cool temperate conditions with some winter snow may have been possible but temperatures were probably not cold enough to allow extensive ice at sea level (Francis, 2000; Pole et al., 2000). Spore and pollen assemblages recovered from the erratics reflect *Nothofagus*-podocarpaceous conifer-Proteacea vegetation with other angiosperms growing in temperate climate conditions (Askin, 2000). Oligocene and younger erratics show a major drop in species richness, which is also noted in sequences recovered in CIROS-1 and the Cape Roberts Project (CRP) cores (Mildenhall, 1989; Raine and Askin, 2000).

Fossil invertebrate remains recovered from the erratics include a humerus shaft from a pseudodontorn (giant bony-toothed sea bird) (Jones, 2000), a probable crocodile tooth (Willis and Stilwell, 2000) and teeth from two species of shark (Long and Stilwell, 2000). The small but significant record of East Antarctic invertebrate fauna indicates a temperate to cool temperate marine environment.

8.2.4. Climate Evidence from Drilling on the Antarctica Margin

The onset of glaciation in Antarctica is not yet well constrained, largely because no cores have yet been obtained that unequivocally provide a continuous transition from no-ice to ice-sheet scale glaciation. This is because (i) most cores terminated before the base of the glacigenic sediments was reached, (ii) a hiatus exists at the base of the glacigenic strata, as in CRP-3 in the Ross Sea (Barrett and Ricci, 2001a,b) and Ocean Drilling Program (ODP) Site 1166 in Prydz Bay (Shipboard Scientific Party, 2001a,b,c,d), or (iii) the age models based on multiple criteria have been revised several times (e.g. CIROS-1; Wilson et al., 1998a). On land, there is evidence for ice proximal-fjordal sedimentation possibly dating back to Oligocene time in the Prince Charles Mountains (Hambrey and McKelvey, 2000a; McKelvey et al., 2001) and some exposures of the Sirius Group in the Transantarctic Mountains could also be this old (Sugden and Denton, 2004).

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8.2.4.1. Drill cores in the western Ross Sea – CIROS-1 and Cape Roberts

Two drill-holes have been recovered from the western Ross Sea that approach or even cross the Eocene–Oligocene boundary (Fig. 8.2). Drilling was undertaken from a sea-ice platform in spring time, and was characterized by exceptionally high recovery (up to 98 per cent). In the 702 m deep CIROS-1 hole (Barrett, 1989), the lower part of the core was

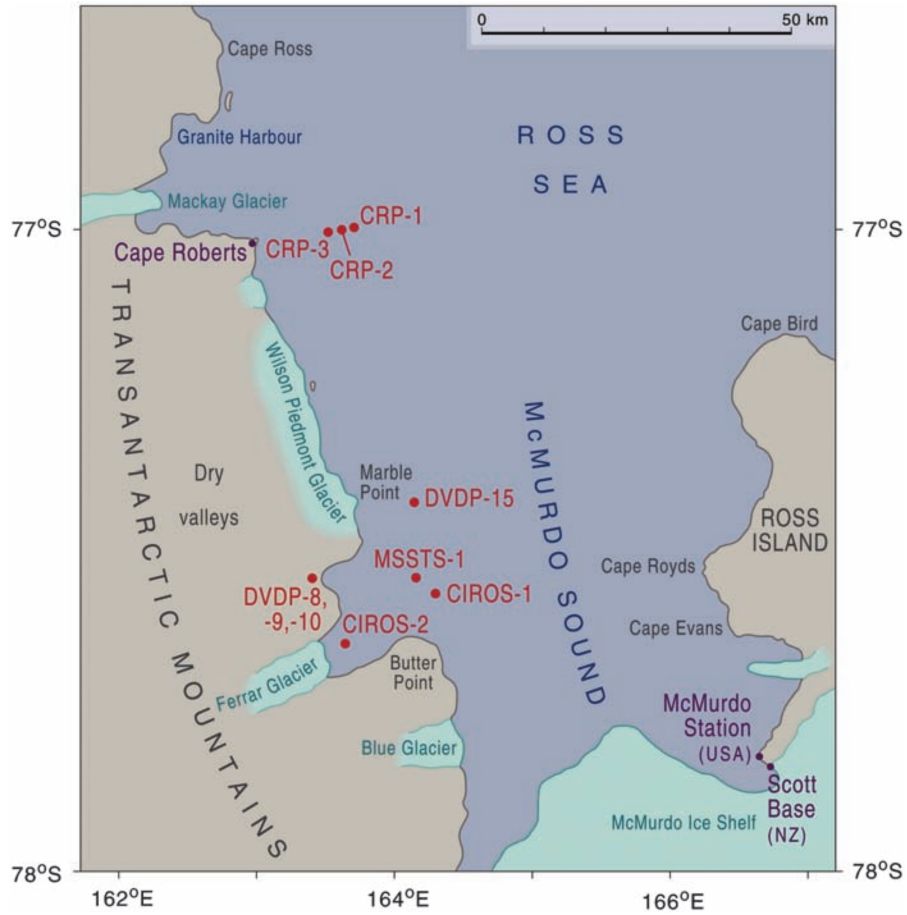

Figure 8.2: Location of the Drill Sites in the Western Ross Sea. CIROS-1 and CRP-3 are those that Bear on the Eocene/Oligocene Question (from Hambrey et al., 2002).

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originally regarded as Late Eocene, with a breccia passing up into mudstone and sandstone. The boundary with the Oligocene was originally placed at about 570 m (Barrett et al., 1989), but magnetobiostratigraphic data (Wilson et al., 1998a) suggest that the Eocene–Oligocene boundary is much higher at about 410–420 m. In either case, there is no obvious lithological transition, these finer grained facies including alternations of weakly stratified sand and mud, with intraformational conglomerate and occasional diamictite. These facies are strongly bioturbated and contain exotic clasts, as well as intraclasts, while some beds are graded. Moving up-core, a major hiatus exists at 366 m, which coincides with the Early/Late Oligocene boundary.

Above is a suite of fining-upwards sandstone beds, followed by alternating massive and stratified diamictites and thin interbeds of sand and mud (Hambrey et al., 1989).

The CIROS-1 core was originally interpreted in terms of depositional setting, ice proximity and water depth (Hambrey et al., 1989). The breccia at the base of the hole is interpreted as a fault brecciated conglomerate. The overlying sandstone/mudstone/diamictite succession is marine, influenced to varying degrees by resedimentation and iceberg rafting. Above the Early/Late Oligocene hiatus, the sandstones were regarded as fluvial, and the diamictites as basal glacial deposits, indicating ice overriding the site. However, Fielding et al. (1997) argued, based on a sequence stratigraphic analysis, that the Late Oligocene diamictite was also glaciomarine. In contrast, Hiemstra (1999) reverted in part to the original view of grounded ice on the basis of microstructural studies. Whichever solution is the correct one, there is no clear evidence for a major environmental shift at the E/O boundary, but there is one towards lower sea level at the Early/Late Oligocene transition.

A record of climate change through the E/O transition has also been determined from the environmental magnetic record in the CIROS-1 core (Sagnotti et al., 1998, 2001). Variations in magnetite were related to the concentration of detrital material into the Victoria Land Basin, influenced by climate and weathering rates on the Antarctica continent (especially of the Ferrar Group). Sagnotti et al. (1998) determined, from changes in the abundance of magnetite, that although there were some cold dry intervals (35–36 and >36.5 Ma) alternating with warm humid climates during the Late Eocene, a stable cold dry climate was not established in Antarctica until the E/O boundary, with major ice-sheet growth occurring at the Early/Late Oligocene boundary. This pattern matches clay mineral history, which shows a shift from smectite-rich to smectite-poor assemblages in Antarctica at the E/O boundary (Ehrmann and Mackensen, 1992; Ehrmann, 1997).

The second core that contains the E/O transition is the Cape Roberts Project Core CRP-3; Barrett (this volume) has synthesized the wealth of data and numerous papers from this and other Cape Roberts cores. The strata were deposited in the same rift basin, the Victoria Land Basin, as CIROS-1. The basin floor comprises Early Devonian sandstone, above which is about 1,500 m of Cenozoic sediment. CRP-3 records a dolerite conglomerate and a basal sandstone breccia at the base of the Cenozic succession, believed to be of latest Eocene age (34 Ma). Above lies nearly 800 m of sandstone with thin beds of conglomerate, all of Early Oligocene age. Diamictite and sandstone occur towards the top of the hole, while outsized clasts are scattered through much of the core. Core CRP-2A almost follows on directly

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above CRP-1 and spans the Late Oligocene/Early Miocene interval. Alternating sequences of mudstone, sandstone, diamictite and conglomerate occur within this part of the Cenozoic record.

The facies in CRP-3 and CRP-2A represent a marine sedimentary record (Barrett, in press). Conglomerate, sandstone and mudstone are typical of the coastal margin of a subsiding sedimentary basin, influenced by iceberg rafting. The diamictite beds additionally record tidewater glaciers that extended periodically beyond the coast. In these respects, the combined Cape Roberts cores represent an expanded record of that from CIROS-1, although there is little evidence of ice-grounding. Nevertheless, the facies fluctuations are thought to reflect glacioeustatic changes in sea level on a wave-dominated coast, in parallel with tidewater glacier fluctuations. In this context, diamicton and sand represent nearshore sedimentation, and this association grades upwards into shelf mud and then to inner shelf/shoreface sand.

A conceptual depositional model for the Late Oligocene/Miocene interval was developed by Powell et al. (2000), based on facies associations and comparison with modern glaciomarine environments, such as those in Alaska and Greenland. This shows that during an advance and still-stand, a grounding-line fan develops, and this is followed by rapid recession until another fan develops. The sequence becomes even more complex when the glacier overrides previously formed fans. Figure 8.3 is a simplified version of this model.

The huge volume of palaeoecological, mineralogical and geochemical data generated by these drilling projects enables us to gain insight into the broader environmental and climatic evolution through the early Cenozoic (Hambrey et al., 2002; Barrett, in press). A temperate glacier regime is suggested for the Early Oligocene, following by cooling into the Miocene, typified by polythermal glaciers. The cold frigid regime of today only began at the end of Pliocene time. The Early Oligocene landscape was characterized by temperate glaciers flowing from the early East Antarctic ice sheet, some terminating in the sea, and others terminating on braided outwash plains (Fig. 8.4).

8.2.5. The Prydz Bay Region

Drilling in Prydz Bay was undertaken by two ship-borne legs of the ODP (Fig. 8.5). In contrast to the western Ross Sea cores, core-recovery rates here were much less satisfactory, hence interpreting depositional processes is more questionable. Nevertheless, plausible scenarios have been derived, albeit lacking precise constraints owing to core loss. Prydz Bay represents the continuation of the Lambert Graben, which contains the Lambert

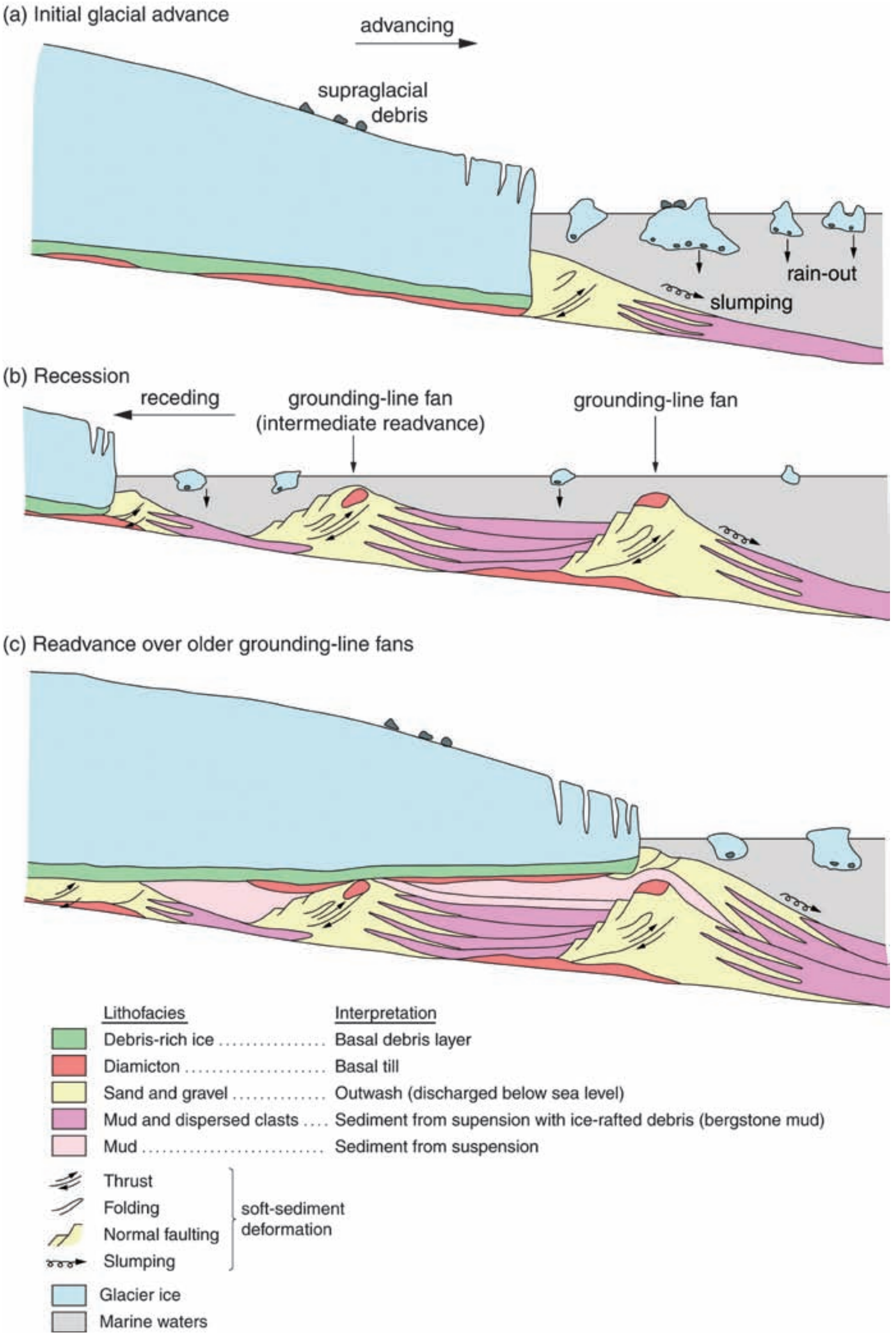

Figure 8.3: Grounding-Line Fan Model of Glaciomarine Sedimentation for Late Oligocene/Early Miocene Time (from Hambrey et al., 2002; Simplified from Powell et al., 2000).

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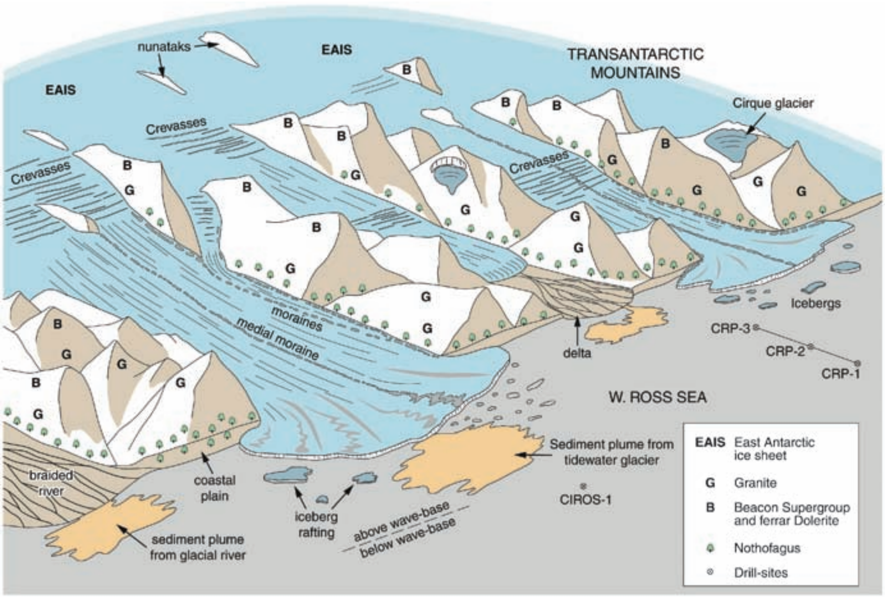

Figure 8.4: Cartoon depicting the Victoria Land Coast in Early/Late Oligocene Time, with Glacier- and River-influenced Coast, and Vegetated Mountainsides and Lowlands. This Scenario is based on a Combination of Sedimentological Evidence and Floral Data from CIROS-1 and Cape Roberts Cores (from Hambrey et al., 2002).

Glacier–Amery Ice Shelf System, an ice drainage basin covering approximately 1 M km^2, draining 13 per cent of the East Antarctic Ice Sheet by surface area. Thus, the record in Prydz Bay provides a signal of the ice sheet as a whole since its inception, and complements the Oligocene (?) to Pliocene uplifted glaciomarine record in the Prince Charles Mountains (see Haywood et al., this volume). Prydz Bay itself is dominated by a trough-mouth fan that prograded during phases of glacier advance to the shelf break. Like the western Ross Sea, large data-sets are available covering all aspects of core analysis from ODP Legs 119 and 188 (Barron et al., 1991; Cooper and O'Brien, 2004; Cooper et al., 2004) and a convenient summary has been provided by Whitehead et al. (2006).

ODP Leg 119 obtained two cores, 739 (480 m) and 742 (316 m), the lower parts of which were loosely dated as Middle Eocene to Early Oligocene. The dominant facies recovered was massive diamictite, with minor stratified diamictite and sand (Hambrey et al., 1991, 1992). Poorly consolidated fine-grained facies may well have been washed away during the drilling, however, since core-recovery rates were less than 50 per cent. A few broken shell

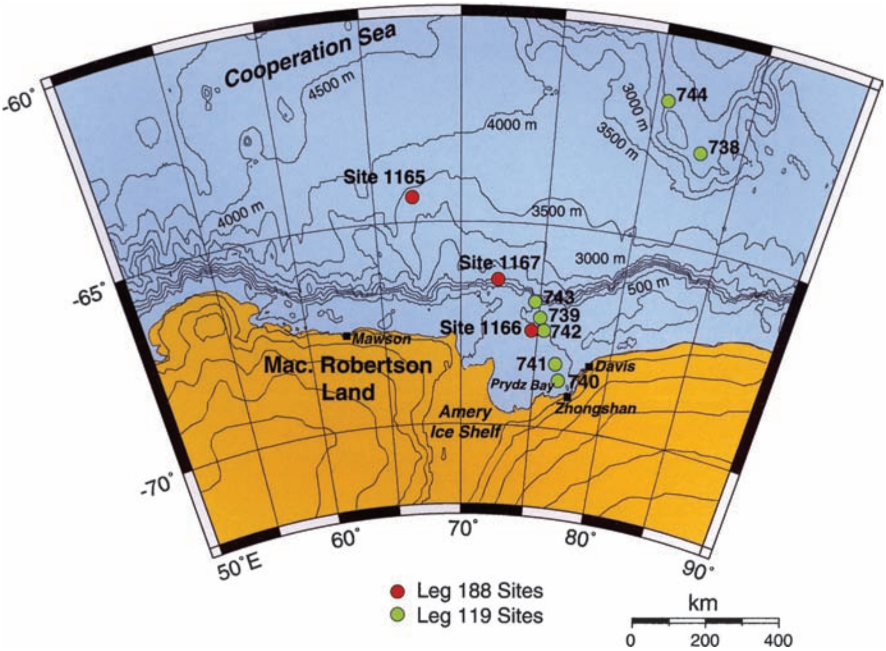

Figure 8.5: Location of Drill Sites in Prydz Bay, East Antarctica, from ODP Legs 119 and 188. Continental Shelf Sites 742 and 1166 include Strata the cross the Eocene/Oligocene Transition. From Shipboard Scientific Party (2001a,b,c,d).

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fragments are present, but there is a dearth of material suitable for precise dating. The base of Site 742 is represented by a zone of soft-sediment deformation. The Oligocene succession forms part of a prograding unit as defined in seismic profiles, but is truncated by a regional unconformity. Above lies a flat-lying sequence of more diamictite, some with preferred clast orientation and overcompaction, of late Miocene to Pliocene age (Cooper et al., 1991).

Leg 188 drilled Site 1166 on the continental shelf near Sites 739 and 742 in order to obtain a more complete record of the E/O transition, but again core recovery was poor (19 per cent). From the base upwards, Late Eocene matrix-supported sand was followed by a transgressive surface and the Late Eocene to Early Oligocene graded sand and diatom-bearing claystone with dispersed granules. These facies were capped, above an unconformity, by ‘‘clast-rich clayey sandy silt’’ (diamicton/ite) of Neogene age (Shipboard Scientific Party, 2001a,b,c,d).

The interpretation of the Leg 119 facies is as follows: the deformed bed at the base of Site 742 may represent the first stages of glaciation, with the ancestral Lambert Glacier extending across the continental shelf for the first time. Then the bulk of the recovered facies in Sites 739 and 742 (diamictite) records deposition from the grounding-line of a tidewater glacier margin, by debris rain-out and submarine sediment gravity flow beyond the shelf break, conditions which characterize much of Early Oligocene time. The overlying Miocene succession is quite different, and represents successive advances across the already prograded shelf to produce sub-glacial, ice-proximal and ice-distal facies in alternation, forming the flat-lying sequence (Hambrey et al., 1991). Leg 188's Site 1166 begins at the base with Late Eocene sand of fluvio-deltaic character, and is inferred to be pre-glacial. The overlying Late Eocene to Early Oligocene sand and claystone represent shallow marine and open marine conditions, respectively, but in a proglacial setting as indicated by ice-rafted granule-sized material. The Neogene strata that lie unconformably above represent full glacial conditions.

Combining the sedimentary and seismic records, along with bathymetric data from the over-deepened Lambert Graben, a conceptual model of erosion and deposition has been developed (Fig. 8.6). Considering the whole Prydz Bay region in 3D, it is apparent that Early Oligocene time was characterized by sedimentation at the shelf break. A major change then took place in the Late Miocene to mid-Pliocene following the development of a fast-flowing ice stream that now occupied a more constrained channel, leading to the growth of the trough-mouth fan (Shipboard Scientific Party, 2001a,b,c,d; Taylor et al., 2004).

Comparing data-sets from the western Ross Sea and Prydz Bay, it is apparent that ice first reached the continental shelf edge in earliest Oligocene time in Prydz Bay, while the Victoria Land coast was influenced by iceberg rafting. Late Oligocene time saw repeated ice expansions recorded in the CIROS-1 core by a succession of basal tills, and probably by the major unconformity in Prydz Bay. Subsequently, frequent expansions took place to the shelf edges in both areas, resulting in a fragmented stratigraphic record.

8.3. Climate Signals from the Terrestrial Realm – Fossil Plants and Palynomorphs

8.3.1. Plant Macrofossils

An important source of palaeoclimate information on land comes from fossil plants (both macro- and microfossils) (Fig. 8.7). The fossil plant record suggests that during the Late Palaeocene to Early Eocene, moist, cool

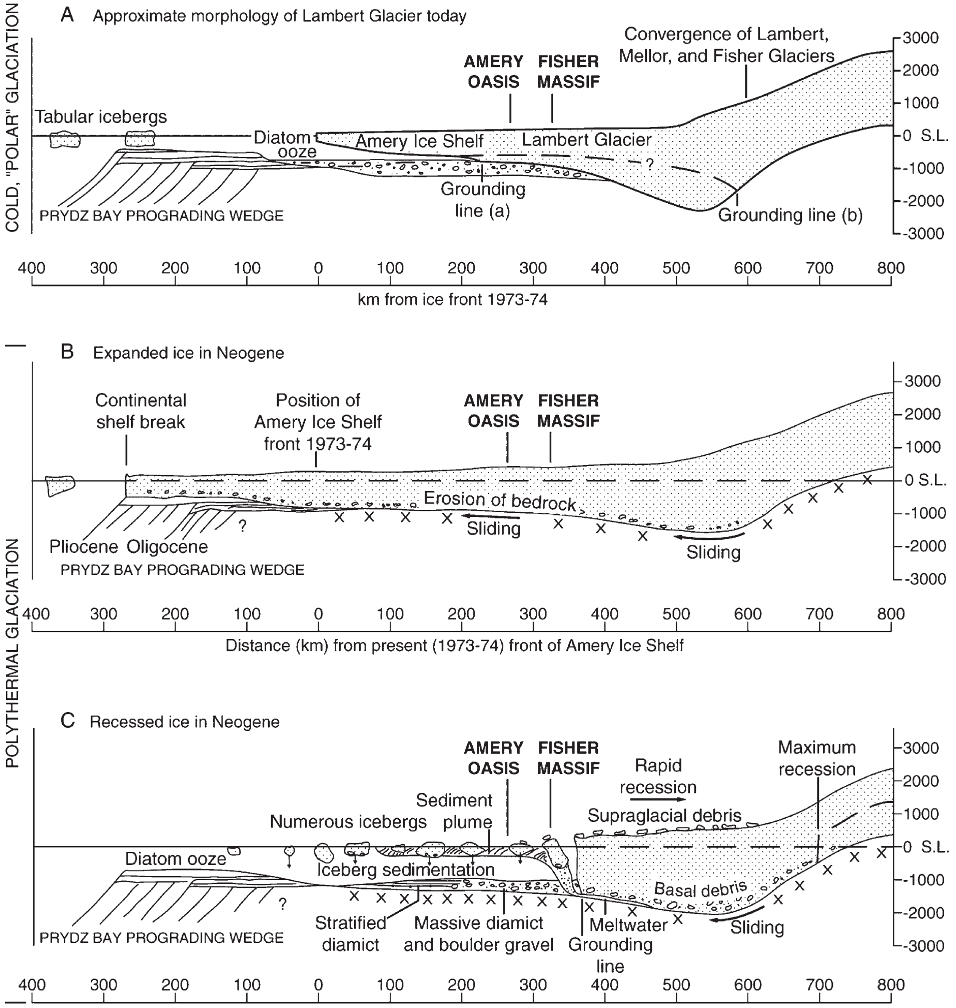

Figure 8.6: Conceptual Model of Styles of Glacial Sedimentation during Phases of Advance and Recession of the Ancestral Lambert Glacier (from Hambrey and McKelvey, 2000b). (A) "Cold Polar" Glaciation as occurs Today. (B) Expanded Ice in the Neogene Period. (C) Recessed Ice in the Neogene Period. Both B and C are characterized by Polythermal Glaciation (Comparable to the High Arctic today), and may also have been a Feature of the Palaeogene Period; However, Data are too Limited to Define the Palaeoclimate for This Earlier Period. AU:7

Stratigraphic age	***Flora***	***Composition***	***Vegetation model***	***Modern analogue***	***Inferred palaeoclimate***
King George Island, South Shetland Islands:					
Late Eocene	Cytadela/Platt Cliffs	Ferns *Nothofagus*-type (mostly small) and other dicotyledon types ?Podocarpaceae	*Nothofagus* forest with well-developed undergrowth, where ferns (including tree ferns) are important	Recent fern-bush communities of southern oceanic islands, e.g. Gough and Auckland Islands	Warm (frost-free, MAT 11.7-15°C) Moist (MAP 1220-3225mm)[1]
	Petrified Forest Creek	*Fagus-Nothofagus* Araucaria			
Late Palaeocene-Middle Eocene	Mount Wawel	*Equisetum* Ferns Several *Nothofagus* spp. (microphyllous leaves) and a few other angiosperms Podocarpaceae	*Nothofagus-Podocarpus* dominated communities	Patagonian-Magellanic rainforests, Chile	Cool temperate (MAT 5-8°C) Moist (600-4300mm)
	Dragon Glacier Moraine	*Equisetum* Ferns Angiosperms (dominated by *Nothofagus* spp.) Conifers (Araucariaceae, Cupressaceae, Podocarpaceae)	Temperate rainforest	Valdivian rainforests, Chile	MAT *c.*10°C MAP *c.*1000mm

[1] Although the small leaves suggest this may be too warm (Francis, 1999).

Figure 8.7: Antarctic Peninsula Late Palaeocene–Late Eocene Fossil Floras: Composition, Modern Analogue Vegetation and Palaeoclimate Interpretations. Refer to Text for Relevant References. MAT: Mean Annual Temperature; MAP: Mean Annual Precipitation.

Stratigraphic age	***Flora***	***Composition***	***Vegetation model***	***Modern analogue***	***Inferred palaeoclimate***
Late Palaeocene-Middle Eocene (continued)	Fossil Hill	40 taxa: Ferns Mixed broadleaf angiosperms (dominated by *Nothofagus* spp.) Conifers (podocarp, araucarian, cupressacean)	Rainforest – mixed neotropical and subantarctic elements[2]	Tropical Latin America and southern South American rainforests	Temperate – warm temperate[3] Slightly above 10 C with a low annual temperature range (MAT *c*.9 C), Abundant precipitation (*c*.1000mm)
	Collins Glacier	Ferns Angiosperms (including *Nothofagoxylon* spp., *Weinmannioxylon eucryphioides*, *Myceugenelloxylon antarcticus* Conifers (*Araucaria, Cupressinoxylon*, *Podocarpoxylon fildesense*)	Cool temperate rainforest	Low-mid altitude Valdivian rainforests, Chile	Cool temperate
Seymour Island, James Ross Basin:					
Early-latest Late Eocene	La Meseta Fm.	Ferns, angiosperms and conifers including: *Nothofagus* spp. (notophyllous at older Middle Eocene locality), Dilleniaceae, Myricaceae, Myrtaceae, Lauraceae, Proteaceae, Podocarpaceae, Araucariaceae	Cool – cold temperate rainforest	Valdivian and Magellanic rainforests, Chile	Cool to cold temperate (MAT 11-13°C or 10.8°C) with abundant rainfall (MAP 1000-3000mm) Seasonal.

[2] Perhaps altitude related?

[3] The *Nothofagus* leaves in the Fossil Hill flora are much bigger than their modern relatives suggesting a warmer and more humid climate during the Middle Eocene.

Figure 8.7: (*Continued*).

temperate rainforests were present, similar to modern low to mid-altitude Valdivian rainforests in southern Chile. These forests were dominated by *Nothofagus* and conifer trees, with ferns, horsetails and some less-prominent angiosperm groups. Assemblages of Eocene fossil plants and palynomorphs signal a generally warm climate during the Early Eocene but conditions deteriorated throughout the Antarctic Peninsula through the latter part of the Eocene when cold, seasonal climates developed (Francis, 1991; Francis and Poole, 2002; Francis et al., 2004a,b, 2008; Poole et al., 2005).

Fossil plant assemblages of younger age have been found in glacial sediments in the Transantarctic Mountains, representing vegetation that grew on Antarctica under icehouse conditions. These include the Sirius Group flora from the Transantarctic Mountains (the date of which is problematic, Francis and Hill, 1996; Wilson et al., 1998a,b; Askin et al., 1999), and a new flora of Miocene age discovered in the Dry Valleys (Ashworth et al., 2008). Surprisingly, leaf fossils, all be they single leaves, have also been discovered in drill cores within Oligocene and Miocene glacial sequences. The early glacial world of Antarctica was clearly not totally barren of vegetation.

AU :21

Palaeogene fossil plants have been discovered from around the Antarctic margin in outcrop, in sea-floor cores and in glacial erratic boulders from the Antarctic Peninsula (King George Island and Seymour Island) and Ross Sea (McMurdo Sound and Minna Bluff) regions. The collections consist of compressions, impressions and petrifactions of leaves, seeds, flowers and wood indicating a high southern latitude flora of variable diversity but dominated by fossils comparable to modern *Nothofagus* (the southern beech) and conifer trees. Ferns, horsetails and additional significant Southern Hemisphere angiosperm families, including the Proteaceae, Myrtaceae and Lauraceae, are also represented. The macrofossil record has largely been described from isolated collections, even of single leaves in cores, and although reasonably dated as a whole, a comprehensive understanding of stratigraphic relationships between the floras, particularly in the Antarctic Peninsula region, is at present hampered by differing stratigraphic interpretations.

8.3.2. Antarctic Peninsula: King George Island

Many macrofloras have been discovered on King George Island in the South Shetland Island group, north of the Bransfield Strait in the Antarctic Peninsula region, currently dated between Late Palaeocene and Late Eocene in age. The floras may have lived at a palaeolatitude of ~62°S, similar to its present-day location (Lawver et al., 1992). The stratigraphy is complex AU :22 because Birkenmajer (1981, 1989, 1990) and Birkenmajer et al. (1986) erected many local formations in comparison to a simpler scheme created by Smellie et al. (1984). No single stratigraphic scheme exists and so the relationship between the floras is confusing. The stratigraphic framework used here includes both schemes (also reviewed by Hunt, 2001). Leaf macrofloras, currently understood to be of Late Palaeocene to Middle Eocene in age, have been described in varying completeness from the Admiralty Bay and Fildes Peninsula areas of the island.

In the Admiralty Bay area, the Middle Eocene Mt. Wawel Formation (Point Hennequin Group) contains the macroflora deposits collectively known as the Point Hennequin Flora with individual localities named Mount Wawel and Dragon Glacier Moraine floras (Zastawniak et al., 1985; Birkenmajer and Zastawniak, 1989a; Askin, 1992; Hunt, 2001; Hunt and Poole, 2003). The Mount Wawel flora comprises macrofossils of *Equisetum* AU :23 (horsetail), ferns and several *Nothofagus* species as microphyllous leaves (a leaf-size category of 2.5–7.5 cm long), in addition to a few other angiosperms and Podocarpaceae. The Dragon Glacier Moraine flora is similar, the angiosperm leaves being dominated by *Nothofagus* and the conifers including Araucariaceae and Cupressaceae, in addition to Podocarpaceae. The Middle Eocene Petrified Forest Creek flora from the Arctowski Cove Formation and the Late Eocene Cytadela flora from the Point Thomas Formation are both within the Ezcurra Inlet Group. The former is a wood flora requiring revision, but intermediate *Fagus*–*Nothofagus*-type species are recorded. The Cytadela leaf flora includes ferns (including a *Blechnum*-affinity species), mostly small *Nothofagus*-type leaves with pinnately veined leaves of other dicotyledonous types and possible Podocarpaceae (Birkenmajer and Zastawniak, 1989a; Askin, 1992; Birkenmajer, 1997). Birkenmajer and Zastawniak (1989a) considered this flora to be Eocene–Oligocene boundary in age.

In this region, therefore, *Nothfagus*-dominated forests were the norm in the Middle to Late Eocene with ferns and tree ferns becoming increasingly important. Estimated mean annual temperatures of 5–8°C are slightly cooler than those on Seymour Island to the east during the Middle Eocene, and the vegetation was similar to the southernmost Patagonian–Magellanic forests of southern Chile (Zastawniak et al., 1985; Birkenmajer and Zastawniak, 1989a; Askin, 1992; Hunt, 2001). By the Late Eocene, vegetation was more comparable to the recent fern-bush communities of southern oceanic islands (e.g. the Auckland Islands), interpreted from the Cytadela and Petrified Forest Creek floras (Birkenmajer and Zastawniak, 1989a; Askin, 1992; Birkenmajer, 1997). However, mean annual temperature estimates of

11.7–15°C appear too high, especially considering the small-sized leaves (Francis, 1999).

In the Fildes Peninsula area in the southwest of King George Island, the Fildes Peninsula Group contains the contemporary Middle Eocene Collins Glacier and Rocky Cove floras within the Fildes Formation, and the diverse Late Palaeocene–Middle Eocene Fossil Hill flora (Fossil Hill Formation). The latter is a leaf flora containing 40 recognized taxa, including mixed broadleaf angiosperms (with large-leaved *Nothofagus* species), conifers (podocarp, araucarian and cupressacean) and ferns (Birkenmajer and Zastawniak, 1989a,b; Li, 1992; Haomin, 1994; Francis, 1999; Reguero et al., 2002). Neotropical and sub-Antarctic elements appear to be mixed perhaps indicating a collection derived from communities at different altitudes (Li, 1992), although this mixed signature may be a feature of early Tertiary polar vegetation (Francis et al., 2004a,b). The *Nothofagus* leaves are much larger than their modern relatives, suggesting a warmer and more humid climate during the early part of the Eocene. Estimates of mean annual temperature suggest $>10\,^{\circ}\mathrm{C}$ (from 40 per cent entire-margined leaves) and a small annual temperature range (Li, 1992).

Fossil leaves remain undescribed from the Rocky Cove flora; however, wood from this locality has been identified as *Nothofagoxylon antarcticus* (Shen, 1994; Hunt, 2001). The Collins Glacier deposit is primarily a wood flora that includes wood of both coniferous (*Cupressinoxylon* sp. and *Podocarpoxylon fildesense*) and angiospermous [*Nothofagoxylon* spp., *Weinmannioxylon eucryphioides* (Cunoniaceae) and *Myceugenelloxylon antarcticus* (Myrtaceae)] affinity (Hunt, 2001; Poole et al., 2001, 2005). The mean annual temperature had dropped to 9°C by the Middle Eocene and precipitation appears to have increased; however, the latter is probably due to changes in the environmental setting rather than climate change because of no immediate change in angiosperm wood from a semi-ring to ring porous condition (Poole et al., 2005).

8.3.3. Antarctic Peninsula: Seymour Island

The Late Eocene La Meseta Formation of the Seymour Island contains floras that document climate change. Leaves and wood of both angiosperm and conifer affinity occur with fern fossils and a flower (Case, 1988; Reguero et al., 2002; Francis et al., 2004a,b). Middle Eocene *Nothofagus* leaves were found to be notophyllous (a leaf-size category 7.5–12.5 cm long). Angiosperm fossils affiliated to families including Nothofagaceae, Dilleniaceae, Myricaceae, Myrtaceae, Elaeocarpaceae, Moraceae, Cunoniaceae, Winteraceae and Lauraceae have been described (Gandolfo et al., 1998a,b; Reguero et al., 2002; Francis et al., 2004a,b). Doktor et al. (1996) also described leaves affiliated with Podocarpaceae, Araucariaceae, Nothofagaceae and Proteaceae. Gothan (1908) was the first to describe fossil wood from Seymour Island, which has subsequently been re-examined by several authors and identified as having both angiosperm and coniferous affinities (Francis, 1991; Torres et al., 1994; Brea, 1996, 1998; Francis and Poole, 2002; Reguero et al., 2002).

Decrease in leaf sizes during the Late Eocene suggests that climate deteriorated towards the end of the Eocene, as observed in studies of the La Meseta Formation by Case (1988) and Reguero et al. (2002). Gandolfo et al. (1998a,b) suggested a MAT of 11–13°C for the Cucullaea I Allomember; early Late Eocene. Further climate data were provided by leaf margin analyses of a Late Palaeocene flora (Cross Valley Formation) and of the early Late Eocene Cucullaea 1 flora, which indicate a decrease in floral diversity and a change from mean annual temperatures of 14°C in the Late Palaeocene to 11°C in the early Late Eocene, with signs of freezing winters in the Late Eocene (Francis et al., 2004a,b).

8.3.4. Ross Sea Region: McMurdo Sound

Only two E/O floras have been described from East Antarctica – a significant, but *ex situ*, Middle–Late Eocene flora within glacial erratic boulders found at Minna Bluff (mentioned above), and a single *Nothofagus* leaf within the Early Oligocene strata of the CRP-3 core in McMurdo Sound (Cantrill, 2001; Florindo et al., 2005). An additional single *Nothofagus* leaf was also found in the CIROS-1 core, originally thought to be Oligocene in age (Hill, 1989), but after recent refinement of the age model, it is now considered Early Miocene (Roberts et al., 2003). The McMurdo erratic flora comprises leaves, wood, seeds and *Araucaria*-type scale-leaves (Pole et al., 2000). Two species of *Nothofagus* leaves were found, some of which are interpreted as deciduous based on the interpretation of plicate vernation (venation with distinct folds). Fossil wood was identified as *Araucarioxylon*, *Phyllocladoxylon* and *Nothofagoxylon* (Francis, 1999, 2000). The CRP-3 *Nothofagus* leaf (compared to *Nothofagus beardmorensis* from the Sirius Group; Francis and Hill, 1996; Cantrill, 2001) was small and also had plicate vernation, but differs to the leaf in the CIROS-1 core thought to be of *N. gunnii* affinity, an alpine species from Tasmania (Hill, 1989; Francis, 1999).

The floras in the erratics indicate the presence of large forest trees comparable to those in South American araucarian (emergent) – *Nothofagus* forests in the Valdivian Andes of Chile or the *Phyllocladus*- and *Nothofagus*-dominated cool, sclerophyll forests of temperate New Zealand and Tasmania. Francis (1999, 2000) suggested a cool temperate climate, with a mean annual temperature of <13°C from the wood flora, considering some winter snow likely, but temperatures not cold enough to allow extensive ice to form at sea level. Single *Nothofagus* leaves found in the CRP-3 and CIROS-1 cores, from Early Oligocene and Early Miocene intervals, respectively, indicate, in conjunction with the palynomorphs, a cold temperate and periglacial climate at those times. The CRP-3 Early Oligocene vegetation is compared to low *Nothofagus* woodland in the Magellanic region of southern Chile (Francis and Hill, 1996; Cantrill, 2001). Cantrill (2001) compared it to *N. beardmorensis*, known from the Sirius Group and considered to have a minimum requirement of −22°C (which is probably conservative) with several weeks at least 5°C during the growing season. The Early Miocene *Nothofagus* leaf is different and may have come from a small- to medium-sized tree in a sub-alpine rainforest or shrub community probably <1 m tall and living in exposed conditions (Hill, 1989; Francis, 1999).

8.3.5. Environmental Signals from Palynomorphs

While macrofossils of plants, found usually in distinct levels in geological sections, allow a fair understanding of the biodiversity and ecology of fossil vegetation at specific times, palynological assemblages can deliver a higher resolution picture of vegetation and climate change through time, especially due to their presence in drill cores. In addition, pollen and spores can be recovered from areas in which macrofossils are unknown (e.g. Prydz Bay). Palynological studies have been undertaken on various Palaeogene sections in West and East Antarctica, but complete recovery is rare, especially over the crucial time interval spanning the E/O boundary when climate dramatically deteriorated.

Cranwell (1959) carried out the earliest palynological studies in the Antarctic realm on a single sample of probable Palaeogene age from Seymour Island in the Antarctic Peninsula area. Subsequently, several early Tertiary stratigraphic sections on Seymour, Cockburn and King George Islands, and cores from the South Orkney Islands (ODP Leg 113, Site 696) and South Scotia Ridge (Bruce Bank, Eltanin Core IO 1578-59) have been subjected to detailed palynological analyses (Mohr, 1990; Askin, 1991, 1997; Grube and Mohr, 2004). Data used to reconstruct Tertiary vegetation and climate in East Antarctica have been derived from drilling campaigns in southern McMurdo Sound in the Ross Sea (CIROS-1, CRP-2/2A and CRP 3: Mildenhall, 1989; Askin and Raine, 2000; Raine and Askin, 2001; Prebble et al., 2006) and in Prydz Bay (MacPhail and Truswell, 2004). ODP Leg 189, in the Tasman Sea, cored the E/O boundary (Grube and Mohr, 2008). The glacial erratics of Eocene-aged sediments in southern McMurdo Sound have also yielded terrestrial palynomorphs.

Pollen assemblages in Antarctic Palaeogene strata contain many taxa comparable to those still found today in southern high latitudes, including areas of southern South America, Tasmania, Australia, New Zealand and New Caledonia. They vary quite substantially in the abundance of their major components, and consist of moss and fern spores, gymnosperm and angiosperm pollen. Ferns were species-rich until the Middle to early Late Eocene (Mohr, 2001) and include genera that live today under humid subtropical conditions, such as *Cnemidaria* (Mohr and Lazarus, 1994). At Bruce Bank (c. 46–44 Ma), fern spores dominated the assemblage, at intervals comprising more than 50 per cent of the sporomorphs. During the Late Eocene, and even more so during the Oligocene, fern diversity and abundance dropped dramatically. During the Early Miocene, some of the taxa seem to return, but recycling of older Eocene spores cannot be completely ruled out, a problem prevalent in glaciogene sediments all around the Antarctic.

Gymnosperms were important components of the southern high latitude Palaeogene forests. Cycads seem to have been present until the Middle Eocene (*Cycadopites*), while Araucariaceae (*Araucaria*), Cupressaceae and especially Podocarpaceae pollen (*Podocarpus*, *Phyllocladus*, *Lagarostrobus*, *Dacrydium* and *Microcachrys*) play a major role in the palyno-associations well into the Oligocene.

Angiosperms were relatively species-rich in the Palaeogene Antarctic pollen spectra with a dominance of various types of *Nothofagidites* (pollen comparable to that of extant *Nothofagus*, the southern beech), but diversity declined during the Late Eocene. Middle to Late Eocene assemblages from the Antarctic Peninsula area (ODP Leg 113, Site 696) show, in earliest sections, relatively large amounts of angiosperm pollen with a clear dominance of *Nothofagidites*, while in assemblages of latest Eocene and Early Oligocene age, moss spores become more common, and, except for *Nothofagidites*, almost no angiosperms are registered. The McMurdo Sound cores (CIROS-1, CRP-2/2A and CRP-3, Late Eocene to Oligocene) and glacial erratics (Middle to Late Eocene) are characterized by prominent *Nothofagidites*, particularly *Nothofagidites lachlaniae* and the *N. fusca* group. Various *Podocarpus* taxa are also abundant (Askin, 2000; Raine and Askin, 2001). In a relatively short sequence of CRP-3 (glaciomarine cycle 26) of Early Oligocene age, *N. fusca*-type, *N. flemingii* and *N. lachlaniae* contribute each about 23 per cent of the total count (Prebble et al., 2006). In glaciomarine cycle 11 of Late Oligocene age, *N. fusca*-type pollen dominates with about 50 per cent, followed by *N. flemingii* and *N. lachlaniae*. In Prydz Bay sections dated Late Eocene, *Nothofagidites* is clearly dominant at 41–57 per cent of sporomorphs; the second largest group are conifer pollen that reach in a few samples up to 50 per cent and more. Fern spores comprise 6–20 per cent and cryptogams 3–6 per cent (Macphail and Truswell, 2004) (Figs. 8.8 and 8.9).

During the warmer periods of the Palaeogene, the following families have been identified from pollen: Aquifoliaceae (includes holly), Casuarinaceae (she-oak), Cunoniaceae/Elaeocarpaceae, Epacridaceae (southern heath, now included within Ericaceae), Euphorbiaceae (spurge), Gunneraceae, Liliaceae, Myrtaceae, Nothofagaceae with the four subgenera (*N. brassi*, *N. fusca*, *N. menziesii* and *N. antarctica*), Olacaceae, Proteaceae (*Gevuina*/*Hicksbeachia*, *Adenanthos*, *Carnarvonia*, *Telopea* and *Beauprea*; Dettmann and Jarzen, 1991), Restionaceae (rush), Sapindaceae (soapberry) and Trimeniaceae (Prebble et al., 2006). In Prydz Bay sections (Late Eocene) and within the La Meseta Formation on Seymour Island (Eocene), *Fischeripollis* and

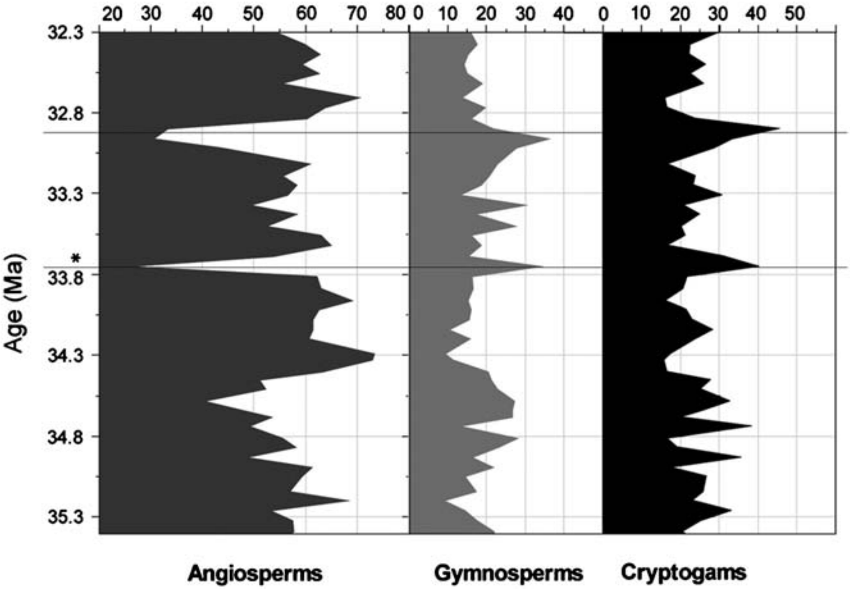

Figure 8.8: Relative Abundance (Per Cent) of Major Plant Groups in ODP Leg 189, Site 1168. Asterisk: Eocene/Oligocene Boundary.

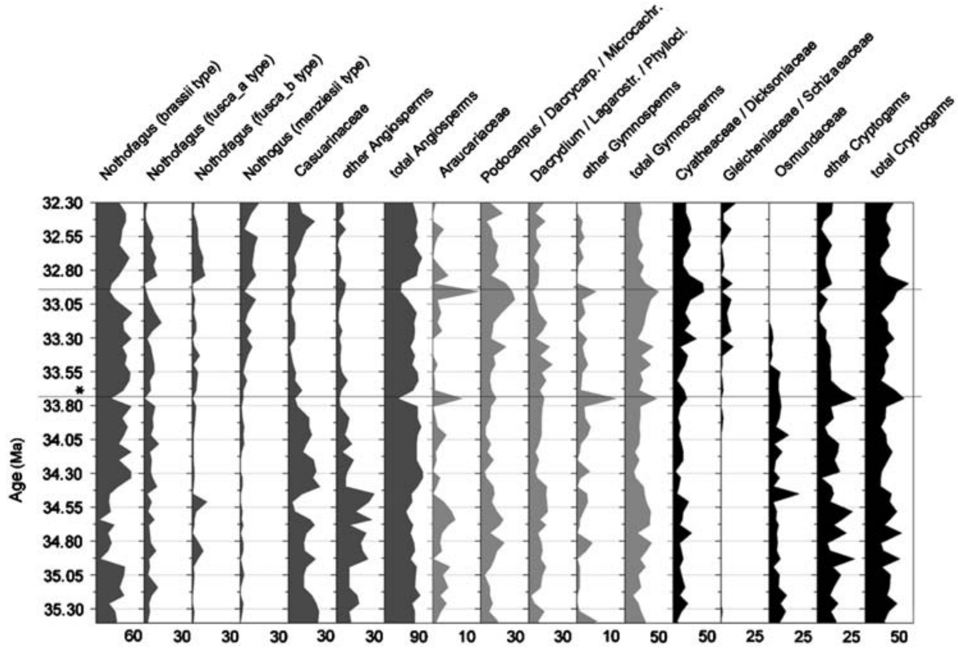

Figure 8.9: Relative Abundance (Per Cent) of Sporomorphs in the Tasman Sea Samples from ODP Leg 189, Site 1168. Asterisk marks Eocene/Oligocene Boundary; Lines represent Palynological "Events".

Droseridites, that belong to Droseraceae (sundew), which are today restricted to moors or damp sites, are excellent ecological markers.

During the Late Eocene to Oligocene, Apiaceae, Asteraceae (daisy), possibly Campanulaceae (bellflower), Caryophyllaceae (carnation), Chenopodiaceae (now in the Amaranthaceae), Onagraceae (willowherb; *Corsinipollenites*) and perhaps Gramineae (grasses) seem to play a role as members of the local vegetation (Mildenhall, 1989; Mohr, 1990; Askin, 1997; Askin and Raine, 2000; Raine and Askin, 2001; Prebble et al., 2006). During the Early Oligocene, a low shrub or closed *Nothofagus*-podocarp forest of small stature may have developed, occupying warmer sites on the Antarctic continent (Prebble et al., 2006). In colder phases, a tundra-like vegetation, evidenced by moss spores, few but relatively diverse herb pollen and a few *Nothofagidites* pollen, derived possibly from dwarfed southern beech, may have grown near the coast.

Palynological studies by Grube and Mohr (2008) of cores from ODP Leg 189, Site 1168 in the Tasman Sea show an intriguing response (or lack of) to E/O climate change. The abundance-time-chart (Figs. 8.8 and 8.9) for the

Tasman Sea samples shows that during the latest Eocene, the pollen flora was dominated by the Nothofagaceae (especially the evergreen type *Brassospora*), with araucarian and podocarp conifers (gymnosperms) and typical fern families (cryptograms). Near the E/O boundary itself, there is a short peak in the occurrence of araucarian and some other gymnosperm pollen, as well as an increase in ferns, in response to a decline in Nothofagaceae. Surprisingly, however, there was no sustained change in terrestrial pollen after this that might reflect a major change in climatic regime. Vegetation typical of latest Eocene composition seems to have been restored during the earliest Oligocene, the only significant changes being the decrease in Casuarinaceae angiosperms, and the gradual replacement of Osmundaceae ferns by Schizaeaceae and Gleicheniaceae. There is a slight decline in angiosperm pollen, which Grube and Mohr (2008) interpret as a gradual response to long-term cooling.

The pollen record in Figs. 8.8 and 8.9 also highlights a further short-live episode of vegetational change at about 32.9 Ma. The pattern is similar to that at 33.7 Ma, with an increase in araucarian conifer pollen and in fern spores at the expense of the angiosperms, especially the Nothofagaceae – does this represent a later episode of cold climate? The pollen diagram in Fig. 8.9 also hints at cyclical changes, possibly at intervals of 0.8 m.y or even 0.4 m.y.

8.4. Environmental Changes Documented by Marine Microfossils

Early Palaeogene marine microfossil associations from the circum-Antarctic realm are typically characterized by the dominance of largely endemic Antarctic organic walled dinoflagellate cysts (dinocysts) and siliceous groups like diatoms and radiolarians over calcareous microfossils (e.g. Brinkhuis et al., 2003a,b; Stickley et al., 2004; Warnaar, 2006). The characteristic endemic dinocyst assemblages reported from Middle and Late Eocene deposits around Antarctica are often referred as the "Transantarctic Flora" (cf. Wrenn and Beckmann, 1982). In fact, Late Cretaceous to (early) Palaeogene organic walled dinoflagellate cysts (dinocysts) from the circum-Antarctic realm are comparatively well known, notably from southern South America, from James Ross and Seymour islands, but also from southeastern Australia and New Zealand, from erratics along the Antarctic margin, from the Ross Sea continental shelf (CIROS; CRP) and from several ocean drill sites (see, e.g. Haskell and Wilson, 1975; Wilson, 1985, 1988; Askin, 1988a,b; Wrenn and Hart, 1988; papers in Duane et al., 1992; Pirrie et al., 1992;

Mao and Mohr, 1995; Hannah, 1997; Truswell, 1997; Hannah et al., 2000; Levy and Harwood, 2000a,b; Guerstein et al., 2002).

Meaningful chronostratigraphic calibration of (sub-)Antarctic dinocyst events was a classic problem due to the general absence of other age-indicative biotas and/or magnetostratigraphy or other means of dating in sections in which dinocysts are encountered. The first integrated Oligocene to earliest Miocene biomagnetostratigraphy, including dinocysts, was achieved only relatively recently on the basis of successions drilled during the CRP (e.g. Hannah et al., 1998, 2000). Even more recently, the first magnetostratigraphically calibrated Late Maastrichtian to earliest Oligocene dinocyst succession was established on the basis of records drilled during ODP Leg 189, offshore Tasmania (e.g. Brinkhuis et al., 2003a,b; Sluijs et al., 2003; Huber et al., 2004; Stickley et al., 2004). Building on these studies, Warnaar (2006) produced higher resolution records for the ODP 189 holes, and (re)analysed critical intervals from other circum-Antarctic sites like 696, 739, 1090 and 1166.

In the following section, the dinocyst record for the Eocene and Oligocene is documented and the implications for our understanding of palaeoceanography at this time are discussed.

8.4.1. Palaeocene–Middle Eocene Dynocysts

Circum-Antarctic early Palaeogene dinocyst associations were recovered from the Tasmanian Gateway (e.g. Brinkhuis et al., 2003a,b), New Zealand (e.g. Wilson, 1978, 1984, 1988; Willumsen, 2000; Crouch, 2001) and Seymour Island (e.g. Askin, 1988a,b; Elliot et al., 1994). The associations are characterized by nearly identical composition and stratigraphic succession. While circum-Antarctic endemic taxa were present at least since the Maastrichtian (e.g. Riding and Crame, 2002; Brinkhuis et al., 2003a), the taxa often referred as the "Transantarctic Flora" established itself in the early Palaeogene in an otherwise largely cosmopolitan assemblage. Since the late Early Eocene, the influence of the "Transantarctic Flora" (constituted by species such as *Deflandrea antarctica*, *Octodinium askiniae*, *Enneadocysta partridgei*, *Vozzhennikovia* spp., *Spinidinium macmurdoense* and *Arachnodinium antarcticum*) increases until the middle Late Eocene.

AU :24

While the younger part of the circum-Antarctic Middle Eocene is comparatively less well studied, with records only from New Zealand (e.g. Wilson, 1988; Strong et al., 1995), DSDP Sites 280 and 281 (Crouch and Hollis, 1996), Seymour Island (Wrenn and Hart, 1988), the Scotia Sea (Mao and Mohr, 1995), ODP Site 1172 (Brinkhuis et al., 2003a) and southern Argentinean successions (e.g. Guerstein et al., 2002), a broad similarity is apparent. The younger part of the Southern Ocean Middle Eocene appears to be characterized by several important last occurrences (LOs), including those of *Membranophoridium perforatum*, *Hystrichosphaeridium truswelliae*, *Hystrichokolpoma spinosum* and *H. truncatum* (cf. Wilson, 1988; see Brinkhuis et al., 2003b).

8.4.2. Late Eocene–Early Oligocene Dinocysts

At the Tasmanian Gateway, the early Late Eocene dinocyst distribution forms a continuation of the Middle Eocene pattern. "Transantarctic Flora" species predominate, and final acmes of *Enneadocysta partidgei*, the *D. antarctica* group and *S. macmurdoense* are recorded (Fig. 8.10). Important first occurrences (FOs) in this phase include those of *Schematophora speciosa*, *Aireiana verrucosa*, *Hemiplacophora semilunifera* and *Stoveracysta ornata*. Towards the middle Late Eocene, FOs of *Achomosphaera alcicornu*, *Reticulatosphaera actinocoronata* and *Alterbidinium distinctum* and the LO of *S. speciosa* appear important for interregional correlation, as is the FO of *Stoveracysta kakanuiensis*. *Vozzhennikovia* spp. continues to be a common constituent of the associations (Sluijs et al., 2003).

Typically, sediments representing the E/O transition are barren of organic microfossils in all ODP Leg 189 records; dinocysts briefly reappear in the Early Oligocene (assigned to Chron C11-1r; Stickley et al., 2004). In this single productive sample thus far from the Early Oligocene, virtually all Transantarctic Palaeogene dinocysts have disappeared (only a single, poorly preserved, probably reworked specimen of *E. partridgei* was recovered; Brinkhuis et al., 2003b). The association in this sample is characterized by the abundance of taxa more typical for Tethyan waters, including an occurrence of *Hystrichokolpoma* sp. cf. *Homotryblium oceanicum* (e.g. Brinkhuis and Biffi, 1993; Wilpshaar et al., 1996; Brinkhuis et al., 2003a).

Some of the Late Eocene dinocyst events have previously been reported from the South Australian margin, e.g. from the Browns Creek section (Cookson and Eisenack, 1965; Stover, 1975). For example, the ranges of *S. speciosa*, *A. verrucosa*, *H. semilunifera* and *S. ornata* appear useful for regional and even global correlation. Many of the "Browns Creek" Late Eocene dinocysts have been recorded from locations around the world, also in otherwise well-calibrated sections in central and northern Italy, including the Priabonian Type Section (Brinkhuis and Biffi, 1993; Brinkhuis, 1994). It appears that these index species have slightly earlier LOs in this region than they have in Italy (Tethyan Ocean), if the records of Cookson and

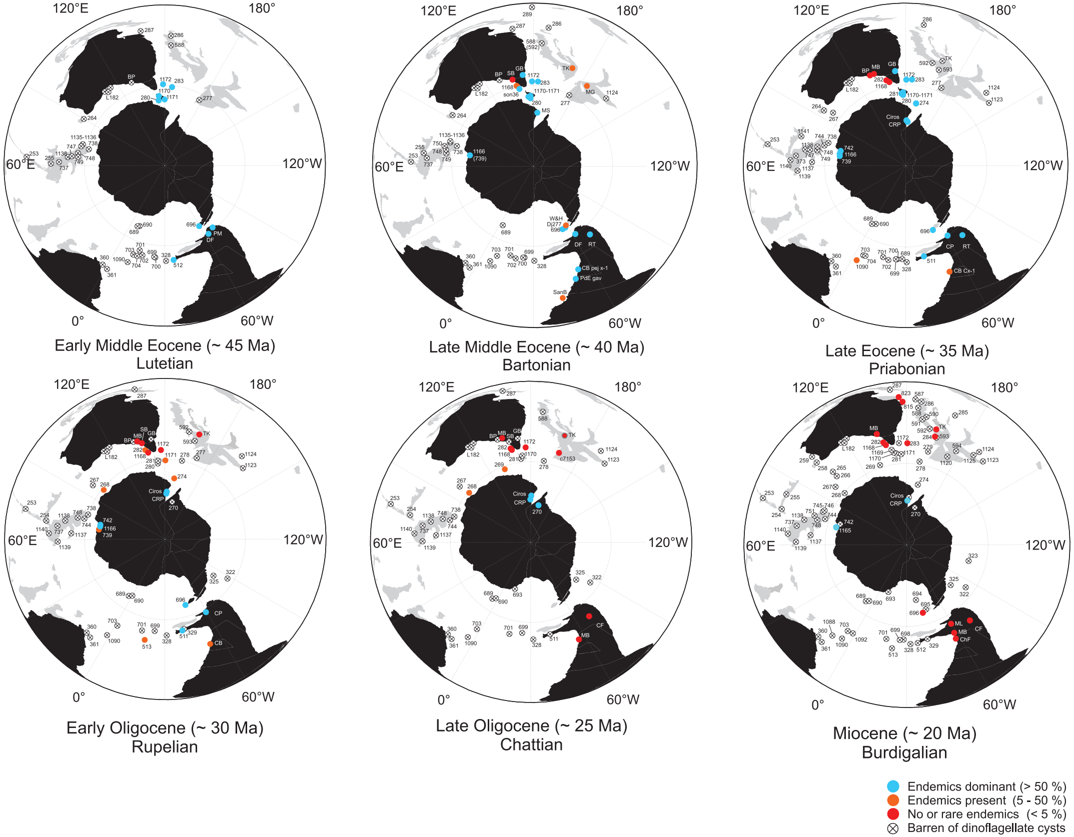

Figure 8.10: Circum-Antarctic Geographical Distribution Maps (Late Paleocene–Miocene) Showing Dinocyst Endemism. Maps derived from the Ocean Drilling Stratigraphic Network (ODSN). Black Areas indicate (Continental) Blocks that are mostly Sub-Aerial. Note that Several Blocks Shown in Black were Partly Submerged (e.g., the Ross Sea, the Southern Australian Margin and Parts of Argentina). Shaded Areas indicate mostly Submerged (Continental) Blocks (e.g., Brown et al., 2006).

Eisenack (1965) and Stover (1975) are combined with more recent nannoplankton and magnetostratigraphic studies from the same section (Shafik and Idnurm, 1997). This aspect may be related to the progressive global cooling during the latest Eocene (Fig. 8.11).

In the rare sediments covering the Eocene–Oligocene transition, more specifically the Oi-1 event, from the Weddell Sea and near the Drake Passage, dinocysts are typically not preserved (Gradstein et al., in press). However, in the oldest Oligocene sediments bearing dinocysts, the Transantarctic Palaeogene dinocysts (dominant in the latest recovered Eocene sediments) are replaced by cosmopolitan taxa (Gradstein et al., in press). This suggests that the changes in dinocyst associations in this area were at least broadly similar compared to those at the Tasman Sector. In contrast, it seems that sediments covering the Eocene–Oligocene transition are preserved at Prydz Bay (ODP Site 739). Here a gradual change is observed from the typical Transantarctic Palaeogene dinocysts to the taxa typically found in (post-) Oligocene near-Antarctic records (Warnaar, 2006). However, the (cosmopolitan) taxa that are useful for correlation, as mentioned above, are apparently not present in the Prydz Bay records.

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8.4.3. Palaeoceanography

During the early Palaeogene, Antarctica was less glaciated than it is now, or not glaciated at all (e.g. Zachos et al., 2001; Pagani et al., 2005; see above). South America and Australia were still not fully separated from the Antarctic continent (e.g. Livermore et al., 2005; Brown et al., 2006), which prevented the development of a (proto-) Antarctic Circumpolar Current (ACC). It was hypothesized that during the early Palaeogene, warm ocean currents from lower latitudes could reach and warm Antarctica. The opening and subsequent deepening of critical conduits (i.e. Drake Passage and Tasmanian Gateway) towards the end of the Eocene have long been thought to have played a central role in ACC establishment and Antarctic cooling (e.g. Kennett et al., 1975a,b; Kennett, 1977, 1978; Murphy and Kennett, 1986). However, recent advances through ocean drilling (e.g. ODP Leg 189) and coupled Global Circulation Model (GCM) experiments suggest that Eocene Southern Ocean surface circulation patterns were fundamentally different than previously thought, and that the opening and deepening of oceanic gateways were of little climatic consequence (Sloan and Huber, 2000, 2001; Huber et al., 2004; Warnaar, 2006). Instead, it is nowadays argued that the “greenhouse–icehouse” transition was caused by changes in greenhouse

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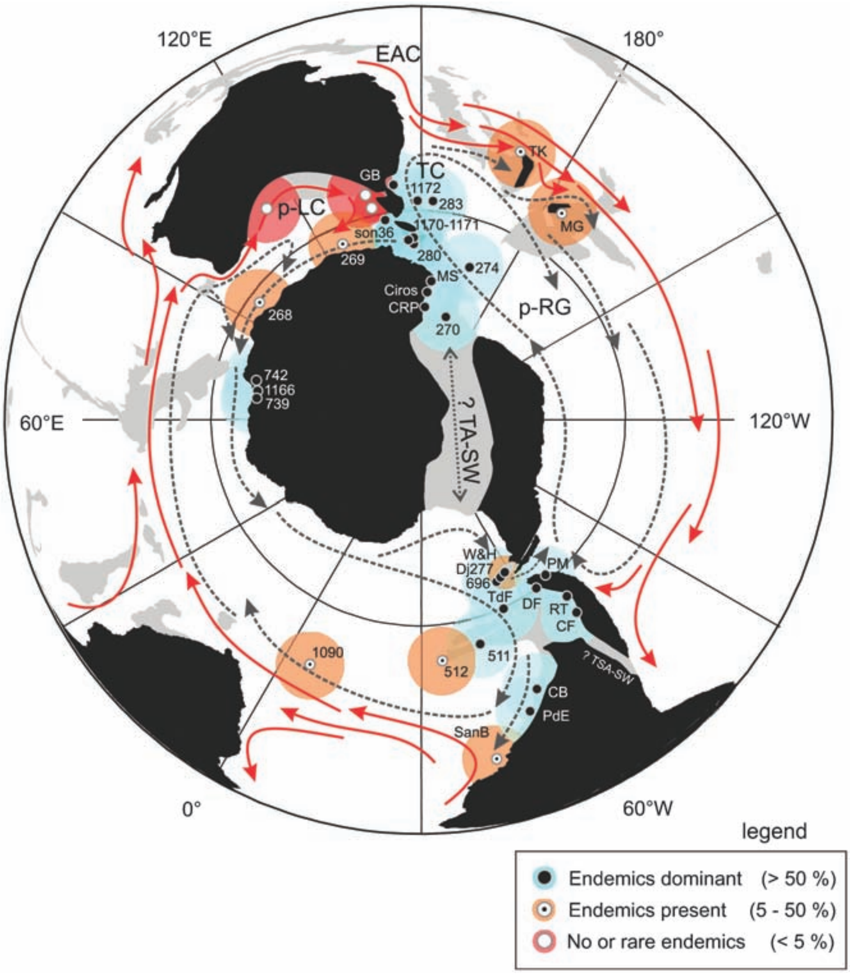


Figure 8.11: Generalized (Middle Eocene) Dinocyst Geographical Distribution Map overlain with the Ocean Circulation Pattern inferred from GCM Results. Maps derived from the Ocean Drilling Stratigraphic Network (ODSN). Shaded Areas indicate mostly Submerged (Continental) Blocks (e.g., Brown et al., 2006). *Abbreviations*: TS-SW, Trans-Antarctic Seaway (Hypothetical; See Wrenn and Beckmann, 1982); TSA-SW, Trans-South American Seaway (Hypothetical; See Kohn et al., 2004); EAC, East Australian Current; p-LC, proto-Leeuwin Current; p-RG, proto-Ross Gyre; TC, Tasman Current.

gas concentrations, rather than oceanographic changes (e.g. DeConto and Pollard, 2003a,b; Huber et al., 2004; Pagani et al., 2005).

The high degree of endemism in the circum-Antarctic marine microfossil associations denies the existence of a southward-bound, warm, proto-East Australian Current as proposed by Kennett et al. (1975b) and Kennett and Exon (2004), according to Huber et al. (2004) (Fig. 8.11). For example, a colder northward flowing western boundary current, designated the "Tasman Current" (see Huber et al., 2004) existed off southeast Australia. GCM experiments indicate that the Eocene Southern Ocean, including the southern Pacific, was dominated by clockwise gyres (Sloan and Huber, 2000, 2001; Huber et al., 2004). Moreover, several studies show that the Tasmanian Gateway had already been open to neritic water depths (i.e. <200 m) since at least the Middle Eocene (Langford et al., 1995; Stickley et al., 2004). Deepening to bathyal water depths (i.e. 200–4,000 m) occurred during the early Late Eocene (~35.5 Ma). The Drake Passage had possibly been open to (upper) bathyal water depths by the Middle Eocene (Eagles et al., 2006; Scher and Martin, 2006). Both tectonic events thus seem to have occurred too early to be related to the Antarctic glaciation in the Early Oligocene (i.e. the Oi-1 stable-isotope event (33.3 Ma), e.g. Miller et al., 1991; Zachos et al., 2001).

Given the similar continent–ocean configuration, a corollary of the GCM experiments is that the Palaeogene circum-Antarctic surface circulation should not have been fundamentally different from the Cretaceous situation. If this were the case, and as long as substantial equator–pole temperature gradients existed, then it may be expected that, throughout the Late Cretaceous to early Palaeogene time interval, circum-Antarctic waters were consistently dominated by endemic biota, particularly in environments influenced by the proposed western boundary currents. Both hypotheses were recently tested by Warnaar (2006) by mapping distribution patterns of circum-Antarctic dinocysts through Palaeogene times and comparing them with coupled GCM results.

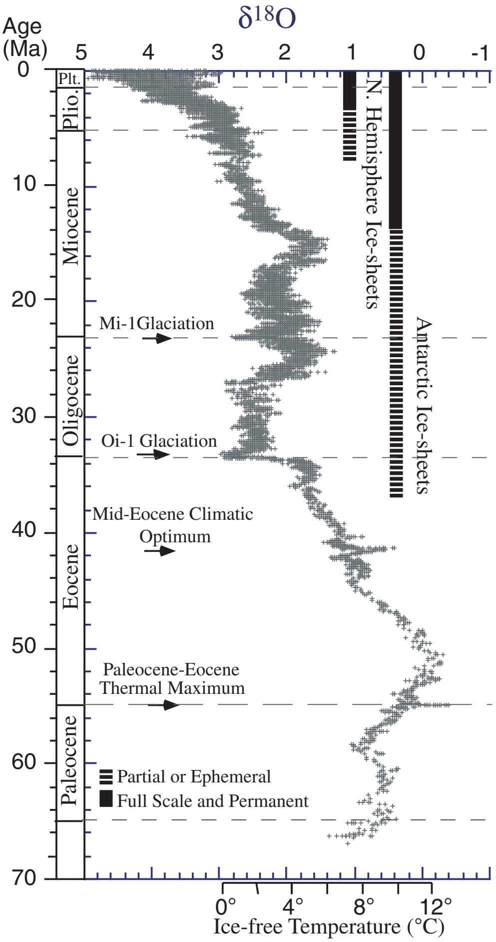
Warnaar (2006) conceived a model termed the "*refrigerator trap*" wherein it is hypothesized that cosmopolitan and endemic dinoflagellates were taken through the cold and darkness along the Antarctic continent, transported by the proto-Ross Gyre. Conceivably, taxa normally living in warmer waters (e.g. the East Australian Current) that were trapped in the gyre were unable to survive such conditions. It is conceivable that the endemic taxa (notably taxa of the "Transantarctic Flora" and bi-polar *Phthanoperidinium echinatum* group) were specifically adapted to tolerate cold conditions (4°C to the freezing point), prolonged darkness and possibly seasonal sea ice.

8.5. Evolution of Ocean Temperatures and Global Ice Volume During the Eocene to Oligocene from the Ocean Isotope Record

The evolution of climate during the Eocene and Oligocene can be determined from the deep-sea isotope and trace element records of ocean temperatures and ice volume. Earlier isotope work suggests that the primary transition from greenhouse to icehouse world took place during the Late Eocene and Early Oligocene, with large, permanent ice sheets appearing on Antarctica at 34 Ma (Zachos et al., 1992, 1996, 2001; Miller et al., 1998; Coxall et al., 2005). This transition was preceded by a period of long-term cooling which initiated near the Early–Middle Eocene boundary, roughly 50 Ma, following a sustained period of Early Eocene warmth. The Eocene cooling trend was not monotonic, but followed a somewhat step-like pattern with several reversals, the most substantial of which was the Middle Eocene climatic optimum (MECO) (Bohaty and Zachos, 2003). By the Late Eocene, the climate on Antarctica appears to have cooled sufficiently to allow for the formation of small, ephemeral ice sheets, a state that persisted until ~34 Ma, when most of east Antarctica became glaciated with a large ice sheet (Fig. 8.12). From that time forward, the ice sheet was a permanent feature of Antarctica. For the remainder of the Oligocene, this ice sheet waxed and waned, most likely in response to orbital forcing (Naish et al., 2002). AU :34

The long-term cooling trend that facilitated the formation of continental ice sheets has been attributed to either changes in palaeogeography or the concentration of greenhouse gases. Geographical isolation of the Antarctic continent with the tectonic widening of ocean gateways is often cited as one means of driving long-term cooling (see above). A second possible mechanism is a decline in atmospheric p_{CO_2}. Recent modelling studies (see below) suggest that the former would have had little impact on heat fluxes and mean annual temperatures on Antarctica, while the latter would be a more effective way of cooling the continent (Huber and Sloan, 2001). However, the record of Eocene p_{CO_2} has until recently (Pagani et al., 2005) lacked sufficient resolution to fully test this possibility.

Recent investigations of marine cores have largely focused on improving two aspects of the Eocene and Oligocene climate reconstructions: (1) the resolution of proxy records of ocean temperature and ice volume, which has promoted the development of high-resolution, orbitally tuned records, and (2) quantifying changes in ice volume, which has spurred the development and application of palaeotemperature proxies. These studies have benefited in part from efforts of the ODP to recover highly expanded, stratigraphically



Zachos et al., 2001; Billups et al., 2003; Bohaty & Zachos, 2003; Lear et al., 2004

Figure 8.12: Global Compilation of Oxygen Isotope Records for the Cenozoic (after Zachos et al., 2001). Solid Bars span Intervals of Ice-Sheet Activity in the Antarctic and Northern Hemisphere. AU:8

intact sediment sequences spanning the Eocene–Oligocene. The latest high-resolution records show that ice volume and ocean temperatures varied in a periodic fashion with power concentrated in the long eccentricity and obliquity bands (Fig. 8.13; Coxall et al., 2005). The latter is consistent with the presence of a large polar ice sheet. Prior to the Late Eocene, however, power in the obliquity band is relatively weak (Palike et al., 2001), suggesting little to no ice volume on Antarctica. These records have also revealed

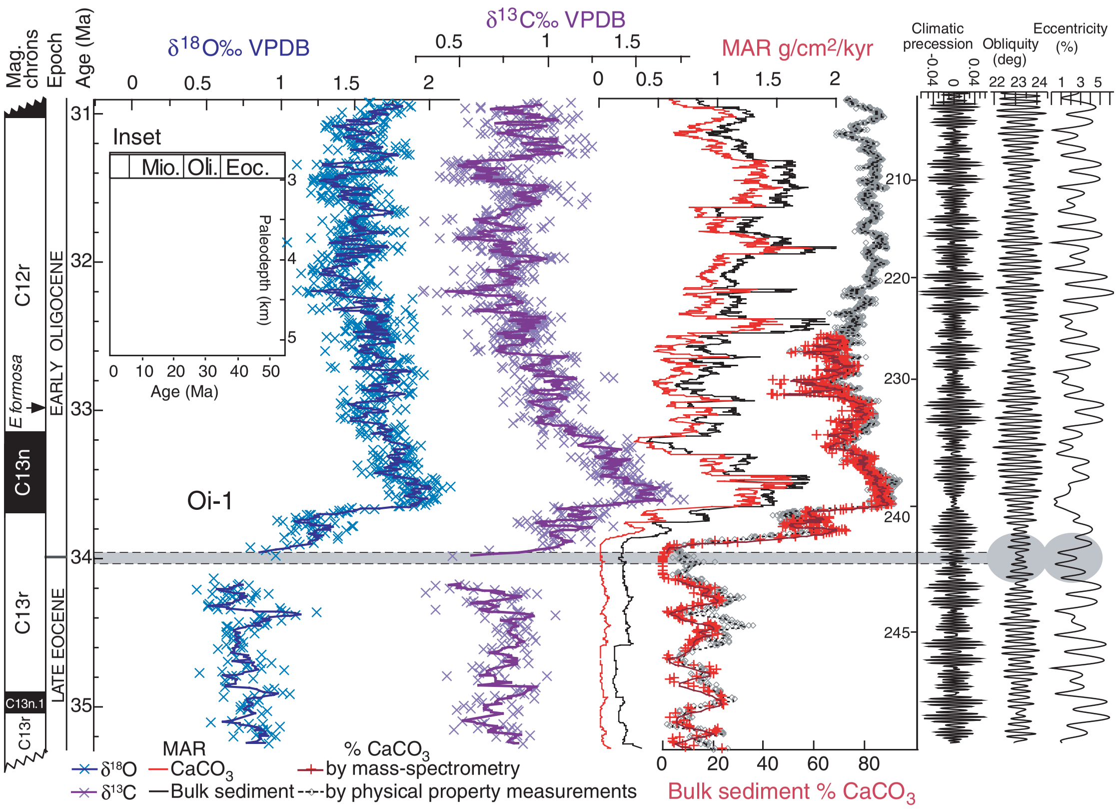


Figure 8.13: High-Resolution Isotope and Per Cent $CaCO_3$ Records of the Eocene–Oligocene Boundary (Coxall et al., 2005). Age Model is derived from Orbital Tuning.

distinct climatic variability coherent with the lower frequency components of obliquity with periods of 1.25 m.y.

The recent development of seawater temperature proxies, Mg/Ca and TEX_{86} (Schouten et al., 2003), has improved estimates of ice volume from oxygen isotope records. In particular, the first low-resolution benthic Mg/Ca records suggest that much of the $\delta^{18}O$ increase just after the E–O boundary (33.4 Ma) was the result of a substantial increase in ice volume as deep-sea temperature was fairly constant (Lear et al., 2000; Billups and Schrag, 2002). Though still controversial, the magnitude of ice-volume increase would have exceeded that of the present-day Antarctic ice-sheet. The multi-proxy approach has been applied to resolving Antarctic climate evolution for the time preceding the E–O boundary, specifically the Middle and Late Eocene, and initial results suggest appearance of small, ephemeral ice sheets at ~38 Ma (Florindo et al., in review). The ice sheets appear to appear/disappear in a cyclical fashion, increasing in size with each cycle until 33.4 Ma when they expand and become permanent.

AU :35

8.6. Connection of CO_2 and Ice-Sheet Inception at the Eocene–Oligocene Boundary – Computer Modelling

While the onset of major, continental-scale glaciation in the earliest Oligocene (Oi-1 event) has long been attributed to the opening of Southern Ocean gateways (Kennett and Shackleton, 1976; Kennett, 1977; Robert et al., 2001), recent modelling studies suggest declining atmospheric CO_2 was the most important factor in Antarctic cooling and glaciation.

As the passages between South America and the Antarctic Peninsula (Drake Passage), and Australia and East Antarctica (Tasmanian Passage) widened and deepened during the late Palaeogene and early Neogene (Lawver and Gahagan, 1998), the ACC and Polar Frontal Zone (APFZ) presumably cooled the Southern Ocean by limiting the advection of warm subtropical surface waters into high latitudes (Kennett, 1977). While the opening of the Tasmanian gateway does broadly coincide with the earliest Oligocene glaciation event (Oi-1) (Stickley et al., 2004), the tectonic history of the Scotia Sea remains equivocal. Estimates for the timing of the opening of Drake Passage range between 40 and 20 Ma (Barker and Burrell, 1977; Livermore et al., 2004; Scher and Martin, 2006), clouding the direct "cause and effect" relationship between the gateways and glaciation.

A number of ocean modelling studies have shown that the opening of both the Drake and Tasmanian gateways reduces poleward heat convergence in the Southern Ocean and cools sea surface temperatures by up to several degrees (Mikolajewicz et al., 1993; Nong et al., 2000; Toggweiler and Bjornsson, 2000). More recent, coupled atmosphere–ocean GCM simulations suggest a more modest effect, however. Huber et al. (2004) showed that the Tasmanian Gateway likely had a minimal effect on oceanic heat convergence and sea surface temperatures around the continent, because the warm East Australia Current does not travel any further south if the gateway is open or closed. The gateway's effect on East Antarctic climate and snowfall was also shown to be minimal, pointing to some other forcing (perhaps decreasing atmospheric CO_2 concentrations) as the primary cause of Antarctic cooling and glaciation.

The recent development of coupled climate–ice sheet models capable of running long ($>10^6$ years), time-continuous simulations of specific climate events and transitions (DeConto and Pollard, 2003a) has allowed simulations of the Oi-1 event that account for decreasing CO_2 concentrations, orbital variability and prescribed changes in ocean transport (DeConto and Pollard, 2003b; Pollard and DeConto, 2005). These simulations support the conclusions of Huber et al. (2004) as to the likely importance of CO_2, by showing that even if significant, tectonically forced changes in ocean circulation and heat transport had occurred around the Eocene–Oligocene boundary, they would have had only a small effect on temperature and glacial mass balance in the Antarctic interior. Therefore, Southern Ocean gateways could only have triggered glaciation if the climate system was already close to a glaciation threshold. Considering the sensitivity of polar climate to the range of CO_2 concentrations likely to have existed over the Palaeogene–Neogene (Pagani et al., 2005), CO_2 likely played a fundamental role in controlling Antarctica's climatic and glacial sensitivity to a wide range of forcing mechanisms. This conclusion is supported by a number of additional modelling studies exploring the role of orbital variability (DeConto and Pollard, 2003b), mountain uplift in the continental interior (DeConto and Pollard, 2003a), geothermal heat flux (Pollard et al., 2005), Antarctic vegetation dynamics (Thorn and DeConto, 2006) and Southern Ocean sea ice (DeConto et al., in press) in the Eocene–Oligocene climatic transition.

The results of these studies can be summarized as follows. The timing of glaciation on East Antarctica was shown to be sensitive to orbital forcing, mountain uplift and continental vegetation, but only within a very narrow range of atmospheric CO_2 concentrations around 2.8 times modern levels, close to the model's glaciation threshold. Once the glaciation threshold is approached, astronomical forcing can trigger sudden glaciation through non-linear height/mass balance and albedo feedbacks that result in the growth of a continental-scale ice sheet within 100 kyears (Fig. 8.14).

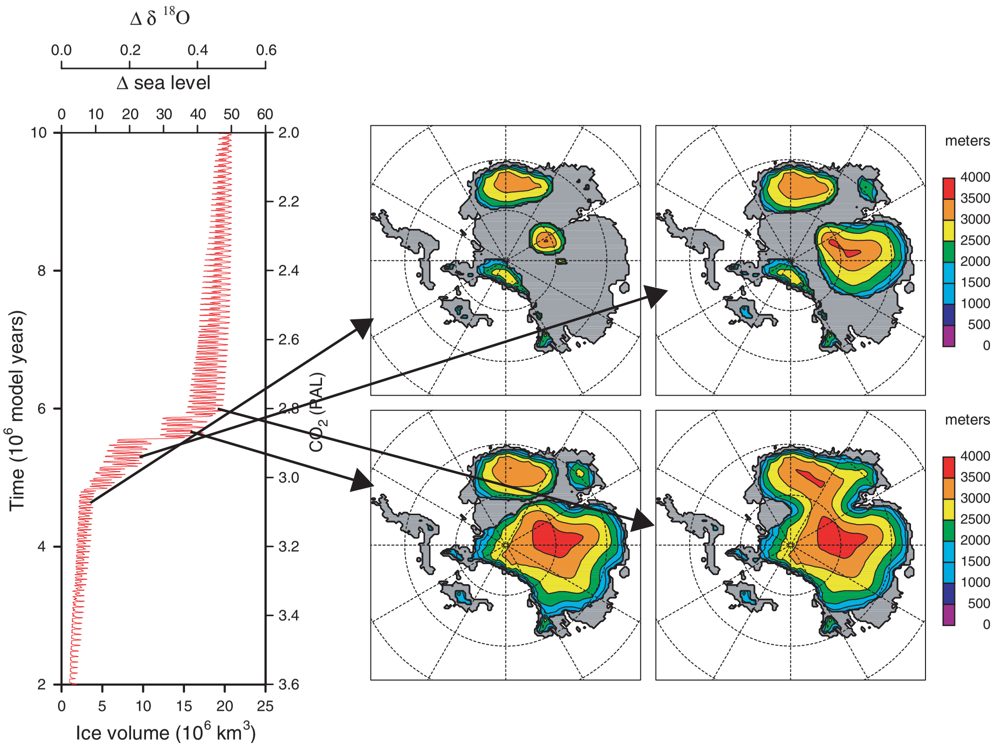


Figure 8.14: Ice Volume (Left) and corresponding Ice-Sheet Geometries (Right) simulated by a Coupled GCM-Ice Sheet Model in response to a Slow Decline in Atmospheric CO$_2$ and Idealized Orbital Cyclicity across the Eocene–Oligocene Boundary. The Sudden, Two-Step Jump in Ice Volume (Left Panel) corresponds to the Oi-1 Event. The Left Panel Shows Simulated Ice Volume (Red Line), extrapolated to an Equivalent Change in Sea Level and the Mean Isotopic Composition of the Ocean (Top). Arbitrary Model Years (Left Axis) and Corresponding, Prescribed Atmospheric CO$_2$ (Right Axis) are also Labelled. CO$_2$ is Shown as the Multiplicative of Pre-Industrial (280 ppmv) Levels. Ice-Sheet Geometries (Right Panels) Show Ice-Sheet Thickness in Metres. Black Arrows correlate the Simulated Geometric Evolution of the Ice Sheet through the Oi-1 Event (Modified from DeConto and Pollard, 2003b).

The timing of glaciation appears to be insensitive to both expanding concentrations of seasonal sea ice and changes in geothermal heat flux under the continent; however, a doubling of the background geothermal heat flux (from 40 to 80 $\mathrm{mW\,m^{-2}}$) does have a significant effect on the area under the ice sheet at the pressure-melt point (where liquid water is present), which may have had some influence on the distribution and development of sub-glacial lakes and subsequent ice-sheet behaviour.

While these modelling studies have certainly improved our understanding of the importance of atmospheric CO_2 concentrations relative to other Cenozoic forcing factors, several important model-data inconsistencies remain unresolved. For example, long, time-continuous GCM-ice-sheet simulations of an increasing CO_2 (warming) scenario show strong hysteresis once a continental ice sheet has formed (Pollard and DeConto, 2005). In these simulations, orbital forcing alone is not sufficient to produce the range of Palaeogene–Neogene ice-sheet variability (~50–120 per cent modern Antarctic ice volumes) inferred by marine oxygen isotope records and sequence stratigraphic reconstructions of eustasy (Zachos et al., 2001; Pekar and DeConto, 2006; Pekar et al., 2006), pointing to the importance of additional feedbacks (possibly related to the marine carbon cycle and atmospheric CO_2) in controlling Cenozoic ice-sheet variability. Furthermore, several recent isotopic analyses of deep-sea cores imply ice volumes during the peak Oligocene and Miocene glacial intervals that are too big to be accommodated by East Antarctica alone (Lear et al., 2004; Coxall et al., 2005; Holbourn et al., 2005). This suggests that either our interpretations of the proxy data are faulty, or episodic, bipolar glaciation occurred much earlier than currently accepted. These, among other unresolved controversies related to the climatic and glacial evolution of the high southern latitudes, will be the focus of future ACE modelling exercises and model-data comparisons.

8.7. Summary

Although no one single source of evidence or locality yields the complete story of climate change and the onset of glaciation in Antarctica, piecing together information from a range of sources, as presented above, does provide a picture of how climate cooled and glaciation became established on the continent. A summary is presented below (for references, see text above). Paradoxically, the present ice sheet hides its own history, certainly at the highest latitudes in the middle of the continent, so most of what is known about past environments is from the lower latitude marginal sites.

8.7.1. Early–Middle Eocene Polar Warmth

Evidence from fossil plants, sediments and isotopes indicates that the Late Palaeocene and Early Eocene experienced warm climates at high latitudes, at least on the margins of Antarctica where strata of this age crop out. Climates appear to have been warm and wet, seas were warm and plants flourished in a frost-free environment. The oldest record of glacial activity (if the dating is correct in this problematic region) is of valley-type tillites of Middle Eocene age on King George Island, indicating the presence of alpine glaciers. However, floras of mid-Eocene age from King George and Seymour islands suggest warm to cool temperate climates, generally moist and probably frost-free. The ocean isotope record also suggests that climates were generally warm until the Middle Eocene, although the climate trend was towards cooling.

8.7.2. Late Eocene Cooling

A variety of sources, particularly fossil plants, suggest that during the early Late Eocene, climates cooled but perhaps not to the extent of significant ice build-up. The Late Eocene sediment record in the Ross Sea region (McMurdo Erratics, magnetic and clay mineral record) and in the Prydz Bay area could be indicative of cold climates but the coastal/open marine shelf and fluvial-deltaic environments in these two areas, respectively, do not show signs of the presence of significant ice. By the latest Eocene, however, glacial deposits are apparent. Glacial deposits on Seymour Island, close to the E/O boundary, may indicate the presence of valley glaciers in that region, situated at about 65°palaeos. AU :36

The ocean record, especially marine microfossils, provides information about the climate and currents in the oceans at critical times during this interval. South America and Australia were still not separated from Antarctica during the Early Eocene, so the ACC was unable to develop. Instead, warm equatorial currents may have fed warmth to the continent. Palaeoceanographic changes related to the deepening of the Tasman gateway and the opening of the Drake Passage are still debated but do not seem to have been strictly related to Early Oligocene climate cooling and intensification of glaciation, so these oceanographic changes are not considered to be major drivers of polar cooling. Recent atmosphere–ocean modelling has also shown that changes in oceanography related to tectonic events were not likely to have driven the climate cooling that led to glaciation.

8.7.3. Latest Eocene/Earliest Oligocene Glaciation

By the E/O boundary times, there is no doubt that ice was present on Antarctica. In the Ross Sea region, drill cores show evidence of relatively uniform marine sedimentation through the latest part of the Eocene and into the Oligocene but sediments include exotic clasts indicative of iceberg rafting. There does not appear to be a major environmental shift at this time but more of an intensification of cooling. In the Prydz Bay region, tidewater glaciers were present in the early Oligocene, with ice reaching the continental shelf edge. In the oceans, the oxygen isotope record and other geochemical indicators signal a strong cooling at the boundary, the Oi-1 event, which has been interpreted as a time of major build-up of ice.

Even though climates were cold, vegetation was able to persist but by this time the higher diversity and warmth-loving plants of the Early and Middle Eocene forests had disappeared, to be replaced by vegetation that was dominated by several species of the southern beech, *Nothofagus*. Along with mosses, a few ferns and some podocarp conifers, southern beech trees probably grew as shrubby tundra-like vegetation in the most hospitable areas.

8.7.4. Oligocene Ice Sheets

A hiatus at the Early/Late Oligocene boundary marks a change to fluvial conditions, with grounded ice or possibly glaciomarine conditions. A distinct drop in sea level is noted at this time. Facies changes and diamictite beds in the Cape Roberts core are indicative of the periodic expansion of tidewater glaciers, typical of a temperate glacier regime with glaciers flowing from the young East Antarctica ice sheet across the continental margin.

Throughout the Oligocene, glaciation seems to have waxed and waned. A distinct record of glacial activity is recorded by glacial sediments in the Polonez Cove Formation on King George Island of mid-Oligocene age (Krakowiak Glacial Member). At that time, ice was grounded on a shallow marine shelf. More extensive ice sheets may have been present further south in the Weddell Sea region, from which clasts of rock from the Transantarctic Mountains may have been derived to be incorporated as exotic clasts in the Polonez Cove Formation. Sediments without a glacial signature that overlie these glacial deposits suggest a phase of climate warming and glacial retreat until the next glacial pulse in the Miocene.

Why did the climate cool during the Eocene and Oligocene, causing such a major change in Antarctic environments? The influence of palaeoceanographic changes is now considered less critical; instead, coupled climate–ice sheet modelling indicates that it was changing levels of atmospheric CO_2 that controlled Antarctica's climate. Factors such as mountain uplift, vegetation changes and orbital forcing all played a part in cooling the polar climate, but only when CO_2 levels reached critical threshold levels (2.8 times present-day levels) did orbital forcing tip Antarctica into its icy glacial world.

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