CLIMATE OVER PAST MILLENNIA

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Received 20 October 2003; revised 4 February 2004; accepted 17 February 2004; published 6 May 2004.

[1] We review evidence for climate change over the past several millennia from instrumental and high-resolution climate "proxy" data sources and climate modeling studies. We focus on changes over the past 1 to 2 millennia. We assess reconstructions and modeling studies analyzing a number of different climate fields, including atmospheric circulation diagnostics, precipitation, and drought. We devote particular attention to proxy-based reconstructions of temperature patterns in past centuries, which place recent large-scale warming in an appropriate longer-term context. Our assessment affirms the conclusion that late 20th century warmth is unprecedented at hemispheric and, likely, global scales. There is more tentative evidence that particular modes of climate variability, such as the El Niño/Southern Oscillation and the North Atlantic Oscillation, may have exhibited late 20th century behavior that is anomalous in a long-term context. Regional conclusions, particularly for the Southern Hemisphere and parts of the tropics where high-resolution proxy data are sparse, are more circumspect. The dramatic differences between regional and hemispheric/global past trends, and the distinction between changes in surface temperature and precipitation/drought fields, underscore the limited utility in the use of terms such as the "Little Ice Age" and "Medieval Warm Period" for describing past climate epochs during the last millennium. Comparison of empirical evidence with proxy-based reconstructions demonstrates that natural factors appear to explain relatively well the major surface temperature changes of the past millennium through the 19th century (including hemispheric means and some spatial patterns). Only anthropogenic forcing of climate, however, can explain the recent anomalous warming in the late 20th century. *INDEX TERMS:* 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology; 3309 Meteorology and Atmospheric Dynamics: Climatology (1620); 1620 Global Change: Climate dynamics (3309); 3394 Meteorology and Atmospheric Dynamics: Instruments and techniques; 3354 Meteorology and Atmospheric Dynamics: Precipitation (1854); *KEYWORDS:* climate

Citation: Jones, P. D., and M. E. Mann (2004), Climate over past millennia, *Rev. Geophys.*, *42*, RG2002, doi:10.1029/2003RG000143.

CONTENTS

8755-1209/04/2003RG000143*$15.00*
Paper number 2003RG000143

1. INTRODUCTION AND MOTIVATION

[2] Climate varies on all timescales, from years to decades to millennia to millions and billions of years. This climate variability can arise from a number of factors, some external to the climate system others internal to the system [*Bradley*, 1999; *Ruddiman*, 2001]. Any improvements in our ability to predict future climate changes are most likely to result from a better understanding of the workings of the climate system as a whole. Such an understanding can, in turn, best be achieved through a greater ability to document and explain, from a fundamental dynamical point of view, observed past variations. Instrumental meteorological records can only assess large-scale (hemispheric and global) climate changes over roughly the past 100–150 years, while more regionally limited data (e.g., in Europe) are available back to the early 18th century. Improved knowledge of past large-scale climate changes should enable recent, potentially anomalous, climate change to be placed in a longer-term context. Information from other data sources and/or climate models is thus necessary for insights into climate variability on multicentury and longer timescales. In this review we document the range of information available from paleoclimatic data sources and modeling studies to inform our understanding of climate variability on centennial to millennial timescales. We emphasize climate variations during the past few millennia (the late Holocene), the period that is most relevant to assessing the potential uniqueness of recent climate changes and the projected changes in the 21st century.

1.1. Climate of the Late Holocene

[3] The Pleistocene climate epoch was marked by major "glacial/interglacial" oscillations in global ice volume that occurred on timescales of tens to hundreds of millennia. These oscillations appear to have been governed by similar timescale changes in the distribution of insolation over the Earth's surface related to long-term changes in the orbital position of the Earth relative to the Sun. The most recent glacial period culminated about 21,000 years ago (the Last Glacial Maximum) when continental ice sheets extended well into the midlatitudes of North America and Europe. Global annual mean temperatures were probably about 4°C colder than the mid-20th century, with larger cooling at higher latitudes and little or no cooling over large parts of the tropical oceans [e.g., *Ruddiman*, 2001]. This glacial period terminated somewhat abruptly, ~12,000 years ago. The climate appears to have exhibited less dramatic, but nonetheless significant, variability on centennial and millennial timescales during the subsequent interglacial period, referred to as the "Holocene," within which we currently reside. During the early millennia of the Holocene, atmospheric circulation, surface temperature, and precipitation patterns appear to have been substantially altered from the present day, with evidence, for example, of ancient lakes in what is now the Sahara [*Street and Grove*, 1979]. During the "mid-Holocene" of 5000–6000 years ago, surface temperatures appear to have been milder in some parts of the globe, particularly in the extratropics of the Northern Hemisphere during summer [see *Webb and Wigley*, 1985]. This has given rise to the use of the descriptor "Holocene Optimum" sometimes used to characterize this period. There is still considerable uncertainty, however, with regard to the relative global, annual mean warmth at this time, because much of the evidence for warmer conditions comes from the extratropics and appears biased toward warm season conditions. The orbitally induced insolation changes likely favored warmer high-latitude summers but cooler winters and slightly cooler tropics, with any net hemispheric- or global-scale changes representing a subtle competition between these seasonally and spatially heterogeneous changes [*Hewitt*, 1998; *Kitoh and Murakami*, 2002; *Liu et al.*, 2003] and seasonally specific (e.g., vegetation-albedo) feedbacks [e.g., *Ganopolski et al.*, 1998]. Recent modeling studies suggest that mid-Holocene surface temperatures for annual and global means may actually have been cooler than those of the mid-20th century, even though extratropical summers were likely somewhat warmer [*Kitoh and Murakami*, 2002].

[4] Extratropical summer temperatures appear to have cooled over the subsequent four millennia [e.g., *Matthes*, 1939]. This period, sometimes referred to as the "Neoglacial" because it was punctuated with periods of glacial advance (and of glacial retreat) of extratropical mountain glaciers [e.g., *Matthes*, 1939; *Porter and Denton*, 1967], was reminiscent of, though far more modest than, a full glacial period during the Pleistocene epoch. Orbital influences likely influenced large-scale climate over this time frame through an interaction with the monsoon and *El Niño–Southern Oscillation* (ENSO) phenomena [e.g., *Joussaume et al.*, 1999; *Bush*, 1999; *Liu et al.*, 2000; *Clement et al.*, 2000]. (Italicized terms are defined in the glossary, after the main text.) Evidence regarding the actual nature of changes in ENSO over past millennia [*Sandweiss et al.*, 1996, 2001; *Tudhope et al.*, 2001; *Koutavas et al.*, 2002; *Stott et al.*, 2002] is controversial [*DeVries et al.*, 1997; *Trenberth and Otto-Bliesner*, 2003]. This is due to the paucity of evidence, uncertainty concerning the influences in the proxies used, or the inability of the proxy records to resolve true (interannual) El Niño variability. Recent evidence also suggests that human-induced land use changes influencing methane and CO_2 production may have begun to influence the climate over the past several millennia [*Ruddiman*, 2003].

1.2. Past One to Two Millennia

[5] When we restrict our attention to the more limited interval of the past one to two millennia, a period that can be referred to as the "late Holocene" [*Williams and Wigley*, 1983], the principal boundary conditions on the climate (e.g., Earth orbital geometry and global ice mass) have not changed appreciably. The variations in climate observed over this time frame are likely therefore to be representative of the natural climate variability that might be expected over the present century in the absence of any human influence. Placing modern climate change, including recent global-scale warming, in a longer-term context can thus help

establish the importance of anthropogenic forcing (human-generated greenhouse gas concentration increases and aerosol production) on recent past and future climate changes.

[6] Much recent work has focused on the changes during this shorter time interval, over which widespread high-resolution, precisely dated proxy records are available for large regions of the Northern Hemisphere and some parts of the *Southern Hemisphere* (SH) [*Jones et al.*, 2001a; *Mann*, 2001a, 2002a; *Mann and Jones*, 2003]. This paper reviews the evidence for changes over the past one to two millennia. Details of the available proxy climate data and how they are assimilated into multiple-proxy-based (multiproxy) climate reconstructions are discussed in sections 2 and 3. Interpretations of the proxies, in terms of climate change over the past one to two millennia, are discussed in section 4.

1.3. Comparison of Models and Observations

[7] Evidence of past climate change should not be interpreted in isolation. Comparison of climate model simulations and empirical paleoclimatic data can greatly enhance our understanding of the climate system. Such studies help to assess the role of external forcing, including natural (e.g., volcanic and solar radiative) and anthropogenic (greenhouse gas and sulphate aerosol) influences, and natural, internal variability (e.g., natural changes in the El Niño–Southern Oscillation and century- or millennial-scale natural oscillations in the coupled ocean-atmosphere system). Models can help us determine how we might have expected the climate system to have changed given past changes in boundary conditions and forcings, which we can compare to inferences derived from paleoclimatic data.

[8] Internal variability generated in coupled ocean-atmosphere models can be verified against the long-term variability evident in proxy-based temperature reconstructions of the past millennium [*Mann et al.*, 1995; *Barnett et al.*, 1996, 1999; *Jones et al.*, 1998; *Braganza et al.*, 2003; *Covey et al.*, 2003; *Bell et al.*, 2003]. Natural external forcing by volcanoes, the Sun [*Lean et al.*, 1995; *Crowley and Kim*, 1996, 1999; *Cubasch et al.*, 1997; *Overpeck et al.*, 1997; *Mann et al.*, 1998a; *Damon and Peristykh*, 1999; *Free and Robock*, 1999; *Rind et al.*, 1999; *Crowley*, 2000; *Gerber et al.*, 2003; *Bertrand et al.*, 2002; *Shindell et al.*, 2001, 2003; *Waple et al.*, 2002; *Bauer et al.*, 2003], and, particularly during the 19th and 20th centuries, human land use changes [*Govindasamy et al.*, 2001; *Bauer et al.*, 2003] all appear to have played significant roles. Comparisons of climate models with estimated forcing changes and proxy climate reconstructions can therefore provide constraints on the sensitivity of the climate to radiative forcing [*Crowley and Kim*, 1999; *Crowley*, 2000; *Gerber et al.*, 2003; *Bertrand et al.*, 2002; *Bauer et al.*, 2003]. Additionally, *general circulation model* (GCM) simulations of climate can provide details of the spatial response of the climate to forcing [*Cubasch et al.*, 1997; *Waple et al.*, 2002; *Shindell et al.*, 2001, 2003]. Coupled climate-carbon cycle models can provide additional constraints on the sensitivity from paleoclimate temperature reconstructions by comparing observed and modeled CO_2 variability prior to modern anthropogenic influences on atmospheric CO_2 [*Gerber et al.*, 2003]. Most of these studies suggest an equilibrium climate sensitivity in the range of 1.5°–4.5°C for a doubling of CO_2, consistent with other evidence [e.g., *Cubasch et al.*, 2001, and references therein]. Details of past climate forcing and model/data comparison are discussed in section 5. Section 6 addresses future directions, and section 7 concludes.

2. PALEOCLIMATE "PROXY" DATA

[9] As widespread, direct measurements of climate variables are only available about one to two centuries back in time, it is necessary to use indirect indicators or "proxy" measures of climate variability provided by natural archives of information present in our environment to reconstruct earlier changes. These natural archives record, by their biological, chemical, or physical nature, climate-related phenomena. Additionally, information is provided by written archives from historical documents. Some proxy indicators, including most sediment cores, low accumulation ice cores, and preserved pollen, cannot record climate changes at high temporal resolution. These indicators generally have poor chronologies because of uncertain radiometric dating methods or questionable "age model" assumptions (e.g., the assumption of constant stratigraphic rates between marker or "dated" horizons). Such proxy indicators are thus only useful for describing climate changes on centennial and often longer timescales. High-resolution, annually and/or seasonally resolved proxy climate records (historical documents, growth and density measurements from tree rings, corals, annually resolved ice cores, laminated ocean and lake sediment cores, and speleothems) can, however, describe year-by-year patterns of climate in past centuries [*Folland et al.*, 2001a; *Jones et al.*, 2001a; *Mann*, 2001a]. This review concentrates on these higher-resolution proxy records. We make note of the strengths of each proxy source but emphasize the potential weaknesses and caveats, which are more central to current debates taking place in the paleoclimatological literature, particularly with respect to interpretation of implied decadal-to-century timescale variability.

[10] All proxy data are indirect measurements of climate change, and they vary considerably in their reliability as indicators of long-term climate. For a reliable reconstruction of past changes from proxy data it is essential that reconstructions based on these indirect climate indicators be "calibrated" and independently "validated" against instrumental records during common intervals of overlap. All reconstructions are based on statistical regression models, so they implicitly assume long-term stationarity in the nature of a proxy's response to climate. We discuss where this might be relevant for the interpretation of specific proxy sources. Temporal calibration exercises are possible only with high-resolution (annual or seasonally resolved or, perhaps, decadally resolved) proxy data. Less temporally resolved climate indicators such as pollen, nonvarved sediment cores, lake level reconstructions, glacial moraine

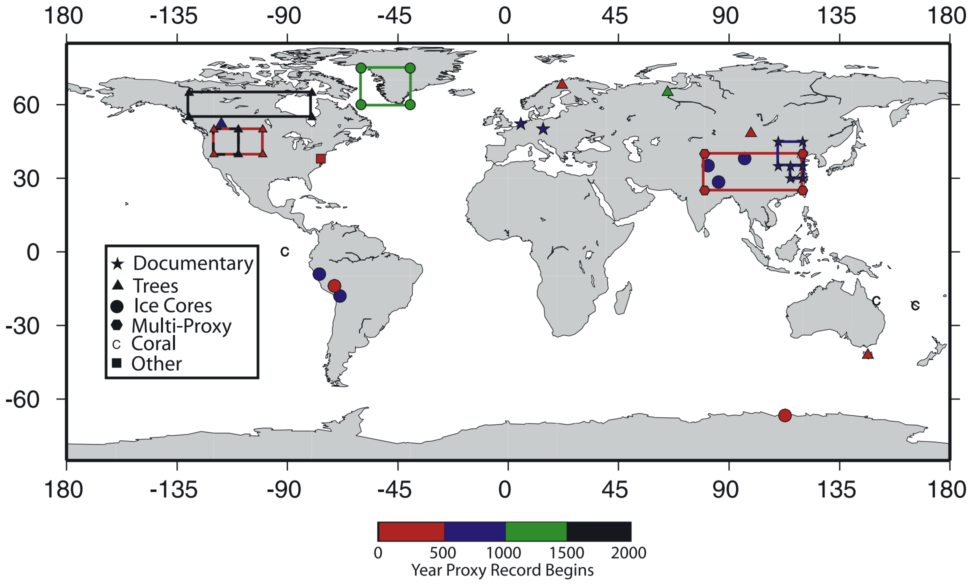


Figure 1. Map of available high-resolution (annual or decadal) proxy indicators with a verifiable temperature signal (see Figure 4) available over much of the past two millennia. The sources of all the data are given in Table 1. The boxes represent the regions of multiproxy (as opposed to single-location proxies) composite series (e.g., for the ice core composite series from Greenland).

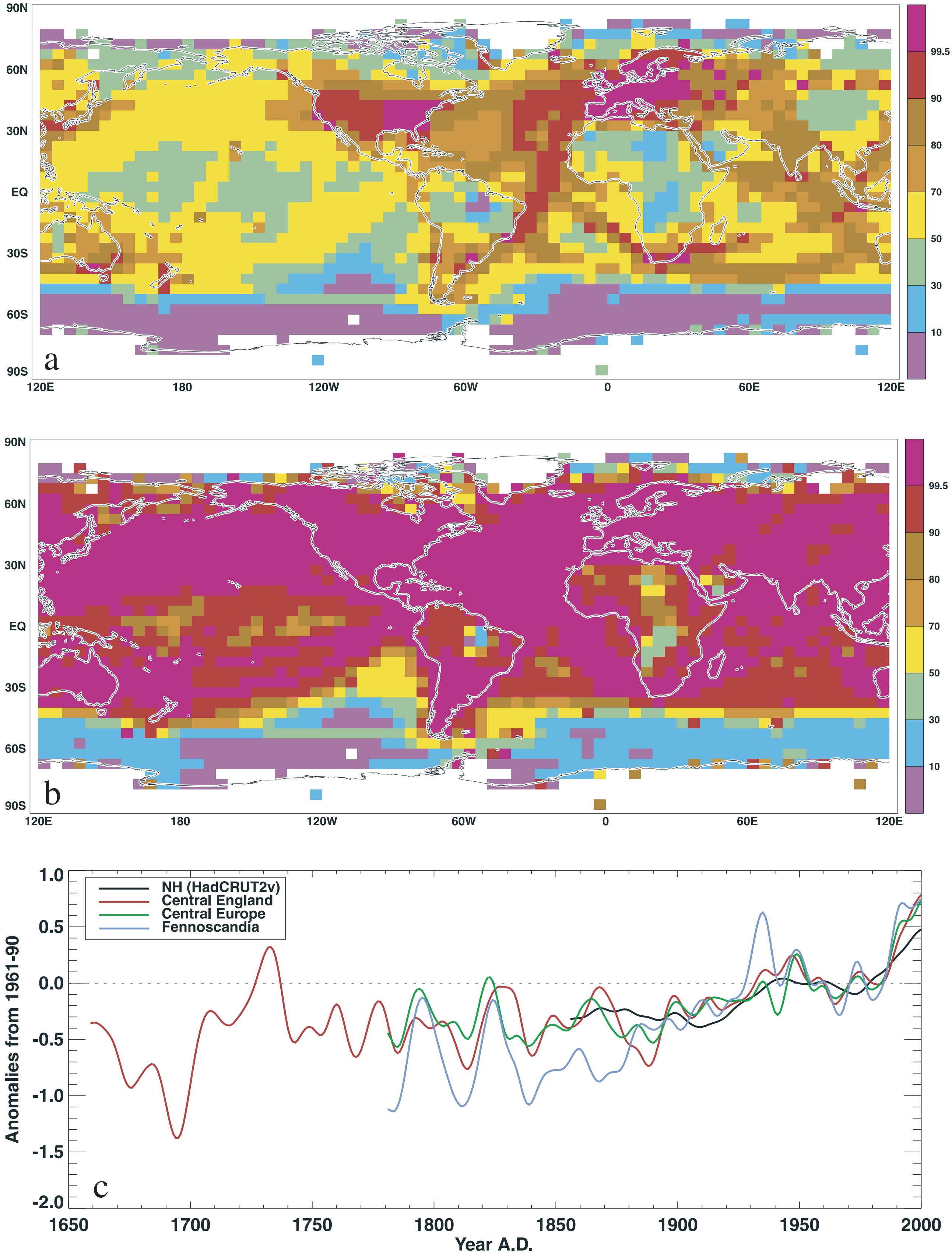
evidence, and terrestrial and ice borehole data must use less direct "spatial" methods of calibration. A brief discussion of the two basic approaches (temporal and spatial) to calibration is given in section 3.1. Such "coarser" proxies are valuable for the insight they provide into broad patterns of climate variability on longer timescales. However, attempts to reconstruct high-frequency climate patterns over the past few centuries to millennia have been restricted to annually to decadally resolved indicators, such as tree rings, corals, ice cores, lake sediments, and the few available multicentury documentary and instrumental series that are available in past centuries (Figure 1).

2.1. Instrumental Climate Data

[11] Instrumental records are by far the most reliable of all available climate data. They are precisely dated, require either no explicit calibration, or employ physically based calibrations (e.g., a mercury thermometer). These data, which include thermometer-based surface temperature measurements from the ocean and land regions, *sea level pressure* (SLP) measurements, continental and oceanic precipitation measurements (including drought indices), sea ice extents, and winds and humidity estimates, are, however, only available on a widespread basis back to ~1850. Moreover, random, systematic, and time-varying biases may exist to some extent in all instrumental data sources [*Jones et al.*, 1999]. Such biases include possible residual urban warming biases in thermometer-based temperature measurements and under-catch issues in precipitation (particularly of snowfall) gauge estimates [e.g., *Folland et al.*, 2001a].

[12] Of primary interest from the point of view of recent global warming and past global temperature trends is the instrumental surface temperature data set. This is based on a compilation of marine (ship-based ocean *sea surface temperature* (SST) observations) and terrestrial (station *surface air temperature* (SAT) measurements) sources. Several compilations provide gridded monthly mean estimates of temperatures on a large-scale basis back through the mid-19th century [e.g., *Jones et al.*, 1986; *Hansen and Lebedeff*, 1987; *Vinnikov et al.*, 1990; *Jones and Briffa*, 1992; *Briffa and Jones*, 1993; *Wigley et al.*, 1997; *Peterson et al.*, 1998; *Hansen et al.*, 1999, 2001; *Jones et al.*, 1999, 2001b]. Coverage becomes increasingly sparse, however, in many parts of the world in earlier decades (see examples in Figures 2a and 2b). A number of studies have used statistical methods to infill the instrumental surface temperature record back through the mid-19th century [*Smith et al.*, 1996; *Kaplan et al.*, 1997, 1998; *Folland et al.*, 2001a, 2001b; *Rutherford et al.*, 2003; *Rayner et al.*, 2003]. The statistical methods employed make use of spatial covariance information from the relatively complete (e.g., post-1950) measurements to interpolate earlier missing data. While useful from the point of view of providing globally com-

Figure 2. Maps and time series of 5° by 5° grid box instrumental temperature data: (a) percent coverage of the Hadley Centre/Climatic Research Unit version 2 (HadCRUT2v) data set [*Jones and Moberg*, 2003] for the 1856–2002 period, (b) percent coverage over the 1951–2002 period, and (c) 20-year smoothed annual average values for the Northern Hemisphere (from HadCRUT2v [*Jones and Moberg*, 2003]), central Europe, Fennoscandia, and central England. The construction of these European series is detailed by *Jones et al.* [2003c] for continental Europe and by *Parker et al.* [1992] for central England.



90N
60N
30N
EQ
30S
60S
90S
120E
180
120W
60W
0
60E
120E
99.5
90
80
70
50
30
10
a
b
c
NH (HadCRUT2v)
Central England
Central Europe
Fennoscandia
Anomalies from 1961-90
1.0
0.5
0.0
-0.5
-1.0
-1.5
-2.0
1650
1700
1750
1800
1850
1900
1950
2000
Year A.D.

Figure 2

plete climate fields, there are many issues to consider with respect to assumptions of stationarity of spatial climate relationships [e.g., *Schneider*, 2001; *Mann and Rutherford*, 2002; *Rutherford et al.*, 2003; *Pauling et al.*, 2003].

[13] Longer instrumental SAT measurements are available on a restricted regional basis for parts of Europe and North America back to the mid-17th century [e.g., *Bradley and Jones*, 1993, 1995b; *Jones*, 2001; *Camuffo and Jones*, 2002; *Chenoweth*, 2003; *Moberg et al.*, 2003]. Many of the European series have been extensively homogenized in a number of recent studies (see references given by *Jones and Moberg* [2003]). Here in Figure 2c we compare smoothed temperature estimates for the *Northern Hemisphere* (NH), central Europe, Fennoscandia, and central England, assessing whether the NH average can be extended back with regional-scale data. Not only is there limited agreement between the NH and the European regions, except on the longest of timescales, but there are also clear differences between the regions. The principal conclusion that can be drawn from Figure 2c is that NH averages clearly cannot be inferred from a single region. Despite this, tentative extensions of the Northern Hemisphere annual mean instrumental record have been attempted based on scaling composites of the sparse instrumental data available back to the mid-18th century against the more complete record of the mid-19th through 20th centuries [*Mann et al.*, 2003a]. These estimates include additional early series from a few locations in eastern North America, India, and Siberia in addition to the European data, so they provide a meaningful, though more highly uncertain, estimate of hemispheric temperature changes.

[14] Measures of the atmospheric circulation are provided by SLP estimates, which are also available, albeit on an increasingly sparse basis, back through the mid-19th century [*Trenberth and Paolino*, 1980; *Allan et al.*, 1996; *Basnett and Parker*, 1997]. Globally infilled reconstructions of SLP have been achieved using the same methods as those discussed above for surface temperature data [e.g., *Kaplan et al.*, 2000]. Regional atmospheric circulation indices such as the *North Atlantic Oscillation* (NAO) index and *Southern Oscillation Index* (SOI), which describe important modes of large-scale atmospheric variability, have been estimated farther back in time based on key long SLP station records. The NAO has been estimated back through the early 19th century, with prospects for instrumental extension (using station pressure data alone) back through the late 17th century [*Jones et al.*, 2003b], while the SOI has been estimated back to the mid-19th century [*Können et al.*, 1998]. A few longer SLP records are available two or more centuries back in parts of Europe [e.g., *Luterbacher et al.*, 2002a, 2002b, and references therein], India [*Allan et al.*, 2002], and Japan [*Können et al.*, 2003].

[15] Finally, as climate change may be altering patterns of hydroclimatic (moisture) variability during the 20th century [e.g., *Karl and Knight*, 1998], it is important to consider the longer-term nature of moisture variations. Land-based precipitation measurements have been amalgamated into gridded data sets [*Hulme*, 1992; *Dai et al.*, 1997; *Hulme et al.*, 1998] and extended to drought indices [*Dai et al.*, 1998; *Cook et al.*, 1998], using gridded temperature and precipitation databases, for all of the 20th century. Some long instrumental precipitation series in Europe, North America, and India are available back through the 18th century [*Bradley and Jones*, 1995b; *Jones*, 2001; *Slonosky*, 2002]. Drought indices based on station precipitation and temperature measurements have been extended back into the mid-19th century at some locations in the United States [*Zhang et al.*, 2004] and Europe [*Jones*, 2001].

2.2. Historical Documentary Records

[16] These sources include records of frost dates, droughts, famines, the freezing of water bodies, duration of snow and sea ice cover, and phenological evidence (e.g., the dates of flowering of plants and the ranges of various species of plants). All can provide insight into past climate conditions [*Wigley et al.*, 1981; *Bradley and Jones*, 1995a; *Bradley*, 1999; *Luterbacher et al.*, 2002a, 2002b]. Documentary evidence is, however, generally limited to regions with long written traditions, such as Europe [e.g., *Bradley and Jones*, 1995a; *Le Roy Ladurie*, 1971; *Martin-Vide and Barriendos*, 1995; *Pfister et al.*, 1998; *Rodrigo et al.*, 1999; *Ogilvie and Jónsson*, 2001; *Brázdil et al.*, 2004], eastern Asia [*Wang and Zhao*, 1981; *Zhang and Crowley*, 1989; *Wang et al.*, 2001], and, more recently, North America [*Bradley and Jones*, 1995a; *Druckenbrod et al.*, 2003; *Overland and Wood*, 2003]. However, logs from Spanish galleons crossing the Pacific Ocean during the 16th–19th centuries provide possible insights into variations in the strength of the prevailing winds [*Garcia et al.*, 2001], and documentary information from South America enables a chronology of El Niño [*Quinn and Neal*, 1992; *Ortlieb*, 2000] to be developed over the past few centuries. Human accounts (e.g., through artistic depictions) of mountain glacier retreats and advances during past centuries provide evidence of climate change on more low-frequency timescales. Some of this type of evidence is more anecdotal, and we discuss this, with examples, at the end of this section. We emphasize here issues in the development of long index series from documentary sources.

[17] Despite the wealth of data available in some regions, documentary sources alone are not useful for detailing truly global-scale climate variations and must, furthermore, be interpreted with caution, as they are not equivalent in their reliability to actual instrumental measurements of meteorological variables. Historical documentary information often emphasizes extreme conditions (i.e., heat waves and cold snaps) [see, e.g., *Pfister*, 1992]. This may provide a highly misleading sense of the true seasonal or annual mean climate anomalies that are of interest in the context of discussions of climate change. It has been shown [e.g., *Thompson and Wallace*, 2001] that change in the phase of the North Atlantic Oscillation/*Arctic Oscillation* (NAO/AO) pattern of winter atmospheric circulation variability (see section 4.4.3) can result in roughly a tripling (Orlando, Florida) to nearly a tenfold (Paris, France) increase in the incidence of cold winter temperature extremes, even in the context of far more modest changes (1° or 2°C) in seasonal

mean temperatures in these regions. It is likely [*Shindell et al.*, 2003] that such atmospheric circulation changes are therefore responsible for many of the reports of extreme cold spells in Europe during the 17th–19th centuries [e.g., *Pfister*, 1992], despite evidence indicating more moderate mean temperature changes [*Mann et al.*, 2000a; *Shindell et al.*, 2003].

[18] Historical documentary climate reconstructions based on combinations of information from different observers may be further biased as the qualitative standards used to reflect climate conditions are likely to be observer-dependent. Does the "wettest summer in living memory" have the same meaning to observers living at least a century apart? Not only is linguistic knowledge vital but so are changes in the meaning of currently used words. The word "gale" took on its present usage just before Francis Beaufort developed his wind force scale (~1810) used ever since by mariners. Before ~1750, however, every wind above force 4 was recorded as a "gale" by English mariners, generally with a qualifier. The current word "breeze" entered the English vocabulary (probably from Catalan) during the later half of the 18th century.

[19] Development of long series therefore needs to be undertaken with care and in a consistent and repeatable way. The principal keys to developing long series of reliable climatic measures are the selective use of contemporarily reported material [*Ingram et al.*, 1981] and rigorous assessment of reporters (e.g., using content analysis [*Baron*, 1982]). It becomes more difficult to verify the fewer records available in earlier periods. In the High Medieval Age (A.D. 1000–1250 [*Bradley et al.*, 2003]), chroniclers in Europe often included reports from neighboring regions and countries, and with even small cities using different calendars and saints' days, the potential for misrecording of events is rife (generally through duplication). All protagonists (exemplified by *Bell and Ogilvie* [1978]) of documentary evidence stress the need to consult the original sources and not to use late 19th or 20th century compilations (such as those of *Easton* [1928] and *Britton* [1937]). Many highly respected climatologists pioneering the use of documentary climate histories in the 1960s [e.g., *Lamb*, 1965; *Bryson*, 1962] sometimes used less than reliable sources of information taken from such questionable compilations [see, e.g., *Ogilvie and Farmer*, 1997]. A most recent and very comprehensive review of the documentary data archive has been provided by *Brázdil et al.* [2004]. Useful additional discussions are given by *Mikami* [1999] and *Jones et al.* [2003a].

[20] Historical documentary series must be considered "proxy" climate records as they require independent calibration against instrumental climate data for any reliable, quantitative interpretation in terms of, for example, seasonal/annual temperature or precipitation/drought variations. This is often extremely difficult, and various approaches have been introduced [*Pfister*, 1984, 1992; *van Engelen et al.*, 2001; *Jones et al.*, 2003a]. It is clear that the quality of the preinstrumental part of any documentary series must be of a lesser quality than the modern instrumental part [*Jones et al.*, 2003a], and users should be mindful of this, particularly with respect to calibration exercises (see section 3.1), when developing multiproxy compilations.

[21] We have thus far emphasized the work of historical climatologists who have combined detailed information and rigorous statistical techniques in order to develop long, continuous, and well-replicated series. Despite these extensive research efforts, anecdotal evidence concerning the last millennium based on factually dubious beliefs is still rife. We note three specific examples that are often misrepresented in terms of their relevance to past climate: (1) the freezing of the River Thames in London in past centuries, (2) the cultivation of vines in medieval England, and (3) the settlement of Iceland and southwestern Greenland about 1000 years ago. It is not unusual to find any one of these examples (often referred to as anecdotal information), and indeed, in some cases, all three simultaneously [*Cutler*, 1997], incorrectly presented as evidence of large-scale warmth or cold in popular accounts of past climate change. Examples can also be found in the peer review scientific literature [*Soon and Baliunas*, 2003; *Soon et al.*, 2003] (see also the discussion given by *Mann et al.* [2003a]).

[22] 1. River Thames freeze-overs (and sometimes frost fairs) only occurred 22 times between 1408 and 1814 [*Lamb*, 1977] when the old London Bridge constricted flow through its multiple piers and restricted the tide with a weir. After the bridge was replaced in the 1830s, the tide came farther upstream, and freezes no longer occurred, despite a number of exceptionally cold winters. The winter of 1962/1963, for example, was the third coldest in the *central England temperature* (CET) record (the longest instrumental record anywhere in the world extending back to 1659 [*Manley*, 1974; *Parker et al.*, 1992]), yet the river only froze upstream of the present tidal limit at Teddington. The CET record clearly indicates that Thames (London) "frost fairs" provide a biased account of British climate changes (let alone larger-scale changes, see Figure 2c) in past centuries.

[23] 2. Monks in medieval England grew vines as wine was required for sacramental purposes. With careful husbandry, vines can be grown today, and indeed, vineyards are found as far north as southern Yorkshire. There are a considerably greater number of active vineyards in England and Wales today (roughly 350) than recorded during medieval times (52 in the *Domesday Book* of A.D. 1086). Vine growing persisted in England throughout the millennium. The process of making sparkling wine was developed in London (by Christopher Merret) in the 17th century, fully 30 years before it began in the Champagne region of France. Thus the oft cited example of past vine growing in England reflects little, if any, on the relative climate changes in the region since medieval times.

[24] 3. Iceland was settled mainly from Norway and the northern British Isles beginning ~A.D. 871. The further migration to SW Greenland approximately one century later, by a small group of Icelanders, was the result primarily of a political and economic need to leave Iceland [*Ogilvie and Jónsson*, 2001]. Climate was not a factor in their decision despite claims otherwise that still appear in

the literature [*Soon and Baliunas*, 2003; *Soon et al.*, 2003]. The SW Greenland settlements survived for many centuries, but in the mid-14th century the more marginal and more northerly located Western Settlement was abandoned. There were a number of reasons for this, including culture and economic factors. However, it seems likely that climate did play a part in the abandonment. The focus of their economy on animal husbandry denied them the advantages of hunting marine and other mammals that ensured the survival of their Inuit neighbors. A series of unusually late springs and cold summers, for example, may have helped to make a marginal situation untenable [*Barlow et al.*, 1997]. The more southerly Eastern Settlement survived to around the mid-15th century [*Buckland et al.*, 1996]. Related myths exist for the North American continent. *Overland and Wood* [2003], for example, have recently demonstrated that, despite past claims that the extreme cold of the "*Little Ice Age*" (LIA) impeded the navigation of a Northwest Passage in the Canadian Arctic during the early 19th century, an exhaustive study of 19th century explorer logs for the region yields no evidence of unusually cold conditions.

2.3. Tree Ring Records

[25] Tree ring or "dendroclimatic" proxy climate indicators can provide information regarding past seasonal temperatures or precipitation/drought, based on measurements of annual ring widths and/or maximum latewood densities [*Fritts*, 1976, 1991; *Fritts et al.*, 1971; *Schweingruber*, 1988; *Jacoby and D'Arrigo*, 1989; *Briffa et al.*, 1992a, 1992b, 1998b; *Hughes and Funkhouser*, 1999; *Stahle et al.*, 1998a, 1998b; *Cook et al.*, 1999; *Briffa and Osborn*, 1999, 2002; *Briffa et al.*, 2001]. While tree ring data are the most widespread source of annually resolved proxy climate information and exhibit some of the strongest statistical relationships with instrumental climate records [*Jones et al.*, 1998], they are limited to where trees can be cross-dated and chronologies developed. They are found in most subpolar and midlatitude terrestrial regions. Generally only extratropical species are useful for climate reconstruction, though there are some exceptions (e.g., tropical teak in Indonesia/Southeast Asia [*Jacoby and D'Arrigo*, 1990; see also *Stahle*, 1999; *Stahle et al.*, 1999]). Tree ring reconstructions offer the advantage of potentially being quite long (e.g., several millennia) [*Hughes and Graumlich*, 1996; *Cook et al.*, 2000; *Briffa and Osborn*, 2002].

[26] The relative strengths and weaknesses of different "standardization" methods for preserving climate variability on multicentennial timescales from tree ring measurements derived by combining shorter length "segments" is actively discussed in the literature [*Cook et al.*, 1995; *Briffa et al.*, 1996; *Esper et al.*, 2002; *Briffa and Osborn*, 2002; *Mann and Hughes*, 2002]. Reconstructions developed in the last decade use conservative techniques such as *regional curve standardization* (RCS) [*Briffa et al.*, 1992a] and *ageband decomposition* (ABD) [*Briffa et al.*, 2001] to develop series that capture more low-frequency climate variability. These approaches are now widely used but they are not applicable for all sites as large numbers of sample series are required to maintain a similar age structure to the chronology over time [*Cook et al.*, 1990; *Mann and Hughes*, 2002]. RCS and ABD approaches are unlikely to be the last word on "standardization," and site selection is still very important [*Fritts et al.*, 1971; *Schweingruber*, 1988]. The emphasis on standardization underscores the importance of longer-timescale information in reconstructions. As instrumental data will always be too short to adequately address this issue, low-frequency variability should be compared, where possible, with information from other proxy sources in the region.

[27] During the most recent decades, there is evidence that the response of tree ring indicators to climate has changed, particularly at higher latitudes and more so for density than ring width measurements [*Briffa et al.*, 1998a]. One suggested source for this behavior is "CO_2 fertilization," the potential enhancement of tree growth at higher ambient CO_2 concentrations. Though it is extremely difficult to establish this existence of this effect [*Wigley et al.*, 1988], there is evidence that it may increase annual ring widths in high-elevation drought-stressed trees [*Graybill and Idso*, 1993]. Recent work making use of climate reconstructions from such trees has typically sought to remove such influences prior to use in climate reconstruction [*Mann et al.*, 1999; *Mann and Jones*, 2003]. Other factors have been suggested as possible explanations for apparent anomalous tree ring/climate relationships [see *Briffa et al.*, 1998a], including the changing seasonality of the climate itself [*Vaganov et al.*, 1999; *Biondi*, 2000; *Druckenbrod et al.*, 2003]. The potential existence of such nonstationary relationships introduces an additional caveat in the use of tree ring data alone for climate reconstruction, since changes in environmental factors in the past could have introduced similar, unknown changes in tree ring response to climate.

[28] Carbon and oxygen isotopes from tree rings have been considered as potential climate proxies, but a straightforward interpretation of the measurements has proven elusive [*White*, 1989]. While there has been some promising recent work in this area [*Leavitt et al.*, 1995; *Robertson et al.*, 1997; *Anderson et al.*, 1998], tree ring isotopes have not been used in large-scale climate reconstruction. Intertree variability needs to be assessed for isotopic measurements. Isotopic analyses require considerably greater effort than traditional tree ring analyses based on ring width and density measurements. As this added effort should be cost-effective relative to the usefulness of the results, it still needs to be demonstrated that fewer numbers of individual cores/trees need be isotopically analyzed to obtain an equivalent climate signal, when simpler methods are available.

2.4. Coral Records

[29] Corals can also provide useful proxy climate information (see *Dunbar and Cole* [1999] for a review). Proxy reconstructions from corals generally rely on geochemical characteristics of the coral skeleton such as temporal variations in trace elements (e.g., Sr/Ca chemical ratios) or stable isotopes (e.g., oxygen isotopes) or, sometimes, on density or

variations in fluorescence [e.g., *Isdale et al.*, 1998] of the coral's aragonite structure. Because they offer information regarding tropical and subtropical maritime regions, corals are complementary in terms of the spatial climate information provided by tree ring data. Corals can be precisely (annually or even seasonally) dated and their environment can be continuously sampled over the full year, providing a potentially more uniform window of climatic information than other more seasonally specific proxies. Because of their potential to sample climate variations in ENSO-sensitive regions [e.g., *Cole et al.*, 1993; *Dunbar et al.*, 1994], a modest network of high-quality coral site records can resolve key patterns of tropical climate variability [*Evans et al.*, 1998, 2002]. However, though recent work has shown some promise in moving toward the possibility of millennial-length reconstructions based on the combination of overlapping multidecadal and century-length fossil corals [*Cobb*, 2002], very long, multicentury records are rare. The longest continuous coral series extend back to the early 16th century [see *Dunbar and Cole*, 1999], but "snapshots" of 50–150 years in duration have been used to assess ENSO variability during the past millennium [*Cobb*, 2002; *Cobb et al.*, 2003a] and various other intervals for the whole Holocene [e.g., *Tudhope et al.*, 2001, and references therein]. Such an approach has important limitations. A fossil coral must be growing at similar depths to its modern counterparts (specifically those corals used to calibrate against the instrumental record), and any inferred climate changes must be measured against the ability of individual modern corals at that site to accurately reflect 20th century climate changes. Diagenetic alterations in fossil corals can be extremely difficult to detect and could significantly bias fossil coral-based estimates of past climate change. Given these limitations, the most useful estimates come from multiple fossil coral "snapshots" from a given time period.

[30] While corals have demonstrated an excellent ability to describe climate variability on interannual and decadal timescales [*Cole et al.*, 1992, 1993, 1995; *Dunbar et al.*, 1994, 1996; *Slowey and Crowley*, 1995; *Wellington et al.*, 1996; *Quinn et al.*, 1996, 1998; *Crowley et al.*, 1997, 1999; *Lough and Barnes*, 1997; *Felis et al.*, 2000; *Linsley et al.*, 2000; *Cobb et al.*, 2001; *Hendy et al.*, 2002], their reliability on longer timescales has not been adequately established. The effects of temperature and salinity (through ocean water oxygen isotopic composition) involve potentially competing and even conflicting influences on coral oxygen isotopic composition. Estimates of low-frequency variability or long-term changes [e.g., *Winter et al.*, 2000] based on calibration of oxygen isotope ratios against instrumental data that are short (a few decades) and/or dominated by a seasonal cycle (and therefore may reflect a balance of salinity and temperature influences only appropriate to that timescale) are therefore unlikely to yield reliable reconstructions of the variable (e.g., SST) of interest [*Crowley et al.*, 1999; *Juillet-Leclerc and Schmidt*, 2001]. Possible influences of "vital effects" (e.g., nonclimatic influences on the coral's biochemistry) may effect the reliability of climate reconstructions based on single specimens [e.g., *Cardinal et al.*, 2001]. Despite these limitations, recent work [e.g., *Correge et al.*, 2001; *Quinn and Sampson*, 2002; *Hendy et al.*, 2002] has pioneered the use of multiple measurements (e.g., Sr/Ca and oxygen isotope ratios) from the same coral to simultaneously estimate SST and salinity influences, and coral paleoclimatology, while still in its early stages, shows considerable promise [see, e.g., *Evans et al.*, 2002].

2.5. Ice Core Records

[31] Ice cores provide climate information over multiple millennia from the polar regions of both the Northern [e.g., *O'Brien et al.*, 1995; *Fisher et al.*, 1996; *Appenzeller et al.*, 1998; *White et al.*, 1997; *Hoffman et al.*, 2001; *Moore et al.*, 2002; *Vinther et al.*, 2003] and Southern [*Peel et al.*, 1996; *Mayewski and Goodwin*, 1997; *Morgan and van Ommen*, 1997] Hemispheres, as well as alpine (tropical and subtropical) environments [*Thompson*, 1992, 1996; *Thompson et al.*, 2000, 2003]. Ice core information is thus spatially complimentary to that provided by either tree rings or corals, but it is available only over a very small fraction of the global surface. Ice cores can provide several climate-related indicators, including the fraction of melting ice, the rate of accumulation of precipitation, and concentrations of various chemical constituents (including trace gases) that provide information about the atmospheric environment [e.g., *O'Brien et al.*, 1995; *Meeker and Mayewski*, 2002] at the time the ice was deposited. Ice cores also contain cosmogenic isotopes of beryllium and volcanic dust, both providing sources of vital information regarding the past radiative forcing of climate (see section 5).

[32] Of primary use from the point of view of climate reconstruction, however, are stable isotopes of oxygen (and, in some cases, deuterium) recorded in ice cores, which can, in principle, be related to temperature-dependent fractionation processes. To the extent that fractionation dominates at the time of deposition, the oxygen isotope ratios can be related to local temperature conditions, providing a "paleothermometer." However, the added influence of nonlocal fractionation on the isotopic ratios measured by a given ice core means that in practice a simple "paleothermometer" calibration remains elusive [*Hoffman et al.*, 1998]. Annual dating is possible in principle, but in practice it is often quite difficult, and generally, the dating is not precise. Use of stratigraphic markers (like known volcanic dust events), where available, can help anchor the age model [e.g., *Clausen et al.*, 1995]. Stacking of multiple annual-timescale ice core records [e.g., *Fisher et al.*, 1996; *Vinther et al.*, 2003] can reduce the age model uncertainty in chronologies and reduce between-site variability. In regions with low snowfall, 1 year's accumulation may represent a small number of discrete precipitation events. In such cases, annual dating may not be possible, and there is furthermore the potential for a substantial temporal (and seasonal) sampling bias. In regions where accumulation is high (e.g., Law Dome, Antarctica) or where a number of cores can be cross-dated like trees (e.g., Greenland), seasonal resolution is possible. Here well-dated proxy estimates of millennial or greater length are possible [*van Ommen and*

Morgan, 1996, 1997; *Fisher et al.*, 1996; *Vinther et al.*, 2003].

2.6. Speleothems

[33] Speleothems (cave deposits) are formed as a part of the meteoric water cycle, and variations in their growth rates and composition reflect environmental changes on the land surface above the cave [*Lauritzen and Lundberg*, 1999, and references therein]. Various proxies have been studied, including growth rates, isotopic composition, trace elements, organic matter content, and luminescent laminae. Locations are limited, but studies have been undertaken at a number of sites in North America, Eurasia, the tropics, southern Africa, and Australasia. Without replication, uncertainties associated with dating the speleothem (generally by laminae counting and uranium series methods) are an issue despite apparent annual resolution at many sites. Interpretation of the possible proxies is also difficult in terms of a simple climate variable (e.g., temperature) because of the multiple potential influences (e.g., source temperature, seasonal changes in rainfall, groundwater residence time, etc.) on the speleothem's geochemistry and/or morphology [see *Bradley*, 1999]. Despite difficulties in interpreting speleothem records they can often provide information about changes in the hydrological cycle and through this links to the atmospheric circulation [*McDermott et al.*, 1999; *Proctor et al.*, 2000; *Frappier et al.*, 2002] and even past cultural changes [*Polyak and Asmerom*, 2001].

2.7. Varved Lake and Ocean Sediment Records

[34] Annually laminated (varved) lake sediments, like ice cores, provide high-resolution proxy climate information in high-latitude regions where other proxy indicators, such as tree rings, are not available [*Lamoureux and Bradley*, 1996; *Hardy et al.*, 1996; *Overpeck et al.*, 1997; *Wohlfarth et al.*, 1998; *Hughen et al.*, 2001]. Varved sediments formed from the deposition of inorganic sediments (clastic) are controlled by seasonal precipitation and summer temperature, both of which govern the volume of meltwater discharge and sediment load into a closed-basin glacial lake. When the latter process dominates the deposition process, the annual (or even subannual) varve thicknesses provide inferences into summer temperature variations [e.g., *Overpeck et al.*, 1997; *Hughen et al.*, 2001]. In rare cases, annually varved [*Hughen et al.*, 1996; *Biondi et al.*, 1997; *Black et al.*, 1999] sediments can be obtained from marine (coastal or estuarine) environments where deposition is unusually high, and proxy climate information can be obtained from oxygen isotopes or faunal assemblages contained within the sediments. If adequate radiometric dating is available (e.g., through pollen biostratigraphy), nonvarved lake [e.g., *Laird et al.*, 1996, 1998; *Moy et al.*, 2002] and estuarine [*Cronin et al.*, 2003] sediments can still provide useful climate information at decadal timescales in past centuries.

2.8. Boreholes

[35] Boreholes can provide estimates of past temperature variability, though their resolution is intrinsically multidecadal at its finest and decreases back in time. Borehole-derived temperature estimates come from two distinct sources: widespread terrestrial boreholes in tropical, midlatitude, and subpolar environments [e.g., *Beltrami et al.*, 1995; *Harris and Chapman*, 1997, 2001; *Pollack et al.*, 1998; *Huang et al.*, 2000; *Mann et al.*, 2003b] and ice boreholes from polar ice caps [e.g., *Dahl-Jensen et al.*, 1998]. In both cases the subsurface borehole temperature profiles are used to obtain an estimate of *ground* (or ice) *surface temperature* (GST) changes back in time through solution of an appropriate inverse problem. This is generally achieved by inverting the diffusion equation [e.g., *Shen and Beck*, 1991] that describes the downward propagation of surface temperatures into the ground and requires some a priori knowledge and several assumptions. These include the homogeneity of the medium and absence of advective heat transport by fluid flow [see *Shen et al.*, 1995], the thermal properties of the medium, and the nature of the background geothermal heat flux that must be removed to yield an estimate of transient surface-forced temperature variability [see *Harris and Chapman*, 1997; *Pollack et al.*, 1998; *Huang et al.*, 2000; *Harris and Chapman*, 2001].

[36] While borehole temperature profiles, in principle, provide a direct measurement of GST histories, GST potentially differs in some fundamentally important ways from the quantity, surface air temperature (SAT), which is typically of interest in discussions of large-scale climate variability. GST is impacted by changes in seasonal snow cover [e.g., *Zhang et al.*, 2003; *Mann and Schmidt*, 2003; *Stieglitz et al.*, 2003; *Beltrami and Kellman*, 2003], land surface changes [e.g., *Skinner and Majorowicz*, 1999], and various other potential factors (often seasonally specific) unrelated to SAT changes. These complicate interpretation of GST estimates in terms of past changes in overlying SAT [*Folland et al.*, 2001a; *Mann et al.*, 2003b; *Mann and Schmidt*, 2003; *Beltrami and Kellman*, 2003]. The existence of a cold season bias has now been demonstrated in two climate model simulations [*Mann and Schmidt*, 2003; *Gonzalez-Rouco et al.*, 2003], but the implications for past annual mean temperature changes are unclear [*Gonzalez-Rouco et al.*, 2003]. These depend on the actual history of cold season snow cover changes in past centuries, which cannot be recovered in any single model simulation. An additional consideration is the restricted and spatially variable sampling of past GST variations provided by available global borehole networks [*Pollack et al.*, 1998; *Huang et al.*, 2000], which may yield a biased estimate of large-scale mean temperature change [*Briffa and Osborn*, 2002; *Mann et al.*, 2003b]. Despite the various possible sources of bias in estimating large-scale SAT trends in past centuries from borehole data, the apparent discrepancies with other proxy-based estimates appear to be diminished when an attempt is made to account for spatial sampling biases [*Briffa and Osborn*, 2002; *Mann et al.*, 2003b] and when an optimal estimate is made via spatial regression [*Mann et al.*, 2003b; *Rutherford and Mann*, 2004] of the actual SAT signal contained within the borehole GST estimates (Figure 3).

[37] Ice borehole records are less likely to suffer from many of these sources of bias, but a number of assumptions

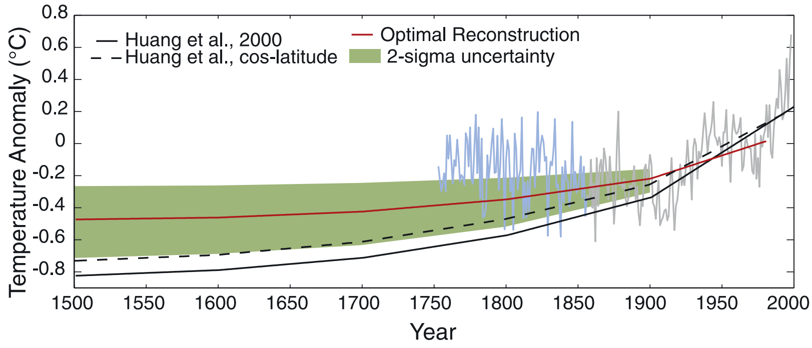


Figure 3. Estimates of Northern Hemisphere mean surface temperature changes over the past five centuries from terrestrial borehole data are shown, based on a simple (unweighted) arithmetic average of available borehole data [*Huang et al.*, 2000], an areally weighted average of the borehole data, and a statistically optimal hemispheric surface temperature estimate (and its 2 standard error uncertainties, shaded region) from the available borehole data [*Mann et al.*, 2003b]. Shown for comparison is the instrumental record of Northern Hemisphere temperature through the mid-19th century (based on HadCRUT2v [*Jones and Moberg*, 2003]) and an extension back through the mid-18th century based on sparser instrumental data [*Mann et al.*, 2003b]. The areally weighted and "optimal" borehole estimates are revised relative to *Mann et al.* [2003b] based on the correction of an error in the normalization of the areal weighting that has been identified in the original analysis [*Rutherford and Mann*, 2004].

when inverting the borehole profile to obtain an estimate of past ice surface temperatures, such as the presence of ice flow over time, will introduce potential chronological error into the record. They provide valuable estimates of past temperature changes in polar environments that are complementary to those obtained from ice core oxygen isotopes [*Dahl-Jensen et al.*, 1998]. Significant differences between ice borehole temperature estimates obtained at relatively nearby sites (Dye3 and Greenland Ice Core Project [see *Dahl-Jensen et al.*, 1998]) suggest that the contribution of local noise or possible biases in isotope records may be substantial. Inferences with regard to past temperature changes from the two locations should be interpreted with appropriate caveats. They are sufficiently reliable, however, to indicate that temporal calibration techniques must be applied to ice core isotopic measurements. Use of the seasonal cycle or spatial variations in isotopic ratios to determine a temperature scale cannot be objectively justified (see also section 3.1).

2.9. Glacial Evidence

[38] The position of glacial moraines (till left behind by receding glaciers) can provide information on the advances (and, less accurately, the retreats) of mountain glaciers in past centuries. This can be interpreted, at least indirectly, as past changes in regional glacial mass balance, which reflect (though in a relatively complex way) past climate variations [e.g., *Grove and Switsur*, 1994]. Simple interpretation of such information in terms of past surface air temperature and precipitation changes is, however, confounded by the delicate balance between local changes in melting and ice accumulation and the variable response timescale of glacial mass balance (which typically increases with the size of the glacier, potentially approaching century timescales for large mountain glaciers). In other words, glacial moraine evidence may be indicative of past climate variability, but it cannot, despite claims sometimes made to the contrary [e.g., *Broecker*, 2001], be uniquely interpreted in terms of specific past temperature or precipitation histories. Both increased winter precipitation (through greater accumulation) and lower summer temperatures (through decreased melting or "ablation") can lead to more positive glacial mass balances. While the response to climate forcing may be close to contemporaneous for small, fast moving glaciers, the inertia of large glaciers dictates that they respond to climate change relatively slowly, with delays of decades or occasionally centuries [*Jöhanneson et al.*, 1989]. A more promising approach to making use of glacial moraine evidence for paleoclimatic interpretation involves the use of forward modeling in which the history of past glacial mass balance is compared to the evolution of a glacial mass balance model using reconstructions from high-frequency proxy indicators (e.g., many tree-based summer temperature reconstructions) or possibly climate model estimates of past and future temperature and precipitation scenarios as boundary conditions [*Oerlemans*, 1992; *Raper et al.*, 1996; *Folland et al.*, 2001a; *Reichert et al.*, 2002].

2.10. Other Proxy Records

[39] Other possible high-resolution proxy indicators are potentially useful but as yet are still under development in terms of their use in climate reconstruction. These include isotopes from molluscs [*Weidman and Jones*, 1994]. In some shallow seas of the North Atlantic region it should prove possible (but with considerable effort) to develop annual-timescale chronologies for the late Holocene based on the clam, *Arctica Islandica*.

[40] In addition, lower-resolution indicators that are difficult to calibrate against instrumental records can, nonetheless, sometimes assist our understanding of long-term climate variability from at least a qualitative point of view. For example, open ocean sedimentation rates may some-

times be sufficiently high that century-scale or at least multicentury variability can be resolved, even if annual "varves" or other means of high-resolution dating are not available [e.g., *Keigwin*, 1996; *Keigwin and Pickart*, 1999; *Anderson et al.*, 2002]. In such cases, however, any quantitative reconstructions of climate variables based on the use of "paleothermometer" calibration approaches (e.g., the "spatial" approach, see section 3.1 and discussion in sections 2.4 and 2.5) are potentially unreliable. Also, in many cases, age determination relies on radiometric methods with relatively poor age control, and uncertainties may be of the order of a century or more. Quantitative estimates of past climate variability from sediment cores should thus be interpreted with caution. Similar arguments hold for any low-resolution reconstructions (sediment cores or pollen data) for which an a priori (e.g., paleothermometer or spatial transfer function) rather than an explicit (i.e., correlation against instrumental data) calibration technique is used to establish a quantitative scale for the variable of interest [see, e.g., *Bradley*, 1999].

[41] One of the most widespread low-temporal-resolution proxies is past pollen records from cores taken on land or from lakes (see European summary by *Huntley and Prentice* [1993]). Inferences about the climate can be made on timescales of 500-1000 years, but, as with varved sediments, they are subject to dating uncertainties. The vegetation composition of a region takes some time to respond to climate change, so interpretation can be complicated by both response time issues and uncertainty in the nature of the underlying climate signal. Faunal remains (e.g., beetles and chironomids) offer the advantage of a contemporaneous response to underlying climate changes but are subject to similar dating uncertainties as pollen [*Atkinson et al.*, 1987]. Another low-resolution indication of past climate is provided by earlier tree line positions (see *Shiyatov* [1993] but see also *Nichols* [1975]). Climatic inferences here might be specific to short periods when trees were able to survive beyond the sapling stage at higher elevations than present. Once established, trees can survive for many hundreds of years, so that indications of warmer conditions may only be representative of those during their early years of life.

3. PROXY CLIMATE RECONSTRUCTIONS

3.1. Calibration of Proxy Data

[42] There are two basic approaches to the calibration of proxies against instrumental climate data. Both are based on linear regression methods but can be distinguished by their use of temporal or spatial relationships. With the temporal approach the strength of the climate "signal" in the proxy is assessed by time series regression against instrumental climate data. Much of the terminology used comes from the field of dendroclimatology, for which there is a long tradition of the development and application of rigorous statistical methods [e.g., *Fritts et al.*, 1971; *Fritts*, 1991; *Cook and Briffa*, 1990]. The first step in the regression process is to assess which part of the year or season the proxy responds to. The quantitative relationship between the proxy and seasonal/annual instrumental data is then determined for a "calibration" period with some instrumental data withheld to assess the veracity of the relationship with independent data (the "verification" period) [*Briffa*, 1995]. If statistical measures are considered adequate, the principle of uniformitarianism [e.g., *Bradley*, 1999] is invoked to reconstruct the climate series from the earlier variability of the proxy.

[43] In the spatial approach (exemplified in palynology) the quality of the proxy climatic signal is determined by assemblages of the proxy (e.g., pollen) in today's climate from a number of locations. A spatial transfer function [e.g., *Berglund et al.*, 1996] is then developed that takes assemblages in earlier times to infer past climate. A slight alternate to this has often been used in ice cores and corals. In this, isotopic ratios (e.g., $\delta^{18}O$ or δD) are regarded as direct measures of temperature with the constants of proportionality determined by the spatial variability of the isotopic ratios and temperature. Time series calibration appears superior, however, to the use of spatial/geographical variations. The latter approach has been shown to be unreliable with respect to isotope/temperature relationships (e.g., the reassessments in polar ice cores with borehole approaches [*Dahl-Jensen et al.*, 1998]) and in the case of corals with respect to calibration using the seasonal cycle [*Crowley et al.*, 1999].

[44] The use of instrumental climate data records is an essential component of high-resolution paleoclimatology, as it provides the quantitative information against which proxy climatic indicators must be calibrated. The longer the instrumental record in the vicinity of the proxy site, the greater the potential for accurate assessment of the fidelity of a proxy reconstruction achieved by local calibration approaches. This holds true particularly with respect to assessments of the stationarity and timescale of response [*Jones et al.*, 1998]. Calibration against local instrumental data is clearly much more of a challenge in data-sparse regions such as the Antarctic and the continental interiors of South America and Africa. Some recent progress has been made in the blending of modern satellite and in situ surface temperature measurements in Antarctica [*Schneider and Steig*, 2002], which should prove useful for the calibration of Antarctic ice core data [*Steig and Schneider*, 2002].

[45] For all annually resolved proxy data (sections 2.2–2.7) the temporal approach provides the most reliable measures of potential proxy performance in the preinstrumental period. Here we reassess all the temperature proxy series we use (those displayed in Figure 4) by comparing the series with instrumental data. The results are given in Table 1. Correlations are calculated for the 1901–1980 period at both the annual and decadal timescale (for annually resolved series) and just the decadal timescale for less well resolved series. The upper limit of 1980 was chosen because many of the proxy series are not available for the most recent decades, while the lower limit of 1901 was chosen because of the increasing sparseness of the instru-

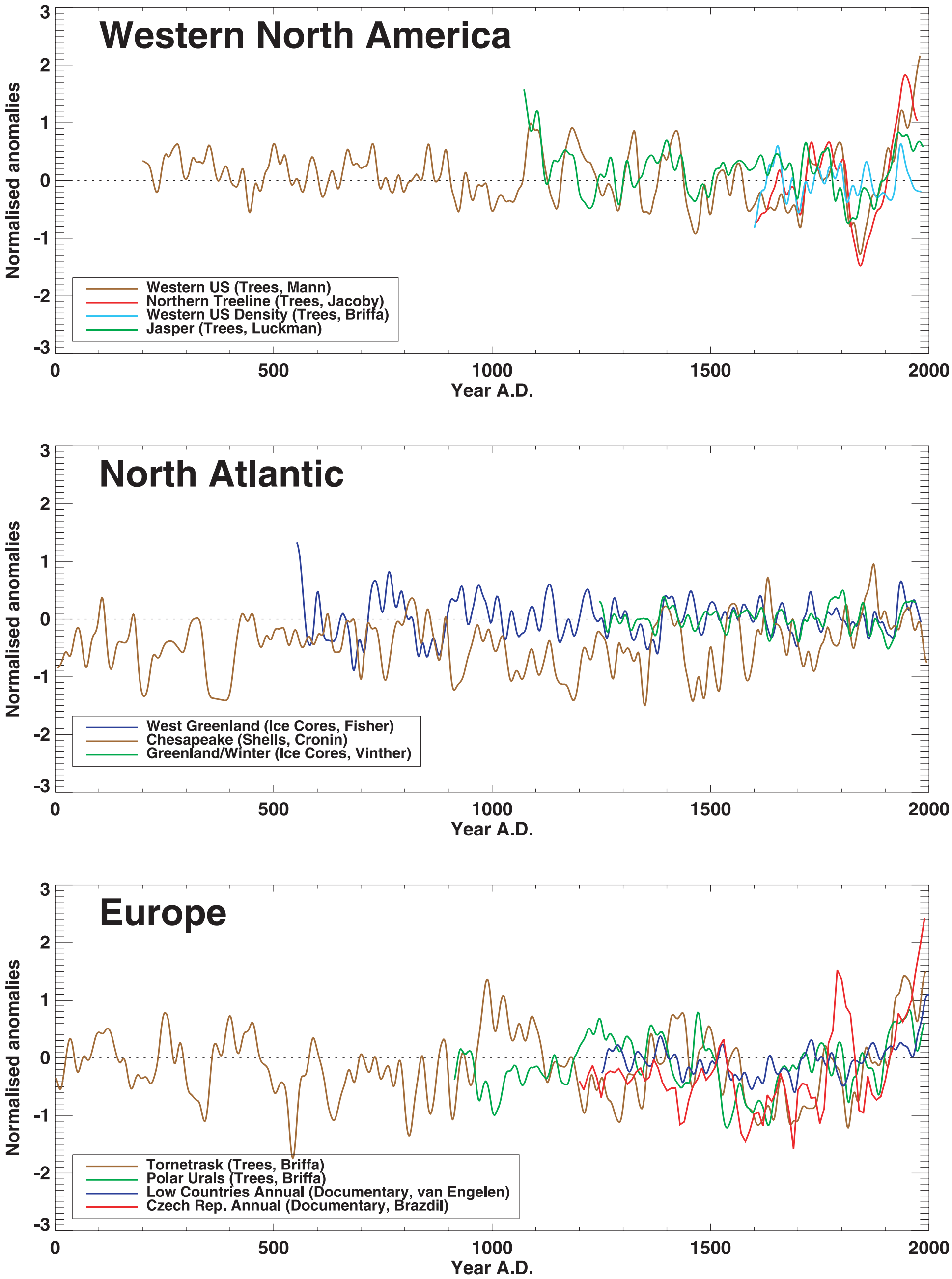


Figure 4. Local and regional proxy temperature reconstructions by continent. Source references for all series are given in Table 1. Each series has been normalized over the period 1751–1950 and then smoothed with a 50-year Gaussian filter. For the decadally resolved data the normalization period is the 20 decades from 1750 to 1949, smoothed using a 5-decade filter.

mental record prior to the 20th century. For the majority of series we compare with instrumental calendar year average temperatures (based on individual or multiple 5° by 5° grid boxes from the *Hadley Centre/Climatic Research Unit version 2* (HadCRUT2v) data set of *Jones and Moberg* [2003]). For some higher-latitude NH tree-based reconstructions we use a more appropriate "growing season" of May to September. Southern Hemisphere series might be improved slightly with a possibly more relevant "year" such as July to June. However, for consistency of comparison we have considered just the two seasonal response windows (principally, for reference to the hemispheric and global composite series and their components, discussed with Figures 5 and 4, respectively). All correlations (with the exception of the Galapagos Islands coral) improve at the decadal timescales, although for some the increase is small and the significance less when the reduced number of degrees of freedom is allowed for. A number of other temperature reconstructions used in earlier multiproxy composites or in review papers [e.g., *Jones et al.*, 1998; *Mann et al.*, 1998a, 1999; *Mann and Jones*, 2003] are not included. This is because they are either less resolved than decadal resolution [e.g., *Dahl-Jensen et al.*, 1998] or correlations with local grid box temperatures are not significant [e.g.,

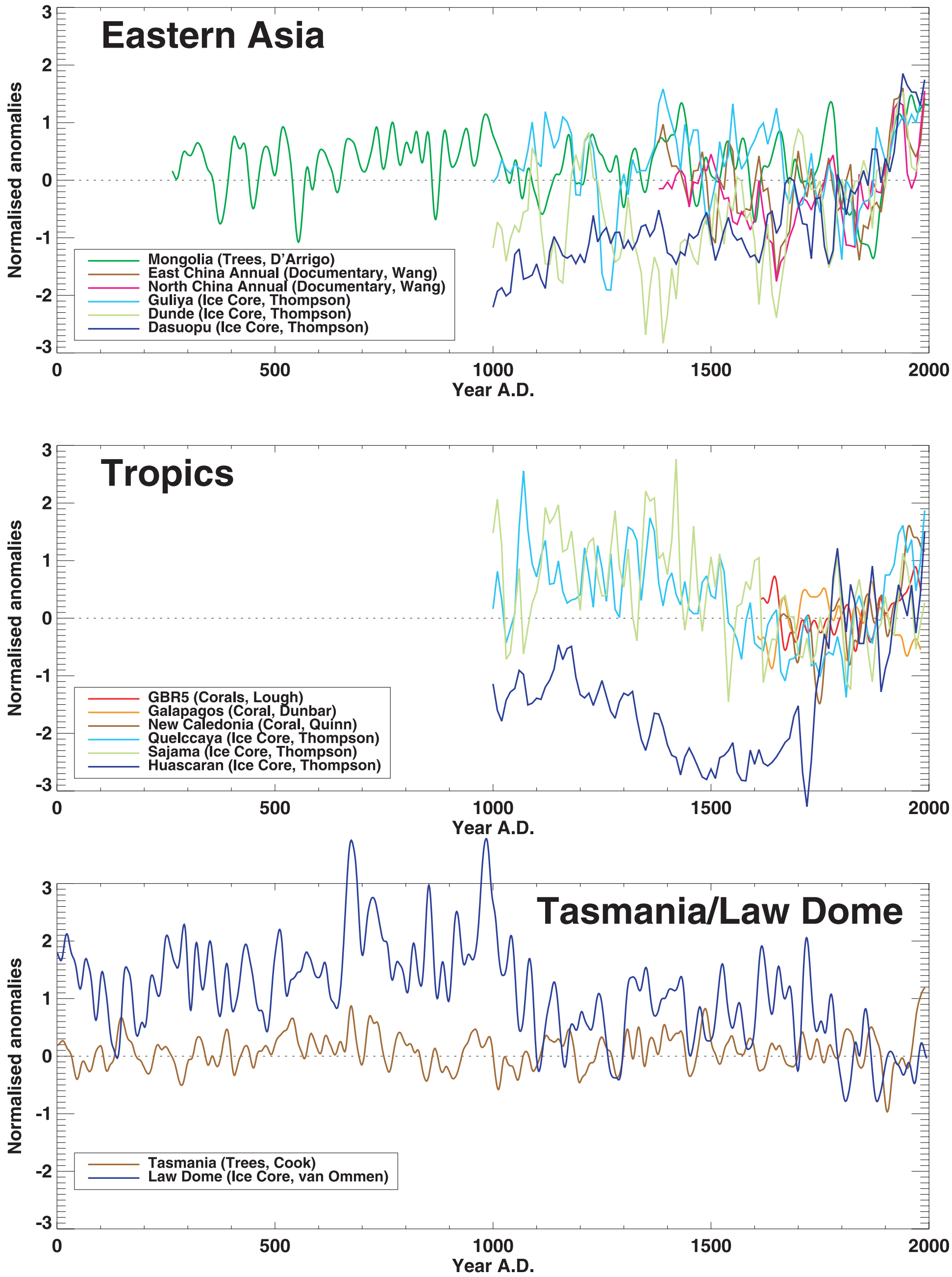


Figure 4. (continued)

Villalba, 1990; *Lara and Villalba*, 1993; *Cook et al.*, 2002b] for the two seasonal response windows used.

3.2. Climate Field Reconstruction Approaches

[46] While individual proxy records can be calibrated against local climate data through the methods discussed in section 3.1, it is sometimes more effective to reconstruct entire (regional, hemispheric, or near global) climate fields through multivariate statistical or *climate field reconstruction* (CFR) approaches. Such approaches have been widely used in applications that infill missing data in instrumental climate fields [*Reynolds and Smith*, 1994; *Smith et al.*, 1996; *Basnett and Parker*, 1997; *Kaplan et al.*, 1997, 1998, 2000; *Schneider*, 2001; *Mann and Rutherford*, 2002; *Rutherford et al.*, 2003; *Rayner et al.*, 2003], but related statistical approaches also have a long history in paleoclimate field reconstruction applications [*Fritts et al.*, 1971; *Fritts*, 1976, 1991; *Kutzbach and Guetter*, 1980; *Guiot*, 1985, 1988; *Briffa et al.*, 1986; *Cook et al.*, 1994; *Mann et al.*, 1998a, 1998b, 1999, 2000a, 2000b; *Mann and Rutherford*, 2002; *Luterbacher et al.*, 2002b; *Evans et al.*, 2002; *Pauling et al.*, 2003; *Zhang et al.*, 2004; *Rutherford et al.*, 2004].

[47] In the paleoclimatic context, CFR seeks to reconstruct a large-scale field, such as surface temperature [e.g., *Mann et al.*, 1998a, 1998b; *Evans et al.*, 1998, 2002; *Zorita et al.*, 2003; *Pauling et al.*, 2003], sea level pressure [*Luterbacher et al.*, 2002b], or continental drought [*Zhang et al.*, 2004]. CFR uses a spatial network of proxy indicators (either of the same type or a "multiproxy" set consisting of

TABLE 1. Correlations Between Proxy Series and Local Instrumental Data[a]

Location/Site	*Proxy Type*	*Reference*	*Annual Correlation (r) (1901–1980)*	*Decadal r[b] (1901–1980)*
Trees				
Tornetrask	density/widths	*Briffa et al.* [1992a]	0.18[c]	0.54[c]
Polar Urals	density/widths	*Briffa et al.* [1995]	0.78[c]	0.85[c]
Mongolia	widths	*D'Arrigo et al.* [2001]	0.25	0.40
Northern tree line	widths	*Jacoby and D'Arrigo* [1989]	0.36[d]	0.71[d]
Western United States	widths (PC)	*Mann et al.* [1998a]	0.20	0.61
Western United States	density	*Briffa et al.* [1992b]	0.64[c]	0.66[c]
Jasper	density/widths	*Luckman et al.* [1997]	0.44[c]	0.49[c]
Tasmania	widths	*Cook et al.* [2000]	0.58	0.79
Corals				
Great Barrier Reef	growth layers	*Lough and Barnes* [1997]	0.18[d]	0.47[d]
Galapagos	isotopes	*Dunbar et al.* [1994]	0.46[d]	*0.12*[d]
New Caledonia	isotopes	*Quinn et al.* [1998]	0.28[d]	0.51[d]
Ice cores				
Greenland	isotopes (six-site average)	*Fisher et al.* [1996]	0.58	0.75
Greenland	isotopes (seven-site average)	*Vinther et al.* [2003]	0.57[d]	0.78[d]
Law Dome	isotopes	T. D. van Ommen et al. (personal communication, 2003)	0.46[d]	0.76[d]
Quelccaya	isotopes	*Thompson et al.* [2003]	–	0.70
Sajama	isotopes	*Thompson et al.* [2003]	–	*0.21*
Huascaran	isotopes	*Thompson et al.* [2003]	–	0.80
Guliya	isotopes	*Thompson et al.* [2003]	–	*0.45*
Dunde	isotopes	*Thompson et al.* [2003]	–	*0.32*
Dasuopu	isotopes	*Thompson et al.* [2003]	–	0.55
Documentary				
Low Countries	historical sources	*van Engelen et al.* [2001]	0.73	0.83
Czech Republic	historical sources	*Brázdil* [1996]	–	0.94
East China	historical sources	*Wang et al.* [2001]	–	0.50
North China	historical sources	*Wang et al.* [2001]	–	0.52
China[e]	mixture	*Yang et al.* [2002]	*0.17*	*0.22*
Shells				
Chesapeake Bay	shells	*Cronin et al.* [2003]	–	0.32

[a]Calendar year averages are given except where stated, with italicized values not significant at the 5% confidence level. All series with the exception of China are plotted in Figure 4. Locations can also be seen in Figure 1. Instrumental data are either for the overlying 5° by 5° grid box (for single-site proxies) or averages of several boxes (for regional or multiproxy series) shown in Figure 1. Generally, for single-site proxies, more than one individual grid box temperature series is used to enable the instrumental series to be both longer and complete.

[b]Decadal correlations are based directly on either decadally resolved proxies or on decadal filtering (Gaussian with 13 terms) of annually resolved proxies.

[c]Value is for May to September season.

[d]Correlation is based on <80 years, because proxy series ends before 1980 or instrumental records are not available in the early 20th century. Number of years is always >65 years, except for Law Dome when it is 24.

[e]As the work of *Yang et al.* [2002] is composed of many of the ice core and documentary series from the region, it is not plotted in Figure 4.

Figure 5. Reconstructions of (a) Northern Hemisphere (NH), (b) Southern Hemisphere (SH), and (c) global mean (GLB) annual temperatures over the past one to two millennia. The expansion in Figure 5a compares a number of different NH estimates over the past 1000 years, while the main plot shows the proxy reconstructions back to A.D. 200 of *Mann and Jones* [2003], updated through 1995 as described in the text. Smoothed (40-year low-passed) versions of these series are shown to highlight the low-frequency variations. We avoid an ad hoc smoothing approach [e.g., *Soon et al.*, 2004] by employing, as in the work of *Mann and Jones* [2003], objectively determined boundary constraints. We make use in each case of the boundary constraint among the three lowest-order constraints [*Park*, 1992] that minimizes the mean square error between the smoothed and raw data [see *Ghil et al.*, 2002; *Mann*, 2004]. The instrumental NH [*Jones et al.*, 1999], SH, and GLB series (red) through A.D. 2003, smoothed in this manner, are shown for comparison. Shown also (yellow shading) is the 95% confidence interval in the reconstruction (i.e., the positive and negative 2 standard error limits of the smoothed reconstructions). The various other (smoothed) NH reconstructions shown in the enlargement to Figure 5a have been scaled by linear regression against the smoothed instrumental NH series over the common interval 1856–1980, with the exception of the "Briffa et al." series, which has been scaled over the shorter 1856–1940 interval owing to a decline in temperature response in the underlying data discussed elsewhere [*Briffa et al.*, 1998a]. The *Crowley and Lowery* [2000] series shown here replaces an incorrect version of the series shown in similar previous comparisons [e.g., *Mann et al.*, 2003a].

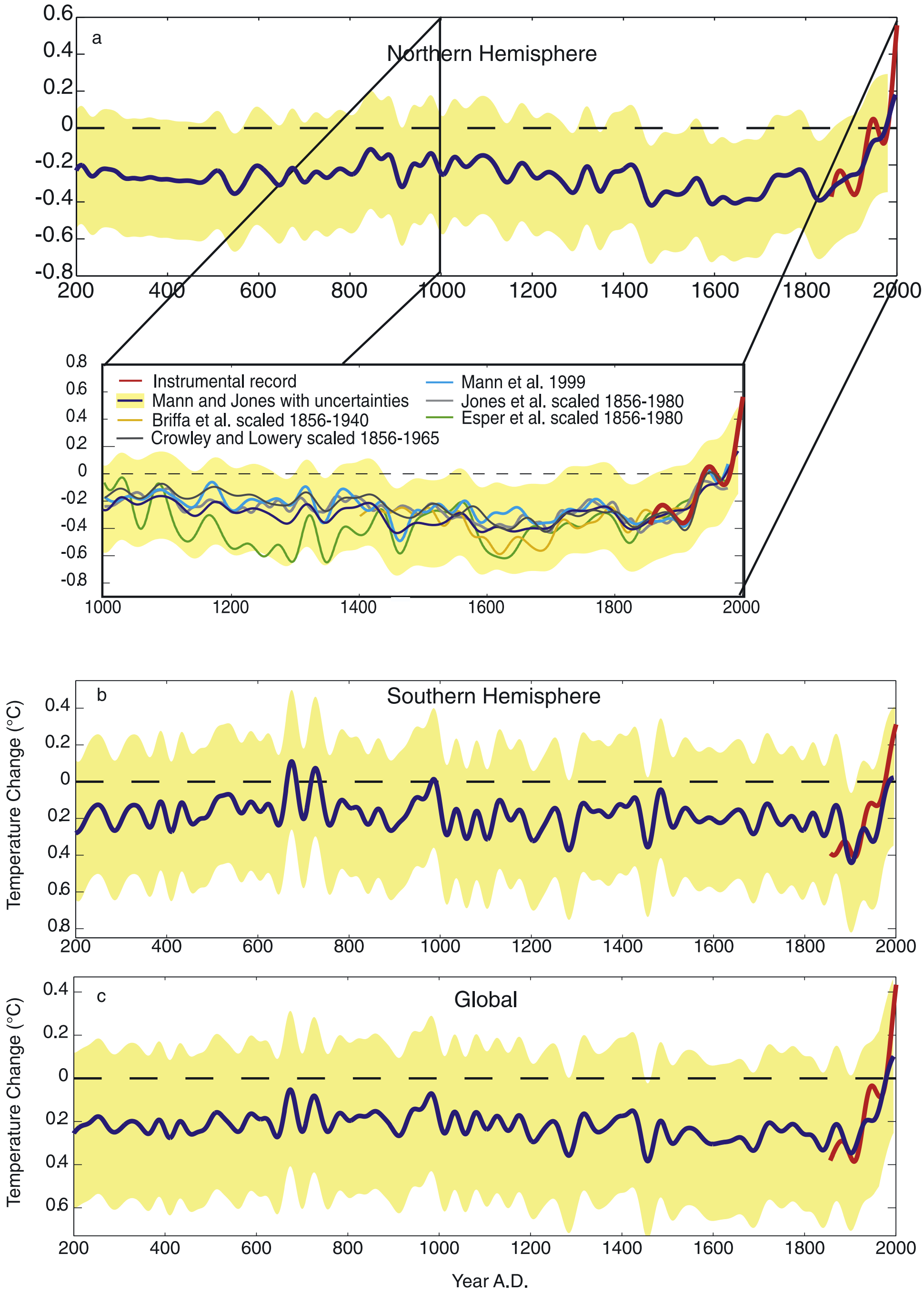


Figure 5

different types of proxy data), performing a multivariate calibration of the large-scale information in the proxy data network against the available instrumental data. As in the simple time series calibration approach described in section 3.1, verification of the skill of the reconstruction through "cross validation" against independent instrumental data is essential [e.g., *Cook et al.*, 1994; *Mann et al.*, 1998a; *Luterbacher et al.*, 2002b]. While various methods for accomplishing such a multivariate calibration exist, most involve the use of empirical eigenvectors of the instrumental data, the proxy data, or both [*Fritts et al.*, 1971; *Guiot*, 1985, 1988; *Briffa et al.*, 1986; *Cook et al.*, 1994; *Mann et al.*, 1998a, 1999, 2000a, 2000b; *Evans et al.*, 2002]. Because the large-scale field is simultaneously calibrated against the full information in the network, there is no a priori local relationship assumed between proxy indicator and climatic variable. All indicators should, however, have been shown to respond to some aspect of local climate during part of the year.

[48] CFR uses a larger number of indicators (including both "original" proxy series and/or large-scale estimates already expressed as temperature anomalies by earlier local/regional-scale calibration) in the reconstruction of a particular climate field (e.g., surface temperature) compared to the "local calibration" approach described in section 3.1. For example, an indicator of convection/rainfall in the central

tropical Pacific indicative of ENSO variability can be used to calibrate the predominant surface temperature patterns associated with ENSO back in time. Large-scale climate fields typically also have relatively few spatial degrees of freedom (on the order of tens) on climatic timescales [*Jones and Briffa*, 1992; *Mann and Park*, 1994; *Weber and Madden*, 1995; *Zwiers and Shen*, 1997; *Jones et al.*, 1997a; *Rutherford et al.*, 2003]. They can often be efficiently reconstructed from a relatively modest network of proxy indicators [e.g., *Bradley*, 1996; *Evans et al.*, 1998; *Mann and Rutherford*, 2002; *Zorita et al.*, 2003; *Pauling et al.*, 2003]. The quality of the reconstruction, however, may depend not just on the number/density and quality of proxy indicators but also on the specific locations available, since this determines whether or not key large-scale patterns of variance (e.g., those associated with ENSO) are likely to be captured. This is especially true when relatively small numbers of indicators are available [*Bradley*, 1996; *Zwiers and Shen*, 1997; *Evans et al.*, 1998; *Mann and Rutherford*, 2002; *Rutherford et al.*, 2003; *Zorita et al.*, 2003; *Pauling et al.*, 2003]. Objective estimates of errors and self-consistent uncertainties in the reconstructions can be obtained from calibration diagnostics within CFR techniques [*Kaplan et al.*, 1998; *Mann et al.*, 1998a, 1998b, 1999; *Schneider*, 2001; *Rutherford et al.*, 2003, 2004]. Such uncertainty estimates do not take into account uncertainties in the instrumental temperature data against which the proxy indicators are calibrated nor the possibility of expanded uncertainties in the proxy data prior to the calibration/cross-validation periods. The former are generally negligible compared to other contributions toward uncertainty [e.g., *Folland et al.*, 2001b]. The latter contribution may in some cases be sizable, suggesting the circumspect use, for example, of tree ring series that represent a composite of a decreasing number of contributing trees in the chronologies back in time or of certain historical documentary series when earlier information is less reliable than the more recent "instrumental" part of the series.

[49] CFR approaches arguably invoke stronger assumptions regarding the stationarity of relationships between proxy indicators and large-scale patterns of climate variability than the "local calibration" approach described in section 3.1. For example, changes over time in the base state of the tropical Pacific climate could affect the relationship between convection in the central tropical Pacific recorded by a salinity-sensitive coral indicator in the western Pacific and the deduced pattern of SST variation in the eastern tropical Pacific. Alternatively, it might seem reasonable to assume that nonstationarity in the eigenstructure of the 20th century climate record due to anthropogenic forcing of climate could bias reconstructions of the past based on calibration of proxy indicators against modern instrumental data. Experiments with both control and anthropogenically forced model simulations, however, suggest that such nonstationarity of relationships are unlikely to impact large-scale CFR approaches to reconstructing the variability during recent past centuries [*Rutherford et al.*, 2003]. It is likely that these underlying stationarity assumptions do not hold during the more distant past (see section 1.1). During the early to mid-Holocene and, even more likely, the Pleistocene, orbital and glacial boundary conditions on the climate were fundamentally different from their current state, and the structure of the prevailing atmospheric circulation itself was probably fundamentally altered from its present-day configuration [*Yin and Battisti*, 2001]. Clearly, more work remains to be done in using climate model simulations [e.g., *Rutherford et al.*, 2003; *Zorita et al.*, 2003] to explore the relative strengths and weaknesses of various alternative approaches to the CFR problem.

3.3. Multiproxy Reconstructions

[50] As discussed in section 2, the various high-resolution (annually or decadally resolved) proxy indicators (historical documents, tree rings, corals, ice cores, and varved sediment records) are often complementary in terms of their sampling of different regions of the globe (tropics versus extratropics versus polar regions and terrestrial versus marine environments) and in their sampling of seasonal variations (warm season versus cold season versus annual). Moreover, as also discussed in section 2, each proxy type has limitations and/or potential biases (e.g., with respect to timescale) that are generally specific to each type. It is thus arguably advisable to employ a "multiproxy" approach to take advantage of the complementary strengths of each of the diverse sources of proxy information described in section 2 in proxy-based climate reconstruction. Such an approach (see reviews by *Folland et al.* [2001a], *Jones et al.* [2001a], and *Mann* [2001a]) has been taken in the reconstruction of climate indices, such as the NAO [*Luterbacher et al.*, 1999; *Cullen et al.*, 2001; *Cook et al.*, 2002a; *Cook*, 2003], the *Antarctic Oscillation* (AAO) [*Jones and Widmann*, 2003], and the SOI [*Stahle et al.*, 1998b] and Niño3 indices [*Mann et al.*, 2000a, 2000b; *Evans et al.*, 2002] of ENSO. Such an approach has also been taken for reconstructions of NH [*Bradley and Jones*, 1993; *Overpeck et al.*, 1997; *Jones et al.*, 1998; *Mann et al.*, 1998a, 1999, 2000b; *Crowley and Lowery*, 2000; *Mann and Jones*, 2003] and, more tentatively, SH [*Jones et al.*, 1998; *Mann et al.*, 2000b; *Mann and Jones*, 2003] and global mean temperature [*Mann and Jones*, 2003].

[51] A number of distinct methods have been employed in assimilating the multiproxy data into seasonal or annual climate reconstructions. In many cases, multiproxy data are incorporated into the reconstruction of a large-scale climate measure based on CFR techniques discussed in section 3.2. [e.g., *Mann et al.*, 1998a, 1998b, 1999; *Luterbacher et al.*, 2002a]. CFR need not be based solely on a multiproxy indicator network. It can also be based on a spatially distributed set of "local calibrations," or it can use larger-scale data sets of a single widespread proxy (e.g., a spatially distributed network of tree ring densities such as in the large-scale summer temperature reconstructions of *Briffa et al.* [2001] or tree ring widths as in the U.S. continental drought reconstructions of *Cook et al.* [1999] and *Zhang et al.* [2004] or coral oxygen isotopes as in the tropical Pacific

SST pattern reconstructions by *Evans et al.* [2002]). In either case the spatial reconstructions can be averaged to yield large-scale diagnostics. For example, large-scale surface temperature pattern reconstructions can be averaged over the eastern equatorial Pacific to yield an estimate of the Niño3 index of ENSO [*Mann et al.*, 2000a, 2000b; *Evans et al.*, 2002] or over the entire Northern Hemisphere [*Mann et al.*, 1998a, 1999] or extratropical Northern Hemisphere [*Briffa et al.*, 2001] to yield estimates of hemispheric mean temperature.

[52] Other hemispheric mean temperature reconstructions have employed a simpler *composite plus scale* (CPS) approach. Here various temperature-sensitive proxy records are normalized and composited (perhaps after applying some weighting factor, e.g., based on area represented or modern correlations with colocated instrumental records, see Table 1). The average is then simply scaled against the available temporally overlapping instrumental record (which could be the annual average or the average for a specific part of the year, such as the growing season) to yield a hemispheric reconstruction [*Bradley and Jones*, 1993; *Overpeck et al.*, 1997; *Jones et al.*, 1998; *Crowley and Lowery*, 2000; *Esper et al.*, 2002; *Mann and Jones*, 2003]. In this simpler approach it is essential that each proxy record or local climate reconstruction is clearly shown to be representative of local temperature variations. In contrast, CFR-based reconstructions (see section 3.2), series not correlating with any pattern or combination of patterns of large-scale climate, will receive zero weight. Although this step is implicit within CFR, it is also best done explicitly before beginning with CFR [see *Rutherford et al.*, 2004].

[53] The similarity of reconstructions based on the simpler CPS approach to those determined based on different variants of CFR suggests that the CPS approach can yield a reliable reconstruction. Uncertainty estimates for CPS reconstructions can be obtained from the sampling statistics of the reconstruction with respect to the predictors used (e.g., a bootstrap [*Esper et al.*, 2002]) or from calibration-resolved variance statistics [*Mann and Jones*, 2003]. It is important with the CPS approach to examine the sensitivity of the resulting reconstruction to the scaling (instrumental series and time interval) employed [*Mann and Jones*, 2003]. Slightly different choices of scaling can, in some cases, yield a vastly different reconstruction [*Esper et al.*, 2002] (note, e.g., the quite different alternative scalings of this reconstruction provided by *Briffa and Osborn* [2002] and *Mann* [2002a]).

[54] Most previous work [*Bradley and Jones*, 1993; *Overpeck et al.*, 1997; *Jones et al.*, 1998; *Mann et al.*, 1998a, 1999, 2000b; *Crowley and Lowery*, 2000] has emphasized the reconstruction of only the Northern Hemisphere average temperature over the past 1000 years for which adequate proxy data exist. The recent availability of long records with reliable multicentury and millennial-scale variability has allowed preliminary extensions over nearly the past two millennia for both hemispheres and the globe [*Mann and Jones*, 2003]. Comparisons of different estimates of NH and SH mean (and regional) temperature changes over the past one to two millennia, based on different data sources and statistical approaches, are discussed in section 4.

4. CLIMATE RECORD OF THE PAST TWO MILLENNIA

4.1. Inferences From Instrumental Data and Model Simulations

[55] Since the mid-19th century, instrumental annual mean temperatures have warmed by between 0.6°C and 0.9°C (the exact value depending on the period chosen and how the trend has been estimated) for both the NH and SH averages [*Folland et al.*, 2001b]. Warming has not occurred monotonically, however, with two principal phases in the early and later decades of the 20th century (1915–1940 and since 1975). Slight cooling is seen between these dates, more apparent in the continental interiors of the NH and in the Arctic. Patterns of warming and cooling rates indicate extensive spatial, seasonal, and temporal variability with only 10–20% of the Earth's surface (individual 5° by 5° grid boxes) showing local statistical significance over the periods of strong warming in the global average [*Jones et al.*, 1999; *Jones and Moberg*, 2003]. Climate model-based detection and even attribution of trends are therefore much more likely at the large spatial scales (typically the first few low-order *empirical orthogonal functions* (EOFs) in a *principal components analysis* (PCA)) than at individual grid box scales [see, e.g., *Mitchell et al.*, 2001, and references therein].

[56] As the 20th century has experienced the greatest changes in radiative forcing of any century in the last few millennia (see section 5), we should not expect spatial and seasonal synchroneity to be any greater in earlier periods. The instrumental record also tells us that to begin to understand the course of hemispheric-scale changes in earlier centuries we need to aggregate many proxy indicators from as many locations as possible. There are a number of long instrumental records from western Europe, which extend back to the late 17th century. Climatologists who have assessed these records have established relevance, however, only for past climate changes at local-to-regional space scales [e.g., *Manley*, 1974; *Jones*, 2001] (see also discussion of Figure 2c in section 2.1).

[57] The effective number of spatial degrees of freedom in large-scale temperature changes is significantly reduced as the timescale lengthens from years to decades and to centuries [*Mann and Park*, 1994; *Jones and Briffa*, 1996; *Jones et al.*, 1997a]. So, in the absence of any timescale resolvability limitations in the proxy data used, one should expect greater fidelity on longer timescales [see, e.g., *Zorita et al.*, 2003] with multiproxy compilations. Millennial integrations of GCMs (both forced and unforced) and statistical exercises with proxy and/or instrumental data enable various spatial coherence measures to be estimated and the potential of limited proxy networks to be assessed [*Bradley*, 1996; *Evans et al.*, 1998; *Mann and Rutherford*,

2002; *Rutherford et al.*, 2003; *Zorita et al.*, 2003]. However, results are likely to be somewhat model-dependent or data set-dependent. The relevance of model-based analyses is influenced by the GCM's ability to simulate reality and the reliability and completeness of past forcing histories. However, GCMs provide data sets that are globally complete and potentially multimillennial in length, enabling many of the issues of representativeness of currently available proxy networks to be addressed [*Bradley*, 1996; *Rutherford et al.*, 2003; *Zorita et al.*, 2003].

4.2. Large-Scale Temperature Changes

[58] The development of most paleoclimatic disciplines in Europe and North America (with emphasis on the North Atlantic environs) has led to a "simplistic" (and, indeed, incorrect) picture of past global-scale climate variability. There was, in this simplistic picture, a *Medieval Warm Period* (MWP) (extending from ~A.D. 950 to 1200 [*Lamb*, 1965, 1977]) followed later by a Little Ice Age (from ~A.D. 1450 to 1850 [*Grove*, 1988]). Even a cursory inspection of regional records of past climate variability [*Williams and Wigley*, 1983; *Bradley and Jones*, 1993; *Hughes and Diaz*, 1994] indicates a complex pattern of past regional variations (see Figure 4) that rarely, if ever, follows the actual pattern of hemispheric or global mean variations. Insights from the instrumental record (section 4.1) tell us that any inferences about hemispheric-scale variations are likely to be biased if based on any one region of the world, such as the North Atlantic/Europe.

[59] Figure 4 reveals both markedly different courses of temperature change in the different "continental" regions as well as marked within-continent variability. Of the various proxy sources, ice cores are the most variable in their temporal histories, with some showing marked multicentury variability. The lesser variability evident in the two Greenland series, which are based on multicore averages, suggests that the enhanced variability in some ice cores may be a result of small sample size. Certain features stand out in the various panels, suggestive of continental-scale temperature changes over the last 500 years. Examples are a cool early 19th century in North America; cool 16th, 17th, and 19th centuries in Europe with a milder 18th century; cool 19th century in eastern Asia; and a cool period from 1650 until 1750 in much of the tropics (with the possible exception of the central and eastern tropical Pacific, as discussed in section 4.4.1). All "regions," with the exception of the North Atlantic and the Tasmania-eastern Antarctica region, indicate the warmest conditions during the second half of the 20th century. The markedly diverse behavior among the South American ice cores makes interpretation difficult, though it is likely that ice core oxygen isotopes in this region are governed to some extent by nontemperature influences [*Hoffmann et al.*, 1998].

[60] Recent studies (e.g., for the MWP [*Hughes and Diaz*, 1994] and the LIA [*Bradley and Jones*, 1993]) have assessed these periods in more detail, covering far wider spatial domains and using more data than were available in the 1960s when the terms (as presently understood) were coined. Both conclude that the periods were much more complex than these two terms imply, and the timing of the events varied regionally, if they can be discerned at all [see *Folland et al.*, 2001a]. A recent study promoting the "simplistic" notions of the MWP and LIA [*Soon and Baliunas*, 2003] (hereinafter referred to as SB03), however, compels us to stress two points (in addition to those discussed is section 3 [see also *Mann et al.*, 2003a]). First, it is essential to assess each proxy series for sensitivity to past temperature variability and not, as in the SB03 study, to equate hydrological (wet/dry periods) influences with temperature influences. Second, it is also essential (e.g., through the compositing of records) to distinguish between regional anomalies, which often cancel in a hemispheric mean, and not, as in the SB03 study, to equate the existence of asynchronous warm/cold anomalies in different regions with the existence of hemispheric warm/cold anomalies. Discussion in section 2.1 (e.g., for Figure 2c) underscores this important point.

[61] SB03 asked and attempted to answer the following (paraphrased) questions based on an interpretation of various proxy indicators of climate change in past centuries: Were there discernible climate anomalies indicative of warm, dry, or wet conditions during the MWP (defined by SB03 before looking at any of the series as the period A.D. 900–1300)? Were there discernible climate anomalies indicative of cold, dry, or wet conditions during the LIA (defined similarly as the period A.D. 1300–1900)? While the authors interpreted their "yes" answers to these questions as providing evidence for the mild MWP and the cold LIA, they could also have classified the MWP as "cold" and the LIA as "warm" by these same criteria. As the various panels in Figure 4 indicate, the period A.D. 1300–1900 (LIA as defined by SB03) contains a discernible warm period in almost all series, although rarely at the same time. Similarly, the period A.D. 900–1300 (MWP as defined by SB03) contains a discernible cold period in almost all series, though, again, rarely at the same time. Similarly, the warm and cold periods that do occur during their MWP and LIA periods, respectively, only clearly coincide during some of the centuries discussed with reference to Figure 4.

[62] The use of these two terms (MWP and LIA) is still quite widespread in the literature, despite the lack of any accepted definition [*Bradley and Jones*, 1992; see also *Bradley et al.*, 2003]. As discussed in section 3.3, hemispheric and/or global average estimates of climate fields (e.g., Northern Hemisphere mean temperature) must average, in some way, the variations estimated for different regions. Any true global or hemispheric definitions of such terms [e.g., *Folland et al.*, 2001a] must measure the cancellation of regionally distinct, often opposing, climate trends. Most paleoclimatologists developing regionally specific climate reconstructions of past centuries conventionally label their coldest interval as the "LIA" and their warmest interval as the "MWP" provided they fall within the widely defined (often overlapping in some definitions) ranges available in the literature. The logical absurdity of this convention is cleverly satirized by *Cobb et al.* [2003b],

who use the terms "Little Warm Age" and "Medieval Cool Period" to define the apparent pattern of cold (early) and warm (later) over the past millennium evident in their reconstructions of past climate trends in the central tropical Pacific. Indeed, if the development of paleoclimatology had taken place in the tropical Pacific, Africa, Australia, New Zealand, or Latin America, the paleoclimatic community would almost certainly have adopted other terminology embracing different epochs and/or features (e.g., the "Extended Dry Period").

[63] Paleoclimatology should be striving to determine the true, potentially quite regionally and temporally complex, pattern of past climate variability, without any preconceived "pigeonholing" of new data implied by the use of terms such as the LIA or MWP. Estimates of global or hemispheric mean quantities based on the assimilation of networks of proxy data (e.g., section 3.3) afford our best opportunity to establish the course of hemispheric-scale climate history over the past millennium and beyond. This is comparable to the way global climate changes have been established during the "instrumental" era of the past century or so (see, e.g., section 4.1). Indeed, the only way to define periods in the past at the hemispheric/global scale is to use appropriate large-scale averages.

[64] As the number of available regional climate estimates becomes fewer farther back in time, the uncertainties in any estimates of global or hemispheric mean changes must increase. Even for the instrumental period, a greater density of temperature records would lead to improvement in the spatial and seasonal details of our knowledge [e.g., *Weber and Madden*, 1995]. Additional measurements, however, would be unlikely to change our assessment of decadal mean trends [*Jones and Moberg*, 2003] owing largely to the decreased spatial degrees of freedom at decadal or longer timescales (see, e.g., section 4.1). For these reasons, decadal and lower-frequency variations in past centuries are more likely to be resolved from relatively modest global networks of proxy data [*Bradley and Jones*, 1993; *Bradley*, 1996; *Mann et al.*, 1998a; *Mann and Jones*, 2003]. The markedly fewer proxy records for the SH compared to the NH available to reconstruct surface temperature changes in past centuries leads to a less certain state of knowledge for SH or global mean temperatures. This is ameliorated on longer timescales by the greater ocean/land ratio for the SH, implying less year-to-year variability and fewer spatial degrees of freedom than for the NH. We thus emphasize here conclusions from NH hemispheric temperature reconstructions of past centuries, on decadal and longer timescales, though we show reconstructions for both hemispheres and the globe (Figure 5).

[65] These reconstructions indicate relatively modest variations in the NH (Figure 5) as well as the "continental" scale (see panels in Figure 4) over the past one to two millennia prior to the marked warming of the 20th century. The range of NH temperature variability over the past two millennia is quite small, with the warmest and coldest years having a difference of less than 1.5°C and decades having a difference of less than 1.0°C. NH mean temperatures appear, for example, to have been slightly warmer (a couple of tenths of a degree Centigrade) during the period A.D. 800–1400 relative to the later, cooler period from A.D. 1400 to 1900 and the earlier period A.D. 200–800. However, even the early interval of relative warmth does not approach in magnitude the hemispheric warmth of the late 20th century. These estimates indicate the 6th, 15th, 17th, and 19th centuries as having been the coldest for the Northern Hemisphere. The 20th century has seen the greatest temperature change within any century in the past two millennia (0.6°–0.9°C) compared to less than approximately ±0.2°C for any other century. Numerous recent studies have now established the late 20th century as the warmest multidecadal period of the past millennium [*Jones et al.*, 1998; *Mann et al.*, 1999; *Crowley and Lowery*, 2000; *Mann et al.*, 2003a, 2003b]. Recent extensions back to A.D. 200 suggest that late 20th century warmth is likely unprecedented over at last roughly the past two millennia for the Northern Hemisphere [*Mann and Jones*, 2003]. Uncertainties are greater, because of the paucity of Southern Hemisphere proxy data, for the Southern Hemisphere and globe, but it is also likely that late 20th century warmth is unprecedented in these contexts as well. These observations are consistent with evidence that radiative forcing has been considerably enhanced by anthropogenic influences during the 20th century. The role of natural and anthropogenic radiative forcing of climate over the past one to two millennia is discussed in section 5.

[66] Available proxy data networks become increasingly sparse during the most recent decades because many of the key series were obtained during the 1970s and 1980s and have not been updated to the present. Typically, multiproxy climate reconstructions are calibrated and/or cross validated against instrumental data available through the 1980s [e.g., *Mann et al.*, 1998a; *Cook et al.*, 2002a; *Evans et al.*, 2002; *Mann and Jones*, 2003]. Upon successful cross validation the proxy reconstructions are then typically compared, within the context of self-consistently estimated uncertainties, to more modern instrumental data available through the end of the 20th century. This enables late 20th century warmth to be compared with earlier conditions [e.g., *Mann et al.*, 1998a, 1999, 2001; *Folland et al.*, 2001a; *Mann and Jones*, 2003]. Such comparisons have incorrectly been argued by some [*Soon et al.*, 2003] to suggest that multiproxy data do not show evidence of anomalous late 20th century warmth. We address this latter claim by extending the proxy-based hemispheric and global temperatures of *Mann and Jones* [2003] through 1995 based on the available proxy information. The hemispheric series of *Mann and Jones* [2003] were based on weighted composites of eight and five regional temperature reconstructions available through 1980 for the NH and SH, respectively. As in some instrumental estimates of global mean temperature trends [e.g., *Jones et al.*, 1999], a global mean series was defined for simplicity as the average of the two hemispheric (NH and SH) series. A weighted global mean penalizing the greater uncertainty in the SH reconstruction would yield a global mean series more similar to the NH series than this

unweighted global series. We have extended the "standard" hemispheric and global series of *Mann and Jones* [2003] (which ends in 1980) to 1995. We do this by making use of the restricted proxy series available subsequent to 1980 (for the NH, five of the eight series that were used are available through 1984; four are available through 1990, and three are available through 1995; for the SH, four of the five series are available through 1984, two are available through 1990, and one is available through 1995). For purposes of calculating the composites any proxy series terminating prior to 1995 are extended forward by persistence of the final available value. A similar result is obtained through 1990 if, instead, only available values are used. After 1990, there are too few series to develop a meaningful weighted-mean composite of available values alone.

[67] These updated estimates are shown in Figure 5. The conclusion that hemispheric and global mean temperatures during the late 20th century are anomalous in at least a nearly two-millennial context is observed to hold from the proxy reconstructions alone, independent of any inferences from the instrumental record. The additional comparison of the proxy reconstructions against the instrumental record of the late 20th century (which resolves the full 20th century trends), taking into account the uncertainties in the reconstructions, nonetheless, allows such conclusions to be established in a more statistically robust manner.

[68] Some reconstructions of hemispheric temperature changes based solely on extratropical, continental (tree ring) data suggest wider swings in temperature, including greater cooling during the 17th–19th century than is evident in either the instrumental, model, or other proxy-based estimates [*Esper et al.*, 2002]. Despite this greater variability the *Esper et al.* [2002] reconstruction, like other reconstructions, reveals the late 20th century warmth to be anomalous in a greater-than-millennial context [*Cook et al.*, 2004]. The use of an unusually liberal tree ring standardization technique has been argued to produce greater low-frequency variability evident in the reconstruction [*Esper et al.*, 2002], though potentially also introducing spurious, nonclimatic low-frequency variability [*Mann and Hughes*, 2002]. The greater-amplitude variability apparent in this reconstruction, however, appears to result from a combination of the exclusive use of warm season temperature-sensitive tree ring chronologies [see *Mann*, 2002a] and, equally important, restricted spatial sampling. *Rutherford et al.* [2004] show that the extensive network of extratropical Northern Hemisphere tree ring reconstructions of warm season temperatures used by *Briffa et al.* [2001], once spatially masked for the much smaller number (14) of regions used by *Esper et al.* [2002], yields a remarkably similar long-term mean history. A recent study [*McIntyre and McKittrick*, 2003] claims that revisions of the data and methods used in the *Mann et al.* [1998a, 1999] reconstruction shown here in Figure 5 yield a "corrected version" exhibiting 15th century NH mean temperatures warmer than those of the latter (1970s–1980s) 20th century. This claim is so clearly at odds with every other reconstruction discussed in this section (particularly *Esper et al.* [2002] and *Huang et al.* [2000]) that it should be dismissed on this basis alone. However, a careful analysis (M. E. Mann et al., Critical flaws in a recent criticism of the *Mann et al.* [1998] study, submitted to *Climate Change*, 2003) of the *McIntyre and McKittrick* [2003] result reveals that their anomalous 15th century warmth results from their elimination of over 70% of the 15th century proxy data used by *Mann et al.* [1998a]. Also, their reconstruction, unlike that of *Mann et al.* [1998a] or *Mann and Jones* [2003], fails independent pass cross-validation tests. Their result can thus be dismissed as spurious on this basis also. Indeed, the statistical methodology used by *Mann et al.* [1998a] has been independently validated in applications to data from extended model simulations [*Zorita et al.*, 2003], and their NH mean reconstruction has been closely reproduced with an independent method of CFR [*Rutherford et al.*, 2004] applied to the same multiproxy network.

[69] Regional temperature trends typically exhibit quite different behavior from that behavior evident for the hemispheric mean (compare Figures 4 and 5). While the coldest centuries of the past millennium for the NH were the 17th and the 19th, the coldness of the 17th century was more evident, for example, in Europe, while the coldness of the 19th century was more apparent in North America. The 18th century was markedly milder, particularly in the 1730s and 1760s in Europe. Though European instrumental temperatures, for example, indicate two distinct cold phases in the 17th and the 19th centuries (as do the long European proxy records shown in Figure 4), the hemispheric reconstructions suggest a relatively steady, long-term cooling. While modest hemispheric temperature changes appear to be tied largely to simple energy balance response to radiative forcing changes (see section 5.2), regional temperature variations, which are often greater in their amplitude, are dominated by internal or externally generated changes in atmospheric circulation associated with ENSO, the NAO, and other patterns of variability (see sections 4.4 and 5.3).

4.3. Large-Scale Hydroclimatic/Moisture Changes

[70] Though not as relevant, for example, for the issue of the detection of anthropogenic warming, a knowledge of the range of variability in hydroclimatic variables such as seasonal precipitation and drought in past centuries is clearly important from a societal, as well as a scientific, point of view. Moreover, past drought and precipitation patterns can also aid our understanding of past changes in atmospheric circulation [e.g., *Wanner et al.*, 2000; *Luterbacher et al.*, 2002a]. There has been considerable recent interest in the use of paleoclimate proxy indicators to reconstruct past large-scale patterns of precipitation and drought [*Mitchell et al.*, 1979; *Meko*, 1981; *D'Arrigo and Jacoby*, 1991; *Hughes and Brown*, 1992; *Graumlich*, 1993; *Meko et al.*, 1993; *Stine*, 1994; *Martin-Vide and Barriendos*, 1995; *Hughes and Graumlich*, 1996; *Laird et al.*, 1996, 1998; *Fisher et al.*, 1996; *Woodhouse and Overpeck*, 1998; *Swetnam and Betancourt*, 1998; *Stahle et al.*, 1998a, 2000; *Ren*, 1998; *Cole and Cook*, 1998; *Hughes and Funkhouser*, 1999; *Cook et al.*, 1999; *Rodrigo et al.*, 1999; *Verschuren et al.*, 2000; *Druckenbrod et al.*, 2003; *Zhang et al.*, 2004].

[71] Traditional research has emphasized North American drought [e.g., *Mitchell et al.*, 1979; *Meko*, 1981; *D'Arrigo and Jacoby*, 1991; *Graumlich*, 1993; *Laird et al.*, 1996, 1998; *Cook et al.*, 1997, 1999; *Stahle et al.*, 2000]. Past reconstructions suggest that the range of North American drought variability observed during the 20th century may not fully represent the range of drought evident in earlier centuries [*Laird et al.*, 1996, 1998; *Woodhouse and Overpeck*, 1998; *Hughes and Graumlich*, 1996; *Hughes and Funkhouser*, 1999; *Stahle et al.*, 2000]. The 20th century dust bowl still stands out as the most extreme summer drought of recent centuries [*Cook et al.*, 1999; *Stahle et al.*, 2000; *Zhang et al.*, 2004]. Extended drought reconstructions [*Stahle et al.*, 2000] appear, however, to indicate that a 16th century "megadrought" over the United States, western Canada, and northwestern Mexico exceeded in scale and magnitude the 20th century dust bowl event. *Stine* [1994] argues for exceptional drought in the Sierra Nevada of California prior to the 15th century, while *Swetnam and Betancourt* [1998] argue that recent spring wetness in the American Southwest is greater than that observed in at least the last thousand years. Long-term changes in North American summer drought [e.g., *Cole and Cook*, 1998] may also be tied to changes in ENSO (see section 4.4.1).

[72] There is also a sizable body of research into past precipitation patterns in Europe [*Pfister*, 1992; *Martin-Vide and Barriendos*, 1995; *Rodrigo et al.*, 1999; *Wanner et al.*, 2000]. Long precipitation or drought reconstructions have been developed for central Europe [*Pfister*, 1984, 1992], the Czech Republic [*Brázdil*, 1996; *Brázdil et al.*, 2004], Germany [*Glaser*, 2001], and Iberia [*Rodrigo et al.*, 1999]. Changes in precipitation patterns in Europe in past centuries, such as the dry conditions in central Europe and the European Alps during the late 17th and early 18th centuries [*Pfister*, 1992; *Wanner et al.*, 2000], may reflect changes in the large-scale atmospheric circulation (e.g., the NAO) influencing the region (see section 4.4.3).

[73] In all other parts of the world, evidence of moisture variability has not been brought together in a rigorous quantifiable context, but tentative, qualitative interpretations may, nonetheless, be possible in some cases. Past precipitation and/or drought changes are available in various regions of the tropics, including the tropical Pacific [*Cole et al.*, 1993; *Linsley et al.*, 1994; *Dunbar et al.*, 1996; *Hendy et al.*, 2002] and tropical South America [*Thompson et al.*, 1988]. *Hodell et al.* [1995] provide lake sediment isotope evidence for pronounced drought between A.D. 800 and 900 in the Yucatan Peninsula of Mexico. Reconstructed lake levels in equatorial east Africa (Kenya) [*Verschuren et al.*, 2000] suggest dry conditions during the earlier part of the millennium (11th–13th centuries) and wet conditions during the later centuries (e.g., 14th–19th centuries), with peak wet conditions during the mid-17th to the mid-18th centuries.

[74] Precipitation changes in Asia in past centuries, as qualitatively inferred from proxy data sources, have often been interpreted in terms of changes in the Asian monsoon. Ice accumulation on the Dunde Ice Cap in the Tibetan Plateau [*Thompson*, 1996] appears to have been reduced in the first half of the last millennium compared to the latter half, suggestive of a decrease in summer monsoonal precipitation during the latter centuries of the past millennium. By contrast, historical documentary evidence [*Zhang and Crowley*, 1989] suggests relatively dry summer conditions (and possible increases in winter snow cover) in China during the A.D. 1400–1900 period, and pollen evidence [*Ren*, 1998] appears to indicate increased summer precipitation in northeastern China from A.D. 1000 to 1340. Evidence for changes in the Asian monsoon is available from sediment core foraminiferal evidence of monsoonal wind-induced upwelling in the Arabian Sea back through the 11th century [*Anderson et al.*, 2002; *Gupta et al.*, 2003]. Owing to differing temporal resolutions of the records, uncertainties in the (radiocarbon based) age model of the sediment core evidence, and possible nonclimatic secular influences on the interpretation of the historical ship data, it is difficult to draw, as yet, any internally consistent picture of past changes in the Asian monsoon.

4.4. Regional Circulation Patterns

[75] Proxies generally respond directly to temperature and precipitation influences that are only indirectly linked to the atmospheric circulation itself. Despite this, considerable interest in various circulation "indices" has encouraged reconstructions of their histories back in time from proxy, historical, and long instrumental records. Since the influence of the circulation measure on surface climate varies with time, proxy-based reconstructions may additionally be influenced by this variability [*Jones et al.*, 2003b]. To minimize this influence, the proxies used should ideally be located in regions where there is a particularly strong influence from the circulation feature. In sections 4.4.1–4.4.3 we emphasize reconstructions over past centuries of ENSO, the *Pacific Decadal Oscillation* (PDO), and the NAO (or its close relative, the Arctic Oscillation (AO)) [e.g., *Folland et al.*, 2001a]. Such reconstructions are shown in Figure 6a for the SOI and Figure 6b for the NAO.

[76] Preliminary work has also been undertaken with respect to the reconstruction of other atmospheric circulation or climate indices such as the 10- to 12-year timescale Atlantic Decadal Mode [*Black et al.*, 1999] and the 50- to 70-year timescale Atlantic Multidecadal Oscillation [*Mann et al.*, 1995; *Delworth and Mann*, 2000; *D'Arrigo et al.*, 2003a]. *Black et al.* [1999] also show evidence for a multidecadal spectral peak in tropical Atlantic variability. A preliminary reconstruction of the Southern Hemisphere counterpart to the AO, the AAO, has recently been developed [*Jones and Widmann*, 2003]. Newly available high-resolution ice core isotope records that form the International Trans-Antarctic Scientific Expedition project [*Steig and Schneider*, 2002] should allow for improved reconstructions of this index in the future.

4.4.1. ENSO

[77] A number of reconstructions of the time evolution of ENSO over the past few centuries have been attempted in recent work [see, e.g., *Folland et al.*, 2001a; *Jones et al.*,

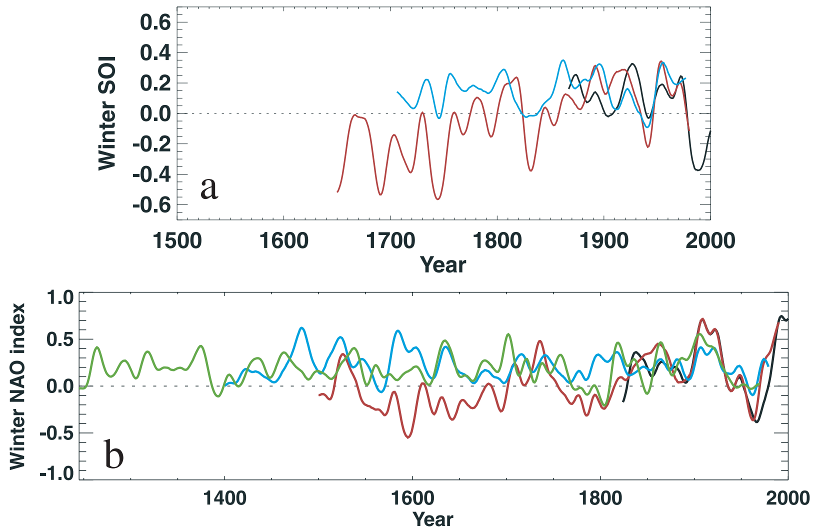


Figure 6. Reconstructions of two major atmospheric indices: (a) the Southern Oscillation Index (SOI) and (b) the North Atlantic Oscillation (NAO). For the SOI, two ENSO series (red for 1650–1980 [*Mann et al.*, 2000b] and blue for 1706–1977 [*Stahle et al.*, 1998b]) are rescaled to the observed mean and variance over 1867–1977 for the boreal winter (October to March) season. The observed data (black) for the SOI are based on monthly average sea level pressure data from Tahiti and Darwin, using the method of *Ropelewski and Jones* [1987] with 1961–1990 as the normalization period. All three series are smoothed with a 30-year Gaussian filter. For the NAO, three series (blue for 1400–1979 [*Cook*, 2003], red for 1500–1990 [*Luterbacher et al.*, 2002a], and green for 1250–1970 [*Vinther et al.*, 2003]) were regressed against the boreal winter (December to February) NAO index [*Jones et al.*, 1997b] for the 1901–1970 period. Winters are dated by the year of the January and smoothing uses a 30-year Gaussian filter.

2001a; *Mann et al.*, 2000a, 2000b; *Evans et al.*, 2002]. These include a boreal winter season (December, January, and February) SOI reconstruction based on ENSO-sensitive tree ring indicators [*Stahle et al.*, 1998b], a multiproxy-based reconstruction of the annual [*Mann et al.*, 2000a] and boreal cold season (October-March) [*Mann et al.*, 2000b] Niño3 index (constructed by averaging reconstructed global SST anomalies over the region 5°N to 5°S, 150°W to 90°W in the eastern equatorial Pacific). A tropical coral-based Niño3/3.4 reconstruction has also been attempted [*Evans et al.*, 2002]. These various ENSO index reconstructions, based on partially or largely independent data and entirely independent methodologies, share significant common variance (typically more than 30% during their respective cross-validation periods [see, e.g., *Folland et al.*, 2001a; *Evans et al.*, 2002]), suggesting a relatively consistent history of El Niño in past centuries. These quantitative reconstructions have been shown, moreover, to correspond well with the more qualitative evidence available regarding the history of El Niño events in past centuries from historical documents [*Quinn and Neal*, 1992; *Ortlieb*, 2000]. The large 1982/1983 and 1997/1998 warm events seem to be outside the range of variability of the past few centuries [*Folland et al.*, 2001a, Figure 2.28]. The uncertainties in the reconstruction are sizable, however, and the reconstructions may tend to systematically underestimate the amplitudes of some of the largest events such as the 1877/1878 El Niño [*Folland et al.*, 2001a]. Only a richer network of ENSO-sensitive proxy indicators is likely to reduce the uncertainties to the point where more confident conclusions can be reached. Some ENSO reconstructions (see Figure 6a) show relatively little low-frequency variability [*Stahle et al.*, 1998b] (the absence of which is likely due to the traditional processing of tree ring data employed in the Stahle et al. study). Others show more pronounced low-frequency variability, including a multidecadal "warm event" phase from the mid-17th to mid-18th century [*Mann et al.*, 2000b; *Jones et al.*, 2001a]. This latter observation is also consistent both with coral evidence for a relative absence of cooling [*Hendy et al.*, 2002] and even warming [*Cobb et al.*, 2003a] in the tropical Pacific during this period and with evidence discussed in section 4.3 for wet conditions in equatorial east Africa at this time, possibly implying [e.g., *Ropelewski and Halpert*, 1987] warm event ENSO conditions. Fossil coral evidence, though incomplete, allows insights into changes in both the interannual variability and mean state of an ENSO-dominated region of the tropical Pacific over roughly the past millennium [*Cobb*, 2002; *Cobb et al.*, 2003a]. Such evidence suggests La Niña-like conditions in the earlier centuries of the past millennium (the 10th–13th centuries) and El Niño-like conditions later on (e.g., the 17th century). Evidence regarding even longer-term (i.e., millennial scale) variability in ENSO is even more tenuous. Significant millennial-scale ENSO-like variability has been inferred from coarsely resolved records [*Koutavas et al.*, 2002; *Stott et al.*, 2002], but such inferences have been called into question because of the inability of low-resolution proxy records to distinguish between true El Niño (interannual variability) changes, and changes in various aspects of the tropical Pacific mean state [*Trenberth and Otto-Bliesner*, 2003].

[78] There is also evidence of significant multidecadal and century-scale variability in the frequency domain character of ENSO in past centuries [*Stahle et al.*, 1998b; *Mann et al.*, 2000a; *Urban et al.*, 2000; *Jones et al.*, 2001a; *Mann*, 2001b]. Some studies suggest, for example, a relative absence of interannual variability, in the face of persistent decadal variability, during the early and middle 19th century [*Mann et al.*, 2000a; *Urban et al.*, 2000]. Nonstationarity is evident on multidecadal timescales in the extratropical teleconnections of ENSO during past centuries [*Cole and Cook*, 1998; *Mann et al.*, 2000a].

4.4.2. PDO

[79] A number of proxy-based reconstructions of decadal and multidecadal Pacific climate variability [*Mann et al.*, 1995; *Minobe*, 1997; *Mann et al.*, 2000a; *Linsley et al.*, 2000; *Biondi et al.*, 2001; *Correge et al.*, 2001] and, in some cases, a PDO index specifically [*Evans et al.*, 1998; *Biondi et al.*, 2001] have been produced from combinations of tropical and/or extratropical tree ring, coral, and ice core proxy data. *Moore et al.* [2002] infer changes in North Pacific winter climate associated with the *Pacific North American* (PNA) atmospheric circulation pattern from changes in accumulation in a northwestern North American ice core. These indicate a large positive trend toward increased accumulation since the mid-19th century, consistent with a trend toward the positive phase of the PNA and PDO pattern, with concomitant accelerated warming in northwest North America over the same time interval.

4.4.3. NAO/AO

[80] A number of studies have attempted to analyze the behavior of the boreal winter or cold season NAO (or the related AO) over the past three to six centuries. Recent work [see, e.g., *Folland et al.*, 2001a; *Jones et al.*, 2001a, 2003b; *Mann et al.*, 2000a, 2000b] compares various composites of high-resolution proxy data or combinations of proxy and instrumental data [*D'Arrigo et al.*, 1993; *Cook et al.*, 1998; *Appenzeller et al.*, 1998; *Cullen et al.*, 2001; *Luterbacher et al.*, 1999, 2002a; *Slonosky et al.*, 2000; *Jones et al.*, 2001a; *Mann*, 2001a, 2001b, 2002b; *Cook et al.*, 2002a; *Cook*, 2003; *Vinther et al.*, 2003]. Several, though not all [e.g., *Luterbacher et al.*, 2002a], of the NAO reconstructions suggest that the trend in recent decades toward an enhanced positive phase of the winter NAO is relatively unusual, if not unprecedented, in recent past centuries (see Figure 6b). This conclusion seems consistent with insights from model-based analyses of the natural variability of the NAO [*Osborn et al.*, 1999] and proposed mechanisms for anthropogenic forcing of the NAO/AO [*Shindell et al.*, 1999]. However, as in the case of the El Niño reconstructions discussed in section 4.4.1, sizable uncertainties prevent any definitive conclusions being drawn with regard to the anomalous nature of the recent positive NAO trend. The lack of skillful cross validation of most NAO reconstructions against the earliest instrumental data [*Schmutz et al.*, 2000; *Jones et al.*, 2003b] calls into question the long-term reliability of current proxy-based NAO reconstructions. Use of long calibration periods, however, appears to yield an NAO reconstruction that cross validates well with the earliest available instrumental data [*Cook et al.*, 2002a; *Cook*, 2003], but the lowest-frequency variability in the mean value may not be reliable, owing to the use of traditionally standardized tree ring data. Other reconstructions do contain enhanced low-frequency variability, indicating, for example, a prominent tendency toward the negative phase of the NAO during the 17th century [*Wanner et al.*, 1995; *Slonosky et al.*, 2000; *Luterbacher et al.*, 2002a]. A preliminary reconstruction of the summer AO, as distinct from the cold season NAO, has also recently been attempted by *D'Arrigo et al.* [2003b].

[81] The NAO may have played a prominent role in cold season extratropical NH surface temperature changes in past centuries, potentially explaining (through a tendency for the negative phase) the enhanced regional winter cooling during the 17th century in Europe [*Shindell et al.*, 2001] and, perhaps (through its positive phase), the enhanced regional warming during medieval times. This interpretation is also consistent with century-scale temperature variations in Greenland estimated from ice borehole data [*Dahl-Jensen et al.*, 1998]. Less quantitative, but nonetheless compelling, qualitative evidence for a predominant NAO-like pattern of variability in past centuries comes from oxygen isotopes and foraminiferal evidence [*Keigwin*, 1996; *Keigwin and Pickart*, 1999], Mg/Ca paleo-SST estimates [*deMenocal et al.*, 2002], and diatom evidence [*Jansen and Koc*, 2000] from North Atlantic sediment cores, as well as Greenland ice cores [*O'Brien et al.*, 1995]. Forward modeling of glacial mass balance in Europe in past centuries [*Reichert et al.*, 2002] suggests that the inverse relationship between glacier expansion and recession in northern and southern/central Europe in past centuries is consistent with NAO precipitation forcing. Alkenone-based SST reconstructions from sediment cores throughout the eastern North Atlantic and the Mediterranean [*Rimbu et al.*, 2003] and lake deposit evidence from eastern North America [*Noren et al.*, 2002] suggest a dominant NAO signature in the climate variability of the North Atlantic and neighboring regions on millennial timescales throughout the late Holocene.

5. PHYSICAL EXPLANATIONS OF PAST OBSERVED CHANGES

[82] Climate simulations and comparisons of their results with empirical paleoclimate reconstructions can greatly inform our understanding of the factors governing climate change in past centuries and millennia. On timescales of the past one to two millennia, in particular, it is likely that external forcing, including natural (e.g., volcanic and solar radiative) and anthropogenic (greenhouse gas and sulphate aerosol) influences, and natural, internal variability (e.g., internally generated changes in ENSO

and century or millennial-scale natural oscillations in the coupled ocean-atmosphere system) have played the most important roles. Models can help us determine both how the climate system might have been expected to change given estimates of past changes in forcings and how much variability might be expected to have arisen from internal variability. The insights gained from models can then be compared against those gleaned from paleoclimatic data in the hope of developing a consistent synthesis of theoretical and empirical considerations [see, e.g., *Mann et al.*, 2001].

5.1. Forcing Factors

[83] External forcing by changes in the distribution of insolation associated with Earth orbital changes clearly dominates climate variability on multimillennial timescales [e.g., *Ruddiman*, 2001]. Any understanding of changes in the large-scale circulation of the atmosphere, including monsoon and ENSO influences during the mid-Holocene, for example, almost certainly requires (e.g., section 1.1) taking into account changes in Earth orbital geometry. On timescales of the past one to two millennia, however, it is likely that external forcing due to natural changes in solar irradiance and explosive volcanism and anthropogenic influences from land use changes, greenhouse gas concentrations, and more recently sulphate aerosols represent the dominant forcings of climate variability. Of course, the histories of these forcings are not known precisely, but like the climate itself, they must be imperfectly reconstructed from proxy information sources. Empirical analyses employing simple linear correlations or multivariate regressions between forcing series and climate reconstructions can provide insights into the relative roles of such forcings in past centuries [e.g., *Lean et al.*, 1995; *Mann et al.*, 1998a, 2000a; *Beer et al.*, 2000; *Waple et al.*, 2002]. Model simulations, however, driven with estimated forcings (see section 5.2) and comparisons of these model simulations with empirical paleoclimate reconstructions (see section 5.3) will likely yield more detailed physical insights.

[84] Reconstructions of forcings (at the global or NH scale) over the past one to two millennia are shown in Figure 7. In Figure 7 all forcings are expressed in W m^{-2} to aid comparisons. In these units this is equivalent to a "top of the atmosphere" forcing that a climate model (see section 5.2), in its crudest sense, responds to. Each type of climate model will, however, "see" the various forcings in different ways. GCMs, for example, are provided a specific spatial pattern for a forcing (particularly important for volcanoes), and sulphate aerosol forcing may be fully coupled with a chemistry model, while simpler models (e.g., *energy balance models* (EBMs)) are likely provided with estimates of global or hemispheric mean radiative forcing.

5.1.1. Solar Irradiance Forcing

[85] An estimate of solar irradiance variations, available back to 1610, is provided by variations of observations of sunspots recorded by humans. Under certain assumptions (with respect to the underlying solar physics) this can be calibrated against modern satellite measurements to yield an estimate of solar irradiance variability over the past four centuries [*Lean et al.*, 1995]. Various reconstructions for the 20th century are currently available, but although they qualitatively agree, some significant differences exist [*Lean*, 2000; *Lean et al.*, 2002; *Foukal*, 2002; *Willson and Mordvinov*, 2003; *Solanki and Krivova*, 2003]. Solar irradiance variations over the past few millennia can be reconstructed from isotopic information recorded in tree rings (^{14}C) and in ice cores (^{10}Be), both known to be produced because of the interaction of solar radiation with the upper atmosphere [see *Bard et al.*, 2000; *Crowley*, 2000]. To the extent that these alternative sources of information agree qualitatively, a splicing of the two estimates can be used to yield a quantitative estimate of solar irradiance variability over the past few millennia [e.g., *Crowley*, 2000] and potentially longer. There are, of course, substantial uncertainties in these reconstructions due, for example, to assumptions regarding the nature of the calibration of low-frequency solar variability based on a very short ($\sim$25 years) satellite irradiance record [see *Lean et al.*, 1995; *Crowley*, 2000; *Lean*, 2000; *Lean et al.*, 2002; *Foukal*, 2002; *Willson and Mordvinov*, 2003; *Solanki and Krivova*, 2003]. Fortunately, solar radiative forcing is an intrinsically global forcing, and regional variability in the forcing need not be taken into account. Three solar forcing series, used by climate modelers, are shown in Figure 7, plotted as anomalies from the mean value of the "solar constant" of 1365.6 W m^{-2}, calculated for 1611–1998 by *Lean et al.* [1995]. To allow for direct comparison with other forcing factors, however, the solar forcing estimate has been reduced by a factor of 4 relative to

Figure 7. Estimates of natural and anthropogenic radiative forcings over the last couple of millennia used by climate models to simulate the climate over the period, (a) forcings used by *Crowley et al.* [2003], (b) solar and volcanic forcing used by *Ammann et al.* [2003], and (c) solar and volcanic forcing series used by *Bertrand et al.* [2002] (their series TSI_L and VOLC_C, respectively). All forcings are expressed in W m^{-2} (see text in section 5.1 for more discussion) and represent global averages in Figures 7a and 7c and NH averages in Figure 7b. In Figures 7b and 7c the greenhouse gas (GHG) and sulphate aerosol (Aer) forcing will be similar to that used in Figure 7a. All solar forcing series are expressed as anomalies from the mean value of 1365.6 W m^{-2} for the 1611–1998 period (see text for discussion of the allowance for only a quarter of the surface being seen and also for albedo). Details of the extension of the solar series before visually based observations began in the early 17th century are given by *Bard et al.* [2000] and *Crowley* [2000]. Over this period the solar forcing in Figure 7a is slightly smaller than the other two as it applies the background trend not to the Maunder Minimum period but to the ^{10}Be estimates for the earlier Spörer Minimum. Volcanic forcing is converted to W m^{-2} by multiplying the aerosol optical depth estimates made from ice cores by -21 [*Hansen et al.*, 2002]. Volcanic forcing dips below -7 W m^{-2} in either 1258 (Figures 7a and 7b) or 1259 W m^{-2} (Figure 7c) to -9.1, -11.9 and -8.3 W m^{-2} in Figures 7a, 7b, and 7c, respectively.

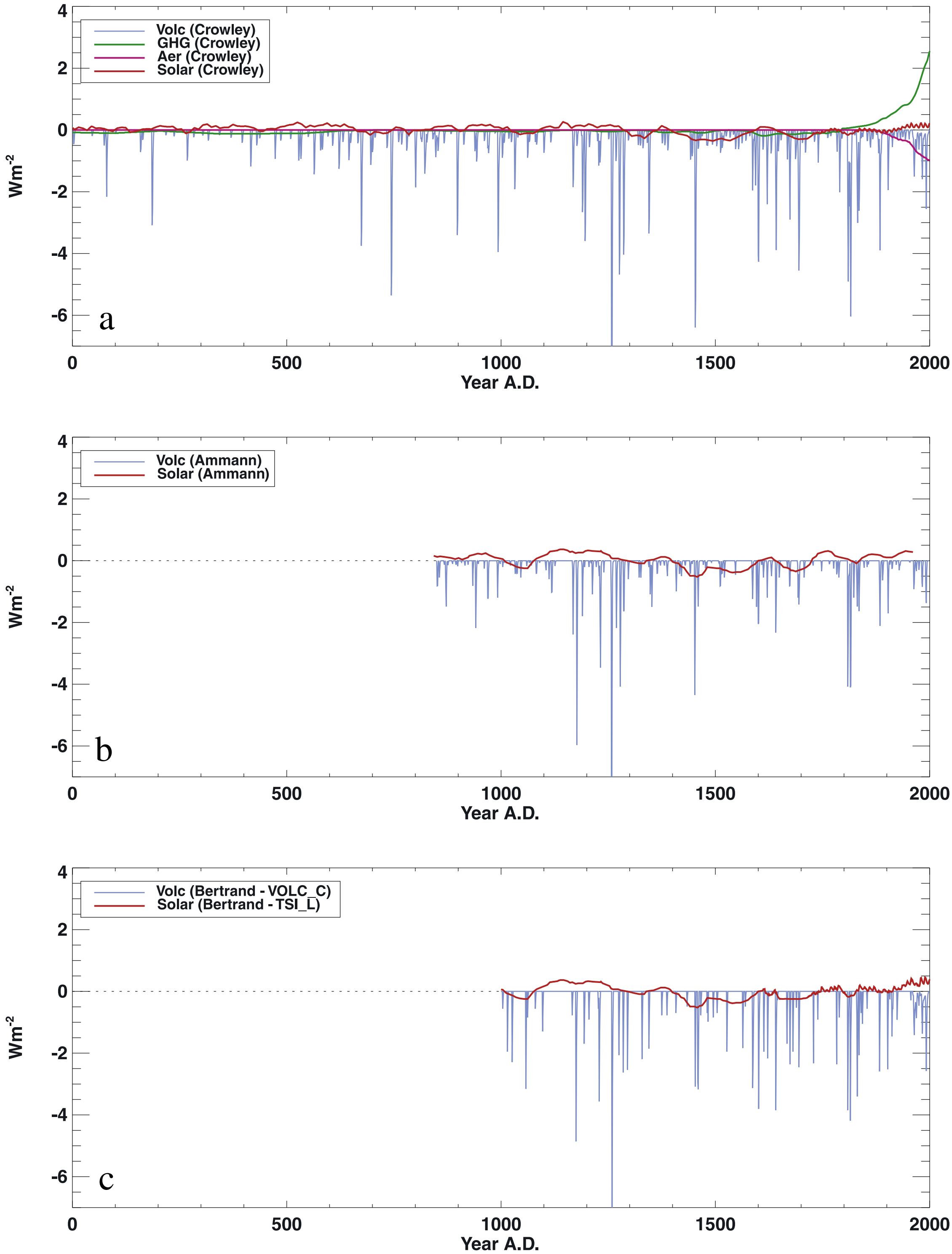


Figure 7

the solar constant to account for the smaller fraction of the Earth's surface that is actually exposed to the Sun on the average. An additional reduction of ~30%, the approximate fraction of incoming insolation reflected back to space because of the albedo of the Earth and atmosphere, is also applied. While solar irradiance is clearly the largest of the forcings in terms of its mean magnitude, its variability over time (in these reconstructions) is often relatively small when compared with volcanic and greenhouse gas forcing.

5.1.2. Volcanic Aerosol Forcing

[86] Volcanic forcing is typically estimated from sulphate aerosols contained in annual ice core layers back in time [*Robock*, 2000]. Greater concentrations of trapped sulphates are typically indicative of larger eruptions. There will then be a greater potential for the aerosols that have resided in sufficient concentrations in the lower stratosphere to have had a large-scale (negative) radiative influence on surface climate (i.e., by reflecting radiation back to space that

otherwise would have been incident on the Earth's surface). Ice core volcanic indices have been developed for the last two millennia by *Robock and Free* [1995, 1996], and extended back farther and modified by *Crowley et al.* [2003], by averaging records from polar ice cores. An additional estimate has been developed by *Ammann et al.* [2003]. In some cases, global volcanic radiative forcing estimates are derived from ice core data under the assumption that those eruptions that appear in ice cores in both poles are most likely to represent climatically relevant (i.e., global) radiative forcings [e.g., *Crowley*, 2000]. Estimation of the relevant radiative forcing is complicated, however, by the fact that tropical and high-latitude eruptions likely have very different residence times in the atmosphere (influencing their radiative forcing potential). The season and latitude of the eruption [*Robertson et al.*, 1998; *Robock*, 2000] are also important. In all cases, certain assumptions must be made regarding the relationship between the sulphate aerosol deposited at the surface and the extent and duration of a significant stratospheric dust veil. These assumptions are highly uncertain and can only be partially tested for a few recent eruptions [*Stenchikov et al.*, 1998]. Volcanic series are generally expressed as aerosol optical depths, but for use by some types of climate models they must be converted to forcing [see *Hansen et al.*, 2002]. This latter fact underscores our previous point that different models "see" a given forcing differently. Some modern GCMs, for example, are given a size distribution of the particles and incorporate radiative transfer calculations involving scattering and absorption of the particles [see *Stenchikov et al.*, 1998; *Ammann et al.*, 2003]. Figure 7 shows the extended volcanic forcing series from *Crowley et al.* [2003], compared to that of *Ammann et al.* [2003] and *Bertrand et al.* [2002]. All three volcano forcing series in Figure 7 show strong similarities, but the choice of ice cores used to define the specific volcanic dust veil chronology clearly influences the resulting series. Many papers focusing on 20th century climate change have taken note of the dearth of volcanic eruptions between 1915 and 1960. It is also equally noteworthy in the context of longer-term climate changes, that there were relatively few eruptions between A.D. 1000 and 1500 relative to the period 1500–2000, provided events have been faithfully recorded in polar ice cores. While we focus in our comparisons of different model simulations in section 5.2, for simplicity, on hemispheric mean responses to volcanic forcing, many models simulate spatial and sometimes seasonally specific responses.

5.1.3. Greenhouse Gas Forcing

[87] Greenhouse gas concentrations are available from instrumental measurements back through the mid-20th century. Prior to that time, they must be estimated from the relevant trace gas concentrations (i.e., CO_2, CH_4, and N_2O) trapped in ice cores. Typically, these trace gas concentrations are globally well mixed on the timescales of interest (decades and longer) so that measurements from ice cores in either hemisphere suffice. Because it takes some time (several years to several decades) for enough snow and ice to accumulate before the ice is sealed off from the atmosphere, trace gas concentrations typically represent smoothed-out estimates of the instantaneous concentrations that existed at the time a given annual layer was deposited. Ice cores from sites with relatively high seasonal deposition rates such as Law Dome, Antarctica, however, can yield greenhouse gas concentrations over the past millennium with at least multidecadal temporal resolution [*Etheridge et al.*, 1996]. The primary trend (i.e., the increase in CO_2 concentrations, from roughly 275 ppm during the preindustrial period of the 18th century to values of $\sim$320 ppm by the mid-20th century) is fairly robust. There are differences, however, between different ice core estimates in the smaller, preindustrial estimates of a bit less than 10 ppm amplitude in preceding centuries [e.g., *Gerber et al.*, 2003]. Like solar radiative forcing, well-mixed greenhouse gases represent a global radiative forcing. However, unlike solar radiative forcing, greenhouse gas concentrations can represent both a response to and a cause of climate variability, complicating the use of longer-term greenhouse gas concentrations as a pure forcing of climate. Prior to industrialization, it is likely that the relatively small estimated variations in CO_2 and CH_4 rather than representing a forcing of temperature changes instead reflect alterations that occurred in terrestrial carbon uptake due to temperature changes [*Gerber et al.*, 2003]. There is, however, some recent evidence [*Ruddiman*, 2003] that there may have been a significant human influence on greenhouse gas concentrations (CO_2 and CH_4) due to cultivation and land use changes over a much longer time period (the past several millennia) than is typically assumed (i.e., the past two centuries).

5.1.4. Sulphate Aerosol Forcing

[88] Sulphate aerosol forcing is not considered important prior to the 20th century but must be included in modeling climate changes over the past century. Sulphate aerosol forcing tends to cool the climate, particularly so on regional scales. Compared to greenhouse gas forcing, sulphate aerosol forcing is far more uncertain, principally because of limited understanding of the radiative properties of the aerosols and their effects on clouds. This forcing is also regionally specific and must be estimated from past fossil fuel use (see, e.g., *Crowley* [2000, and references therein] for further discussion).

5.1.5. Land Use Change Forcing

[89] *Ramankutty and Foley* [1999] have estimated anthropogenic land use changes from historical evidence from 1700 to 1992 taking into account large-scale deforestation trends, conversion of land for agricultural purposes, and other factors. Using these evaluations, *Bauer et al.* [2003] and *Govindasamy et al.* [2001] both estimate that a roughly 0.4°C net cooling of Northern Hemisphere annual mean temperature during the past few centuries could have occurred because of these changes.

5.2. Modeling of Past Climate Changes

[90] Theoretical models of the climate system driven with the best available estimates of the changes in external forcing discussed in section 5.1 can provide important

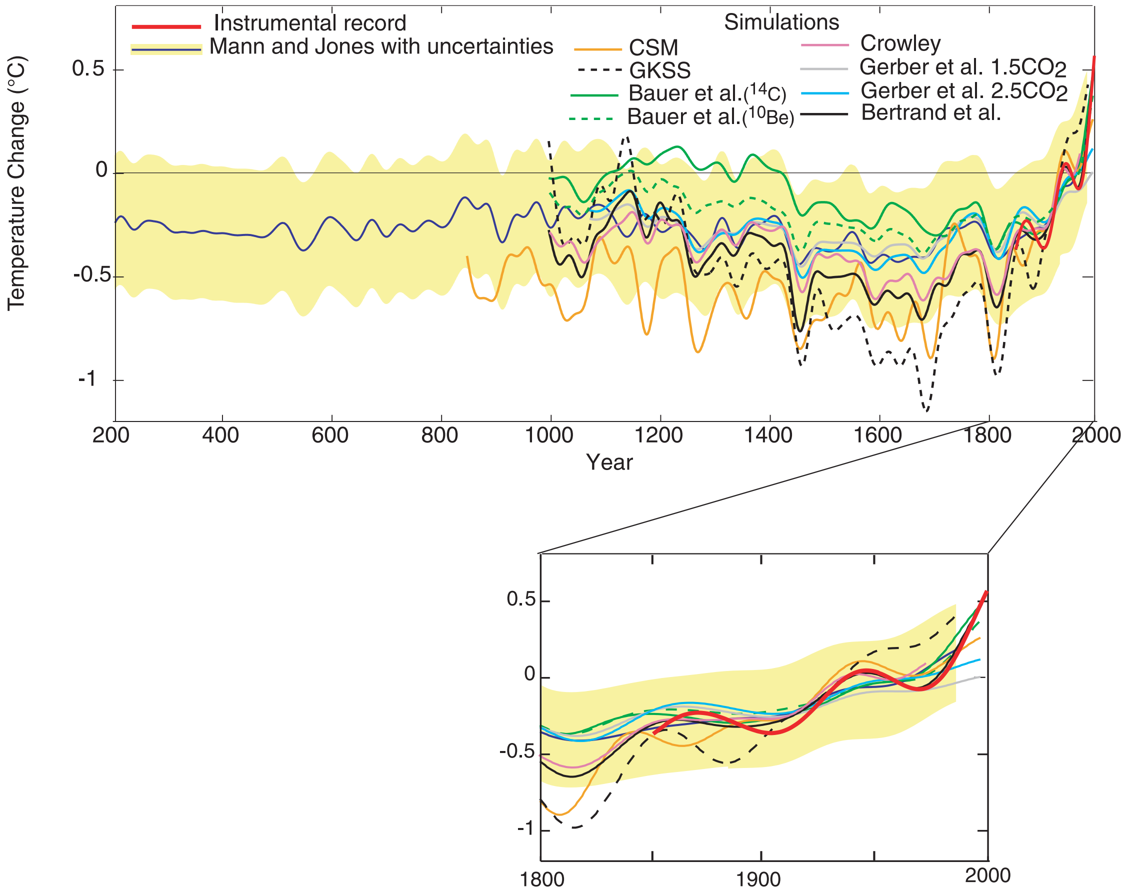


Figure 8. Model-based estimates of NH temperature variations over the past two millennia. Shown are 40-year smoothed series based on the convention used for Figure 5. The simulations are based on varying radiative forcing histories (see Figure 7 and associated discussion), employing a hierarchy of models including one-dimensional energy balance models [*Crowley*, 2000], two-dimensional reduced complexity models [*Bauer et al.*, 2003; *Bertrand et al.*, 2002; *Gerber et al.*, 2003], and full three-dimensional atmosphere-ocean general circulation models (GKSS, Geesthacht, Germany [*Gonzalez-Rouco et al.*, 2003; H. von Storch et al., personal communication, 2003] and Climate System Model (CSM) (C. Ammann et al., personal communication, 2003). Shown for comparison is the instrumental NH record (1856–2003) and the proxy-based estimate of *Mann and Jones* [2003] through 1995 with its 95% confidence interval (see Figure 5). Models have been aligned vertically to have the same mean over the common 1856–1980 period as the instrumental series (which is assigned zero mean during the 1961–1990 reference period). The expansion provides a view of the changes over the past two centuries.

insights into the factors governing climate changes in past centuries. The modeling approach is limited, however, by the imperfectly known history of these changes in forcing, which are typically estimated from indirect and inexact sources (section 5.1). Moreover, the approach only estimates the forced component of climate change; it cannot assess the possible role of internal dynamics (e.g., natural changes in ocean circulation) in the actual climate changes in past centuries, which are, in part, unpredictable from a model simulation. Climate models range from the simplest EBMs to the most complex coupled *atmosphere-ocean general circulation models* (AOGCMs). They are limited by the faithfulness of the physics contained in the model and the accuracy of parameterizations of processes that are not explicitly resolved by the model, such as small-scale convective processes. Simulations of climate changes over the past few centuries have been performed using EBMs [*Crowley and Kim*, 1996; *Free and Robock*, 1999; *Crowley*, 2000], two-dimensional or "intermediate complexity" climate models [*Bertrand et al.*, 2002; *Bauer et al.*, 2003] and coupled climate-carbon cycle models [*Gerber et al.*, 2003], atmospheric three-dimensional GCMs coupled to mixed layer ocean models [*Shindell et al.*, 2001, 2003], and fully coupled three-dimensional AOGCMS [*Cubasch et al.*, 1997; *Gonzalez-Rouco et al.*, 2003]. The results of these simulations (Figure 8) yield remarkably similar conclusions (at hemispheric and global scales) regarding the factors controlling climate changes in past centuries, given the diverse range of models employed.

[91] Unforced, "control" climate model simulations can also be used to provide insights into the role of internal climate processes on the climate variability of the past few centuries to millennium [*Barnett et al.*, 1996, 1999; *Jones*

et al., 1998; *Delworth and Mann*, 2000; *Cubasch et al.*, 2001; *Covey et al.*, 2003; *Braganza et al.*, 2003; *Bell et al.*, 2003]. Though limited by the faithfulness of the internal variability of the model (e.g., the existence of realistic ENSO variability), current generation models seem to simulate such details of the actual climate quite well [*Cubasch et al.*, 2001].

5.3. Model/Data Intercomparison

[92] A number of important conclusions regarding the factors governing climate variability in past centuries can be obtained by a comparison of model simulation results and insights from empirical climate reconstructions. Unforced model simulations, for example, clearly indicate that 20th century warming is outside the range of internally generated climate variability on multicentury and millennial time-scales [e.g., *Osborn et al.*, 1999; *Cubasch et al.*, 2001]. Model simulations indicate a pattern of internal multidecadal coupled ocean-atmosphere variability (involving interactions between the NAO and meridional overturning ocean circulation) that compares well with evidence from proxy-based surface temperature reconstructions [*Delworth and Mann*, 2000]. The associated temperature pattern, however, projects weakly on to hemispheric mean temperature.

[93] Figure 7 compares various estimates of radiative forcings histories over the past one to two millennia, while Figure 8 compares the results of simulations of externally forced changes in NH temperature averages based on these forcing histories with reconstructions of NH mean temperatures over the same time period. Though direct comparisons are complicated by the fact that different models assimilate the forcing in different ways, the proxy reconstructions of NH mean temperatures over the past one to two millennia are remarkably consistent (Figure 8) with the range of results obtained in forced model simulations [*Free and Robock*, 1999; *Crowley and Kim*, 1999; *Crowley*, 2000; *Gerber et al.*, 2003; *Bertrand et al.*, 2002]. Most of the model simulations lie within the estimated 2-standard error uncertainties of the reconstruction, and the reconstruction is roughly at the center of the cluster of various model estimates. Comparisons of modeled and reconstructed NH variations suggest a primary role of volcanic forcing at the hemispheric mean scale prior to 19th and early 20th centuries (after which anthropogenic forcing increasingly dominates hemispheric mean temperature trends [see, e.g., *Mann et al.*, 1998a, 2000b; *Hegerl et al.*, 2003]). The milder first half of the millennium is partly due to the reduced level of volcanism compared to later centuries. The likely importance of past volcanism over the millennium implies that late 20th century warmth is all the more remarkable, given the level of explosive volcanism, which is currently near its highest state of intensity, comparable only with the 17th and 19th centuries. Solar variability appears to play a significant, though somewhat lesser, role, with the relative roles of the forcings somewhat dependent on variable assumptions made in scaling proxy estimates of forcings to actual radiative forcing estimates [*Crowley*, 2000; *Gerber et al.*, 2003; *Bertrand et al.*, 2002; *Bauer et al.*, 2003]. During the 19th and 20th centuries, human land use changes appear to have played a potentially significant role in the large-scale radiative forcing of the climate. Simulations that include this forcing [*Bauer et al.*, 2003] match well to the actual NH mean temperatures during the 19th and early 20th centuries, while simulations without this forcing [*Crowley*, 2000; *Gerber et al.*, 2003; *Bertrand et al.*, 2002] over-predict temperatures at these times. Simulations that do not include land use changes in the 19th and 20th centuries may exhibit an artificially cold pre-19th century mean temperature relative to empirical estimates when, as in our case (Figure 8), the model simulation results have been aligned vertically to have the same mean as the instrumental temperature record during the late 19th and 20th centuries. The coupled model simulations appear to indicate greater variability in NH temperature than the other simulations.

[94] Model-data comparisons can provide some potential insight into the sensitivity of the climate system to radiative forcing [e.g., *Crowley and Kim*, 1999], although the substantial uncertainties in both forcing and response have limited any precise quantitative conclusions. Preliminary coupled climate-carbon cycle model simulations of the past millennium [*Gerber et al.*, 2003] indicate that the larger-amplitude century-scale variability evident in some temperature reconstructions [*Huang et al.*, 2000; *Esper et al.*, 2002] is inconsistent with constraints provided by comparison of modeled and observed preanthropogenic CO_2 variations. This reinforces the evidence for relatively modest (significantly less than 1°C) variations prior to the 20th century evident in other hemispheric temperature reconstructions. Comparisons of these reconstructions and model simulation results thus suggest a moderate (rather than low or high) equilibrium climate sensitivity of roughly 2°–3°C for a doubling of CO_2 concentrations [e.g., *Cubasch et al.*, 2001].

[95] Natural radiative forcing may play a more prominent role in explaining regional variability in surface temperature changes in past centuries, owing to the seasonal and regional details of the large-scale atmospheric response to forcings. Such details have been investigated with GCM simulations of forced past climate changes [*Cubasch et al.*, 1997; *Shindell et al.*, 2001, 2003]. Some simulations indicate a strong NAO- or AO-like component to forced (solar and volcanic) century-scale past climate variations [*Shindell et al.*, 2001, 2003] somewhat consistent with empirical reconstructions (section 4.4.3). The seasonal and spatial details of the response to volcanic forcing [*Shindell et al.*, 2003], which indicates large continental summer cooling but a tendency for a dynamically induced, offsetting winter warming, imply that regionally and seasonally restricted proxy data may provide a biased estimate of actual large-scale annual temperature changes [*Mann*, 2002a]. Such details need to be taken into account in drawing inferences from reconstructions based on proxy data with different spatial and seasonal emphases [*Rutherford et al.*, 2004]. Volcanic radiative forcing of changes in ENSO in past centuries [*Adams et al.*, 2003] may explain preliminary empirical evidence (section 4.4.1) for a prevalence of La Niña-like conditions during the 11th–14th centuries and El Niño-like conditions during the 17th century [*Cobb*, 2002; *Cobb et al.*,

2003a]. This underscores the principle that hemispheric mean temperature reconstructions based on extratropical data alone may bias estimates of the true hemispheric mean temperature variations in past centuries [e.g., *Mann*, 2002a; *Rutherford et al.*, 2004].

[96] The modest amplitude of variations in the Northern minus Southern Hemisphere temperature difference or ''interhemispheric contrast'' over the past one to two millennia [*Mann and Jones*, 2003] casts some doubt [see, e.g., *Crowley and Kim*, 1994] on the proposition that hemispheric mean temperature variations over this time frame have been driven by changes in the thermohaline circulation [e.g., *Bond et al.*, 2001], either externally or internally generated. Recent modeling evidence suggests that atmospheric influences, rather than ocean circulation influences, are likely to dominate climate variability in the extratropical Northern Hemisphere [*Seager et al.*, 2002]. Correlations that are evident between diagnostics of the ocean circulation and proxies of solar irradiance variability on centennial to millennial timescales [*Bond et al.*, 2001; *Waple et al.*, 2002; *Andrews et al.*, 2003] may result from solar influences on the North Atlantic (NAO/AO) atmospheric circulation variability discussed in section 4.4.3 through atmospheric forcing of moderate-amplitude variations in the meridional overturning ocean circulation [*Delworth and Dixon*, 2000]. The associated forced variations in oceanic heat transport in combination with the expected response to past radiative forcing [*Shindell et al.*, 2001, 2003] might explain tentative proxy evidence for a modest basin-wide component of North Atlantic SST cooling on multicentury and millennial timescales [*deMenocal et al.*, 2000; *Bond et al.*, 2001] superimposed on larger regional variations more indicative of an NAO/AO pattern [*Shindell et al.*, 2001].

6. FUTURE DIRECTIONS: WHERE DO WE GO FROM HERE?

[97] Section 4 indicates that the broadest features of the ''so-called'' MWP and LIA can be seen in the reconstructions of large-scale mean temperature variations over the past millennium, but such reconstructions show considerably greater detail that defies the use of these simplistic terms. This detail will continue to be tuned and honed in the future as the paleoclimate research community incorporates results from a greater variety of proxies from diverse locations. As the detail of our knowledge improves, the ''MWP'' and ''LIA'' are increasingly likely to be regarded as overly restrictive terms. Following the conventions of researchers studying instrumental climate changes, we encourage paleoclimatologists to use specific calendar dates (e.g., specific decade or century time markers) rather than ill-determined descriptors such as the ''Little Ice Age'' or ''Medieval Warm Period'' to describe climate changes over the last few millennia. In the instrumental period, warming in the 20th century is generally qualified as occurring during the early to middle and/or the later decades [*Jones and Moberg*, 2003]. This interpretation is only possible with hemispheric and global average series and would be impossible to determine from series at local or regional scales. In a similar vein the climate of the recent two millennia at the largest of scales can only be described using composite series of the type displayed in Figure 5.

[98] Likely expansion and improvements in the network of available proxy series will increasingly in coming years lead to viewing the restrictive terms such as the MWP and LIA as obsolete. We anticipate the field of paleoclimatology moving toward data sets that can be used to develop spatial maps (with associated errors) for each season of the last one to two millennia, using techniques that can easily accommodate continual revision to our best available estimates as new series are developed. Particularly at regional scales, such revisions will arise from the addition of specific new high-quality proxy series. The method used to combine the series into multiproxy composites is of secondary importance, given the similarity of the CFR and CPS approaches at the hemispheric/global scales. CFR variants are preferred, however, as they provide estimates of the spatial patterns of past changes. Such spatial estimates of past change are particularly important in comparisons with integrations from AOGCMs and for insight into the role of climate processes such as ENSO and the NAO in past variations.

[99] Our understanding of the past should adapt as new evidence becomes available rather than have to conform to outdated concepts, such as the notion of globally or hemispherically synchronous MWP or LIA periods. A clear priority for future work, given the spatial availability of current proxy series, however, should be the development of many more quality records in the SH and tropical areas, given the obvious dearth (see, e.g., Figure 1) of information available in those regions today. Longer and more detailed records will lead to reduced uncertainties in key quantities, such as global mean and hemispheric mean temperature, and indices of the ENSO phenomenon. As the spatial and temporal details of climate changes during the past millennium are increasingly resolved through expanded and improved networks of multiproxy data, it should soon be possible to use high-resolution reconstructions of the recent past as a template for calibrating networks of longer, lower-resolution proxy data. This latter possibility holds prospects for reconstructing the spatial details of climate changes over several millennia, potentially resolving key details regarding the climate changes of the entire postglacial (''Holocene'') period of the past 10,000 years.

[100] The temperature history of the past few millennia is just one facet of the past changes in the climate system, and it is important not to ignore past changes in other societally relevant climate variables. Efforts should be made in parallel to increase understanding of long-term changes in precipitation and drought patterns and in the links to the controlling circulation influences. As we have discussed (section 2), no single proxy climate record can provide the whole story of past climate change. The various paleoclimatic disciplines need not only to work more closely together but also to become more involved with climate modelers and climate scientists focusing on modern, instrumental climate records. Only by cooperation between the different disciplines can

climatologists improve our understanding of the largest and most complex system on the planet, the global climate. From the climate-modeling perspective, expanded histories of the past will provide details that will test and improve both understanding and the climate models.

7. CONCLUSIONS AND SUMMARY

[101] The purpose of this review has been to detail not only our current best knowledge about the course of climate change over the past few millennia but also to express how best to report advances in the field in the future. The emphasis on the past few millennia restricted the types of proxy information that can be used to those providing high-resolution (annual or decadally resolved) exactly dated information. We therefore described in detail instrumental records, documentary material, tree rings, corals, and ice cores, as these have a proven track record of producing the long, verifiable reconstructions that are required to extend our knowledge of past climatic changes. Less well resolved (in a temporal sense) proxies can provide more qualitative information about the past, but their use is restricted by both dating uncertainties and less rigid methods of calibration against instrumental climate records. Use of these proxies in multiproxy composites has thus been far more limited.

[102] In describing past changes we focus not just on proxy reconstructions of past temperature history but also on associated changes in a number of other fields such as precipitation and drought patterns and atmospheric circulation diagnostics, as well as the complementary changes in these variables in climate model integrations. Our review reaffirms that the warmth of the late 20th century has been unprecedented at the NH and, likely, at global scales in at least a roughly two-millennium (1800 years) context. The 20th century has seen the greatest temperature change within any century in the past two millennia (0.6°–0.9°C compared to less than approximately ±0.2°C for any other century). The coolest centuries appear to have been the 6th, 15th, 17th, and 19th centuries. Regional conclusions, particularly for the Southern Hemisphere and parts of the tropics where high-resolution proxy data are sparse, are more uncertain. There is also evidence, albeit far more tentative, that particular modes of climate variability, such as the ENSO and NAO phenomena may have exhibited late 20th century behavior that is anomalous in a long-term context.

[103] Most earlier studies of the recent past (particularly the last millennium) have interpreted the temperature changes that have occurred in terms of the MWP and LIA, terms introduced by pioneering paleoclimatologists in the 1960s working in the European/North Atlantic sector of the NH. We argue, on the basis of differences between regional and true hemispheric/global temperature trends evident during the instrumental period that past warm/cold periods can only be determined from truly hemispheric- and global-scale series. Inferences from regional data in isolation will clearly provide a biased view of larger-scale changes. Over longer periods (e.g., the past couple millennia), differences are clearly apparent between individual proxy series and the hemispheric/global composites (see our Figures 4 and 5). "Medieval Warm Period" and "Little Ice Age" are therefore restrictive terms, and their continued use in a more general context is increasingly likely to hamper, rather than aid, the description of past large-scale climate changes. We recommend that paleoclimatologists avoid the use of such terms and instead refer to anomalous climate periods by calendar dates, as is the practice in the description of more modern climate changes.

[104] Assessment of the empirical evidence provided by proxies of climate change over the past two millennia, combined with climate modeling efforts to explain the changes that have occurred during the period, indicates that solar and volcanic forcing have likely played the dominant roles among the potential natural causes of climate variability. Neither can explain, however, the dramatic warming of the late 20th century; indeed, natural factors would favor a slight cooling over this period. Only anthropogenic influences (principally, the increases in greenhouse gas concentrations) are able to explain, from a causal point of view, the recent record high level of global temperatures during the late 20th century.

GLOSSARY

Antarctic Oscillation (AAO): Measure of the pressure gradient between the polar and subpolar regions of the Southern Hemisphere; term introduced by *Thompson and Wallace* [2000].

Age-band decomposition (ABD): Tree ring width/density method of standardization introduced by *Briffa et al.* [2001]. Aim is to maintain as much low-frequency variance as possible in the chronologies developed. See also regional curve standardization.

Arctic Oscillation (AO): Measure of the pressure gradient between the Arctic and subtropical highs in the Northern Hemisphere; term introduced by *Thompson and Wallace* [1998]. Whether this index is different from the NAO is still a matter for debate in climatology.

Atmosphere-ocean general circulation model (AOGCM): Fully coupled atmosphere-ocean model of the three-dimensional global climate.

Central England temperature (CET): Longest monthly instrumental record of temperature for any location in the world. Record was developed by *Manley* [1974] and updated [*Parker et al.*, 1992] regularly by the Hadley Centre of the Met Office in the United Kingdom.

Climate field reconstruction (CFR): Approach to reconstructing a target large-scale climate field from predictors employing multivariate regression methods.

Composite-Plus-Scale (CPS): Approach to reconstructing a target index based on compositing of predictors followed by scaling by the amplitude of the target index.

Energy balance model (EBM): Simple climate model consisting of a uniform ocean and atmosphere that respond thermodynamically, but not dynamically, to changes in radiative forcing.

El Niño-Southern Oscillation (ENSO): A natural coupled mode of climate variability associated with both surface temperature variations tied to El Niño (see Niño3/3.4) and atmospheric circulation changes across the equatorial Pacific (see Southern Oscillation Index); term first coined by *Rasmusson and Carpenter* [1982].

Empirical orthogonal functions (EOFs): Spatial pattern tied to a particular mode of time/space variance in a spatiotemporal data set (see principal components analysis).

General circulation model (GCM): Typically refers to a three-dimensional model of the global atmosphere used in climate modeling (often erroneously also called Global Climate Model). This term is often misused and typically requires additional qualification (e.g., as to whether or not the atmosphere is fully coupled to an ocean.) See Atmosphere-ocean general circulation model.

Ground surface temperature (GST): Temperature of the ground surface (beneath any vegetation or snow/ice layer), which is distinguished from the temperature of the overlying surface air. See Surface air temperature.

Hadley Centre/Climatic Research Unit version 2 (HadCRUT2v): The 5° by 5° grid box data set (version 2 with variance correction) temperature data set detailed by *Jones and Moberg* [2003]. Data are updated monthly and are available from http://www.cru.uea.ac.uk.

Little Ice Age (LIA): Term originally introduced in the late 1930s by *Matthes* [1939] to describe the greatest glacial advance of the Holocene period. No space scale is attached to the term, so timing could differ from location to location. In the climatological literature the LIA is generally considered to have been the period from around A.D. 1300 to 1450 until A.D. 1850 to 1900. Variants of these dates abound, and the term is often used by paleoclimatologists and glaciologists without formal dates attached.

Medieval Warm Period (MWP): Also referred to as the Medieval Warm Epoch (MWE). As with the LIA, no date range is universally accepted, but the dates A.D. 900–1300 cover most ranges. Origin is difficult to track down, but it is believed to have been first used in the 1960s (probably by *Lamb* [1965], see also *Bradley et al.* [2003]).

North Atlantic Oscillation (NAO): Measure of the strength of the westerlies across the North Atlantic. Originally defined by G. T. Walker [e.g., *Walker and Bliss*, 1932] as the difference in pressure between Ponta Delgada on the Azores and Stykkisholmur in Iceland. Numerous definitions currently abound, discussed by *Jones et al.* [2003b]. See also Arctic Oscillation.

Northern Hemisphere (NH): The region of the globe between the equator and North Pole. A proper estimate of NH mean temperature should represent an areal average over this entire region.

Niño3 Index (Niño3): SST variations in a region of the eastern equatorial Pacific (90°–150°W, 5°N to 5°S) representative of the El Niño phenomenon (see El Niño-Southern Oscillation).

Principal Components (PCs): Time history tied to a particular mode of time/space variance in a spatiotemporal data set (see principal components analysis).

Principal components analysis (PCA): A procedure by which a spatiotemporal data set is decomposed into its leading patterns in both time (see principal components) and space (see empirical orthogonal functions) based on an orthogonal decomposition of the data covariance matrix.

Pacific Decadal Oscillation (PDO): A pattern of variability in the ocean and atmosphere that appears to be centered in the extratropical North Pacific, which emphasizes decadal, rather than interannual, timescales; term introduced by *Mantua et al.* [1997].

Pacific North American (PNA) pattern: A typical pattern of seasonal and longer-term variability in the atmospheric circulation over North America and neighboring maritime regions; term introduced by *Wallace and Gutzler* [1981].

Regional curve standardization (RCS): Tree ring width/density method of standardization introduced by *Briffa et al.* [1992a]. Aim is to maintain as much low-frequency variance as possible in the chronologies developed. See also age-band decomposition.

Surface air temperature (SAT): The temperature of the air overlying a surface region of the Earth distinguished from the surface temperature of the ground itself (see ground surface temperature).

Southern Hemisphere (SH): The region of the globe between the equator and South Pole. A proper estimate of SH mean temperature should represent an areal average over this entire region.

Sea level pressure (SLP): The air pressure that would be expected to be measured in a particular location if the altitude were reduced to sea level based on standard assumptions of atmospheric composition and physics.

Southern Oscillation Index (SOI): A measure of the difference in sea level pressure between the western (e.g., Darwin, Australia) and central/eastern (e.g., Tahiti) equatorial Pacific, representative of the east-west changes in atmospheric circulation associated with the ENSO phenomenon; term introduced by G. T. Walker [e.g., *Walker and Bliss*, 1932].

Sea surface temperature (SST): The temperature measured from the surface water of the ocean (as measured historically from bucket or ship intake water measurements) distinguished from the temperature of the overlying air.

[105] **ACKNOWLEDGMENTS.** We thank numerous colleagues for very useful discussions and suggestions regarding various aspects of this work, including Caspar Ammann, Cedric Bertrand, Ray Bradley, Keith Briffa, Ed Cook, Tom Crowley, Malcolm Hughes, Anders Moberg, Ellen Mosley-Thompson, Astrid Ogilvie, Tim Osborn, Stefan Rahmstorf, Alan Robock, Scott Rutherford, Gavin Schmidt, Drew Shindell, Simon Tett, Kevin Trenberth, Dennis Wheeler, and Tom Wigley. We thank Kim Cobb and Dave Stahle and two anonymous reviewers for their thoughtful and extensive reviews. We also thank Caspar Ammann, Eva Bauer, Cedric Bertrand, Rudolf Brázdil, Keith Briffa, Victor Brovkin, Martin Claussen, Ed Cook, Tom Crowley, Tom Cronin, Judith Lean, Jürg Luterbacher, Valerie Masson-Delmotte, Anders Moberg, Ellen Mosley-Thompson, Tim Osborn, Wang Shao-Wu, Lonnie Thompson, Aryan van Engelen,

Tas van Ommen, Bo Vinther, Hans von Storch, and Eduardo Zorita for providing data or information regarding proxy/climate model/forcing data used in or referred to in this study. We thank Tim Osborn and Scott Rutherford for technical assistance. P. D. J. acknowledges support of the Office of Science (BER), U.S. Department of Energy, grant DE-FG02-98ER62601. P. D. J. also acknowledges the Climate Change Detection and Attribution Project, a jointly funded effort by NOAA's Office of Global Programs and the Department of Energy's Office of Biological and Environmental Research. M. E. M. acknowledges support for this work by the NSF and NOAA-sponsored Earth Systems History (ESH) program (NOAA award NA16GP2913).

[106] Thomas Torgersen was the Editor responsible for this paper. He thanks three technical reviewers and one cross-disciplinary reviewer.

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