The Climate of the Last Millennium

R.S. Bradley
Climate System Research Center, Department of Geosciences, University of Massachusetts, Amherst, MA, 01003, United States of America

K.R. Briffa
Climatic Research Unit, University of East Anglia, Norwich, NR4 7TJ, United Kingdom

J. Cole
Department of Geosciences, University Arizona, Tucson, AZ 85721, United States of America

M.K. Hughes
Lab. of Tree-Ring Research, University of Arizona, 105 W. Stadium, Bldg. 58, Tucson, AZ 85721, United States of America

T.J. Osborn
Climatic Research Unit, University of East Anglia, Norwich, NR4 7TJ, United Kingdom

6.1 Introduction

We are living in unusual times. Twentieth century climate was dominated by near universal warming with almost all parts of the globe experiencing temperatures at the end of the century that were significantly higher than when it began (Figure 6.1) (Parker et al. 1994; Jones et al. 1999). However the instrumental data provide only a limited temporal perspective on present climate. How unusual was the last century when placed in the longer-term context of climate in the centuries and millennia leading up to the 20th century? Such a perspective encompasses the period before large-scale contamination of the global atmosphere by human activities and global-scale changes in land-surface conditions. By studying the records of climate variability and forcing mechanisms in the recent past, it is possible to establish how the climate system varied under "natural" conditions, before anthropogenic forcing became significant. Natural forcing mechanisms will continue to operate in the 21st century, and will play a role in future climate variations, so regardless of how anthropogenic effects develop it is essential to understand the underlying background record of forcing and climate system response.

Sources of information on the climate of the last millennium include: historical documentary records, tree rings (width, density); ice cores (isotopes, melt layers, net accumulation, glaciochemistry); corals (isotopes and other geochemistry, growth rate); varved lake and marine sediments (varve thickness, sedimentology, geochemistry, biological content) and banded speleothems (isotopes). These are all paleoclimatic proxies that can provide continuous records with annual to decadal resolution (or even higher temporal resolution in the case of documentary records, which may include daily observations, e.g. Brázdil et al. 1999; Pfister et al. 1999a,b; van Engelen et al. 2001). Other information may be obtained from sources that are not continuous in time, and that have less rigorous chronological control. Such sources include geomorphological evidence (e.g. from former lake shorelines and glacier moraines) and macrofossils that indicate the range of plant or animal species in the recent past. In addition, ground temperature measurements in boreholes reflect the integrated history of surface temperatures, with temporal resolution decreasing with depth. These provide estimates of overall ground surface temperature changes from one century to the next (Pollack et al. 1998; Huang et al. 2000).

Proxies of past climate are natural archives that have, in some way, incorporated a strong climatic signal into their structure (Bradley 1999). For some biological proxies, such as tree ring density or coral band width, the main factor might be temperature – or more specifically, the temperature of a particular season (or even just part of a season). Ring density and width can also be influenced by antecedent climatic conditions, and by other non-climatic factors. Similar issues are important in other proxies, such as the timing of snowfall events that make up an ice core, or the rate and timing of sediment transport to a lake. Though we recognize that the details of such relationships are important, proxies are rarely interpreted directly in terms of such very

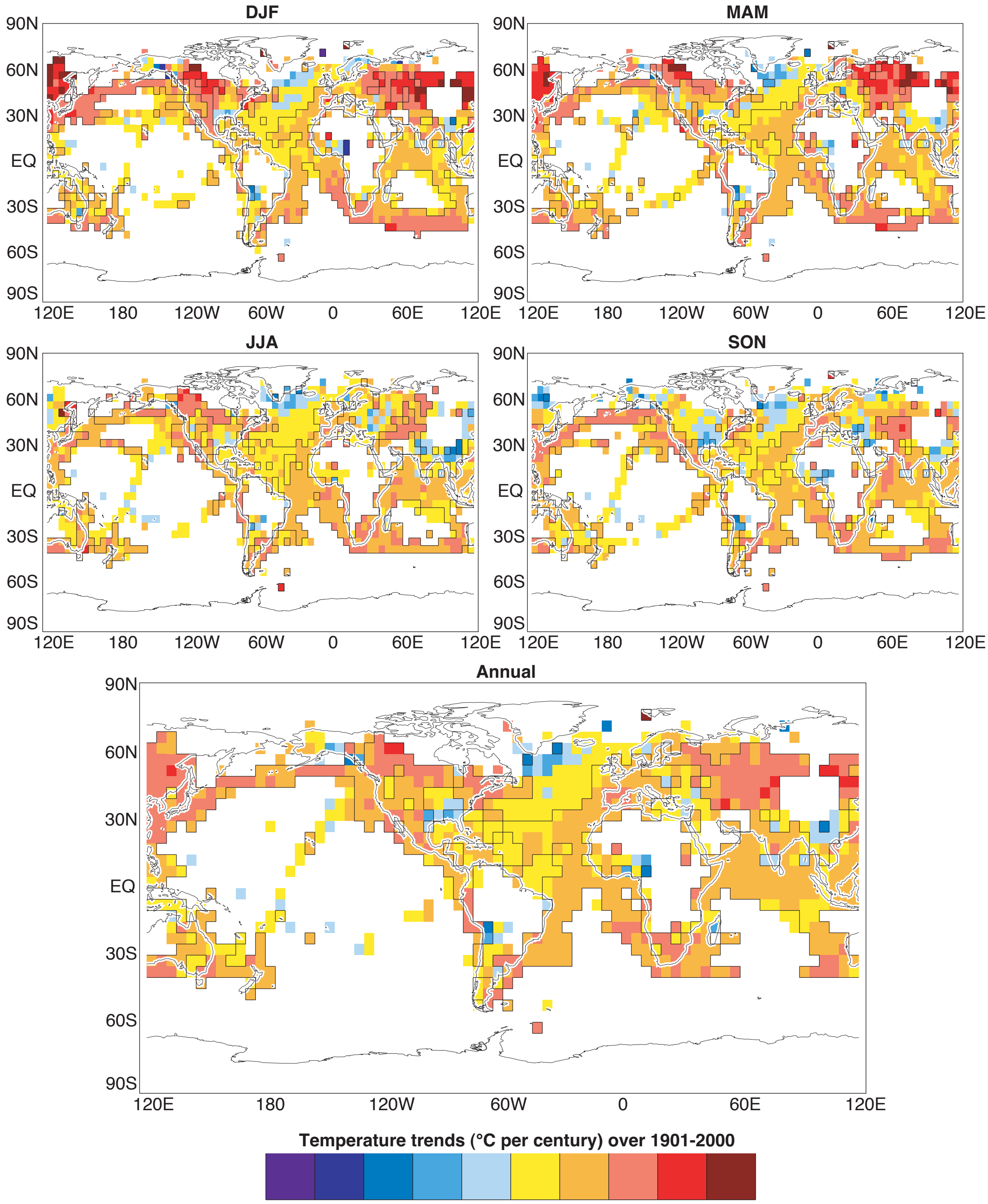

Fig. 6.1. Seasonal and annual trends in surface air temperature, 1901-2000, based on instrumental measurements. Black outlining surrounds those regions with statistically significant trends (at the 95% confidence level).

specific climatic controls, but rather in terms of temperature or precipitation over a particular season. In many cases the main climatic signal in a proxy record is not temperature alone. For example, evidence of a formerly high lake level may indicate higher rainfall amounts and/or a decrease in evaporation related to cooler temperatures. Such issues are grist to the paleoclimatologists' mill and are the subject of numerous studies. Suffice it to say that proxies that optimize a reconstruction of either temperature or precipitation are generally selected, and it is these studies that provide the basis for our review.

Changes in temperature have large-scale spatial coherence, making it easier to identify major variations with relatively few records. Spatially coherent precipitation changes are more local or regional in extent, but they often reflect circulation changes that may have large-scale significance (as, for example, in ENSO-related spring rainfall increases that commonly occur in the southwestern U.S. during strong El Niño events; Stahle et al. 1998). In this chapter, we focus mainly on temperature variations, but precipitation and hydrological variability are examined where there is good evidence for important changes at the regional scale. In particular, we ask two questions regarding each attribute:

- does the 20th century record indicate unique or unprecedented conditions?
- do 20th century instrumental data provide a reasonable estimate of the range of natural variability that could occur in the near future?

First we briefly summarise conditions during the Holocene epoch (the last 10,000 radiocarbon, or ~11,700 calendar years), as a background to climate variability over the last millennium. Then we examine the overall pattern of temperature change during the last 1000 years at the largest (hemispheric) scale, followed by a consideration of climate variability in several large regions. Finally we look at the forcing factors that may have played a role in the variations identified.

6.2 Holocene climate variability

The Holocene epoch follows the last major pulse of glaciation (the Younger Dryas interval) at the end of the last glacial period, and encompasses a period of time before there was any substantial anthropogenic forcing of climate. The Holocene has often been characterized as a period of relatively stable climate, yet there is much evidence to the contrary. In particular, the tropics and sub-tropics witnessed dramatic changes in hydrological conditions during this period. Early to mid-Holocene conditions in the northern deserts of Africa were significantly wetter than in the late Holocene, as revealed by the evidence of extensive early Holocene lake sediments, and fossils of herbivores and aquatic reptiles in areas that are today utterly arid (Petit-Maire and Riser 1983). In fact, evidence for much drier conditions in the late Holocene (after ~4000 calendar years B.P.) is also found across central Asia and into Tibet (Gasse and van Campo 1994). Over much of this region, conditions today are the driest they have been throughout the Holocene. By contrast, lakes of inland drainage on the Altiplano of Peru and Bolivia have expanded and increased in depth from the mid-Holocene to the present. Lake Titicaca, for example, is currently close to its highest level in the Holocene. Similarly, in northern Chile, lacustrine and archeological evidence points to arid conditions from ~8000-3700 years B.P., followed by wetter conditions in the late Holocene (Grosjean et al. 1995). This is comparable to the situation in the western United States, especially Nevada and eastern California (Thompson 1992; Benson et al. 1996; Quade et al. 1998). Furthermore, low latitude hydrological changes in the Holocene were often abrupt (Gasse 2000; De Menocal et al. 2000) (e.g. at ~4200 calendar years B.P. in North Africa and the Middle East, when many freshwater lakes were reduced to swamps and arid lowlands within less than a century). As one might expect, such changes had significant impacts on ecosystems and the people living in those areas, in some cases resulting in complete societal collapse (Weiss et al. 1993; Dalfes et al. 1997; Weiss and Bradley 2001; De Menocal 2001).

A coherent picture is also emerging of a distinctly different pattern of El Niño/Southern Oscillation (ENSO) variability before ~5,000 years B.P. (Clement et al. 2000; Cole 2001). Most data and model results are consistent with a more La Niña-like background state and reduced inter-annual variability during this period (see Chapter 3, Section 3.31 for a complete discussion). For example, paleoclimatic observations indicate reduced incidence of heavy rains in Ecuador, absence of strong annual rainfall extremes in northern Australia, warmer SST along the northern Great Barrier Reef, and attenuated inter-annual variance in the ENSO-sensitive warm pool north of New Guinea (McGlone et al. 1992; Shulmeister and Lees 1995; Gagan et al. 1998; Tudhope et al. 2001). General circulation models forced with early Holocene orbital conditions simulate a cooler eastern/central tropical Pacific, due to intensified trades associated with a stronger Asian monsoon (Otto-Bliesner 1999; Bush 1999; Liu et al. 2000). The latter two references show warming in the westernmost tropical Pacific, much as La Niña brings today. Although these studies disagree on changes in the amplitude of inter-annual variability, a simpler model suggests that precessional forcing should result in weakened interannual ENSO strength. When radiation anomalies are strongly positive in boreal summer, as in the early to mid-Holocene, they result in stronger trade winds that cool the eastern Pacific during autumn, inhibiting the development of warm El Niño events. It is also of interest that Otto-Bliesner (1999) found that ENSO teleconnection patterns were very different at 6,000 B.P. compared to modern; this result cautions against inferring ENSO variability from sites not in close proximity to the tropical Pacific. For example, observations of mid-Holocene climate changes in continental South America and particularly in West Africa may reflect the effects of conditions in the Atlantic rather than in the PacificOcean.

There is a stark contrast between such a picture of major hydrological and circulation changes in the Tropics, with dramatic environmental consequences, and the record of relative stability seen in the well-known ice core accumulation and isotopic records from central Greenland (GRIP and GISP2 sites) (Dansgaard et al. 1993; Meese et al. 1994). However, ice core isotopic and summer melt layer data, from northern Greenland and smaller ice caps around the Arctic, do indicate a general cooling through most of the Holocene (i.e. a decrease in melt layers and lower $\delta^{18}O$), with warmest conditions in the first few millennia of the period (Koerner and Fisher 1990) (Figure 6.2). Furthermore, borehole temperatures at Summit, Greenland point to mean annual temperatures ~3ºC warmer in the early Holocene compared to the last ~500 years (Dahl-Jensen et al. 1998). Thus, the GRIP/GISP2 isotope record appears to be anomalous.

There is also much evidence from other high latitude regions that temperatures generally declined during the Holocene (Figure 6.3). For example, diatom-based SST reconstructions for the Greenland and Norwegian Sea area indicate higher temperatures from ~9000 to ~4000 years B.P. (Koç et al. 1993). In central Sweden trees grew well above the modern altitudinal treeline from 9,000 to ~2,000 calendar years B.P (Kullman 1989) and on the Kola Peninsula, Scots pine grew ~20 km north of the modern polar limit from ~7600 to ~4000 calendar years B.P. (MacDonald et al. 2000a). Similarly, trees grew north of the modern treeline in much of Siberia and in the Mackenzie River delta in early Holocene time (>8000 years ago) (Ritchie 1987; Burn 1997; Macdonald et al. 2000b). There is also evidence of extensive open water conditions in the Beaufort Sea and in the Canadian Arctic islands in the early Holocene, as documented by the skeletons of numerous bowhead whales and other marine mammals (that require relatively ice-free conditions) on raised beaches dating from that period (Dyke and Morris 1990; Dyke and Savelle 2001). At that time, driftwood was carried far into (seasonally ice-free) arctic fiords, whereas pervasive sea ice prevented such movement in the late Holocene. Furthermore, during the last few millennia of the Holocene land-fast ice shelves formed along the shores of the Arctic Ocean on Ellesmere Island where they are still found today (Bradley 1990). This Holocene cooling was not limited to high latitudes. Treeline in the White Mountains of eastern California (37°18'N) was 100 to 150 meters higher than the modern level from the sixth millennium B.C. until roughly 2200 B.C., declining most rapidly after AD 1000 (La Marche 1973). Although caution should be exercised in interpreting treeline movements in such arid regions (Lloyd and Graumlich 1997), the preponderance of evidence supports LaMarche's estimate of a 2°C cooling in warm-season temperatures.

Ice core deuterium (δD) records from Antarctica also indicate a general decline in temperature through the Holocene, with warmest conditions in the first few millennia (Ciais et al. 1992; Masson et al 2001). Studies of deuterium excess in four Antarctic ice cores show an overall increase through the Holocene which is thought to be related to warmer early Holocene sea surface temperatures in the precipitation source regions, the low-latitude oceans of the southern hemisphere (Vimeux et al. 2001).

In the northern hemisphere, late Holocene expansion of glaciers (from mid or early Holocene minima) and the redevelopment of small ice caps accompanied late Holocene cooling (e.g. Nesje and Kvamme 1991; Nesje et al. 2001). It is arguable as to when this period began, but there is much evidence that the onset of this "neoglaciation" occurred ~4000-5000 years B.P. (Porter & Denton 1967; Grove 1988). A series of oscillations in ice extent in mountainous regions around the world characterises the last few thousand years, but the latest of these (within the last few hundred years) was generally the most extensive, indicating the overall severity of climate during this interval. The generic term "Little Ice Age" is commonly used to describe this episode, which is now considered to have occurred during the interval ~A.D. 1250-1880, but with the main phase, after ~A.D. 1550 (cf. Bradley and Jones 1992a; Grove 2001 a,b) (see further discussion below in Section 6.5).

These data all show that there was a period of relatively warm conditions in the first half of the Holocene, in many areas warmer than in the 20th century, after which temperatures generally

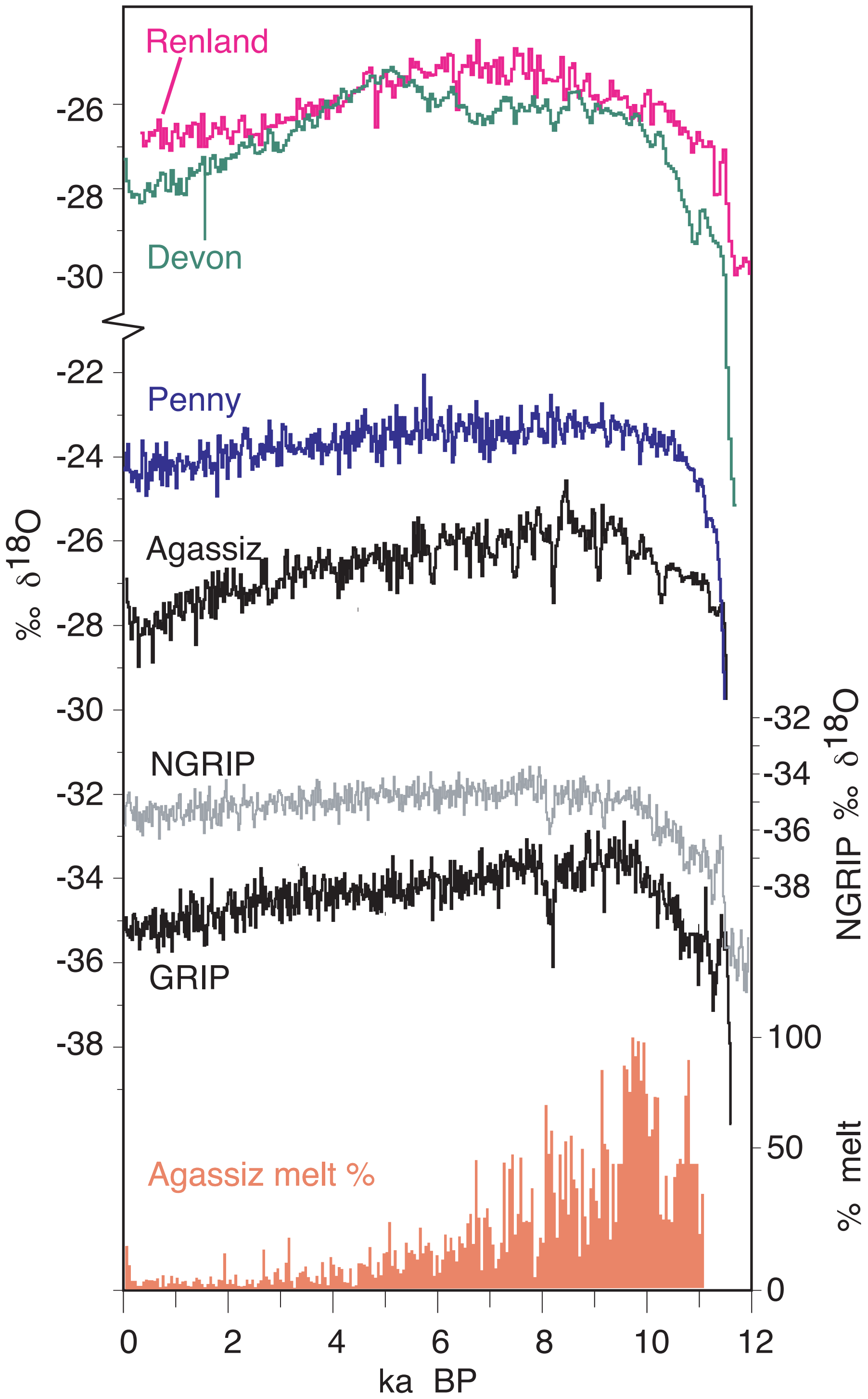

Fig. 6.2. Holocene $\delta^{18}O$ records from Greenland (Renland, GRIP) and Canadian Arctic ice caps (Devon, Penny, Agassiz). Also shown is the record of melt in cores from the Agassiz Ice Cap, Ellesmere Island, Arctic Canada (% of core sections showing evidence of melting and percolation of meltwater into the firn) (after Fisher and Koerner 2002).

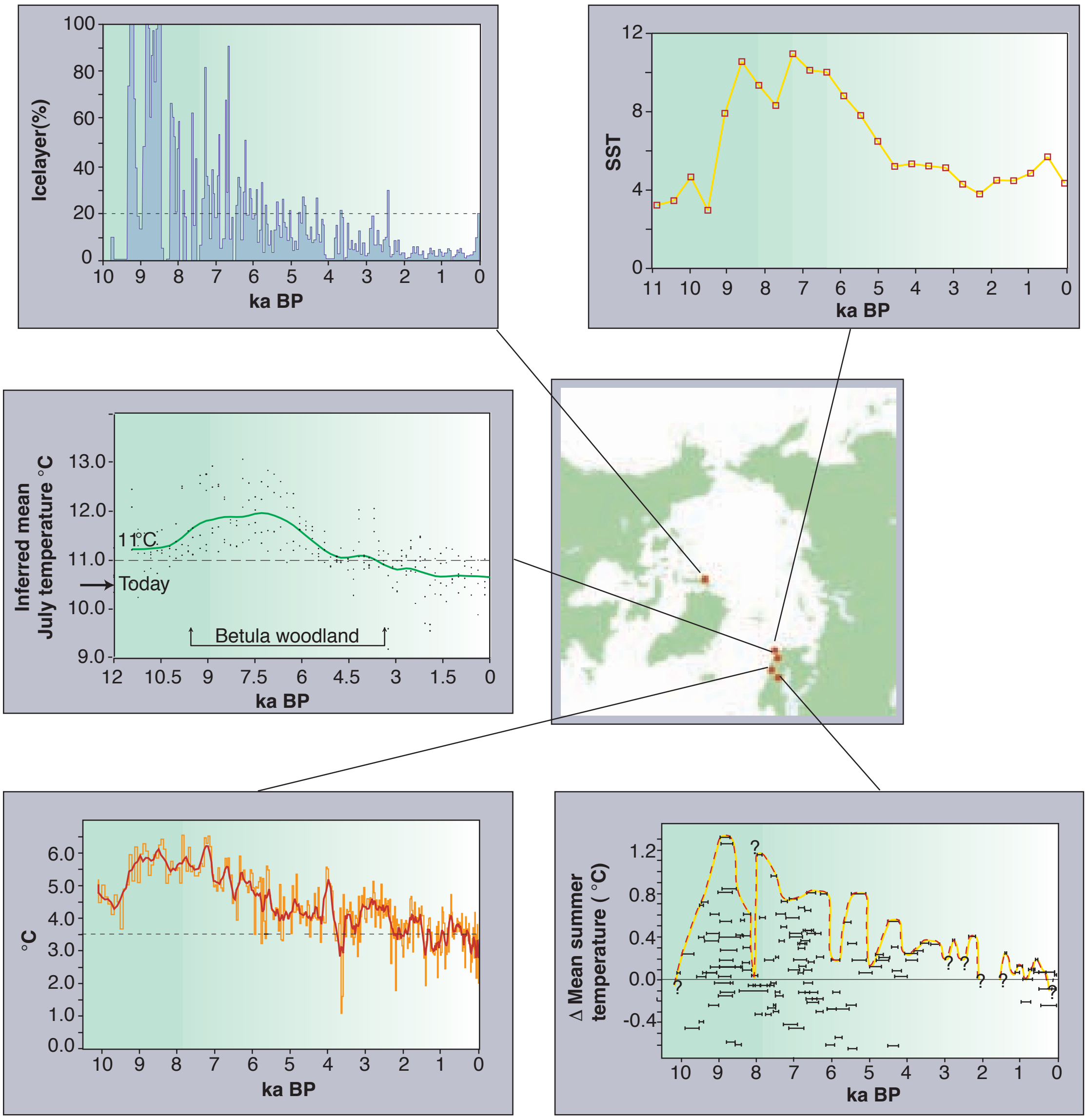

Fig. 6.3. Composite of records showing Holocene temperature changes. Top left: Melt record from Agassiz Ice cap (Koerner and Fisher 1990). Top right: Diatom-inferred SST (Koc et al. 1993). Middle left: Pollen-inferred mean July temperature (Seppä and Birks 2001). Bottom left: summer temepratures from oxygen isotopes (Lauritzen 1996). Bottom right: mean summer temperature from the upper limit of pines (Dahle and Nesje 1996)

declined. The decline was punctuated by centennial-scale warmer and colder episodes, with the most recent cold episode (~A.D. 1550-1850) being the coldest period of the entire Holocene, especially in arctic and sub-arctic regions (Bradley 2000).

6.3 Temperatures over the last millennium

Most high resolution paleoclimate records (i.e. those with annual resolution and a strong climate signal) do not extend back in time more than a few centuries. Consequently, while there are numerous paleoclimate reconstructions covering the period from the 17th century to the present, the number of high resolution millennium scale records is very limited. Continuous records are restricted to ice cores and laminated lake sediments, where the climatic signal is often poorly calibrated, and to a few long tree ring records, generally from high latitudes. Inevitably, this leads to large uncertainties in long-term climate reconstructions that attempt to provide a global or hemispheric-scale perspective. Bearing this in mind, what do current reconstructions tell us about the last millennium?

Figure 6.4 shows a reconstruction of northern

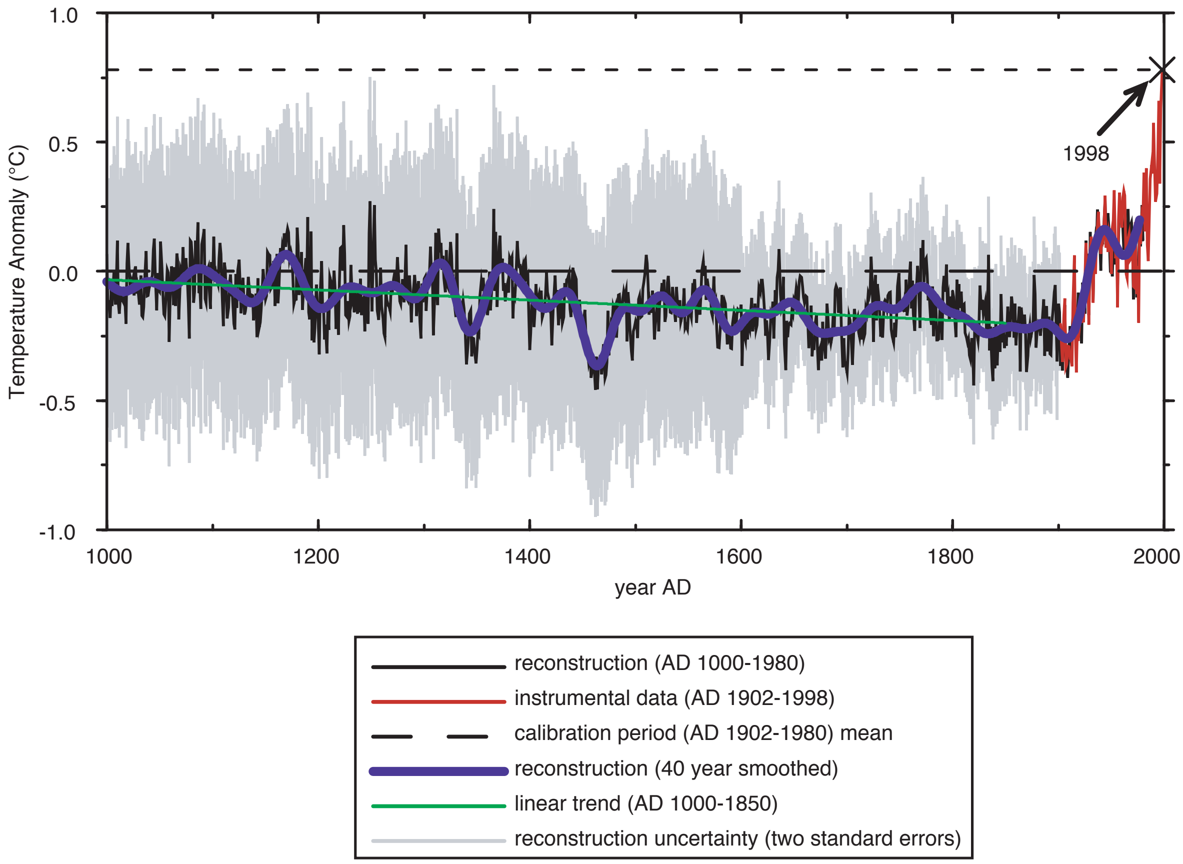

Fig. 6.4. Reconstructed northern hemisphere mean annual temperature with 2 standard error uncertainties (Mann et al. 1999).

hemisphere mean annual temperature for the last 1000 years. This is based on a network of well-distributed paleoclimatic records, the number of which decreases back in time. For the period since A.D. 1400, 397 records were used, but before A.D. 1400 this number drops to 14 (made up of 11 individual records, plus the first 3 principal components of tree ring width, representing a large set of trees in the western United States) (Mann et al. 1998, 1999). These paleoclimatic proxies were calibrated in terms of the main modes of temperature variations (eigenvectors) represented in the instrumental records for 1902-1980. Variations across the network of proxies, for the period before instrumental records, were then used to reconstruct how the main temperature patterns (i.e. their principal components) varied over time. By combining these patterns, regional and hemispheric mean temperature changes, as well as spatial patterns over time were reconstructed (Mann et al. 2000a). To accurately reproduce the *spatial* pattern requires that the proxy data network is extensive enough to capture several of the principal eigenvector patterns. With the data available, regional patterns of temperature variation could only be meaningfully reconstructed for 250 years, although the large-scale (hemispheric) mean temperature could be reconstructed for a longer period. This is possible because the proxy data network, even at its sparsest, exhibits a coherent response to temperature variability at the largest scale (Bradley and Jones 1993; Jones and Briffa 1998). Thus a reconstruction of hemispheric mean temperature back 1000 years is possible, using a quite limited network of data, albeit with ever-increasing uncertainty the further back in time one goes (Figure 6.4). This reconstruction shows an overall decline in temperature of ~0.2ºC from A.D. 1000 until the early 1900s (-0.02ºC/century) when temperatures rose sharply. Superimposed on this decline were periods of several decades in length when temperatures were warmer or colder than the overall trend. Mild episodes, lasting a few decades, occurred around the late 11th and mid-12th century and in the early and late 14th century, but there were no decades with mean temperatures comparable to those in the last half of the 20th century. Coldest conditions occurred in the 15th century, the late 17th century and in the entire 19th century. A critical question in any long-term reconstruction is: to what extent does the proxy adequately capture the true low-frequency nature of the climate record? Given that most of the long-term data used in all paleotemperature reconstructions are from tree rings, it is important to establish that the reconstructed temperature series are not affected by the manner in which biological growth trends in the trees are removed during data processing. This matter is especially critical when individual tree ring records, of differing record lengths (often limited to a few hundred years) are patched together to assess long-term climate changes. Briffa et al. (2001) have carefully evaluated this problem, using a maximum ring density data set that is largely independent of that used by Mann et al. (1998, 1999). By combining sets of tree ring density data grouped by the number of years since growth began in each tree, Briffa et al. provide a methodology that is designed to eliminate the biological growth function problem. They also estimate confidence limits through time (Figure 6.5).

The Briffa et al series shows similar temperature anomalies as Mann et al. in the 15th century (though no sharp decline in temperatures around A.D.1450) but markedly colder conditions from A.D. 1500 to ~A.D. 1800. The early 19th century is also colder in the Briffa et al. series. Their reconstruction thus describes a well-defined minimum in temperatures from ~A.D. 1550-1850 that conforms with the consensus view of a "Little Ice Age" (cf. Figures 6.4 and 6.5) (Bradley and Jones 1992 a,b). Though this period was not uniformly cold and temperature anomalies differed regionally, overall it was significantly below the 1881-1960 mean (by as much as 0.5 ºC for most of the 17th century) in the regions studied. Independent reconstructions derived from borehole temperatures suggest that ground surface temperatures were even colder 400 to 500 years ago (~1 ºC below levels in the 1980s) with temperatures subsequently rising at an increasingly rapid rate (Pollack et al. 1998; Huang et al. 2000). However, borehole-based temperature estimates have large geographical heterogeneity, resulting in a very small signal-to-noise ratio for mean hemispheric and global estimates. Indeed, a large number of borehole records do not capture the upward trend in 20th century air temperature in their respective regions (as discussed further below). Other attempts to assess northern hemisphere temperatures have taken a simpler approach than either Mann et al. (1998) or Briffa et al. (2001), by averaging together normalized paleo-data of various types (Bradley and Jones 1993; Jones et al. 1998) or by averaging data scaled to a similar range (Crowley and Lowery 2000). Such approaches do not provide an estimation of uncertainty, and indeed may lead to rather arbitrary combinations of very diverse data (often having different temporal precision). Nevertheless, the resulting time series from all of these studies are similar, at least for the first 400-500 years of the last millennium. Thereafter, some series indicate especially cold conditions, from the late 16th century until the 19th century, but these estimates are all bracketed by the two standard error confidence limits of Mann et al. (1998) and Briffa et al (2001) (Figure 6.6).

Although all of the reconstructions have much in common, they are clearly not identical. One explanation for the differences may lie in the geographical distribution of data used in each analysis. Each reconstruction represents a somewhat different spatial domain. In the Mann et al. studies, the "northern hemisphere mean" series is the same geographical domain as the gridded instrumental data set available for the calibration period (1902-80). This means that some regions within the northern hemisphere (in the central Pacific, central Eurasia and regions poleward of 70ºN) were not represented. However, because *global* eigenvector patterns were employed, the northern hemisphere mean is influenced by data from low latitudes and parts of the southern hemisphere. By contrast, the study of Briffa et al. is largely based on records from the northern treeline (60-75ºN) where temperatures were particularly low in the 17th century. As the other reconstructions generally do not include data from sub-tropical or tropical regions either, if higher latitudes were colder at that time compared to the Tropics, this may explain why the latter half of the millennium appears colder in those "hemisphere mean" series.

Another reason for the differences in Figure 6.6 may be because each reconstruction represents a somewhat different season. In the Mann et al. (1998, 1999) reconstruction, mean annual temperature data were used for calibration, since data from both hemispheres were used to constrain the eigenvector patterns and data from different regions may have had stronger signals in one season than in another. For example, some data from western Europe contain a strong NAO (winter) signal, whereas data from elsewhere carry a strong spring precipitation signal related to ENSO. Both data sets nevertheless help to define important modes of climate anomalies that themselves capture large-scale annual temperature patterns (Bradley et al. 2000). Other reconstructions are for boreal summer months (April-September), which may also explain some of the differences between the series. If summers were particularly cool in extra-tropical regions in the 17th-19th centuries, relative to temperature anomalies in low latitude regions (equatorward of ~30ºN) this implies an increase in the northern

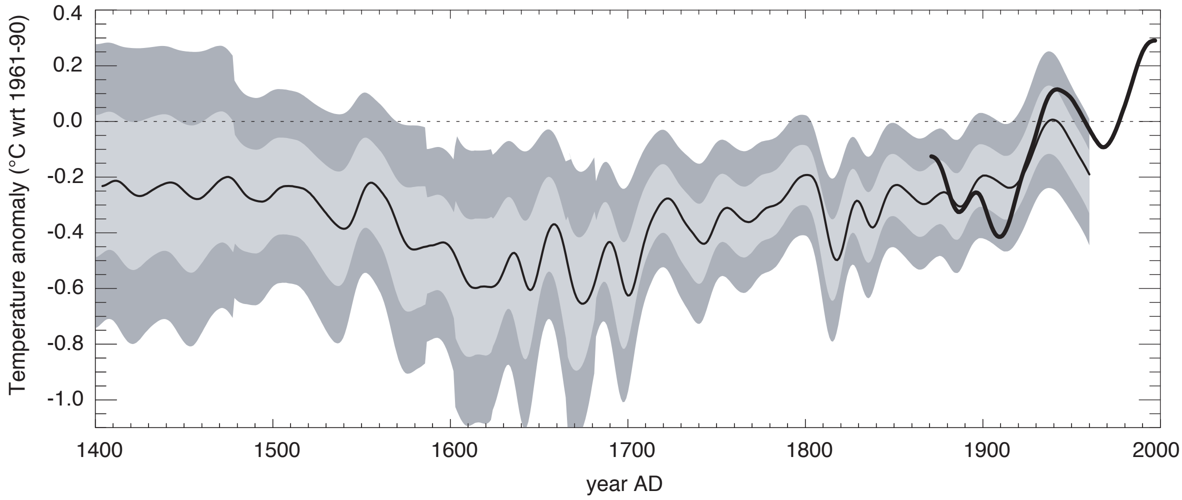

Fig. 6.5. Tree ring density reconstruction of warm-season (April to September) temperature from all land north of 20°N, with the ±1 and ±2 standard error ranges shaded. Units are °C anomalies with respect to the 1961-90 mean (dotted line) and the instrumental temperatures are shown by the thick line. Both series have been smoothed with a 30-year Gaussian-weighted filter (from Briffa et al. 2001).

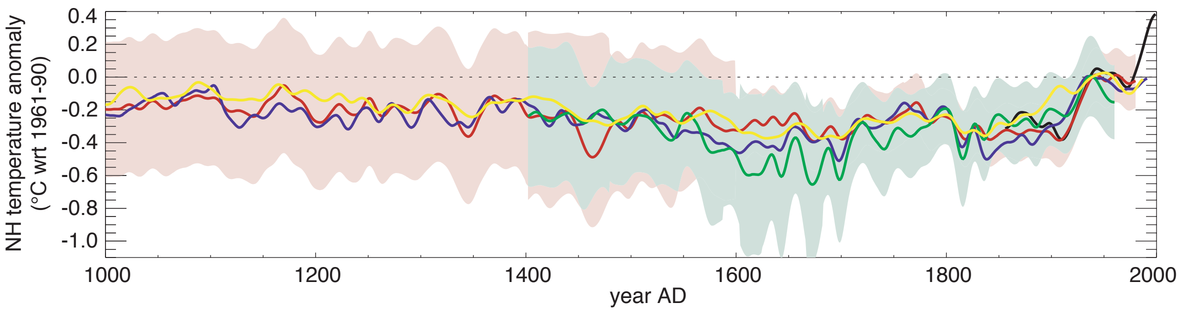

Fig. 6.6. Northern Hemisphere surface temperature anomalies (°C) referenced to the 1961–90 mean (dotted line). Annual-mean land and marine temperature from instrumental observations (black, 1856–1999), and as reconstructed by Mann et al. (red, 1000–1980, with ±2 standard errors shown by pink shading) and Crowley and Lowery (purple, 1000–1987). April-to-September mean temperature from land north of 20°N as reconstructed by Briffa et al. 2001 (green, 1402–1960, with ±2 standard errors shown by green shading), and by re-calibrating the Jones et al. estimate of summer northern hemisphere temperature (by simple linear regression) over the period 1881–1960 (yellow, 1000–1991). All series have been smoothed with a 30-year Gaussian-weighted filter.

hemisphere Equator-Pole temperature gradient during that period.

An independent assessment of temperature changes on the continents over the last few centuries is provided by a network of geothermal measurements in boreholes (Pollack et al. 1998; Huang et al. 2000). The depth profile of sub-surface temperature reflects a balance between heat loss from the surface, heat generated in the deep interior of the earth and the depth-dependent profile of heat diffusivity in the rock substrate. Changes in surface temperature propagate downward into sub-surface rocks, causing slight variations in the temperature depth profile. The depth to which such disturbances can be detected (by inversion of the sub-surface temperature profile) depends on the magnitude and duration of the surface temperature change, and variations of heat diffusivity in the rock, but whatever "signal" is transmitted from the surface, it is strongly attenuated with depth. This method thus provides a time-integrated perspective on paleotemperature with depth (Clow 1992; Beltrami and Mareschal 1995). It is not realistic to reconstruct a long-term annual, or even decadal temperature history from borehole data, but long-term trends or pre-instrumental mean temperatures can be assessed (Harris and Chapman, 2001). Figure 6.7 shows a comparison of century-long trends with data from co-located grid-boxes, derived by Mann et al. (1998). Here, (as in Huang et al. 2000) century-long

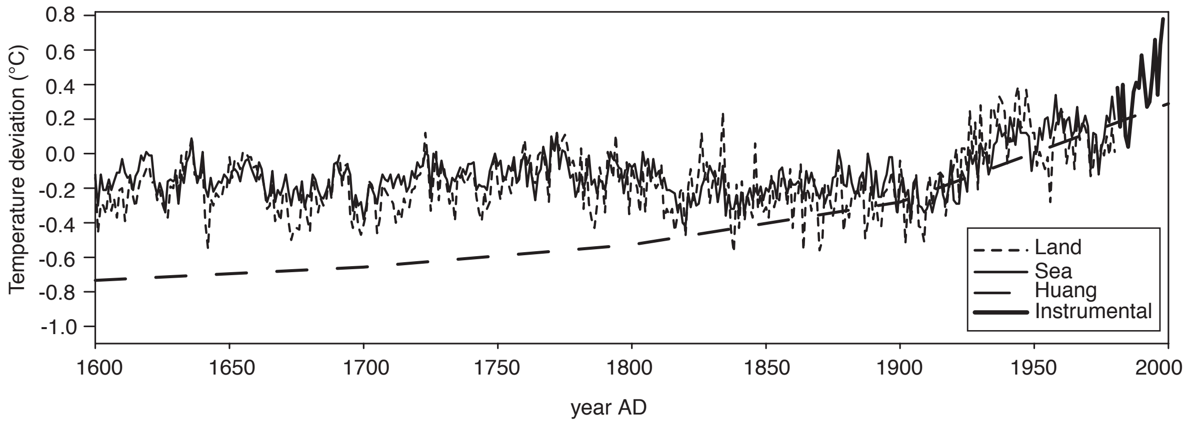

Fig. 6.7. A comparison of century-long ground temperature trends from boreholes with data from co-located grid-boxes, derived by Mann et al. (1998).

trends have been computed for each site, then averaged for all locations for each century and the means were concatenated back in time from the most recent 20th century trend. It is clear that the borehole data point to much greater cumulative warming since A.D. 1600 (and thus a much colder 17th century) than the multiproxy data set. However, borehole data are extremely noisy. The mean hemispheric signal of temperature change is not statistically significant between 1500 and 1900; only the change in the last 100 years emerges above the level of background noise (Mann et al. sub). Furthermore, a comparison of the data on a regional basis shows that whereas borehole data show continuous warming over the last 500 years in *all* areas, the Mann et al. data show overall negative (cooling) trends in the 16th century in Asia and North America, and cooling in all regions in the 19th century.

Why borehole temperatures increase at a rate greater than that indicated by proxy-based temperature reconstructions is not clear, but there are a number of possible reasons for the differences. Ground surface temperatures are not only affected by changes in air temperature (cf. Lewis 1998; Lewis and Wang 1998), but also by alteration of ground cover (e.g. due to land use changes), by changes in snow cover and by the amount of time it takes for near-surface soil moisture to become completely frozen in the winter (until all the moisture is frozen, further penetration of the winter cold wave into the ground is precluded). It is difficult to evaluate the importance of such effects on the diverse borehole data, but a study of data from northwestern North America suggests that ground surface temperatures significantly over-estimate air temperature changes in the 20th century (Skinner and Majorovicz 1999). Given that any major change in land use would likely affect both air and ground surface temperatures in the same way, it is hard to explain all these discrepancies. Nevertheless, it is clear that borehole records commonly do not match observed warming trends in their respective regions in the 20th century, which indicates that using them as a simple proxy for air temperature is problematical. A site-by-site evaluation of the quality of ground temperature data is needed, together with land use histories and snow cover changes, to try and resolve the matter. Once this has been done, combining the valuable low frequency characteristics of borehole data with the higher frequency attributes of annually resolved proxy data should yield better overall assessments of long-term temperature changes (e.g. Beltrami et al. 1995).

6.4 Uncertainties in large-scale temperature reconstructions

All large-scale paleotemperature reconstructions suffer from a lack of data at low latitudes. In fact, most "northern hemisphere" reconstructions do not include data from the southern half of the region (i.e. areas south of 30ºN). Furthermore, there are so few data sets from the southern hemisphere that it is not yet possible to reconstruct a meaningful "global" record of temperature variability beyond the period of instrumental records. For the northern hemisphere records, it must be recognized that the errors estimated for the reconstructions of Mann et al. (1999) and Briffa et al. (2001) are minimum estimates, based on the statistical uncertainties inherent in the methods used. These can be reduced by the use of additional data (with better spatial representation) that incorporate stronger temperature signals. However, there will always be additional uncertainties that relate to issues such as the constancy of the proxy-climate function over time, and the extent to which modern climate modes (i.e., those that occurred during the calibration interval) represent the full range of climate variability in the past. There is evidence that in recent decades some high latitude trees no longer capture low frequency variability as well as in earlier decades of the 20^{th} century (as discussed below in Section 6.8) which leads to concerns over the extent to which this may have also been true in the more distant past. If this was a problem (and currently we are not certain of that) it could result in an inaccurate representation of low frequency temperature changes in the past. Similarly, if former climates were characterised by modes of variability not seen in the calibration period, it is unlikely that the methods now in use would reconstruct those intervals accurately. It may be possible to constrain these uncertainties through a range of regional studies (for example, to examine modes of past variability) and by calibration over different time intervals, but not all uncertainty can be eliminated and so current margins of error must be considered as minimum estimates.

6.5 The Medieval Warm Epoch and the Little Ice Age

Bearing in mind concerns expressed earlier about the uncertainties inherent in paleotemperature reconstructions, what evidence is there for a "Medieval Warm Epoch" (MWE) and a "Little Ice Age" (LIA) during the last millennium?

The original argument for a MWE was made by Lamb (1965) based largely on evidence from western Europe. Much of the evidence he cited was anecdotal and he suggested that temperatures between A.D. 1000 and 1200 were about 1-2ºC "above present values" (probably meaning the 1931-60 average). In revisiting the concept of a MWE, Hughes and Diaz (1996) reviewed a wide range of paleoclimatic data, much of it reported since Lamb's classic work (Lamb 1965). They concluded that "*it is impossible at present to conclude from the evidence gathered here that there is anything more significant than the fact that in some areas of the globe, for some part of the year, relatively warm conditions may have prevailed*". Thus, they found no clear support for there having been a globally extensive warm epoch in the MWE, or indeed within a longer interval stretching from the 9^{th} to the early 15^{th} century. In fact, there is insufficient high resolution proxy evidence to be certain that *global or hemispheric* mean temperatures were higher during the MWE *sensu stricto* than in the 20^{th} century, as data from different regions do not agree on the matter. Consequently, we cannot entirely rule out the possibility of a globally extensive warm episode (or episodes) for at least part of the period from A.D. 1000 to 1200, because of the paucity of high-resolution records (especially from the oceans and the southern hemisphere) spanning that interval. It is interesting that Huang and Pollack (1997) find evidence (in a set of 6144 continental borehole heat flow measurements from around the world) that temperatures were in the range of 0.1-0.5ºC warmer than "present" (the early 1980s -- the mean date of borehole logging) ~700 to 800 years ago. However, more definitive estimates (of timing and amplitude) can not be made and given the large variability of these data, such estimates may not reach statistical significance compared to temperatures in the late 20th century. High-precision borehole temperature measurements at the GRIP (Summit) site on the Greenland Ice Sheet (where the signal-to-noise ratio is relatively large) point to conditions that were 0.5-1ºC above 1970 mean temperature at this one site around A.D. 1000 , but similar data from Law Dome, Antarctica show a temperature minimum at A.D. 1250, followed by warmer conditions in subsequent centuries (Dahl-Jensen et al. 1998, 1999).

High latitude tree ring data from some parts of the northern hemisphere also show evidence of temperatures in Medieval times well above the 20th century mean, at least in summer months (e.g. Briffa 2000), and so do some marine records from the North Atlantic, though the timing can not be precisely resolved (Keigwin 1996; Keigwin and Pickert 1999). There is also strong evidence from early European documentary records that winter temperatures in western Europe were quite mild during at least part of the period A.D. 750-1300 (Pfister et al. 1998). On the other hand, tree ring data from the southern hemisphere paint a different picture, with no clear evidence for a MWE, even when special care has been taken to conserve centennial scale variability (e.g. Villalba et al. 1996). Thus, whether there really were warm episodes of *global* extent in Medieval times and how these compare with late 20th century temperature levels (especially those in the last 20 years of the 20th century) remains an intriguing question that deserves further scrutiny. Until a more extensive set of data is available, the absence of evidence does not necessarily imply evidence of absence. Nevertheless, it must be stated that given the relatively limited evidence that does exist to support Lamb's original contention (Lamb 1965), the burden of proof must rest on demonstrating that his concept of a MWE has validity, rather than trying to show that it does not.

Perhaps of greater significance is that there definitely were significant precipitation anomalies during the period of the MWE; in particular, many areas experienced protracted drought episodes and these were far beyond the range of anything recorded within the period of instrumental records. For example, Stine (1994) describes compelling evidence that prolonged drought affected many parts of the western United States (especially eastern California and the western Great Basin) from (at least) A.D.910 to ~A.D.1110, and from (at least) A.D.1210 to ~A.D.1350. This led him to argue that a better term for the overall period was the "Medieval Climatic Anomaly" (MCA), which removes the emphasis on temperature as its defining characteristic (Stine 1998). The widespread nature of hydrological anomalies during the MCA suggests that changes in the frequency or persistence of certain circulation regimes may account for the unusual conditions during this period, and naturally this may have led to anomalous warmth in some (but not all) regions.

Numerous studies provide strong evidence that cooler conditions characterized the ensuing few centuries, and the term "Little Ice Age" is commonly applied to this period. Since there were regional variations to this cooling episode, it is difficult to define a universally applicable date for the "onset" and "end" of the period, but commonly ~A.D. 1550-1850 is used (Bradley and Jones 1992). However, there is evidence that cold episodes were experienced earlier, and glacier advances were common by the late 13th century in many alpine areas around the North Atlantic and in western Canada (Grove and Switsur 1994; Luckman 1994, 1996, 2000; Grove 2001 a,b). The definitional problem is illustrated in Figure 6.4 which shows temperatures gradually declining over the first half of the last millennium, rather than there being a sudden "onset" of a "LIA". If this reconstruction is accurate, it might explain why different mountain areas registered the onset of this neoglacial episode at different times. As temperatures cooled the threshold temperature for positive mass balance and glacier advances may have been reached in some areas sooner than others, leading to seemingly heterogeneous regional responses. However, by the late 16th century almost all regions had registered glacier advances, and these conditions generally persisted until the mid- to late 19th century making the term "Little Ice Age" ubiquitous and meaningful for this interval. Nevertheless, even within the period 1550-1850 there was a great deal of temperature variation both in time and space (Pfister 1992). Some areas were warm at times when others were cold and *vice versa*, and some seasons may have been relatively warm while other seasons in the same region were anomalously cold. But whatever date one selects for the "onset" of the LIA, there is little doubt that it was firmly at an end by the beginning of the 20th century. The reduction in ice masses accumulated over preceding centuries has continued to the present (in fact, the rate has accelerated) in almost all regions of the world (Dyurgerov and Meier 2000).

No doubt the complexity, or structure that we see in the climate of the LIA is a reflection of the (relative) wealth of information that paleoclimate archives have provided for this period. Having said that, when viewed over the long term this overall interval was undoubtedly one of the coldest in the entire Holocene. If we had similar data for the last 1000 years, our somewhat simplistic concepts of Medieval climatic conditions would certainly be revised and strong efforts are needed to produced a comprehensive paleoclimatic perspective on this time period. Only with such data will we be able to explain the likely causes for climate variations over the last millennium (see Sections 6.11 and 6.12).

6.6 20th century temperatures in perspective

One thing that all reconstructions shown in Figure 6.6 clearly agree on is that northern hemisphere mean temperature in the 20th century is unique, both in its overall average and in the rate of temperature increase. In particular the 1990s were exceptionally warm -- probably the warmest decade for at least 1000 years (even taking the estimated uncertainties of earlier years into account). A caveat to this conclusion is that the current proxy-based reconstructions do not extend to the end of the 20th century, but are patched on to the instrumental record of the last 2-3 decades. This is necessary because many paleo data sets were collected in the 1960s and 1970s, and have not been up-dated, so a direct proxy-based comparison of the 1990s with earlier periods is not yet possible. Nevertheless, confidence that an accurate reproduction of the recent instrumental record would be obtained if all the available paleoclimatic data were updated to the present is provided by Figure 6.8. This shows that a set of proxy data calibrated against the 1902-1980 period of instrumental data captured mean annual temperatures well both during this period and during the preceding 50 years for which an *independent* set of instrumental data is available. The excellent fit over the late 19th century test period provides confidence that an updated set of proxy data would also accurately reproduce recent changes. However, one cautionary note is needed: in the case of tree rings from some areas at high latitudes, the decadal timescale climatic relationships prevalent for most of this century appear to have changed in recent decades, possibly because increasing aridity &/or

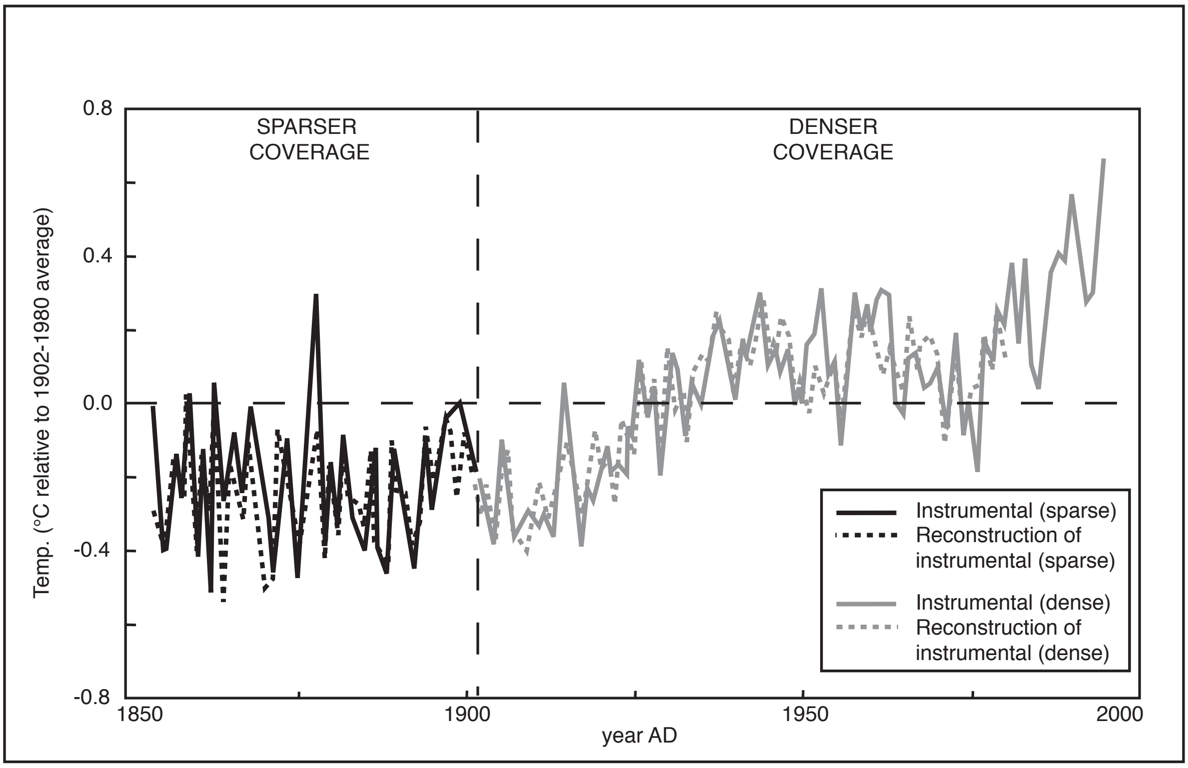

Fig. 6.8. Reconstructed mean annual temperatures for the northern hemisphere for the 19th and 20th centuries, from Mann et al. (2000) compared to the calibration data (1902-1980) and an independent period (1854-1901) for which instrumental data are available.

snowcover changes at high latitudes may have altered the ecological responses of trees to climate (cf. Jacoby and D'Arrigo 1995; Briffa et al. 1998). For example, near the northern tree limit in Siberia, this changing relationship can be accounted for by a century-long trend to greater winter snowfall. This has led to delayed snowmelt and thawing of the active layer in this region of extensive permafrost, resulting in later onset of the growing season (Vaganov et al. 1999). It is not yet known how widely this explanation might apply to the other regions where the partial decoupling has been observed, but regardless of the cause, it raises the question as to whether there might have been other periods in the past when the tree ring-climate response changed, and what impact such changes might have on paleotemperature reconstructions based largely on tree ring data.

In any case, the conclusion that temperatures in the 20th century rose at a rate that was unprecedented in the last millennium, reaching levels by the end of the century that had rarely, if ever, been exceeded in (at least) the preceding 900 years, seems to be an extremely robust result from all large-scale paleotemperature reconstructions, and it is confirmed by borehole heat flow data (cf. Pollack et al. 1998). The change in temperature has led to a major reduction in the mass of alpine glaciers in almost all parts of the world (Dyurgerov and Meier 2000; Thompson et al. 1993; Brecher and Thompson 1998), an increase in permafrost thawing at high latitudes (Osterkamp and Romanovsky 1999; Osterkamp et al. 2000) and at high altitudes (Jin et al. 2000; Isaksen et al. 2001), a reduction in the extent and thickness of Arctic sea-ice (Rothrock et al. 1999; Vinnikov et al. 1999; Wadhams and Davis 2001), later freeze-up and earlier break-up dates of ice on rivers and lakes (Magnuson et al. 2000), an increase in the calving rate of Antarctic ice shelves (Scambos et al. 2000), shifts in the distribution of plant and animal species, latitudinally and altitudinally (Grabherr et al. 1994; Pauli et al. 1996), changes in the phenology of plant leafing and flowering (Myneni et al. 1997) and the storage of significant quantities of heat in the near-surface ocean (Levitus et al. 2000) as well as an overall rise in sea-level driven by the melting of ice on the continents and a steric change, due to the increase in ocean temperature (Warrick and Oerlemans 1990). Thus, regardless of arguments over instrumental versus satellite-based estimates of warming in recent decades (National Research Council 2000) there are multiple indicators of warming in the 20th century that paint a vivid picture of the global-scale environmental consequences of the temperature increase.

Figure 6.6 shows that the overall range in temperature over the last 1000 years has been quite small. For example, the range in 50-year means has only been ~0.5ºC (from the coldest period in the 15th, 16th and 19th centuries, to the warmest period of the last 50 years). Within that narrow envelope of variability, all of the significant environmental changes associated with the onset and demise of the LIA took place. This puts into sharp perspective the magnitude of projected future changes resulting from greenhouse-gas increases and associated feedbacks (Figure 6.9). Even the low end of model estimates involving a scenario of minimal growth in future energy consumption suggests additional temperature increases on the order of 1-2ºC by the end of the 21st century (Intergovernmental Panel on Climate Change, 2001) which would be far beyond the range of temperatures experienced over (at least) the last 1000 years.

The discussion so far has focused exclusively on the northern hemisphere record because there are insufficient data currently available to produce a very reliable series for the entire southern hemisphere. Data from the Mann et al. (1998) reconstruction (back to A.D. 1700) averaged for those parts of the southern hemisphere that were represented in the instrumental calibration period show a similar temporal pattern to that of the northern hemisphere, but generally warmer (less negative anomalies). However, much more work is needed on southern hemisphere proxy records to extend and verify this result. We now turn to selected regional studies where much information has been obtained about major climate systems (modes of climate variability) that affect extensive areas of the world.

6.7 The Tropical Indo-Pacific

The past decade of climate dynamics research has highlighted the important role that tropical climate systems play in orchestrating modern interannual climate variability (see discussion in Chapter 3, Section 3.3). With that recognition has come a substantial effort to understand the variability of those systems on longer time scales. Much of this effort has focused on developing continuous annual reconstructions of sea surface temperatures, or rainfall-induced changes in salinity using coral records from the tropical oceans, along with terrestrial records from adjacent continents that are impacted by ocean-atmosphere variability. Although continuous coral records generally span only the past few centuries, they are providing an emerging picture of coherent decadal and longer-term variability in regions where instrumental records are limited to the past few decades. Terrestrial records suggest that teleconnections to ENSO evolve through time. Radiometrically dated fossil corals are revealing how ENSO responds to changes in background climate and forcing. Paleoclimate records indicate that ENSO is variable on many time scales, sensitive to certain background and forcing changes, and has a changing global signature. Century-scale records from the tropical Indo-Pacific support the global picture of a warming world; most coral records show a trend to unusually warm conditions in the late 20th century, with a few regional exceptions.

6.7.1 ENSO variability

Decadal variations in ENSO frequency, strength, and teleconnections are suggested by twentieth-century instrumental records (Trenberth and Shea 1987). In particular, instrumental records from the tropical Indo-Pacific show a recent shift towards warmer/wetter conditions in 1976. This shift appears as more “El Niño-like” conditions in the equatorial Pacific (Ebbesmeyer et al. 1991; Trenberth and Hurrell 1994; Graham et al. 1994) that set the stage for record El Niño events in 1982-3 and 1997-8. Unlike classic El Niño anomalies, the anomalies associated with this decadal shift reach into the mid-latitudes of both hemispheres (Zhang et al. 1997; Garreaud and Battisti 1999). In the tropical Indian Ocean, warming in 1976 (Terray 1995) coincides with a breakdown of previously established relationships between the Indian monsoon and ENSO and with changes in the relation of the monsoon to Indian Ocean SST anomaly fields (Kumar et al. 1999; Clark et al. 2000).

The change in ENSO in 1976 appears to be highly unusual, given the statistics of ENSO variability in the preceding 100 years of instrumental data. Time series (ARMA) modeling suggests that the 1976 shift should have a recurrence interval of about 2000 years, based on the pre-1976 variability in the Southern Oscillation and assuming statistical stationarity (Trenberth and Hoar 1996). A likely alternative to this "rare event" scenario is that the shift reflects a statistical non-stationarity introduced by a warming background climate (Trenberth and Hoar 1996). Other analyses have suggested, however, that the shift is somewhat less unusual (Rajagopalan et al. 1997) or even that it lacks statistical significance (Wunsch 1999). The ongoing debate over this issue argues strongly that longer records of ENSO are needed.

The 1976 shift appears as unprecedented in many ENSO-sensitive paleoclimate records. It occurs in

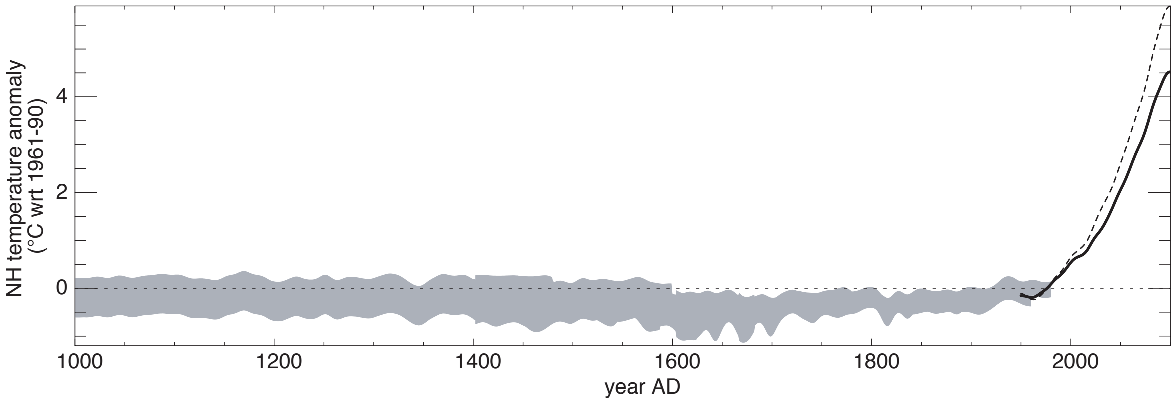

Fig. 6.9. Northern Hemisphere surface temperature (°C anomalies with respect to the 1961–90 mean) reconstructed by Mann et al. (1999) and Briffa et al. (2001), indicating the ±2 standard error ranges of the two reconstructions, compared to that simulated by the Hadley Centre's HadCM3 coupled climate model under increasing greenhouse gas and sulphate aerosol concentrations (SRES scenario A2) from 1950-2099. The dashed line is the model for latitudes higher than 20 degrees north. All data have been smoothed with a 30-year Gaussian-weighted filter.

all the ENSO-sensitive coral records, and in most of these, the post-1976 interval is the warmest/wettest period of the record. In the southwestern US, where El Niño events bring wetter and cooler conditions, the growth of high-elevation thousand-year-old trees shows an unprecedented increase beginning in the mid-1970s. This growth increase is attributed to a combination of increased cool-season precipitation and warmer growing season temperatures (Swetnam and Betancourt 1998). Global temperatures also reflect an upward shift in 1976 which, it is argued, indicates an intensification of the tropical hydrological cycle consistent with model simulations of a climate response to doubled CO_2 (Graham et al. 1994; Trenberth and Hoar 1996). Whether natural or anthropogenic, the mechanisms of this change have received intense scrutiny (Gu and Philander 1997; Zhang et al. 1998; Guilderson and Schrag 1998) and are the subject of ongoing analysis.

Many tropical Pacific records show a significant correlation with ENSO variability in their calibration intervals and can be used to assess how that system has changed through time. ENSO-sensitive records from the equatorial Pacific show clearly that the characteristic time scale of tropical Pacific variability changes over the past 1-4 centuries (Cole et al. 1993; Dunbar et al. 1994; Tudhope et al. 1995; Linsley et al. 2000a; Urban et al. 2000). In the Galapagos (1°N, 90°W), interannual and decadal modes of SST variability appear to change in strength simultaneously, implying a range of time scales for El Niño as well as linkages across these time scales (Dunbar et al. 1994). Shorter ENSO records from Tarawa (1°N, 173°E) and New Guinea (5°S, 146°) indicate shifts in variance among interannual periods that coincide with changes in the strength of the annual cycle (Cole et al. 1993; Tudhope et al. 1995).

At Maiana Atoll (1°N, 172°E), a record that reaches to AD 1840 indicates a long-term trend from cooler/drier to warmer/wetter conditions that is associated with changes in ENSO variance through time (Urban et al. 2000; Figure 3.8). When background conditions are cooler/drier (the mid and late 19th century), variability occurs on a more decadal time scale, in contrast to the dominantly interannual variance of the 20th century. The decadal variance in this record offers a point of comparison for decadal variability in other tropical records; the Maiana record is coherent on decadal time scales with coral records from the equatorial western Indian Ocean and with other ENSO-sensitive records from the Pacific and on adjacent continents (Figure 6.10; Cole 2000). Other ENSO reconstructions that span this interval also note a weakening of interannual variance (Stahle et al. 1998) and stronger decadal variance (Mann et al. 2000b). This widespread signal is thus not a consequence of a changing spatial domain of ENSO impacting a single coral site; it clearly reflects a real change in the time-domain behavior of the system.

Land-based paleoclimate records from North America offer additional clues to past variations in the Pacific and its impacts. Stahle et al. (1998) developed a 272-year Southern Oscillation Index (SOI) reconstruction based on drought-sensitive tree ring chronologies from the southwestern US and northern Mexico. Their reconstruction indicates stronger interannual variance and a tendency for more cool events in the 20th century compared to earlier periods. A network of drought reconstructions over the continental US shows that, although the ENSO-drought link in the southwest is

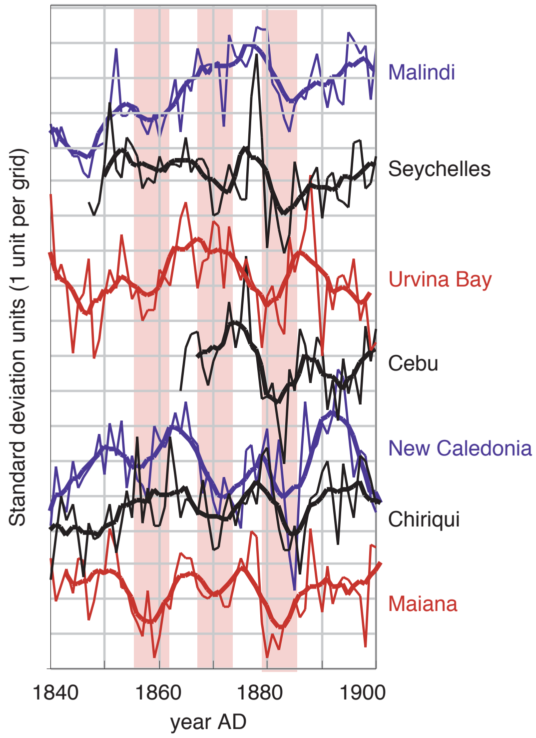

Fig. 6.10. Common patterns of decadal variability in tropical Indo-Pacific coral records during the 19th century. The records shown here all exhibit cool-dry events in the late 1850s, ~1870, and early 1880s (shaded bars). Small age offsets may be real, or may reflect age-model uncertainties. The top three and the bottom record correlate closely with ENSO in their calibration periods, but the remaining records are somewhat removed from ENSO centers of action or have competing climate influences on their $\delta^{18}O$, so do not correlate as strongly to interannual ENSO changes. The fact that they all reflect the decadal variance of the late 19th century suggests similarities with the 1976 shift, whose extent is latitudinally broader than typical ENSO variability.

relatively robust, drought in other regions shows a more variable connection to tropical Pacific variability (Cole and Cook 1998). In the early 20th century, El Niño events bring wetter conditions to the mid-Atlantic region, but by the mid-20th century, El Niño conditions are associated with drought. Cole and Cook (1998) attribute this change to the intensification of decadal variability in the North Pacific, with a downstream influence on mid-Atlantic regional moisture balance. Moore et al. (2001) suggest a similar reversal in the correlation of decadal anomalies in Mt Logan glacial accumulation and tropical Pacific variability.

6.7.2 Century-scale trends in the tropics

Looking beyond ENSO variability, coral isotope records from the tropical Indo-Pacific suggest a consistent pattern of declining $\delta^{18}O$ over the past few centuries, reflecting increasingly warm/wet conditions (Figure 6.11). The rate of change observed between 1895-1989, for those records that span this interval, ranges from 0.3-2.1°C/century. The strongest recent trends are seen at off-equatorial sites (New Caledonia and the Philippines) and in precipitation-sensitive records where salinity changes likely make up a significant part of the coral isotopic trend (Nauru and Maiana). In two instances, the trend over this interval does not indicate warming. At Vanuatu, a strong warming trend that predates 1895 supports the general picture of a relatively warm 20th-century, despite the lack of such a trend between 1895-1989. A slight cooling at Galapagos is consistent with instrumental data from this region and may reflect an ocean dynamical response to greenhouse forcing, whereby warming in the western Pacific strengthens the trade winds, which enhances upwelling of cool water in the easternmost Pacific (Cane et al. 1997). Just as air temperature records from individual sites do not each reflect the global average air temperature change, we do not expect worldwide consistency in oceanic records, but the general pattern supports tropical warming in the 20th century. These records place in context the global analysis of instrumental ocean temperatures (Levitus et al. 2001; Levitus et al. 2000) that shows an unequivocal accumulation of heat in the world ocean since 1955.

Records from tropical high-elevation ice caps provide some of the strongest evidence for tropical warming. Tropical glaciers worldwide are receding rapidly and many have already disappeared (Thompson et al. 1993, 2001; IPCC 2001). Isotopic records place these recent changes in a longer term context (Figure 6.12). Diaz and Graham (1996) have linked the recent ice losses to a systematic warming of the tropical troposphere driven by increasing tropical SST; time series of freezing level indicate a shift upwards in 1976, consistent with the shift in tropical SST (Diaz and Graham 1996; Vuille and Bradley 2000; Gaffen et al. 2000).

6.7.3 Tropical variability in the last millennium

Records that span the past millennium with high resolution are rare in the tropics, but three recently published datasets suggest common patterns at multidecadal-century time scales (Figure 6.13). The depth of Crescent Island Crater Lake, part of Lake Naivasha in Kenya, has fluctuated dramatically over the past millennium, with prolonged wet and dry phases coeval with cultural records of prosperity and hardship, respectively (Verschuren et al. 2000).

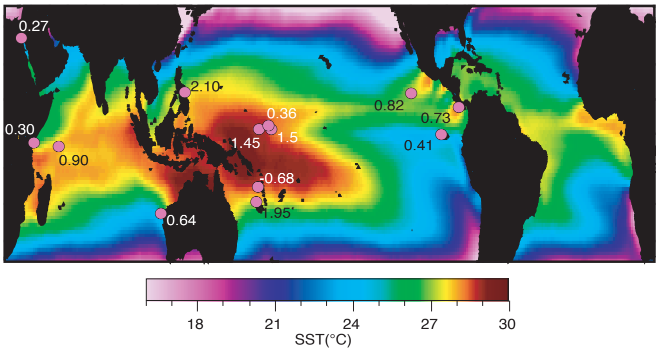

Fig. 6.11. Sites where annual coral isotope records span the interval 1895-1990. Numbers indicate the inferred SST trend in °C per 100 years, assuming all isotopic variability is due solely to SST changes and the slope of the SST-$\delta^{18}O$ relationship is 0.22°C per 1‰. Site sensitivity issues (e.g. depth of coral, influence of rainfall) have not been taken into account in these calculations. Large central Pacific values are almost certainly due to the influence of rainfall on seawater isotopic content. Background colors indicate mean SST field.

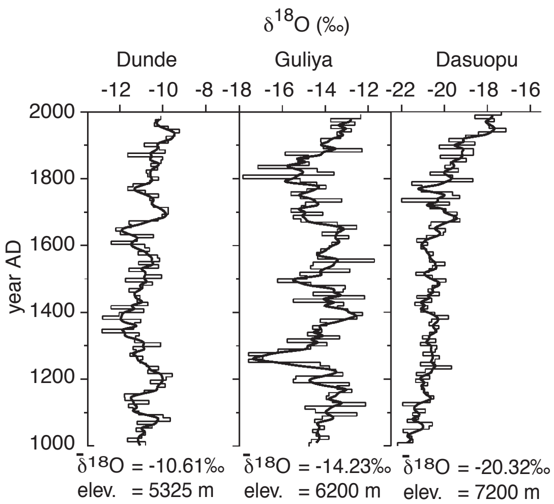

Fig. 6.12. Ice core records of recent isotopic changes in Tibet (upper panel) and South America (lower panel). Unusually enriched values in recent decades at many locations may reflect exceptionally warm conditions.

On the other side of the globe, the strength of the trade winds at the Cariaco Basin (northwestern tropical Atlantic) has waxed and waned at the same time scale; recent decadal variability in this record correlates with both North Atlantic and eastern Pacific SST (Black et al. 1999). Lake sediment records from the Yucatan peninsula (Hodell et al. 2001) and inferred changes in riverine sediment input to the Cariaco Basin (Haug et al. 2001) also indicate marked century-scale oscillations in hydrologic balance. All of these studies point to solar variability as a possible cause for the multidecadal changes (Figure 6.13). High solar radiation corresponds to periods of Naivasha lowstand, weak Atlantic trades, and a drier Yucatan peninsula. The hydrologic responses are opposite those seen in an early Holocene speleothem record from Oman (Neff et al. 2001), which indicates monsoon intensification during high irradiance periods.

Model simulations of temperature change associated with solar forcing suggest some simple mechanisms that may help to explain these observations (Cubasch et al. 1997; Rind et al. 1999). Increased irradiance warms surface air temperature, and the correlation between solar forcing and temperature change is highest in the tropics and subtropics. However, the amplitude of the forced change is greatest at high latitudes, implying a reduced equator-pole temperature gradient when solar irradiance is high. Therefore, during solar irradiance maxima, weaker North Atlantic trades may result from the reduced latitudinal gradient, and the intensified monsoon in Oman could be driven by the enhanced land-ocean temperature gradient. Drier conditions in the Yucatan and Kenya during solar maxima may

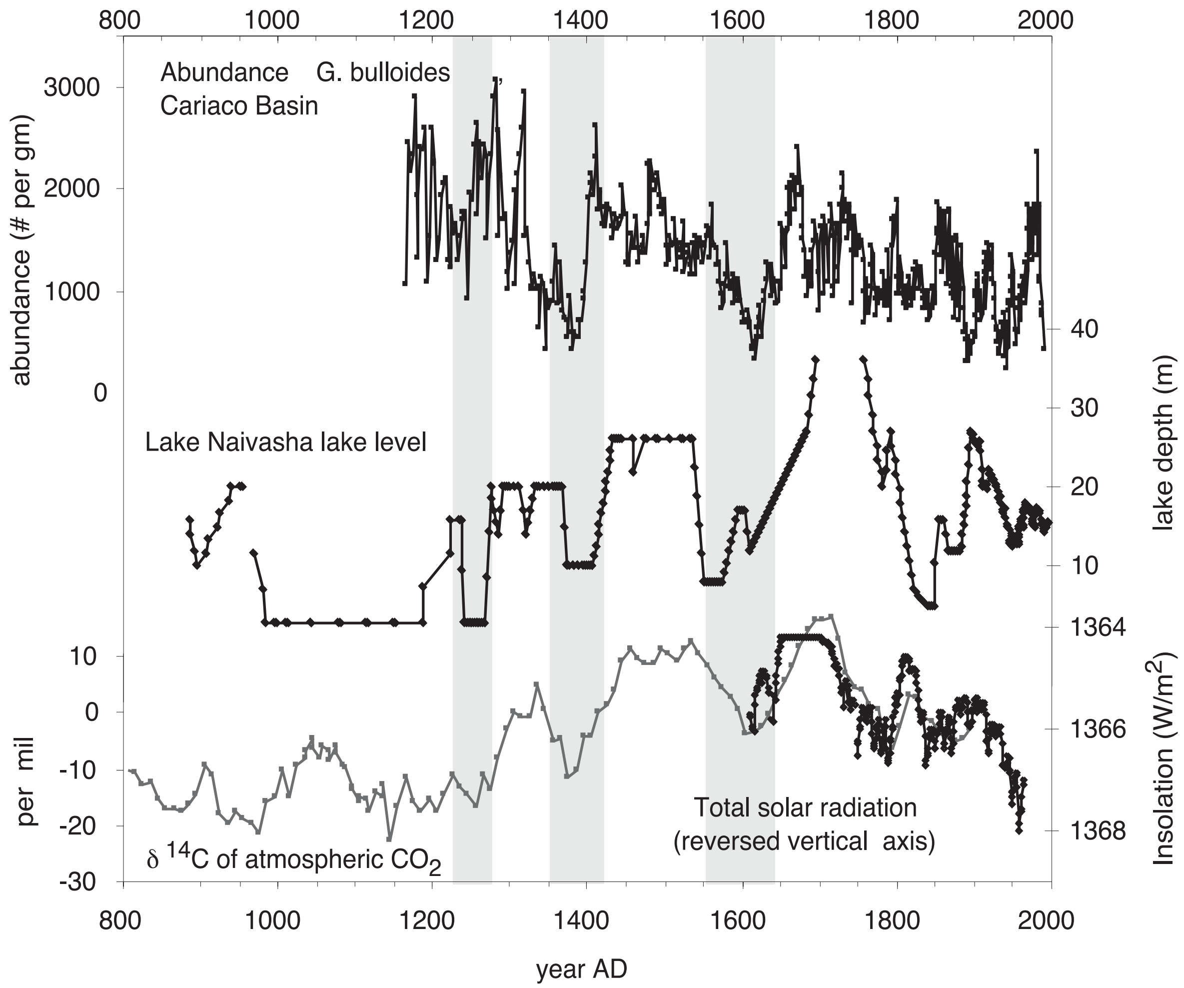

Fig. 6.13. Comparison of records of North Atlantic trade-wind strength (inferred from *G. bulloides* abundance at the Cariaco Basin; Black et al. 1999), Lake Naivasha level (inferred from sedimentological indicators; Verschuren et al. 2000), and solar radiation (inferred from the $\Delta\delta^{14}C$ of atmospheric CO_2 (Stuiver and Reimer 1993) and for the past 400 years from a reconstruction by Lean et al. (1995). Several of the multidecadal changes in these records are coincident (highlighted by grey bars), suggesting the possibility of a common response to radiative forcing on this time scale.

have resulted from the combination of a direct evaporative response to warming and potentially reduced moisture transports from weakened trade winds (both lie to the west of warm ocean basins under easterly trades). Additional well-dated, high-resolution records are needed to confirm and define the geographic dimensions of these relationships, and will undoubtedly add complexity to the preliminary interpretations we present here.

6.8 Hydroclimatic variability in western North America

There is an extraordinary wealth of natural archives of the last millennium's climate variability in western North America, including abundant ancient trees growing under climatic stress, geomorphological features associated with glacier activity and lake level changes, lake and laminated marine sediments. This permits an unusual degree of cross-checking between completely independent natural archives. The annual resolution and century to millennial length of many of these records allows them to yield information on hydroclimatic fluctuations on all time scales from interannual to century and, in a few cases, millennial.

6.8.1 Tree ring networks

Fritts and his colleagues (La Marche and Fritts 1971; Fritts et al 1971; Fritts 1991) used a subcontinental network of tree ring width series to produce annual maps of seasonal climate features such as temperature, precipitation and sea-level pressure back to AD 1602. Their strategy was based on the predominance of moisture limitation as a limiting factor to plant growth in this mainly semi-arid region. These reconstructions then provided the basis for a detailed climatological study of the North American impacts of the El Niño-Southern Oscillation since AD 1602 (Lough and Fritts 1985), as well as of explosive volcanic eruptions (Lough and Fritts 1987), and an early cross-checking of natural archives and historical documents (Fritts et al. 1980). Other work, designed to reconstruct river flow on similar time spans (and focused even more strongly on moisture- sensitive trees), showed the early 20th century to have been anomalously wet on a 400-year time scale in the Upper Basin of the Colorado River (Stockton and Jacoby 1976), and provided rare evidence for a statistical association between the area of drought in the center of the continent and solar and lunar influences (Mitchell et al. 1979).

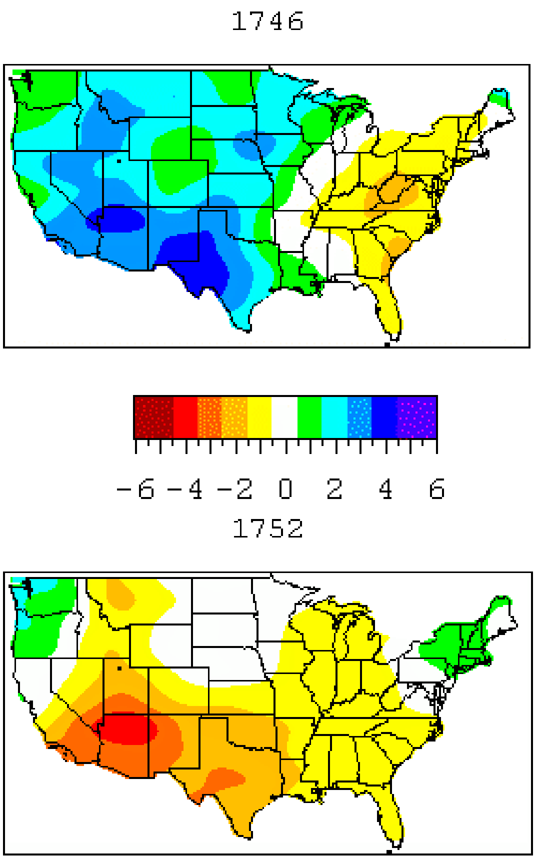

Fig. 6.14. Summer Palmer drought severity index (PDSI) as reconstructed from a continental network of drought-sensitive tree ring width records (Cook et al. 1999). PDSI less than zero represents dry conditions. A.D. 1746 and 1752 were El Niño and La Niña years, respectively, as reconstructed by Stahle et al. (1998). These maps show that summer soil moisture conditions resembled those associated with the same phases of the ENSO fluctuation in the instrumental period.

A network of moisture-sensitive tree rings covering the whole of the lower 48 states of the USA (Meko et al. 1993, Cook et al. 1999) (Figure 6.14) has recently been used for a rigorous test of these findings (Cook et al. 1997). These last authors conclude that although they have not produced physical proof of a solar and/or lunar forcing of drought in the western United States, "the statistical evidence for these forcings appears to be strong enough to justify the continued search for a physical model….". They point out that, alternatively, the close-to-bidecadal drought area rhythm might be caused by unstable ocean- atmosphere interactions.

6.8.2 Interaction between time scales

In work focusing more narrowly on the region between the Rocky Mountains and the Pacific Ocean, from southern Canada to northern Mexico, Dettinger et al. (1998) used a dense network of moisture-sensitive tree ring records to show that the patterns of spatio-temporal variability in precipitation seen in the instrumental record are also evident in the previous two hundred years. They focused on variability in the 'ENSO' timescale, that is 3-7 years, and the 'interdecadal timescale', that is, more than 7 years. Similar patterns of intensity and distribution of precipitation were seen in the 18th and 19th centuries as in the 20th, on both time scales. They did, however, identify multi-decadal variations in regional precipitation that were present in the last 150 years, and before A.D. 1700, but not in the eighteenth and early nineteenth centuries (Dettinger et al. 1998). This feature is also seen in a reconstruction of precipitation in the Great Basin based on six very well replicated lower forest border bristlecone pine chronologies (Hughes and Funkhouser 1998), where fluctuations with a period of around 60 to 80 years appear clearly from AD 200 to 600, from 1300 to 1600, and then after 1850 (Figure 6.15). Minobe (1997) also detected similar oscillations over the past 400 years in tree ring reconstructions from western North America.

What might be the causes of such changes? It is of interest to note that the multi-decadal changes in this reconstruction in the late 19th and 20th centuries appear to correspond to major shifts in the Pacific Decadal Oscillation (PDO) identified by Mantua et al. (1997). The reconstruction of the PDO back to AD 1661 by Biondi et al. (2001), based on tree rings from southern California and northwestern Mexico, shows the same multi-decadal features since the late 19th century as the instrumental record (Figure 6.16). However, as in the work of Dettinger

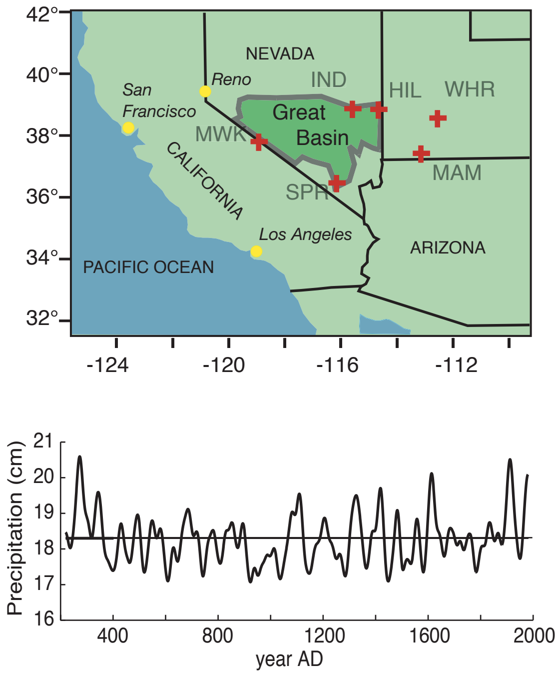

Fig. 6.15. Nevada Division 3 precipitation (July-June) reconstructed from a network of lower forest border stripbark bristlecone pine (after Hughes and Funkhouser 1998). The series has been smoothed with a 50-yr gaussian filter. 1 standard deviation unit equals 4.4cm, mean = 18.3cm. Map shows the location of tree ring sites (red + signs) and of Nevada Division 3 (green line).

et al. (1998), these are absent before the late 19th century. The first 200 years of the Biondi et al PDO reconstruction are marked by a strong bi-decadal fluctuation, uncorrelated with solar radiative forcing. This starts after the circa-1600 transition in frequency domain behavior seen in Figure 6.15. D'Arrigo et al. (subm), using tree ring width and density data from around the Gulf of Alaska, the northwestern conterminous U.S. and southwestern U.S. and northern Mexico to reconstruct the PDO since AD 1700, showed a different pattern. Their reconstruction showed a marked decadal mode of variation, that weakened in the mid-1800s, as also noted by Geladov and Smith (2001) and Villalba et al. (2001). These last authors saw this not only in tree ring chronologies in Northwestern North America but also in Southern South America. Villalba et al. (2001) also point out that the mid-1800s shift from a predominantly decadal to interannual pattern of variability means that "what we know of tropical Pacific variability through the analysis of instrumental records has been based largely on a period of predominant interannual variability". They suggest that there may have been a return to the pre-1850 decadal variability in the last two decades of the 20th century. The case for a strong role for conditions in and over the equatorial Pacific in controlling hydroclimatic variability in western North America is further strengthened by the existence of decadal and multidecadal oscillatory modes common to the tree ring records from both hemispheres. The network of chronologies used by Villalba et al. (2001), which includes, *inter alia*, moisture sensitive records in North and South America, shares the same pattern of Pacific decadal sea surface variability over the past two centuries (Evans et al., 2001) as Sr /Ca ratios from coral at Raratonga (Linsley et al. 2000). This supports the proposition that "Pacific decadal variability is a basin-wide phenomenon originating in the tropics" (Evans et al. 2001). The predictability of the impacts of the ENSO phenomenon in extra-tropical America depends on the phase of the longer-term PDO variation (Gershunov and Barnett 1998), and thus great practical benefit would flow from a better understanding of Pacific decadal variability. Since its frequency domain behavior is not yet well known, it is likely that a more complete definition of the spectrum of variability will depend on natural archives such as those discussed here.

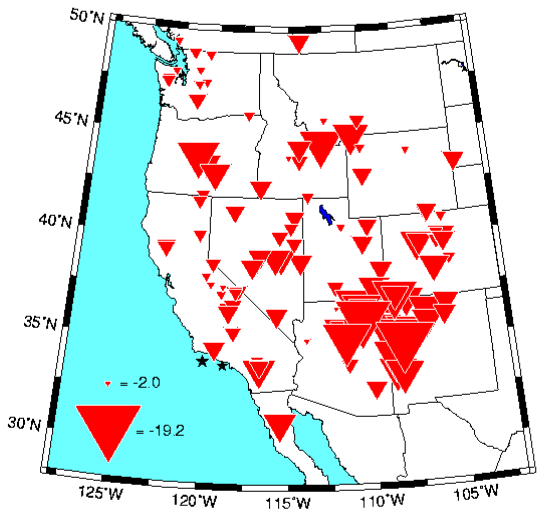

Fig. 6.16. Cumulative severity of A.D. 1561-1600 growth reduction in moisture-limited trees (from Biondi et al. 2000). The location of each symbol indicates the location of a tree ring width index chronology. These are expressed as dimensionless indexes with a mean of 1.0. The size of each symbol is proportional to the sum of all departures for the chronology over the period. The two symbols in the lower left of the map indicate the range of values on the map as percentages. Growth is reduced throughout this region in comparison to the long-term mean.

6.8.3 Extreme and persistent droughts and wet periods

Stahle et al. (2000) pointed out the extraordinary intensity and extent of drought in the middle and late 16th century, from central Mexico to northern Canada and as far east as the Atlantic coast. Grissino-Mayer (1996) identified the sustained drought at this time to be the most severe of the last 2000 years in the Four Corners region of the southwestern U.S. Biondi et al. (1997, 2000) identified major changes at the end of the 16th century in patterns of deposition of laminated sediments in the Santa Barbara Basin, off the central California coast, which are consistent with tree ring records that show a late 16th century dry period was followed by a series of unusually wet decades in the early 17th century (Figure 6.17). The post-1600 wet decades may well be the cause of the lacustrine event at 390 ±90 yr BP indicated by geomorphic data collected in the Mojave River Drainage Basin, southern California (Enzel et al. 1989). This probably coincides with a large flood in AD 1605 ±5 yr that is recorded as a silt layer in the sediments of the Santa Barbara Basin (Schimmelmann et al. 1998).

Giant sequoia tree rings from the western flanks of the Sierra Nevada show variations in the frequency of extreme single-year droughts over the last 2100 years (Hughes and Brown 1992; Brown et al. 1992; Hughes et al. 1996). The incidence of such droughts on a century time scale has varied more than threefold, with highest frequencies in the 3rd and 4th, 8th and 9th, and 15th and 16th centuries. The 20th century frequency was slightly below the 2100 yr mean. Not only tree rings show that there was a greater tendency for droughts to be intense and persistent between A.D. 400 and A.D. 1600 (Figure 6.15). Stine (1994) identified extreme low stands in Mono Lake, a closed basin on the California/Nevada border, that lasted from the early 10th century to the end of the 11th and from the beginning of the 13th to the middle of the 14th. These coincide with droughts seen in the rings of moisture-sensitive bristlecone pine from the neighboring White Mountains (La Marche 1974, Hughes and Graumlich, 1996) and in the Sierra Nevada (Graumlich 1993; Graybill and Funkhouser 1999). Similarly, the rate of change of $\delta^{18}O$ in sediments of Pyramid Lake (250 km north of Mono Lake) corresponds closely to a tree ring based reconstruction of streamflow in the mountains from which the lake is fed (Benson et al. 1999). These two records are, however, sometimes out of phase with those at Mono Lake., indicating a northward shift of storm tracks that resulted in periods of unusually high precipitation in the northern area at the same time as extended droughts prevailed around Mono Lake. Finally, there is evidence for a greater incidence of sustained, severe drought before approximately AD 1600 than since. Records that also support this observation are derived from sediments in Owens Lake, California (Li et al. in press), reconstructions of salinity in lakes on the northern Great Plains (Fritz et al. 2000; Laird et al. 1996), and sand-dune fields (Muhs et al. 1997).

In short, the paleoclimate record provides unequivocal evidence for droughts that far exceed anything in the 20th century, in terms of magnitude and persistence (Woodhouse and Overpeck 1998). A multidecadal drought today would have enormous economic and social impact, even in a well-off nation such as the U.S. The mechanisms by which these droughts are initiated and persist are not known, and therefore prediction is not yet possible. Intensified paleoclimate data collection, in conjunction with model experiments and modern process research, will be needed to document and understand this phenomenon.

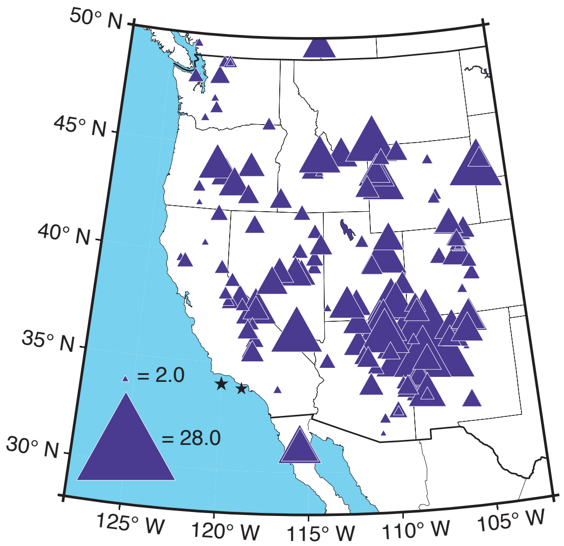

Fig. 6.17. Cumulative excess of A.D. 1601-1640 tree growth in moisture-limited trees (from Biondi et al. 2000). As Figure 6.16, except that growth is enhanced in comparison to the long-term mean.

6.8.4 Ecosystem impacts of climate variability

One unexpected by-product of the development of subcontinental networks of tree ring records has been the discovery of vastly accelerated growth rates in the last two or three decades of the 20th century of trees near timberline in the eastern parts of the American Southwest (Grissino-Mayer 1996; Swetnam and Betancourt 1998). These trees have been growing more rapidly recently than at any time in the last 1000 years, apparently as a result of an unprecedented period of springtime warmth coinciding with the wetter phase of interdecadal variability. The consequences of this acceleration on ecosystems are unknown.

Another complex set of climate-ecosystem interactions has been unraveled. Swetnam (1993) shows that long-term changes in fire frequency, revealed by dendrochronologically-dated fire scars in giant sequoia trees, may be related to century- scale temperature change, presumably through temperature effects on the composition and/or productivity of the forest (Figure 6.18a). There are also strong inter-annual fluctuations in fire frequency superimposed on this long-term variability. These are closely linked to precipitation and drought, which determine the quantity of fine fuels and the moisture content of the fuel in a particular fire season (Figure 6.18b). It would not have been possible to unravel these interactions without the development of high-resolution natural archives of climate and vegetation response. Such multi century histories of fire also have potential to be used themselves as natural archives of past climate (see Chapter 5).

6.9 North Atlantic region

The circum-North Atlantic region has yielded a unique combination of long instrumental climate observations, many documentary records and multiple sources of terrestrial and marine palaeoclimate information. A selection of these data is shown in Figure 6.19, together with two series that represent aspects of atmospheric and oceanic circulation variability. This figure provides a good perspective on the different types of evidence from which we can attempt to distill the spirit of how climate has varied across this region during the last 1000 years.

The Central England Temperature series (CET) spans the last 340 years and is the longest, continuous, homogeneous instrumental record in the world (Manley 1974; Parker et al. 1992; Jones and Hulme 1997). Mean annual temperature values are shown in Figure 6.19a. These are representative of temperature variability over much of western Europe (Jones et al. 2002). The CET can now be compared with a composite temperature record for the Benelux countries recently assembled from direct meteorological observations (for the period after 1700), weather diaries and journals, various account records (e.g. agricultural or river tolls), letters and early annals and chronicles (van Engelen et al. 2001). These data are plotted as a 'pseudo' annual series in Figure 6.19b, which is the average of the separate summer and winter index values that are virtually complete from 1200 onwards.

The CET and Benelux records provide mutual confirmation of the unusual warmth of the 20th century, and particularly of the last 20 years, when viewed against the variability of the last 350 years. However, this warmth is most anomalous in the winter season, even when compared with evidence of unusual winter warmth in the Low Countries during the 11th century and again for a short period around 1200 (note that the earlier, more sparse data and the separate seasonal series are not shown here – but see van Engelen et al. 2001). The Benelux mean annual record and the early CET series clearly exhibit prolonged relative cold through much of the late 16th and the whole of the 17th centuries. This corresponds to what Lamb (1963) described as the 'pessimum period', or the worst part of the Little Ice Age. The early 1400s and early 1500s, the mid 1600s and most of the 18th century were periods of relative warmth. Figure 6.19c provides what is currently the best indication of the year-by-year variability of the winter North Atlantic Oscillation (NAO; see Chapter 3, Section 4.3) over the last 500 years. This was reconstructed on the basis of statistically defined associations between the NAO and a large number of predictors: mainly long surface pressure, temperature and precipitation records distributed over much of western Eurasia (Luterbacher et al. 2001, 2002). The reconstructed index is low during much of the late 16th and 17th centuries. It is high during the mid 19th and early 20th centuries, and in the early 18th and early 16th centuries, coinciding with distinct periods of warmth in England and the Benelux countries. While the NAO index is high also in recent decades, it has not reached the unprecedentedly high levels evident in the temperature records. The early decades of the 1400s were particularly warm in summer while in the mid 1700s the warmth was more in winter.

A long tree-ring-based reconstruction of individual summer temperatures for northwestern Sweden (Figure 6.19d – Briffa et al. 1992) again shows the 20th century to be generally warm, but no more so than the 11th, late 12th, the 15th and the early 16th centuries. The cold of the 17th century is striking in these data. It directly matches a period of high precipitation shown in a record of bog surface-wetness in western Ireland (Figure 6.19e), and one that is representative of general moisture trends over a wider area of the British Isles (Barber et al. 2000).

Moving to the other side of the Atlantic, Figure 6.19, graphs i-m illustrate a group of ice-core-derived proxies in Greenland and Eastern Canada: oxygen isotope ratios (Fisher et al. 1996) and the extent of snow melt that re-froze in the firn (Koerner and Fisher 1990; Fisher et al. 1995). Over the

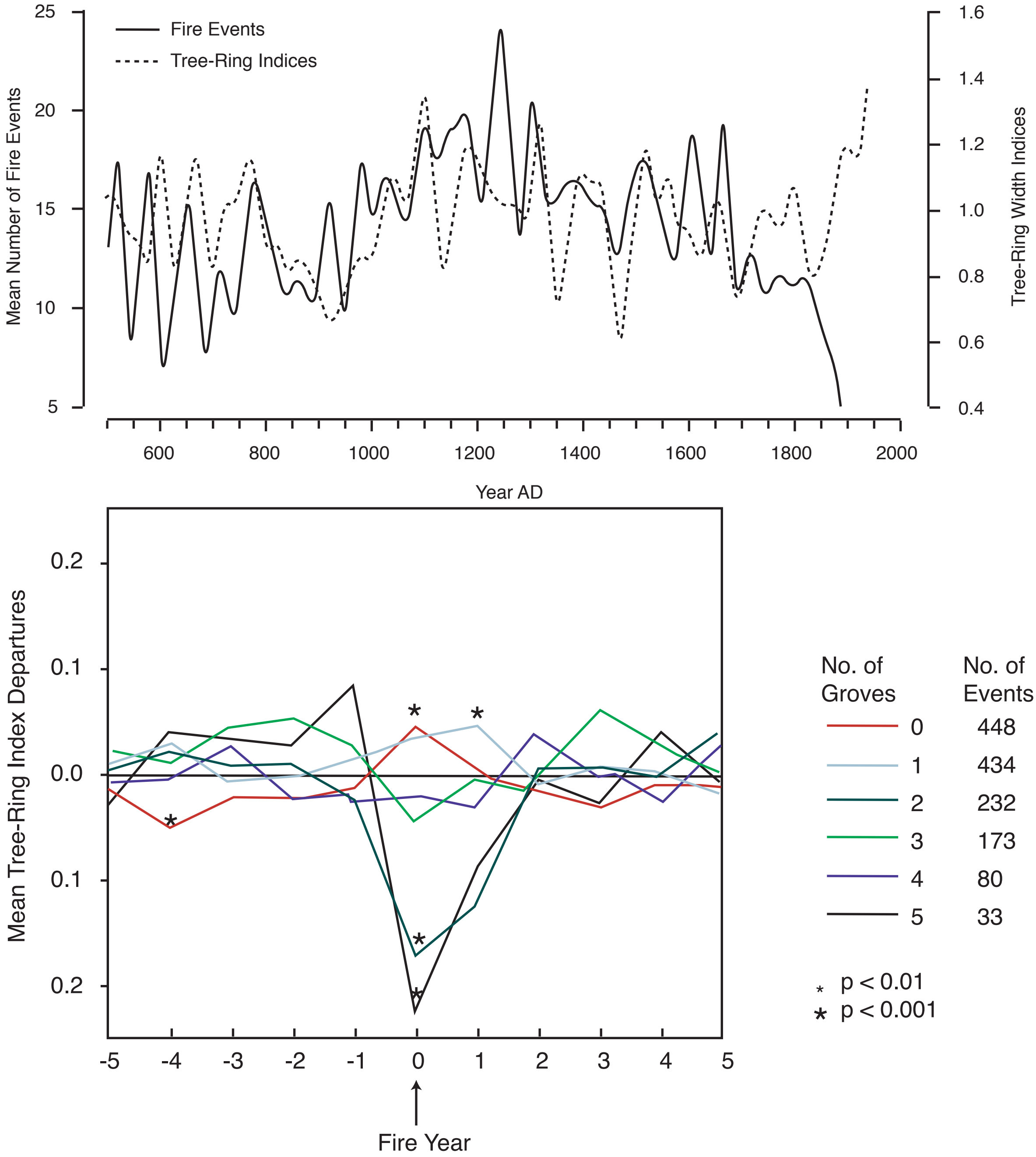

Fig. 6.18. A. 20-year running mean of fire events in five giant sequoia groves compared with tree ring indices of temperature-responsive bristlecone pine from near upper tree limit in the nearby White Mountains (from Swetnam 1993) **B.** Departures of tree ring width index (from A.D. 500-1850 mean) of precipitation-responsive bristlecone pine from near lowest forest border in the White Mountains, California. These are used as an index of regional drought; they are unaffected by fire in the nearby Sierra Nevada. They are shown for 5 years before and after fire in 0 to 5 giant sequoia groves in the Sierra Nevada. Small asterisks p<0.01, large asterisks p<0.001. The most extensive fires are clearly associated with drought years recorded by tree rings (from Swetnam 1993).

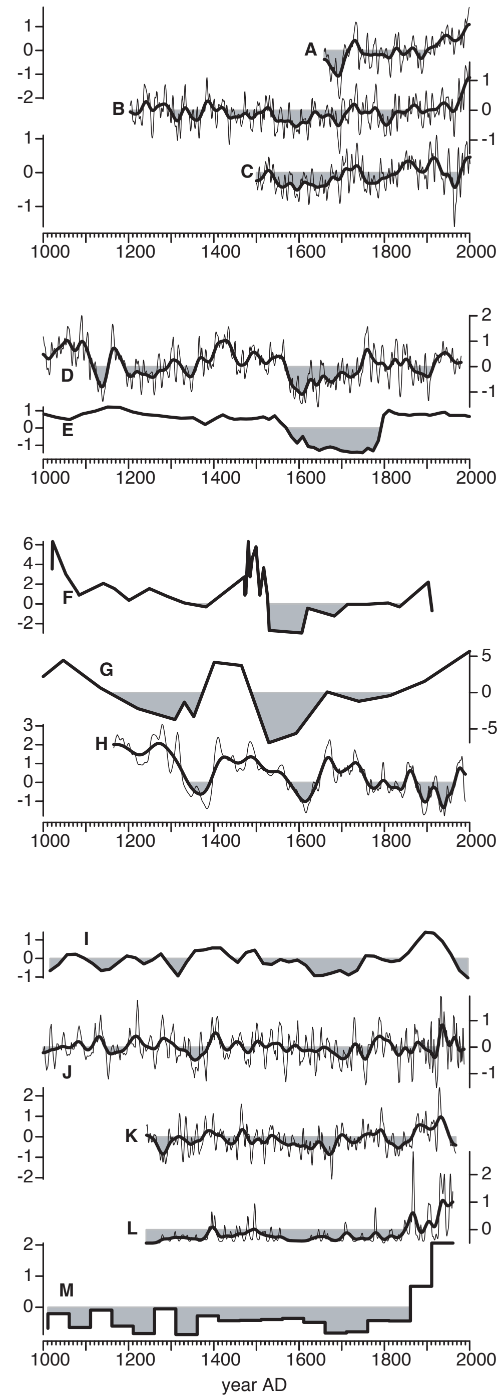

Fig. 6.19. Selected climate-related records for the last millennium around the region of the northern North Atlantic. All of the series are plotted as effective 10-year (thin line) and 50-year (thick line) smoothed and standardized values (with reference to the common base period 1659-1999). **A.** Central England mean annual temperatures (Manley, 1974) **B.** Pseudo annual temperatures for the Benelux countries (from van Engelen et al. 2001); **C**. Reconstructed winter North Atlantic Oscillation indices (Luterbacher et al. 2001); **D**. Warm season (A-S) temperatures reconstructed from tree-ring data (Briffa et al. 1992); **E**. Moisture index based on bog flora (Barber et al. 2000); **F**. SST from *Foraminifera* oxygen isotope composition (Keigwin 1996); **G**. An index of the speed of deep current flow; low values indicate a reduction in NADW production (Bianchi and McCave 1999); **H**. Foraminiferal abundance in the Cariaco Basin, indicative of trade wind intensity and possible changes in temperature in the North Atlantic (Black et al. 1999); **I**. Oxygen isotope data from the North Grip site ice core (Johnsen et al. 2001); **J**. Ice-core-derived oxygen isotopes from the GISP2 and GRIP sites (Johnsen et al. 2001); **K**. Several west Greenland ice-core oxygen isotope series (Fisher et al. 1996); **L**. High-resolution melt-layer data in an ice core from the Agassiz Ice Cap (Koerner and Fisher 1990; Fisher et al. 1995); **M**. A lower-resolution eastern Canadian ice-core melt record.

last millennium, these have a resolution and dating accuracy that is virtually annual, but individual core records provide indications of changing temperatures that are more reliable at decadal than yearly timescales. Appropriate averaging of data from different cores can reduce local non-climate 'noise' (Fisher and Koerner. 1994) and give a stronger signal of local climate (White et al. 1997). The ice melt data provide a more direct indication of (summer) temperature than the isotopic data which have a complex association with regional temperatures, and also reflect changes in atmospheric circulation and shifts in the seasonality of snowfall over the core sites (Dansgaard et al. 1973).

A number of similarities can be discerned by comparing averaged or 'stacked' records from various west and central Greenland sites (Figure 6.19 k and j) (Fisher et al. 1996; Grootes et al. 1993) with a new single isotope record from northern Greenland (Figure 6.19i) (Johnsen et al. 2001) and melt data from the Agassiz ice cap in eastern Canada (Figure 6.19 l) (Koerner and Fisher 1990). There is general evidence of anomalous warmth in the last 100 years. There are indications of stronger and more recent warming in the melt data (the last 50 years) than in the isotope data; the latter show relative cooling in recent decades, as has been observed in Greenland instrumental records, especially in summer (Briffa and Jones 1993). The apparent magnitude of the early 20th century warmth is much greater in the northern Greenland and Canadian records than in the central or west Greenland data.

None of these records indicate decadal or multidecadal changes over the last 1000 years that exceed the mean temperature of the last century. However the data show that there was widespread warmth across eastern Canada and all of Greenland in the late 13th century and again in the late 19th and early 15th centuries. This accords with the warmth indicated by the west European documentary data and the Scandinavian tree-ring record. The middle 14th century was also cool on both sides of the Atlantic (see also Ogilvie and Farmer 1997). However, while the west Greenland evidence (Figure 6.19k) indicates cold conditions through most of the late 16th and the 17th centuries, in phase with western European evidence for the coldest part of the Little Ice Age, the central (and more northern) Greenland data show the cold phase continuing into the 18th century.

In trying to deduce the effect of the ocean on the patterns of temperature changes seen around the North Atlantic we are hampered by the lack of marine paleorecords that can be interpreted with the fineness of resolution that matches the terrestrial data over the last millennium. Figure 6.19g is a record of sediment grain size, a proxy for velocity of bottom-water flow, that can be used to infer the intensity of North Atlantic Deep Water (NADW) flow between Iceland and Scotland near 300 m water depth (Bianchi and McCave 1999). Although of low resolution, this record suggests that there was reduced NADW formation during the 13th century and in the 16th and perhaps early 17th centuries and enhanced production in the 11th and 15th centuries. These data also seem to indicate a strong trend towards enhanced NADW production over the last 200 years but the recent data are very sparse (see also Hansen et al. 2001).

The grain-size record can also be compared with a reconstructed low-resolution (~50 yr) sea-surface temperature history for the Bermuda Rise, at ~34°N 58°W in the Sargasso Sea (Figure 6.19f). This is based on statistically-calibrated counts of the abundance of temperature-sensitive *Foraminifera* in a sediment core (Keigwin 1996) and provides evidence of strong warmth at the beginning of the 11th century, and again in the 15th century, with cooler conditions in the 16th century. In these respects this sea-surface temperature history is similar to the inferred deepwater production record from the south of Iceland, despite the low resolution of both records, and it also shares the major characteristics of temperature change that are evident in the century-timescale features of the west European terrestrial data.

Keigwin and Pickert (1999) have suggested that sustained temperature anomalies such as the cold of the 18th century in Europe and the Sargasso sea region may have been times when the NAO remained in a predominantly negative phase (or the frequency of extreme negative years increased). Figure 6.20 illustrates the spatial association between the NAO and temperature and precipitation variability over the North Atlantic and adjacent land areas in the different warm and cold seasons. These patterns show that northeast Canada and southwest Greenland experience both colder and drier conditions, and northwest Europe is warmer and wetter, when the NAO index is very high, that is when the general westerly flow across the Atlantic is enhanced. At such times, winters are generally warmer and wetter in the southwest USA and over the western North Atlantic and noticeably drier on the Iberian peninsula and in the western Mediterranean.

In summer, though, associations are far less coherent. In high NAO years, there is no anomalous cold over eastern Canada and Greenland and the warm region over northern Europe covers much less of the area south of Scandinavia. Cold conditions over the Mediterranean are virtually absent, and the

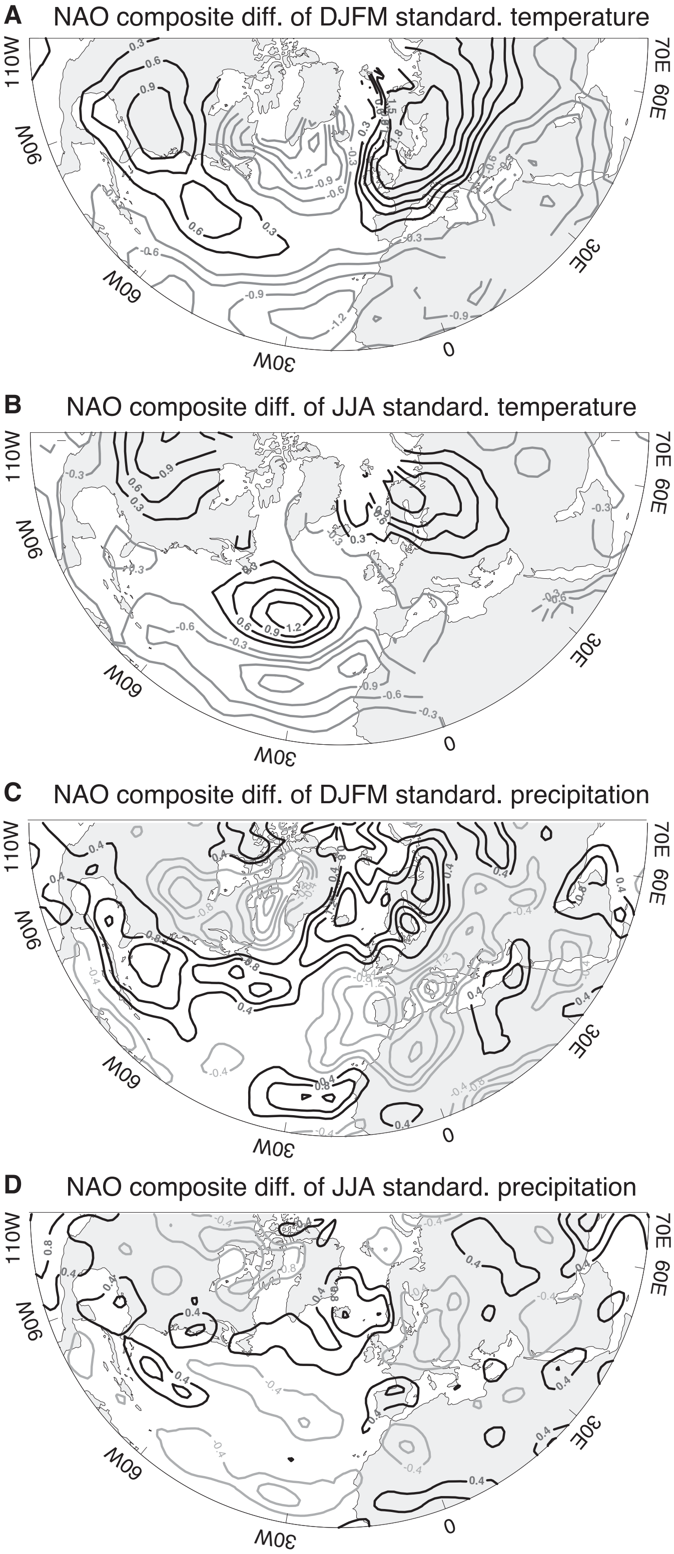

Fig. 6.20. Differences between composites of (**A,B**) temperature or (**C,D**) precipitation from (**A,C**) winters or (**B,D**) summers with positive North Atlantic Oscillation Index (NAOI) and those with negative NAOI. Seasonal temperature and precipitation were standardised to have zero mean and unit variance at each location prior to analysis. Thicker lines indicate positive anomalies.

warm band across the eastern United States and western Atlantic becomes two regions separated by an area with near normal temperatures. Moreover, although the pattern of precipitation associations in summer is similar to that in winter, the magnitudes of the correlations are so low as to be barely perceptible. This implies that the temperature evidence to support an NAO influence on major climate excursions would be found primarily in data that represent winter conditions, with variability that is out of phase on the two sides of the northern Atlantic.

The NAO index series (Figure 6.19c) shows a pronounced low phase in the late 1500s and 1600s that is coincident with cold in the Benelux record and in Fennoscandia (in summer), as well as with the sediment grain-size data and low Sargasso Sea surface-water temperatures.

The last record we shall discuss in this section is shown in Figure 6.19h. This shows high-resolution measurements of foraminiferal abundance in the anoxic, laminated sediments of the Cariaco Basin, off northeastern Venezuela (Black et al. 1999). These variations are interpreted as representing changes in upwelling intensity, which can be used as a proxy for trade wind strength. Stronger trade winds are associated with a southward shift in the ITCZ, which reduces precipitation in the southern Caribbean and enhances it in tropical South America. However, in analyses of recent instrumental data, it has been shown that stronger trades are also linked to cooler conditions in the North Atlantic, at least on an interannual timescale and probably (because of the absence of any lag) through an atmospheric rather than oceanic link. On the multi-decadal timescale it is hard to recognize any consistent associations between the proxy upwelling record and any of the northern temperature proxies. There is a generally positive correlation with the eastern records, and a negative correlation with the Greenland data in the post-1700 period, but these associations appears to be reversed before that date.

6.10 The Southern Hemisphere

Most of the evidence for climate variability in the southern hemisphere over the last millennium comes from tree rings, supplemented by a few ice cores and speleothems, as well as coral records that have already been discussed.

The major dendroclimatic resources of the Southern Hemisphere are found in southern South America, Tasmania and New Zealand. Sub-Saharan Africa has, so far, yielded only a few, relatively short, dendroclimatic records (Dunwiddie and LaMarche 1980; Stahle and Cleaveland 1997). LaMarche and Pittock (1982) first demonstrated the potential of Tasmanian trees as climate records. Huon pine (*Lagarostrobus franklinii*) from a subalpine site in Tasmania has yielded a reconstruction of warm-season temperatures since 1600 BC (Cook et al. 1999b). As might be expected on an island exposed to such a vast expanse of ocean, the range of variability is smaller than in comparable Northern Hemisphere records. Although notable cool and warm periods of decade to century length were recorded during the first two millennia of this reconstruction, the last 1000 years were characterized by variability on shorter timescales. However, the last few decades of the 20th century have been marked by abrupt and major warming that is unique in the last millennium (Cook et al. 1992). The overall reconstruction contains consistent decadal to multi-decadal oscillations, which seem to be primarily associated with sea surface temperatures in a region extending west from Tasmania to South Africa, and in a region of the west Pacific north of the Equator. Extensive dendroclimatological work has been done in New Zealand (for example D'Arrigo et al. 1998; Norton and Palmer 1992; Ogden and Ahmed 1989; Salinger et al. 1994) and although there are species of great longevity, notably kauri (*Agathis australis*), millennial dendroclimatic reconstructions are still awaited.

Many dendroclimatic records, both temperature- and moisture- sensitive, have been developed for Argentina and Chile (Boninsegna 1992; Villalba et al. 1992). They include a 3620-year reconstruction of summer temperature from *Fitzroya cupressoides* at Lenca (Lara and Villalba 1993), which also shows reduced variability compared to Northern Hemisphere reconstructions. In this case it is clear that the reduced variability is in part the result of the standardization methods used to remove non-climatic variation from the ring-width series. A new temperature reconstruction, based on 17 millennia-length chronologies specially constructed to conserve low-frequency variability, shares many features with the Lenca series, but differs in showing a sustained period of cool summers between AD 1500 and 1650 (Lara et al. 1999). The long South American temperature reconstructions show decadal fluctuations rather similar to those in Tasmania (Villalba et al., 1996), but they do not show the sudden warming of recent decades. So far, however, the main connection that has been established between the South American and Australasian chronologies has been through reconstruction of a summer Trans-Polar Index (TPI) back to the mid-18th century, based on the pressure gradient between Hobart, Tasmania and Port Stanley (Falkland Islands/ Islas Malvinas) (Villalba et al. 1997). The anti-phase relationship between pressure at these two locations, or at least its interannual and high-frequency variability, was reconstructed as being stronger in the 20^{th} than in the 19^{th} century, and statistically associated with ENSO on 4-5 year time scales.

The establishment of relatively dense dendroclimatic networks in both North and South America has made possible a number of inter-hemispheric comparisons. Boninsegna and Hughes (2001), for example, showed a much clearer impact of explosive volcanic eruptions on temperatures in North America than South America, perhaps related to the "damping' effect of the extensive oceans of the Southern Hemisphere on such rapid changes. Elsewhere in this chapter we refer to the remarkable common patterns of decadal and multi-decadal oscillations between tree-ring chronologies from Northwestern North America and southern South America (Villalba et al. 2001), which raise fascinating questions about the mechanisms of inter-hemispheric climate linkage. These paleoclimatic records reveal robust teleconnections that can only be glimpsed using instrumental records, yet they have clear relevance to climate variability on time scales of social and economic interest.

Stalagmites from northeastern South Africa (Holmgren et al. 1999; 2001) provide a unique high resolution paleoenvironmental record from an area where few other records exist (cf. Tyson and Lindesay, 1992). Variations in oxygen and carbon isotopes, and in coloration (reflecting humic acids washed into the deposit) all suggest that conditions were relatively cool and dry from ~A.D. 1500-1800. Warmer conditions prevailed from A.D. 900-1100 and ~A.D. 1400-1500. It has been estimated that the temperatures fell rather abruptly, by ~4ºC, from the warmest to the coldest part of the last millennium (i.e., from ~1500 to the 1600s), with temperatures from A.D. 1500-1800 approximately 1ºC below the mean for 1961-90 (Holmgren et al., 2001).

Ice core data from Peru (Quelccaya Ice Cap) clearly indicate a period of low oxygen isotope values (i.e. snow depleted in the heavier isotope, $\delta^{18}O$) from ~A.D. 1520-1880, interpreted as reflecting colder conditions (Thompson et al. 1985; Thompson 1992). It was relatively wet from A.D. 1520 until ~1720 (the wettest period in the interval A.D. 1450-1980), but dry thereafter, from ~1720-1860. Ice core records from Antarctica are quite variable (perhaps not surprisingly, given the area involved and the limited number of records) but generally lowest isotopic values were from the early 1600s until the early 1800s (Mosley-Thompson 1992; Peel 1992). At South Pole, the lowest values were in the 16th century, whereas at Law Dome, nearer the coast in East Antarctica, the lowest values of the last 2000 years were in the late 1700s/early 1800s. Generally, these records fit with the far less resolved picture of glacier fluctuations in South America and the sub-Antarctic region, which show glacier advances during these colder centuries, with recession over the last ~150 years (Clapperton and Sugden 1988). By contrast, several ice cores from the Antarctic Peninsula show lowest oxygen isotope values in the 20th century, following a decline since the mid-19th century, in contrast to observed instrumental temperature observations which suggest warming over this interval (Peel 1992). This highlights the difficulties of using a simple isotopic/temperature interpretation in some locations where shifting moisture source regions, the presence of open water etc, make the climatic signal quite complex.

Overall, the limited data from the southern hemisphere suggest that conditions were generally colder during the interval A.D. 1500-1850, with different regions experiencing minimum temperatures at different times within this period. Not surprisingly in this highly buffered oceanic region, temperature changes appear to have been small, and in some cases changes in hydrological conditions related to shifts in atmospheric circulation may have been more significant than slight changes in temperature. Such changes are registered not only in terrestrial proxies, but also in the marine environment. For example, corals from the southwest Pacific indicate more saline conditions from ~A.D. 1700-1900, as a result of a stronger Hadley circulation and higher evaporation rates in the region (Hendy et al. 2002). This may be a reflection of stronger latitudinal temperature gradients during this interval of time.

6.11 Forcing factors: causes of temperature change in the last millennium

Temperature reconstructions for the last millennium (e.g. Figure 6.4) reveal three characteristic timescales of variability: a long-term cooling trend, from A.D. 1000 to A.D. 1900 on which are superimposed multi-decadal and shorter-term (multi-year) anomalies. Here we consider some possible factors that may have been responsible for the observed changes, and also place those forcing factors in the context of the longer-term (Holocene) climate variations discussed earlier.

6.11.1 Orbital forcing

The pervasive influence of orbital variations (changes in precession, obliquity and eccentricity) undoubtedly played a dominant role in Holocene (and longer-term) climate variability (Figure 6.21). Overall insolation in the northern hemisphere summer season has declined over the last 10,000 years, by ~8% (at the top of the atmosphere) with a small increase in winter values. The effect of such insolation changes under the clear-sky conditions that commonly occur at high elevations, at high latitudes in the early summer (with 24 hour illumination) may explain why the melt record in Greenland and Ellesmere Island ice cores registers such markedly higher values in the early Holocene. The higher summer insolation values may also be responsible for enhanced monsoonal rainfall in continental Africa, though this effect has not yet been convincingly reproduced by GCMs with insolation forcing alone; surface water and soil moisture/vegetation feedbacks seem necessary to explain the magnitude of rainfall changes indicated by paleoclimatic data (e.g. Coe and Bonan 1997; Broström et al. 1998). Higher summer temperatures recorded by many proxy records from the early Holocene are a consequence of these higher radiation receipts. Even though the forcing mechanism for the higher temperatures was unrelated to greenhouse gases, there may be useful analogs in the environmental consequences of this warming that can illuminate how conditions may change in the near future (at least in areas relatively untouched by human activities) as a result of anthropogenic greenhouse gas increases.

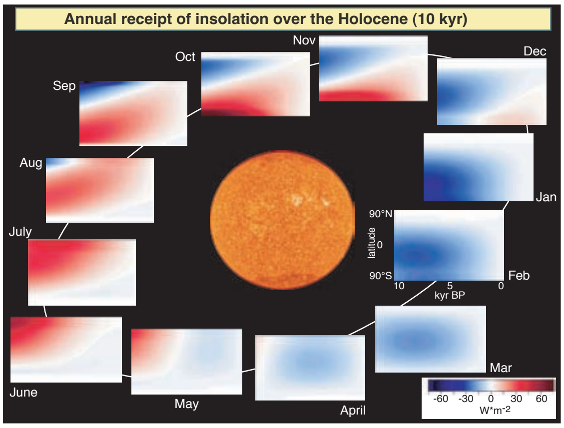

Fig. 6.21. Holocene orbital insolation anomalies (outside the atmosphere) versus month of year at 1000 year intervals. The key to each panel is illustrated in the February case.

On the timescale of the last millennium, orbital forcing is minor compared to the Holocene, but there has been a decline in July insolation of ~5 Wm^{-2} within the Arctic circle over that time (Figure 6.22). In the Inter-Tropical zone, from January-June, insolation increased over the course of the millennium. Overall, the July insolation gradient from 30-60°N increased by ~3% over this interval (which favors a slight equatorward displacement of the sub-tropical high pressure centers).

Orbital effects on insolation receipts were far less significant in the southern hemisphere, with the main features being an overall decline in insolation from August-November (especially in October) and a small increase in summer months (December-March).

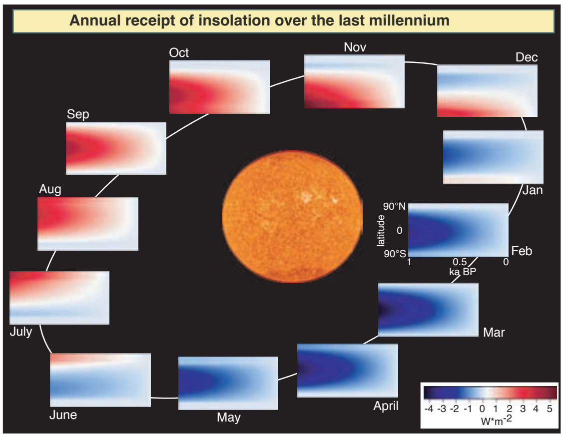

Fig. 6.22. Insolation anomalies (outside the atmosphere) for the past millennium, by latitude and month. The key to each panel is illustrated in the February case

6.11.2 Solar variability

Direct measurements of solar irradiance changes (TSI) from satellites only span the last two solar cycles but these data indicate a variation of ~ 0.1% in TSI over an 11 year solar cycle. Using long-term observations of the sunspot cycle, together with inferences based on the variability of sun-like stars, Lean et al (1992) estimated a change in TSI from the late 17th century/early 18th century solar minimum (Maunder Minimum) to the late 20th century of ~0.24%. To extend such a record further back in time requires the assumption that solar activity variations that affect cosmogenic isotope production in the upper atmosphere are diagnostic of (i.e., are proxies for) changes in TSI. If so, then the record of cosmogenic isotopes (^{14}C, ^{10}Be) in trees and ice cores, respectively, can provide a long-term index of solar activity and TSI changes (Beer et al. 1996; Bard et al. 2000). Figure 6.23 shows the record of TSI as reconstructed by Lean et al. (1992) together with ^{14}C and ^{10}Be data for the past millennium. This suggests there was an overall decline in TSI from the early part of the last millennium to ~AD 1700, followed by an increase to higher levels (the highest of the last 1000 years) in the late 20th century. Analysis of northern hemisphere mean annual temperatures since A.D. 1600 and solar variations showed that the two records were strongly correlated over the period (Lean et al. 1995; Crowley and Kim 1996; Mann et al. 1998). Furthermore, numerous studies of regional climatic anomalies during the Maunder Minimum (~1675-1715) reveal the unusual nature of climate during that period, and its societal impacts (e.g. Borisenkov 1994; Wanner et al. 1995; Barriendos 1997; Alcoforado et al. 2000; Luterbacher et al. 2000, 2001; Xoplaki et al. 2001).

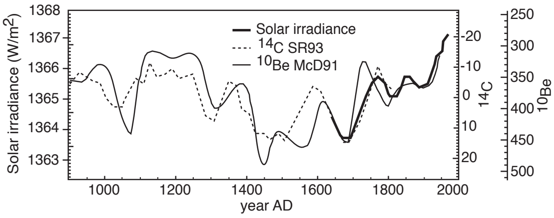

Fig. 6.23. The record of total solar irradiance variations as reconstructed by Lean et al. (1992) together with ^{14}C and ^{10}Be data for the past millennium.

The longer-term record of the ^{14}C anomalies as recorded in the wood of precisely dated trees shows a number of periods with low solar activity, like the Maunder Minimum (Figure 6.24) which may be indicative of episodes of reduced TSI throughout the Holocene. Numerous studies have pointed out correlations between times of solar minima (so defined) and colder episodes, as recorded by a variety of proxies (e.g. Magny 1993; Stuiver & Braziunas 1993). Furthermore, there is growing evidence that solar forcing has played a role in modulating Holocene precipitation variability in the tropics, from northern South America and Yucatan (Black et al. 1999; Hodell et al. 2001; Haug et al. 2001) to east Africa and the Arabian Peninsula (Verschuren et al. 2000; Neff et al. 2001). Indeed, solar variability may also have played a role in mid-continental drought frequency on both short and long timescales (Cook et al. 1997; Yu and Ito 1999). Stuiver et al. (1991) have also noted the similarity between a strong 1470 year periodicity in ^{14}C data and a similar periodicity in isotopic data from GISP2. This is close to the ~1500 year quasi-periodic record of ice-rafted debris in the North Atlantic which has led to speculation that such events are in some way driven by TSI variations affecting the climate system. However, no obvious mechanism for such linkages has been articulated, and oceanic circulation changes may have played a role in both the IRD variations and the ^{14}C anomalies.

6.11.3 Volcanic forcing

Short-term changes in temperature are quite common in the time series of hemispheric mean temperature for the last millennium. There is compelling evidence that these result from gases and aerosols, ejected to high elevations during explosive volcanic eruptions, reducing radiation receipts and leading to lower temperatures (Bradley 1988; Robock 2000). Several studies of tree ring records clearly demonstrate the effects of such changes on tree growth (Briffa et al. 1998; Boninsegna and Hughes 2001). Figure 6.25 shows reconstructed region-wide mean temperatures for North America and Europe since A.D. 1760. This period has sufficient proxy data to enable nine eigenvectors of temperature to be reconstructed, providing reliable information on the spatial patterns of climate variability at these regional scales (Mann et al. 2000). Of particular note is the sequence of years 1834-1838. 1834 was the warmest year in Europe over this period, with overall temperatures ~0.7ºC above the 1902-80 mean. However, in the next four years, temperatures systematically fell to the coldest year of the entire record – 1838 (~0.82ºC below the 1902-80 mean) (Figure 6.27). For marginal agricultural societies, this sequence of cold years was catastrophic. In the central mountains of Norway, for example, harvests were damaged by early frosts, leading to starvation (Nordli 2001) and in Japan there was widespread famine in 1836 (T. Mikami, personal communication). Similarly, dramatic cooling also affected North America (where 1837 was the coldest year of the last ~250 years). This rapid change was due to radiation and circulation anomalies associated with the explosive eruption of Coseguina in January 1835 (Bradley and Jones 1992). Such unpredictable events clearly have a major impact on temperatures over extensive parts of the globe, yet even the largest of these eruptions was small compared to numerous late Holocene eruptions recorded in Greenland ice cores (Zielinski et al. 1994; Zielinski 1995, 2000). If similar magnitude eruptions were to occur today, the rapidity of temperature change would likely have a devastating effect on society, even in a warmer "greenhouse world".

6.11.4 Internal "forcing" factors

All of the "external" forcings discussed above are superimposed on, or modulate "internal" variations of the climate system (Mann and Park 1996; Mann et al. 1995; Trenberth and Hurrell 1994; Gershunov and Barnett 1998; Delworth and Mann 2000). In previous sections, we have discussed the record of two such "internal" modes of climate variability – ENSO and the NAO. Here we consider variations in the North Atlantic thermohaline circulation.

6.11.5 Thermohaline circulation changes

Ice core and marine sediment records reveal that during the last glaciation there were rapid changes

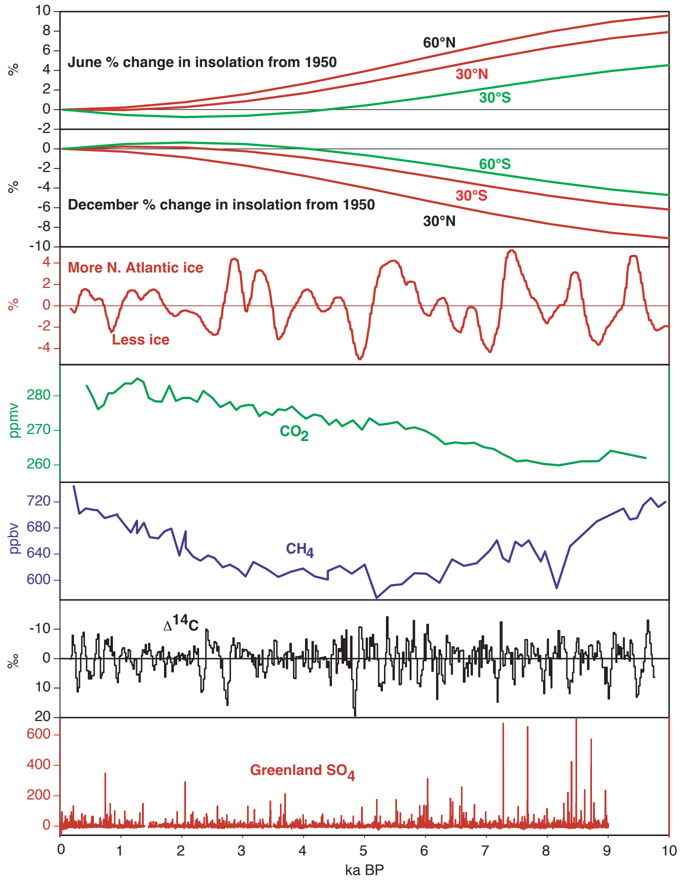

Fig. 6.24. An overview of forcing factors and related large-scale environmental changes in the Holocene. Top to bottom: Insolation anomalies at the top of the atmosphere (as percent departures from A.D. 1950 values) for mid-December and mid-June, for selected latitudes (from Berger and Loutre, 1991); an index of ice rafting in the North Atlantic, based on percentage of ice-rafted grains in ice cores from two sites (data stacked, smoothed and detrended) (from Bond et al. 2001); CO_2 in ice (ppmv) from Taylor Dome. Antarctica (from Indermühle et al. 1999); CH_4 in ice (ppmb) from the GRIP core site, Greenland (from Blunier et al. 1995); ^{14}C anomalies after removing low frequency geomagnetic field variations. Note data are plotted inversely; positive departures imply reduced solar activity and by inference, less irradiance (from Stuiver et al. 1991); volcanic sulfate in the GISP2 ice core from Summit, Greenland, expressed as residuals after removing low frequency variations with a spline function (from Zielinski et al. 1994). Many of these data sets were obtained from the NOAA World Data Centre for Paleoclimatology (Http://www.ngdc.noaa.gov/paleo/data.html)

in the climate of the North Atlantic region, and these have been linked to the cessation of deep-water formation and a concomitant reduction in the influx of warm sub-tropical water in the Gulf Stream and North Atlantic Drift (Chapter 3). Can such changes, albeit on a much reduced scale be responsible for the observed climatic changes in the late Holocene? Alley & Ágústsdóttir (1999), using GISP2 isotope data, argue that there has been a general decline in oceanic heat transport to the North Atlantic over the late Holocene. Their argument rests on an interpretation of seasonal temperature and accumulation trends and "expected" $\delta^{18}O$/temperature relationships. However, a reduction in North Atlantic heat transport would have led to colder and drier wintertime conditions, yet Greenland accumulation data shows no Holocene trend (Meese et al. 1994).

Broecker et al (1999) also argue for late Holocene changes in deepwater formation, but focus attention on the Southern Ocean. Modern geochemical observations of oceanic phosphate, CFC-11 and ^{14}C, appear to be incompatible with the notion of an invariant deep sea ventilation rate. These data can be reconciled if deepwater formation in the Southern Ocean was far higher for several centuries (up to ~800 years) prior to the 20th century. Since overall deepwater formation does not seem to have varied, this implies a parallel reduction in North Atlantic deepwater formation over the same interval, though this interpretation is not supported by WOCE (World Ocean Circulation Experiment) observational data (Ganachaud and Wunsch 2000). Nevertheless, Broecker et al. hypothesise that there may have been a reduction in North Atlantic deepwater formation (and a concomitant reduction of poleward heat transport by the ocean) during the LIA, with opposite conditions during the MWE resulting from an oscillation between deepwater production in the North Atlantic and the Southern Ocean. If such an oscillation has persisted over time, it might explain the ~1500 year periodicity in North Atlantic ice-rafted debris noted by Bond et al. (1997) and the LIA may have just been one of many oscillations within the Holocene, driven either by internal ocean system dynamics, or quasi-regular external forces not yet resolved

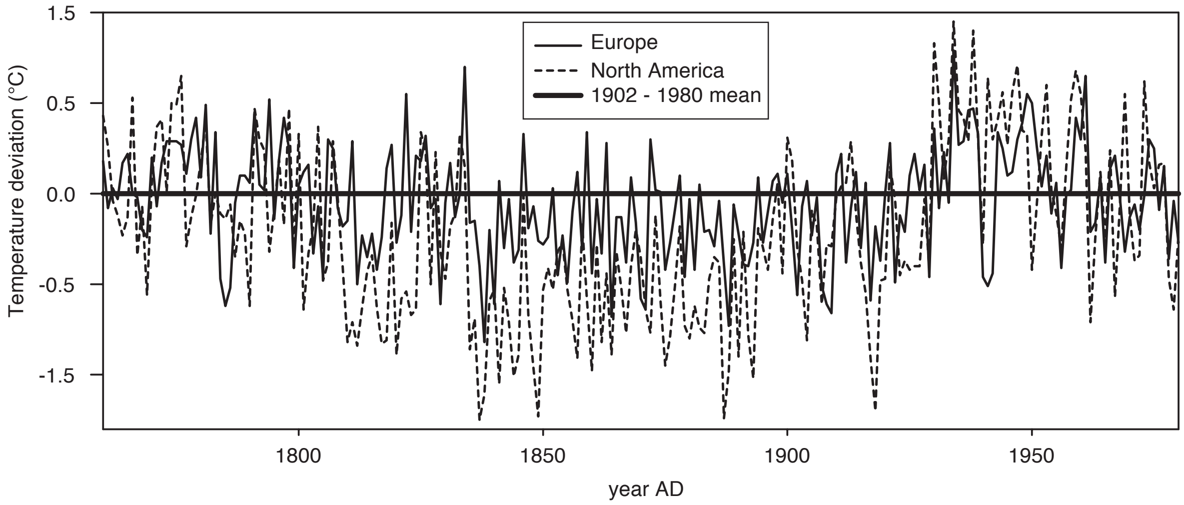

Fig. 6.25. Reconstructed mean annual temperatures for North America and Europe since A.D. 1760 (data from Mann et al. 2000).

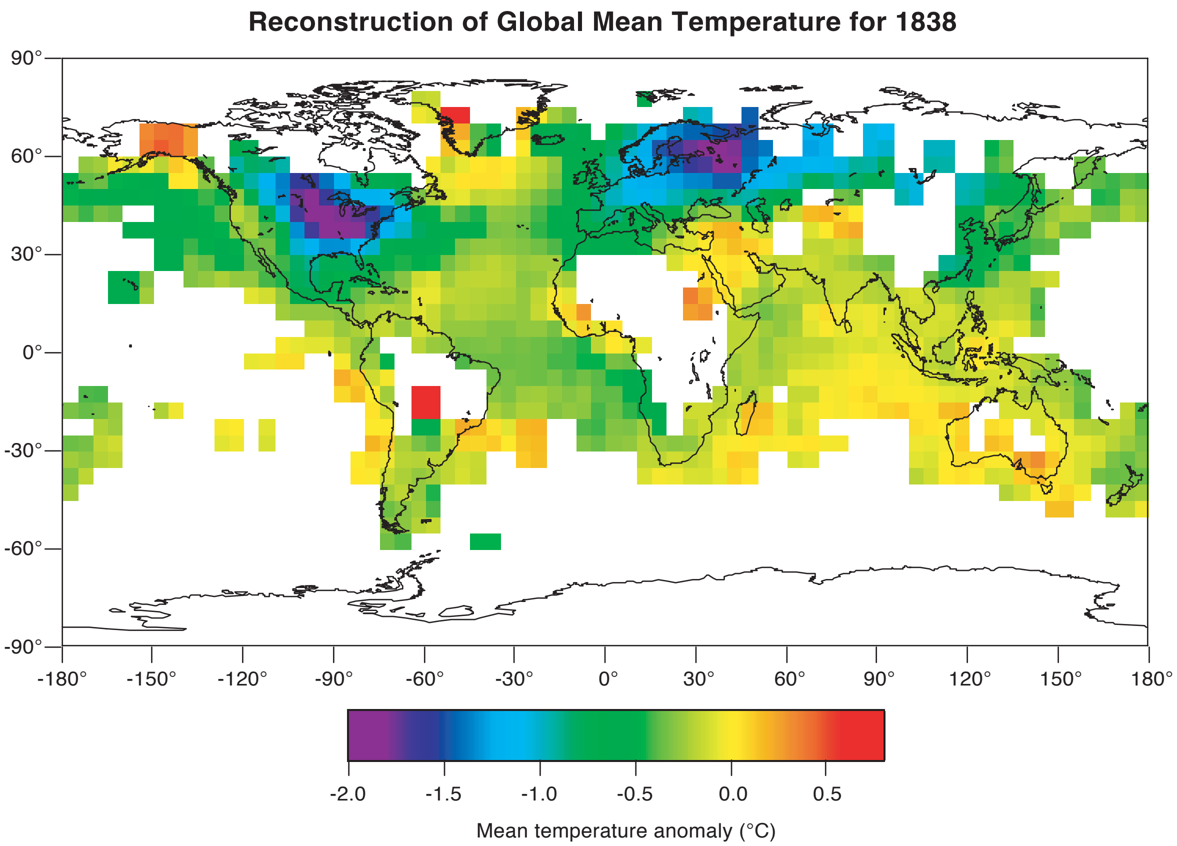

Fig. 6.26. Reconstructed mean annual temperatures in 1838 relative to the mean for 19202-1980 (data from Mann et al. 2000).

6.12 Anthropogenic and natural climate forcings over the past millennium: model results

Present and future global temperature reflects the combination of both anthropogenic forcings and natural ones such as solar irradiance and volcanic aerosols. Early work documented that solar forcing was insufficient to account for the temperature changes in the instrumental record (e.g. Kelly and Wigley 1992; Schlesinger and Ramankutty 1992). Later papers have refined this conclusion by using the paleoclimate record to assess climate sensitivity to radiative changes using correlation analysis (Lean et al. 1995; Bard et al. 2000). Climate model simulations of the recent past, driven by changing radiative forcings, yield temperature records that can be compared to paleo data and used to quantify the sensitivity of climate to radiative change. Refining existing estimates of sensitivity will require improvements on all sides of this question (models, data, and forcings). Model studies have proven very useful in attributing temperature change over the past 100-1000 years to specific forcings. The bottom line from virtually all such investigations is that although solar and volcanic forcing played an important role in pre-industrial climate variability, such natural forcings can explain at most about one-third of the warming of the 20th century, and hence anthropogenic forcings must have played a dominant role in the late 20th-century temperature rise. Here we describe recent modeling efforts to attribute temperature changes seen in paleoclimatic records to specific forcings, and to use these data-model comparisons in assessing climate sensitivity to changes in radiative forcing.

Energy-balance models (EBMs) provide a computationally efficient way to address the response of global temperature to radiative perturbations. EBM simulations of the temperature response to radiative forcings agree well with results from coupled GCMs on hemispheric and global scales (Gates et al. 1995; Raper and Cubasch 1996). However, the sensitivity of an EBM to radiative perturbations is adjustable; the results discussed here therefore can not be used to assess sensitivity, but they do give information on the relative roles of different radiative forcings.

A transient EBM experiment incorporating time-varying solar forcing over the past four centuries (Crowley and Kim 1996) found that solar variability correlates well with pre-industrial decadal-centennial temperature changes but cannot explain 20th century warming. A subsequent EBM analysis combined solar with volcanic and CO_2 radiative forcing, and explored the sensitivity of the exercise to different forcing time series (Crowley and Kim 1999). Using new multiproxy and tree ring temperature reconstructions for the past 600 years, they found that between 18-34% of observed variability was forced by the solar and volcanic changes; the residual variability was spectrally similar to that produced by long unforced GCM runs.

D'Arrigo et al. (1999) diagnosed the response of an EBM to various solar, volcanic and anthropogenic (trace gases plus aerosol) forcing scenarios over the 1671-1973 interval. They compared these results with a tree ring index of northern hemisphere temperature to identify the most successful forcing parameterizations and combinations. The simulations that explain the most temperature variability include anthropogenic forcings, the Dust-Veil Index of volcanic inputs, and any one of four solar reconstructions; these simulate 59-66% of the variance in temperatures derived from instrumental or tree ring data. The model generated a range of variability slightly greater than seen in the data -- cool intervals associated with extreme volcanic eruptions were stronger in the model output, and recent warming was overpredicted. A subsequent study (Free and Robock 1999) used the same model and a similar set of anthropogenic, solar, and volcanic forcings (the latter updated to reduce the impact of strong events) to explore in greater detail the relationship of these forcings to temperature reconstructions of the past 300-400 years. The updated volcanic indices improved the correlation between simulated and observed temperature; volcanism and solar variability had comparable influence on modeled temperatures in this study.

Crowley (2000) used an EBM forced with changing solar, volcanic, anthropogenic greenhouse gas, and tropospheric aerosol forcings, to simulate global temperature change of the past 1000 years. Solar and volcanic inputs explained 41-64% of pre-anthropogenic temperature variations, but anthropogenic radiative forcing had to be included to simulate the dramatic warming of the 20th century. Overall, this exercise suggests that 41-59% of decadal and longer-term temperature variability can be explained as a linear response to solar and volcanic forcings. This model also overestimated 20th-century warmth. The EBM sensitivity was set to 2°C for a doubling of CO_2, and the long-term temperature history agreed with reconstructed temperature histories (Mann et al. 1999; Crowley and Lowery 2000). Naturally, this agreement depends on the sensitivity prescribed – if a 4°C sensitivity had been prescribed, then the long-term temperature trend would be more in agreement with the larger values implied by borehole data (see discussion in Section 6.3).

General circulation models incorporate additional elements of the climate system and allow assessment of spatial patterns of climate change, as well as changes in climate other than temperature (precipitation, circulation, etc). An early GCM study adopted the strategy of imposing a fixed forcing (e.g. a different solar irradiance or volcanic aerosol loading) and assessing the equilibrium response to this forcing (Rind and Overpeck 1993). These authors found that climate would respond to radiative perturbations of reasonable (i.e. observed) magnitudes. Recent work with the HADCM2 model (The Hadley Centre's Second Generation Coupled Ocean-Atmosphere GCM) includes solar, volcanic, and anthropogenic (GHG and aerosol) forcing in assessing the causes of 20th century climate change. Spatial signatures of each forcing were developed and linear combinations of these were calculated to provide the best fit to observations, using optimal fingerprinting techniques (Tett et al. 1999; Stott et al. 2001) and linear correlation analyses (Johns et al. 2001). These studies disagree on the relative role of solar forcing in 20th century warming; depending on methodology and data source, it is either absent or minor. However, they concur that late 20th century warming is dominantly a result of anthropogenic forcing. A follow-on study with HADCM3 (Stott et al. 2000), using the same forcings in a transient mode, supports this conclusion and provides an estimate of the spatial pattern associated with both modeled and observed trends.

The 20th century is a relatively short period for developing confident assessments of the relationship between climate forcing and change, particularly as two of the leading forcings, greenhouse gases and solar irradiance, show positive trends, and solar variability contains substantial low-frequency variability. The combination of multi-century paleoclimatic reconstructions and forced multi-century GCM runs will be a powerful tool for attributing the causes of past change and diagnosing natural climate sensitivity. Such simulations are not yet commonplace, particularly with coupled ocean-atmosphere GCMs, but work in this direction is beginning. Cubasch et al. (1997) used the ECHAM3/LSG coupled ocean-atmosphere model to assess the climatic response to changes in solar forcing (from Hoyt and Schatten 1993) since AD 1700. They found that the 0.35% variations in the solar input time series (peak to trough) produced global mean temperature changes of about 0.5°C, with a pattern of stronger land-sea contrast resulting from increased irradiance. This direct response lagged the forcing by <10 years and was especially clear on the multi-decadal time scale. An inverse response of temperature to solar forcing was also noted in the North Atlantic temperature and thermohaline circulation; high irradiance led to weaker meridional circulation and cooler near-surface temperatures. They concluded that the warming simulated in response to solar variability contributed to 20th century warming, but could not explain it alone, and the pattern of solar warming differed from that generated by anthropogenic forcings in the same model.

Rind et al. (1999) used a slightly more conservative scenario of solar variability (Lean et al. 1995, extended to A.D. 1500 using cosmogenic isotope data) as input to transient simulations spanning the past 500 years. They used an atmosphere-only GCM coupled to a mixed layer ocean that included specified heat transports and diffusion of heat through the bottom of the variable-depth mixed layer. Analysis of the time-dependent temperature variability suggests that temperature response changes with latitude, that responses at different time scales have a preferred spatial expression, and that feedbacks are entrained at preferred time scales. They simulated a similar temperature change as the previous study (0.5°C warming from Maunder minimum to present), but the agreement was fortuitous: in Rind et al., the model was more sensitive, and the solar forcing less extreme. Their analyses indicated that even with a range of potential initial conditions, ocean heat uptakes, and initial solar conditions, the recent warming can not be explained by solar variability alone.

Robertson et al. (2001a) used the same model as Rind et al. (1999), but included a more complete set of radiative forcings. In addition to the extended Lean et al. (1995) solar record, they forced the model with a new, zonally resolved volcanic sulfate record (Robertson et al. 2001b) as well as increasing concentrations of CO_2, CH_4, and tropospheric aerosols (the latter scaled to greenhouse gases on a regional basis). The resulting global temperature history suggests that pre-industrial temperatures were 1.5°C cooler than present. This value agrees with the borehole temperature reconstruction (Huang et al. 2000) but is substantially greater than that indicated by various tree ring based and multiproxy reconstructions (Mann et al. 1999; Briffa et al. 1998; Jones et al. 1998; Crowley and Lowery 2000). On the other hand, the simulated interannual-decadal variability agrees closely with the Mann et al. multiproxy temperature reconstruction. Robertson et al. (2001) argued, however, that the paleoclimatic records used in Mann et al. (1999) may systematically underestimate low-frequency variability, even while capturing the high frequency changes accurately.

The GISS GCM has a sensitivity of 4.2°C for a doubling of CO_2, which places it among the more sensitive of the models used in the IPCC assessment (Houghton et al. 1995; IPCC 2001). The agreement of the Robertson et al. (2001a) transient simulation and the Huang et al. (2000) data supports this high sensitivity. If, however, the multiproxy and tree ring records are correct in yielding a smaller estimated temperature difference, then the model would appear to be overly sensitive. To address the issue of climate sensitivity to increasing greenhouse gases alone, Robertson et al. (2001a) focused on the 1880-1930 interval, when greenhouse gases were rising but solar irradiance was stable. The degree of temperature change over this interval was consistent among the simulations and the observed datasets (including instrumental data) and thus supports the higher sensitivity of the GISS model. It should be noted, however, that the Hoyt and Schatten (1993) solar reconstruction shows an increasing trend over this interval which – if true – would reduce the greenhouse gas sensitivity estimated by Robertson et al. in this exercise. This study clearly highlights the need to improve the accuracy of both temperature reconstructions and forcings over the past several centuries.

Transient GCM studies suggest spatial and temporal fingerprints of solar forcing that can be useful in diagnosing the variability seen in paleoclimatic records. In both the Rind et al. and Cubasch et al. studies, the correlation of solar variability and temperature was strongest in the tropics with a lag of several (5-10) years, even though the amplitude may have been higher in the extra-tropics. The frequency-domain properties of the solar irradiance record vary through time, and those time scales were expressed in the temperature record differently at different latitudes. For example, in the Rind et al. study, 50-80 yr period variability was strongest in the tropics and 20 year variability persisted at all latitudes (even when it disappeared from the forcing!). Neither study found significant 11-year power in the simulated global temperature, likely because the modeled oceanic response time muted the response to forcing at this frequency. Oceanic regions showed a lower correlation to anthropogenic greenhouse forcing and a stronger correlation to solar forcing. These simulations thus provide clues as to where solar forcing may appear most strongly in paleoclimate records (such as the tropical oceans), as well as possible explanations for the absence of solar signals (e.g. the commonly sought-after 11-year periodicity).

6.13 Detecting twentieth century climate change

The detection of 'significant' trends, and being able to attribute them to particular climate forcings, requires knowledge of the expected climate signal and an estimate of natural climate variability (e.g. Santer et al. 1995; Tett et al. 1999). Almost all climate change signal detection and attribution studies to date assume a climate-model-based estimate of natural climate variability. This is a major and relatively untested assumption, and is a source of potential criticism that could be used to detract from all such studies (Bradley et al. 2000). Proxy-based reconstructions of past climate can be used to test this assumption, by providing information about pre-twentieth century variability which can be assumed to have been influenced relatively little by anthropogenic forcing.

There are two distinct approaches that can be followed. The first is to compare the model-based variability estimates (obtained from multi-century control simulations with constant external forcing, or from model integrations under natural forcing changes only) with the reconstructed past climate variability, in terms of levels of variability (standard deviation) or patterns of variability. This may be done at various time scales, encompassing the decadal to century time scales that are most important for climate change detection. This approach has been followed by, for example, Jones et al. (1998) who found different magnitudes and patterns of multi-decadal variability between climate model output and paleodata (but a similar lack of global-scale coherency in both data sets). On the other hand, Collins et al. (2000) found that one climate model simulation exhibited levels of interdecadal variability of summer temperatures across the Northern Hemisphere that were very similar to those reconstructed from tree ring density, unless the tree ring density data were standardised to maintain maximum multi-century variability (Briffa et al. 2001). Thus, so far such comparisons are somewhat inconclusive. Even when model-based variability appears to differ greatly from that derived from paleodata, uncertainty in the paleoclimate reconstructions may be so large that the difference cannot be considered significant. What do we then conclude about the detection studies that have used these model estimates of natural variability?

The second approach circumvents some of these problems. Rather than use the paleoclimate reconstruction to evaluate the veracity of the climate model simulation, it is possible to carry out the same detection study using either the model-based estimate of natural variability or the proxy-based estimate. If the climate change signal is detected significantly in both cases, then it does not matter if the model and proxy estimates are inconsistent.

An example of the second approach is given in

Figure 6.27, (based on Osborn and Hulme 2001) where recent warming trends are compared against the range of trends that have occurred naturally. We base our estimates of natural variability on the AD 1000 to 1900 section of the Mann et al. (1998, 1999) Northern Hemisphere (NH) mean annual temperature reconstruction, and the NH annual temperature simulated in a 1400-year control simulation of the UK Hadley Centre's climate model HadCM2 (Tett et al. 1997). These are compared with observed NH annual temperatures since A.D. 1901. In addition, we compare the observed warming with that simulated by the HadCM2 model, run with prescribed increases of greenhouse gas and sulphate aerosol concentrations (representative of the period from A.D.1901 to present). If we compare twentieth century mean temperature anomalies with the range of mean temperature anomalies occurring in the natural variability estimates, then we find the twentieth century to be very unusual (as already noted in Section 6.3). We do not compare *means* in this example, however, because that would require that the paleoclimate reconstruction is very reliable and homogeneous over multi-century to millennial time scales, a requirement that is both demanding and not easily verified. Instead, we consider in Figure 6.27 decadal to century *trends* of temperature, thus decreasing the reliance on obtaining accurate multi-century to millennial trends. For each of our two natural variability estimates, the distribution of all possible trends of a certain length (ranging from 10 to 100 years) is estimated, and Figure 6.27 shows the 95th percentile (°C/decade) of each distribution. The model and paleo-based natural variability estimates are in excellent agreement for trends longer than 40 years, though the model overestimates the strength of shorter trends relative to the paleoclimate record. This is not a like-with-like comparison, however, because (i) any natural forcings, e.g., volcanic or solar, are absent from the model simulations (and would probably increase the simulated trends at decadal and century time scales, respectively), while (ii) the paleoclimate record has lower variance than observed (because it is does not fully capture the variance of the observed record). If we assume that the residual variance in (ii) can be modeled as a white noise process (e.g., Collins et al. 2000), then this would raise the 95th percentile of the trends, principally at shorter time scales, thus bringing the paleo and model estimates into closer agreement. Mann et al. (1999) show that this white noise assumption is valid for their reconstruction after AD 1600, but before this, their reconstruction does less well at capturing low-frequency climate variability and the residuals are likely to be autocorrelated. Taking autocorrelated residual variance into account would raise the 95th percentile of the proxy-based trends for all time scales.

It is clear, therefore, that the model- and proxy-based estimates of natural temperature trends are quite similar, though some uncertainties need to be taken into account. But do the small differences matter? Figure 6.27 also shows the temperature trends, computed separately from the observed and simulated twentieth century data, over various length periods, all ending in 1999. With the exception of the decadal time scale, all trends lie above the 95th percentile of 'natural' climate variability – indicating a significantly unusual rate of recent warming, regardless of which estimate of natural variability is used.

This approach to detecting significant warming has been applied to other paleoclimate reconstructions, ranging in spatial scale from hemispheric to single site estimates (Hulme et al. 1999; Osborn and Hulme 2001). All hemispheric or quasi-hemispheric cases yield significant warming for recent trends of 30 years and longer, despite larger differences between model and proxy-based natural variability estimates in some cases. For localised sites, however, trends are typically similar to, or below, the 95th percentile of natural variability, thus precluding detection of unusual climate change at small spatial scales.

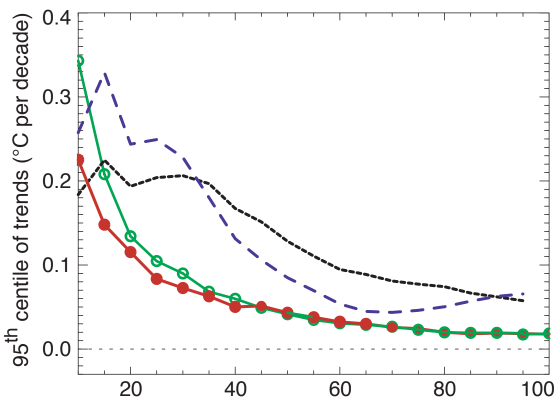

Fig. 6.27. Detection of significant 20th century temperature trends of varying length (all expressed as °C/decade). The purple line shows observed temperature trends from various length periods. The black line is the mean of an ensemble of four simulations from the HadCM2 coupled climate model forced by historical increases in greenhouse gas and sulphate aerosol concentrations. These can be compared against the estimates of the 95thcentile of the various length trends that are possible due to natural climate variability. The green line is computed from a 1000-year control integration of HadCM2, with fixed external forcing. The red line with solid dots is computed from the pre-1900 portion of the 1000-year Mann et al. reconstruction.

6.14 Concluding Remarks

A wide range of proxies for past climate provide an invaluable long-term perspective on global climate variability, and proxies of past forcing allow natural factors affecting climate to be evaluated. Together, these records indicate that recent warming is both unusual and not explicable in terms of natural factors alone. By combining model simulations with paleoclimatic data, a better understanding of climate sensitivity and the climate system response to forcing is emerging. Nevertheless, many uncertainties remain. Paleoclimate research has had a strong northern hemisphere, extra-tropical focus (but even there the record is poorly known in many areas before the 17th century). There are very few high resolution paleoclimatic records from the tropics, or from the extra-tropical southern hemisphere, which leaves many questions (such as the nature of climate in Medieval times) unanswered. Furthermore, much remains to be learned about the mechanisms by which climate is recorded in proxies. Even so, proxy climate records have revealed important, hitherto unknown, features of climate variability, including regime-like fluctuations in the amplitude and frequency of the ENSO phenomenon and its extra-tropical teleconnections, interhemispheric connections over the Pacific Basin, extreme multidecadal droughts in mid-latitudes, and a confounding role for the North Atlantic Oscillation in temperature changes in that basin over the last century. Even so, variability of other major climate systems, such as the monsoons, is poorly documented. Our understanding of regional responses to forcing and how one part of the climate system may lead, or lag, another remains poor. Much more work on these topics remains to be done.

Climate variability over the last millennium provides the essential context for assessing future changes, even as anthropogenic effects become increasingly dominant. It is over the last millennium that modern societies have developed, coping with a wide range of climatic vicissitudes. Much of the world still lives at a subsistence level, very much affected by both inter-annual and interdecadal climate variability. As we confront a world whose population is expected to increase from 6 billion people today to ~9-10 billion by 2060, paleoclimatic research can shed an important light on mechanisms of climate variability at these societally-relevant timescales.

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