Climate Since A.D. 1500
Revised Edition (1995)
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34 Recent developments in studies of climate since A.D. 1500

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34.1 Introduction

Since the first edition of *Climate since A.D. 1500* was published there have been a number of new developments in research which have added to our knowledge of climate fluctuations during this period. Here we provide a brief summary of some of the more important studies which have recently become available. We note also that several volumes have been published recently, which focus to a large extent on paleoclimatic records of the last 500-1000 years. These include Bradley (1991), Diaz and Markgraf (1992), Frenzel *et al.*, (1992), Murray and Overpeck (1993), Martinson *et al.* (1994) and Hughes and Diaz (1994a). In addition, there have been very significant developments in data access, thanks to the establishment of the World Data Center A (WDC-A) for paleoclimatology within the U.S. National Geophysical Data Center (325 Broadway, Boulder, Colorado 80303). Many data sets can now be obtained for a nominal fee on diskette, or accessed via Internet, and transferred using the anonymous FTP protocol. In addition, WDC-A has developed a software program, *Paleovu*, which enables users to display paleoclimate data sets from throughout the world. Further details of these developments can be obtained from WDC-A via their email address (paleo@mail.ngdc.noaa.gov).

34.2 Tree-ring research

Several long tree-ring chronologies have been produced in the last few years. In Europe, new reconstructions from northern Fennoscandia and the northern Urals extend and improve those produced by Briffa and Schweingruber (Chapter 19) and Graybill and Shiyatov (Chapter 20). For the Tornetrask region of northern Sweden, Briffa *et al.* (1990, 1992) have reconstructed April to August summer temperatures back to A.D. 500 using maximum latewood density and ring width data. The reconstruction shows a warm period in the 1400s and the coldest period of the entire reconstruction between 1570 and 1730. There is also evidence of a warm period in the tenth and eleventh centuries. The reconstruction for the northern Urals region has been improved by the use of maximum latewood density data along with ring-width data by Briffa *et al.* (1994a). There is a cold period reconstructed from 1530 to 1650 somewhat earlier than in northern Sweden. The twentieth century in this reconstruction is the warmest of the record (back to A.D. 914) unlike Tornetrask where there were clearly a few warmer centuries earlier. Both reconstructions illustrate the importance of the density data in improving the fidelity of the reconstruction. In the near future many reconstructions from northern Russia are expected to be produced (Shiyatov *et al.*, 1994; Briffa *et al.*, 1994b).

In North America, several new reconstructions have been produced, mainly from the altitudinal and latitudinal tree lines. For example, Graybill and Funkhouser (1994) have reconstructed cool season temperatures from the Sierra Nevada mountains of California and for northern North America, further reconstructions have been made by D'Arrigo and Jacoby (1993). The latter extends the climate record for both Alaska and Labrador. Maximum latewood density data have also been used by Briffa *et al.* (1994c) to reconstruct summer temperatures for Alaska, the Mackenzie Valley and the Quebec/Labrador region. The longest reconstructions for northern North America, which extend to A.D. 1500, indicate that the twentieth century was the warmest of the study periods.

Perhaps the greatest advances in dendroclimatology have taken place in the Southern Hemisphere over the last three years. In South America two millennial-long reconstructions of summer (DJF) temperature have been made using Alerce (*Fitzroya cupressoides*) trees. One is located on the western side of the Andes in south-central Chile (Lara and Villalba, 1993), while the other is on the eastern side in Andean Argentina (Villalba, 1990). Both reconstructions show little evidence of warming during the twentieth century and little century-scale variability over the last 1000 years, though this may be an artefact of the standardization procedure (see Chapter 1, Section 1.5.2). Further studies are in progress to evaluate whether this is the case.

In the Australasian region, temperature reconstructions have been made using ring-width data (from Huon pine, *Lagarostrobos franklinii*) for Tasmania (Cook *et al.*, 1991, 1992) and for New Zealand by Salinger *et al.* (1994) who also reconstruct zonal and meridional pressure indices over New Zealand. The Tasmania reconstruction now extends back to 1261 B.C. and has potential to be extended back further. The twentieth century contains the warmest decades of the last 2000 years. In New Zealand, the reconstructions extend back to 1500 for Stewart Island (D'Arrigo *et al.*, 1994b) and to 1750 for a temperature reconstruction for the two main New Zealand islands (Salinger *et al.*, 1994). The potential for extending the reconstructions further back in time, in both New Zealand and South America, is readily apparent. In both regions they represent the best proxy variable for summer climate reconstruction.

It was also noted in Chapter 1 that few reconstructions have been made from tree-ring data in the tropics. This situation is slowly changing; over the last few years, tentative precipitation reconstructions have been made for Thailand (B. Buckley, personal communication) for Indonesia (Palmer and Murphy, 1993; Jacoby and D'Arrigo, 1990; D'Arrigo *et al.*, 1994a) and for southern Japan (Sweda, 1994). These reconstructions are relatively short, but they do point to the dendroclimatic potential of the regions. Such studies will be important in the future, as for some areas they provide the only means of obtaining pre-instrumental climate information.

34.3 Corals

It was noted in the first issue of this book that coral studies have great potential for paleoclimatic reconstruction, though a number of problems needed to be resolved. In fact, there had been many studies demonstrating excellent correlations between various oceanographic and/or climatic variations, and coral geochemistry or growth rates over several seasonal cycles, but few studies went beyond the period of instrumental records. A number of studies have now extended such analyses into the nineteenth century, or earlier (Table 34.1) providing a unique long-term perspective on climate variability in the tropical ocean-atmosphere system.

Coral growth (in scleractinian corals) involves the accretion of skeletal carbonate. In this process, there is a temperature-dependent fractionation of oxygen isotopes, such that the resulting carbonate is enriched in the heavier (^{18}O) isotope at lower water temperatures (Δ-1°C$\approx\Delta$+0.22‰$\delta^{18}O$) (Fairbanks and Dodge, 1979). Thus, in areas with a strong seasonal cycle in water temperature, a corresponding annual cycle in $\delta^{18}O$ is recorded along the growth axis of the coral (Dunbar and Wellington, 1981). In the Galapagos Islands, for example, warm season coral growth is characterized by relatively low $\delta^{18}O$ values (isotopically light carbonate) (Dunbar *et al.*, 1994). El Niño events, which result in exceptionally high sea-surface temperatures in this region, are identifiable as large negative $\delta^{18}O$ anomalies (Figure 34.1). A 366-year $\delta^{18}O$ record from this region indicates inter-annual temperature changes of up to 2.5°C, with generally cooler conditions in the seventeenth, early nineteenth and early twentieth centuries, and consistently warm sea-surface temperatures for most of the eighteenth century. Singular spectrum analysis of this record indicates a shift to higher frequency climate variability in the ENSO frequency band (from 4-6 to 3-4 years) after ~1840. This is also paralleled by a shift in the lower frequency band, from ~33 years to ~17 years. The analysis also points to a possible solar influence on climate variability in the tropics, since both growth rate and $\delta^{18}O$ demonstrate significant variance at 11 and 22 years, and the amplitude of the $\delta^{18}O$ record varies in phase with that of the 11-year solar cycle. However, the physical mechanism for such a connection is as yet unkown.

In other regions, heavy rainfall amounts (which are generally isotopically light due to convective activity) cause the isotopic composition of the oceanic mixed layer to be lower. Corals growing in such environments record the periods of heavy rainfall as isotopically light carbonate bands. For example in the western Pacific, corals from Tarawa Atoll (Kiribati) record a clear seasonal cycle of $\delta^{18}O$ in response to the strong seasonal rainfall pattern in the

Table 34.1 Long coral-based records of past climate

Site	Latitude	Longitude	Record length	Parameter	Indicator of	Reference
Bermuda	32°N	65°W	~1180-1986	Growth rate	SST/upwelling	Pätzold and Wefer, 1992
Cebu Island, Philippines	10°N	124°E	~1860-1980	$\delta^{18}O$ $\delta^{13}C$	SST and rainfall/cloudiness	Pätzold, 1986
Gulf of Chiriquí, Panama	8°N	82°W	1707-1984	$\delta^{18}O$	Rainfall/ITCZ position	Linsley *et al.*, 1994
Tarawa Atoll, Kiribati	1°N	172°E	1893-1989	$\delta^{18}O$	Rainfall	Cole *et al.*, 1993
Isabela Island, Galapagos Islands	0.4°S	91°W	1587-1953	$\delta^{18}O$	SST	Dunbar *et al.*, 1994
Espiritu Santo Vanuatu	15°S	167°E	1806-1979	$\delta^{18}O$ $\delta^{13}C$	SST and rainfall/cloudiness	Quinn *et al.*, 1993
Great Barrier Reef, Australia	22°S	153°E	1635-1957	$\Delta^{14}C$ $\delta^{18}O$	Oceanic advection and/or upwelling	Druffel and Griffin, 1993

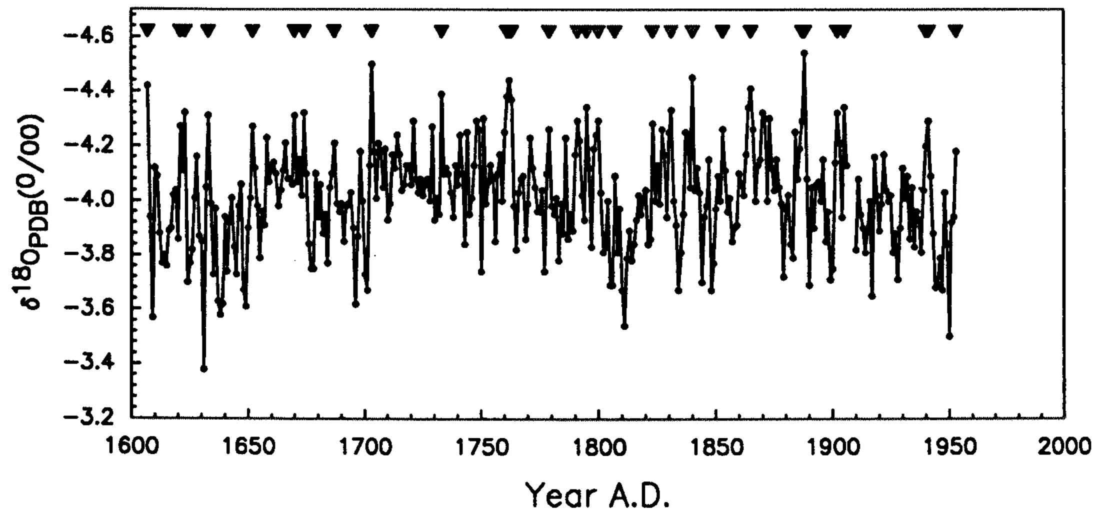

Figure 34.1 Time series of annual $\delta^{18}O$ values for a coral (*Pavona clavus*) from Isabela Island, Galapagos archipelago, Ecuador. More negative values correspond to warmer water temperatures; extreme negative anomalies in this record indicate significant El Niño events, which are indicated by small triangles across the top of the figure (from Dunbar *et al.*, 1994).

region (Cole and Fairbanks, 1990). Furthermore, ENSO events generally result in exceptionally heavy rainfall in this area, which is recorded by unusually low values of $\delta^{18}O$ in the corals (Cole *et al.*, 1993). Similarly, in the Gulf of Chiriqui (western Panama) corals are influenced seasonally by rainfall-induced changes in salinity and surface water $\delta^{18}O$, and hence provide a record of rainfall associated with the strength and position of the ITCZ in this region (Linsley *et al.*, 1994). Over the last 277 years (1707-1984) there is no evidence that the ITCZ position has ever extended well north of its current seasonal range (which would be recognizable as a double rainfall event). Furthermore, the annual range in $\delta^{18}O$ has been relatively constant over the period studied, though there has been a reduction in $\delta^{18}O$ after ~1850, due to either higher sea-surface temperatures, or more precipitation (or a combination of both). There is also some evidence in this record for a periodicity at ~11 years, as noted in the Galapagos record further south.

Corals from the southern end of the Great Barrier Reef, eastern Australia, have been examined in terms of $\Delta^{14}C$ variations and $\delta^{18}O$ for the period 1635-1957 (Druffel and Griffin, 1993). $\delta^{18}O$ (largely a sea-surface temperature signal in this area) declined over the period of record, suggesting an increase in temperature of ~2°C, with an increase in the seasonal range since the early 1700s. Changes in $\Delta^{14}C$ are larger than can be explained by atmospheric changes, suggesting that the record reflects shifts in oceanic circulation (surface water advection, and/or upwelling effects).

Only one other coral record over 200 years in length is currently available. In Bermuda, normalized annual growth rates are inversely correlated with air and sea-surface temperatures. This is related to upwelling of cooler waters which results in enhanced nutrient flux to the photic zone and increased coral growth. An 800-year record of coral growth indicates below average growth from ~1470-1710, and from ~1760 to the end of the nineteenth century (Pätzold and Wefer, 1992). Lowest values were recorded in the 1890s.

All of the coral studies to date provide insights into the strength and variability of ENSO events (both warm and cold events). As one might expect, the record of ENSO strength, as reconstructed by Quinn and Neal (Chapter 32, this volume) is not always reflected in the same way in different parts of the Pacific Basin. Indeed, "revised" ENSO chronologies have been proposed by Hocquenghem and Ortlieb (1992) and Mabres *et al.* (1993) (as well as by Quinn, 1993a, b). As more coral records become available from throughout the tropical oceans (cf. Dunbar and Cole, 1993), a much better picture of the spatio-temporal pattern of ENSO events should become apparent.

34.4 Ice cores

In the last few years, two new deep ice cores have been recovered from Summit site, Greenland by the GISP2 and GRIP projects (Grootes *et al.*, 1993). However, as yet little attention has been paid to the upper Holocene section of the record apart from studies of volcanic sulphate signals (see below). Geochemical analyses by Mayewski *et al.* (1993) provide a comprehensive view of changes in atmospheric chemistry from A.D. 674 to the present. It appears from an increase in sea salt sodium that from ~1587-1914 airflow from lower latitude oceanic regions increased, perhaps indicating more meridional airflow during that interval. Peaks in ammonium levels (related to biomass burning) are seen throughout the record, but are notably absent from ~1650-1750, indicating reduced fire frequency at that time. Melt records and $\delta^{18}O$ from the Agassiz ice cap, Ellesmere Island (Canadian High Arctic) reveal complexities due to wind scouring, but higher summer temperatures appear to have prevailed from ~A.D. 1350-1550 (Fisher and Koerner, 1994). Coldest summer temperatures of the last 500 years were from ~1550-1760 and ~1800-1850.

Ice core records from high elevation sites in central Asia reveal pronounced increases in $\delta^{18}O$ (equated with higher temperatures) over the last few decades (Thompson *et al.*, 1993). Indeed, in the Dunde ice cap ice core record (from western China) the $\delta^{18}O$ values are the highest of the entire record (which is thought to span the last 10,000 years, or more). Other evidence for pronounced warming in tropical mountains in the twentieth-century is shown by marginal recession of the Quelccaya ice cap (Peru) and loss of the annual $\delta^{18}O$ cycle by melting at the summit (5670m), an event apparently unprecedented in the 1500-year record represented by the Quelccaya ice core. Similar observations concerning recent, pronounced ice recession have been made by Hastenrath (1992) for high elevation glaciers in East Africa. In the northern Andes, photographic and historical records document a dramatic reduction in ice cap and glacier extent over the last 100-150 years (Schubert, 1992) and it appears that similar changes have occurred in glaciers from many other parts of the world (Oerlemans, 1994). Thompson *et al.* (1993) note that low latitude, high altitude ice caps show evidence of "Little Ice Age" cooling and twentieth-century warming more strongly than polar ice cap records, suggesting that such sites are especially sensitive to large-scale climatic changes. Indeed, several general circulation models of future greenhouse gas-induced temperature changes also indicate that sites at high elevations (5-7km) in low latitude, continental interiors may experience the largest temperature changes in future decades.

34.5 Forcing factors

The history of large explosive volcanic eruptions is best recorded in remote ice caps where unusually large sulphate signals indicate acidic fallout from the eruption cloud. A detailed record from the Summit site, Greenland provides a 9,000 year chronology of major eruptions, primarily from the Northern Hemisphere (Zielinski *et al.*, 1994). By analogy with the observed fallout distribution patterns from nuclear bomb tests, they estimate that the largest eruptions of the last 2,000 years occurred around A.D. 1259 (possibly El Chichón, Mexico; cf. Palais *et al.*, 1992), ~1604 (Huaynaputina, Peru [1600] and/or Momotombo, Nicaragua [1605]) ~1641 (Awu, Indonesia [1641] and/or Komaga-Take, Japan [1640]) 1831 (Babuyan, Philippines) and 1815 (Tambora, Indonesia). They also note that multiple eruptions in the seventeenth century may have contributed to cooling at that time. This was also suggested by Bradley and Jones (1993) who noted a significant correlation between Northern Hemisphere summer temperatures and the total acidity signal at Crête, Greenland on decadal time-scales. Comparably long and well-dated acidity records from the Southern Hemisphere are not yet available, but Moore *et al.*, (1991) have reconstructed a 700-year history of large volcanic events which led to the deposition of high levels of sulphate over East Antarctica. The A.D. 1600 eruption of Huaynaputina is clearly recorded, as is Tambora (1815) and the explosive event in A.D. 1259, all of which are seen in both Greenland and Antarctica. Interestingly, another large eruption around 1808 is recorded in the Antarctic record, possibly from a source in or around Antarctica. Other large eruptions (of unknown origin) occurred in ~1510 and ~1460, and there are also large signals from the eruptions of Coseguina, Nicaragua in 1835, Krakatoa, Indonesia in 1883 and Agung, Indonesia in 1963.

Estimates of solar irradiance changes over the last 300 years have been made from historical observations of solar activity, calibrated against the (very brief) satellite-based active cavity radiometer (ACRIM) measurements of total irradiance over a single solar cycle (Hoyt and Schatten, 1993). These estimates point to solar irradiance below twentieth-century levels throughout the preceding 200 years, except for the intervals from ~1720-1780 and ~1820-1845. However, the estimated changes in irradiance are small (overall, less than 0.4 per cent in total, spectrally integrated solar radiative output, much of which is in the very short [U-V] wavelengths which do not reach the lower troposphere) and it seems unlikely that such changes alone had a significant impact on global climate. Nevertheless, more work on this topic is needed. There are, for example, intriguing correlations between ^{10}Be records from polar ice cores and various temperature records (Beer *et al.*, 1994). ^{10}Be is a proxy of solar output variations (though complicated by depositional effects brought about by variations in precipitation). Higher levels of ^{10}Be occur when solar activity is reduced; a reduction in the solar wind leads to higher levels of cosmic ray bombardment in the outer atmosphere, and increased production of cosmogenic isotopes such as ^{10}Be and ^{14}C. Hence an inverse correlation between temperature changes (or proxies thereof) and ^{14}C or ^{10}Be may be indicative of a solar-climate connection. Such a relationship has been noted between ^{14}C "wiggles" (anomalies from low-frequency background variations) and the record of worldwide glacier fluctuations (Wigley and Kelly, 1990) though the glacier chronology is poor and the causes of glacier fluctuations are not well documented. At this point, there is not enough evidence to make a convincing case for a (physically) significant connection between solar output changes and temperature. However, in conjunction with other forcings (such as aerosol loading from explosive eruptions) such changes may have played a role in modulating temperature over the last few centuries. Alternatively, solar irradiance changes alone may have led to positive feedbacks within the climate system, amplifying the initial perturbation (Stuiver and Braziunas, 1993). With so much uncertainty on this topic, there is clearly the need for considerably more research into the nature of solar forcing on decadal-to-century timescales.

Another approach to assessing the possible climatic effects of forcing factors, such as solar irradiance and volcanic aerosol load changes, is through the use of general circulation models (GCMs) (e.g. Rind and Overpeck, 1993; Nesme-Ribes *et al.*, 1993; Rind and Lean, 1994). GCMs can also be run without any external forcing to consider the magnitude of internal variations of the ocean-atmosphere climate system on the century-to-millennial timescale. Knowledge of the potential importance of all these factors has been stressed in many modelling and statistical exercises (see e.g. Wigley and Barnett, 1990) to try to detect the enhanced greenhouse effect in instrumental and paleoclimatic data of the last few hundred years. For example, one approach seeks to explain the past using a variety of forcing factors while another attempts to say whether changes during the twentieth century have been unprecedented, in relation to changes over the longer record.

The magnitude of natural internal variability has been assessed using long integrations of GCMs (e.g. Stouffer *et al.*, 1994) and using much simpler energy balance climate models by Wigley and Raper (1990). Both analyses come to similar conclusions from 1,000 years of generated data. These conclusions are that the maximum change likely from this source is one of between 0.2-0.3°C per century, at most only half the rise seen over the last 100 years. This type of study is often taken to lend support to ideas that the enhanced greenhouse effect is smaller than suggested by the Intergovernmental Panel on Climate Change (Houghton *et al.*, 1992). It is equally likely though, that the underlying natural variability could have led to cooler temperatures over the last 100 years, implying a larger greenhouse effect (Wigley and Barnett, 1990). GCMs have also been used to study the short term effects of volcanoes on climate (e.g. Graf *et al.*, 1993). The patterns of change generated by the models appear qualitatively similar to those found in observational data after recent major volcanic events (Robock and Mao, 1992; Kelly *et al.*, 1994; Jones and Kelly, 1994).

Perhaps the most debated use of models in the detection area has been attempts to explain the instrumental and paleoclimatic past. Many years ago these efforts were referred to as "noise-reduction" studies (Wigley *et al.*, 1985) and even today they are still common in the literature (Schoenwiese and Stahler, 1991). Although many are basically curve-fitting exercises (e.g. Friis-Christensen and Lassen's 1991 attempt to explain past temperatures as a result of variations in the length of the solar cycle) they receive wide publicity despite clear problems both with their statistical validity and the physical mechanism causing the relationship with climate. Recent attempts using energy-balance climate models, while much more scientifically based, still have their problems. The best examples of this type of approach are Wigley and Raper (1992), Schlesinger and Ramankutty (1992) and Kelly and Wigley (1992). All conclude that the climate sensitivity is about 2.5°C if the effect of anthropogenic sulphate aerosols is included and about 1.5°C if it is not. The most important variable for explaining the last 130 years of the global temperature record is greenhouse gas increases. Solar irradiance changes inferred from solar cycle length changes are only important when considered in isolation. Although these results may be questioned, proponents of a large solar influence on the climate of the recent past have yet to give a convincing explanation as to why changes in the length of the solar cycle should affect solar irradiance.

Schlesinger and Ramankutty (1992) have extended their analyses to explain the paleoclimatic past (back to A.D. 1600) using greenhouse gas, solar cycle length and sulphate aerosol forcing. Despite problems with the uncertain history of past temperature changes and forcing, such studies, linking paleoclimatic data, forcing-factor history and both simple model and GCM integrations, are likely to become more frequent and important in the future.

34.6 Composite estimates of temperature change

By combining many of the reconstructions available in this volume with several others, Bradley and Jones (1993) produced a series of decadal average temperatures representative of "summer" temperatures for the Northern Hemisphere. The choice of series to include in the average was restricted both by data availability and by the need that each series reflect primarily summer temperatures, and be at least decadally resolved in time. The series which were averaged came from three main regions of the Northern Hemisphere: North America, Europe and eastern Asia. They were restricted also in latitudinal extent to the region 30°-80°N. The series were averaged in normalized form (all with respect to the ten decades, 1860-1959) because although all the individual reconstructions are clearly responsive to summer conditions, many have not been formally calibrated to produce values in degrees Celsius.

The resulting time series back to 1400 provides the best reconstruction of Northern Hemisphere "summer" conditions currently available. Many other review papers have intercompared reconstructions in the past (e.g. Williams and Wigley, 1983), but only one other composite time series for the last few centuries had been published (Groveman and Landsberg, 1979). However, this series combines a number of records that are poorly calibrated in terms of climate response, leading to a composite series that is difficult to interpret. All the series used by Bradley and Jones (1993) reflect summer conditions; furthermore, it was demonstrated that even with the limited number of records used, most of the decadal-scale variance of the overall northern hemisphere temperature over the last century was captured by the composite series. At the present time, some additional reconstructions could be included, but there are still only a few records which would improve coverage in the lower latitudes of the Northern Hemisphere. Potentially, greater strides could be made in producing a composite series for the Southern Hemisphere (not attempted by Bradley and Jones, 1993) since there are now several new series available. However, there are still less than half of the number of series from the Northern Hemisphere, and these are restricted to the 10°-45°S zone.

Overall, the Northern Hemisphere "summer" composite record (Figure 34.2) points to the period from about 1570 to 1730 as the coldest period of the last 500 years, followed by the nineteenth century. The decade 1600-1609 was exceptionally cold. Conditions comparable to the decades from the 1920s onward have not been experienced for several hundred years at least. The record can be viewed in two ways. On the one hand, one could interpret it as showing fluctuations around a lower frequency oscillation (shown schematically in Figure 34.3a) such

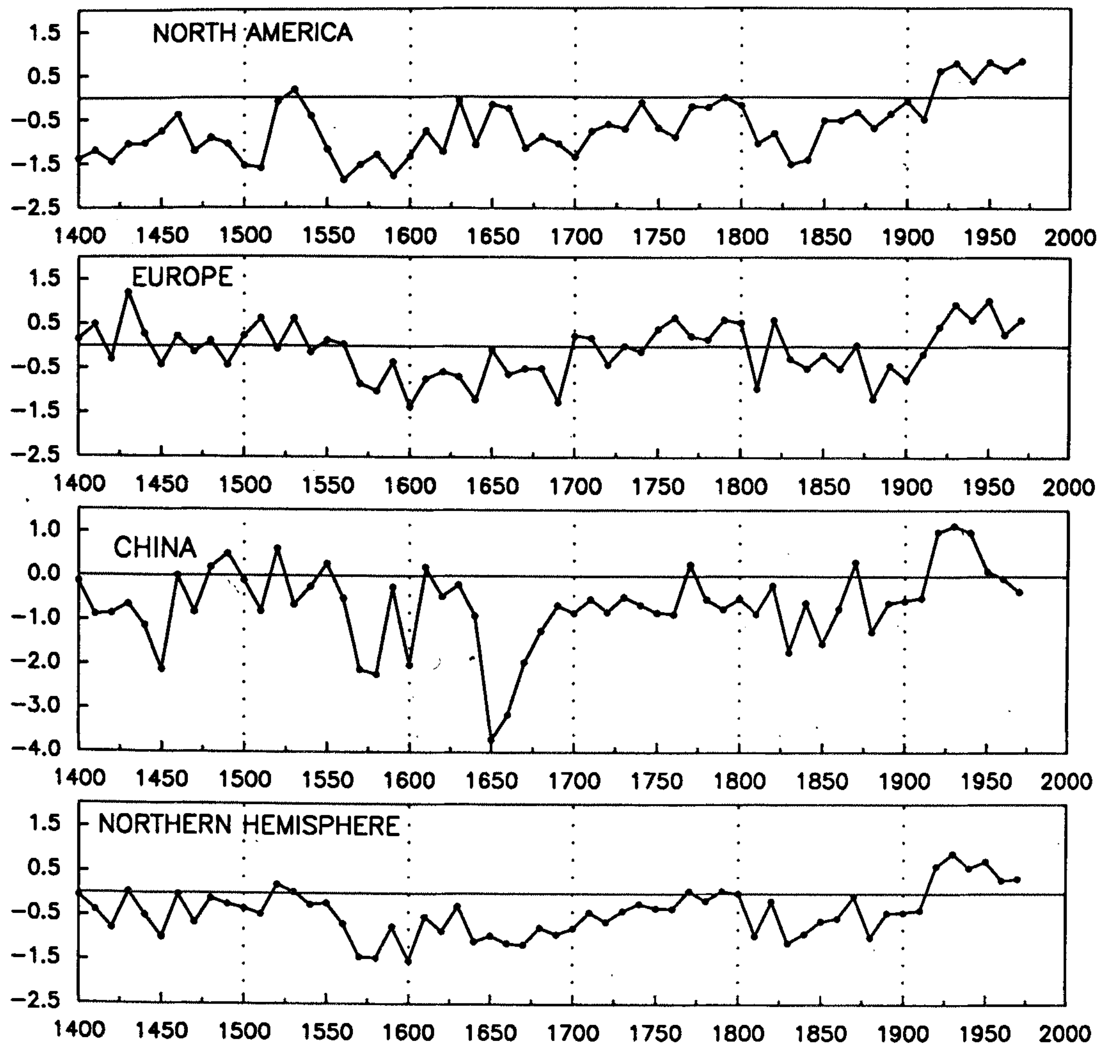

Figure 34.2 Composite summer temperature anomaly series for Europe, North America, East Asia and for the Northern Hemisphere as a whole. These series were derived from normalized decadally averaged anomalies (with reference to the period 1860-1959) (from Bradley and Jones, 1993).

that a gradual rise in temperature took place from about 1600, interrupted by cooler intervals in the nineteenth century. Alternatively, one could argue that temperatures fluctuated around a mean somewhat lower than the 1860-1959 average, punctuated by cooler intervals in the late sixteenth, seventeenth and nineteenth centuries, and then underwent "unprecedented" warming (in the context of the last 500 years) in the twentieth century (Figure 34.3b). Both perspectives provide fuel for arguments over the veracity of anthropogenic greenhouse-gas-induced climatic change. With the former interpretation, recent temperature anomalies would be viewed as normal fluctuations around an underlying low frequency periodicity or trend. With the latter interpretation, recent anomalies would be viewed as unique and probably the consequence of human-induced changes in greenhouse-gas concentrations. To resolve this controversy, further modelling and paleoclimate studies are needed. In particular, additional studies of climate *before* A.D. 1500 are needed to determine if temperatures were indeed higher over the globe (as opposed to just locally) as

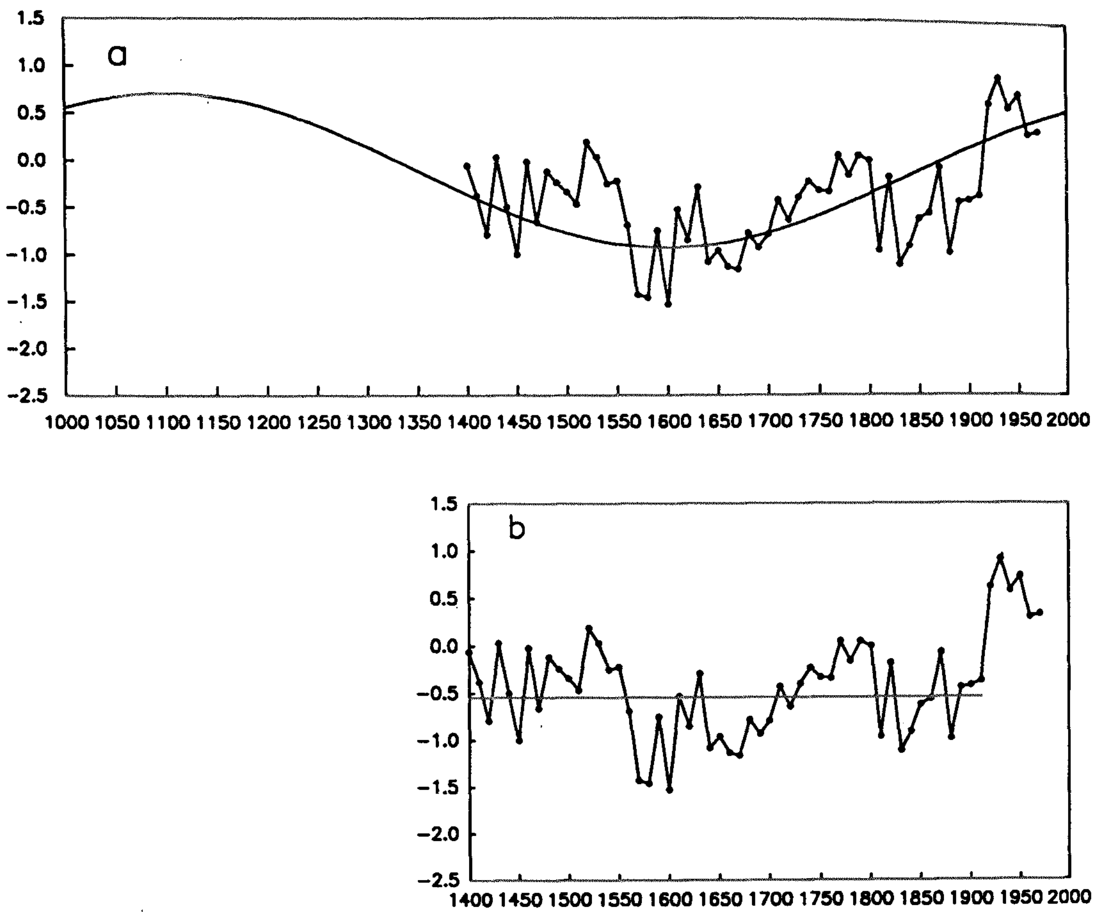

Figure 34.3 Northern Hemisphere normalized summer temperature anomaly series (from lower panel of Figure 34.2) plotted (a) in relation to a hypothetical low frequency variation with a period of ~1000 years, and (b) in relation to a hypothetical lower mean temperature (shown here as a horizontal line, which is the mean anomaly from A.D. 1400-1900). The recent warming is subject to different interpretations depending on the perspective selected (see discussion in text).

Figure 34.3a would imply. Thus, the nature of the so-called "Medieval Warm Period" (Lamb, 1965; Hughes and Diaz, 1994b; Briffa *et al.*, 1994a) assumes increased significance.

Acknowledgement

We thank Ed Cook for helpful comments on this chapter.

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