The History of Climate Dynamics in the Late Quaternary

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3.1 Introduction

Climate variability, defined as changes in integral properties of the atmosphere, is only one small realization of the workings of the much larger earth system. Parts of the other components (ice, ocean, continents) have much slower response times (decadal to millennia). True understanding of climate dynamics and prediction of future changes will come only with an understanding of the workings of the earth system as a whole, and over both the past and present time scales. Such understanding requires, as a first step, identification of the patterns of climate change on those time scales, and their relationships to known forcing. As a second step, models must be developed to simulate the evolution of the climate system on these same time scales. Within the last few decades, a significant number of long time series have become available that describe paleoclimate variability with resolution better than about 1000 years. Global general circulation models lag however, in that they have yet to be successfully integrated for more than a few hundred years. Because of these data and model limitations, the study of past climate change in the geologically recent past (the late Quaternary) has, for the most part, been limited to two basic strategies: (1) detailed description of the mean climate during specific climatic extremes (the "time slice" strategy), and (2) process modeling studies that seek to explain available time series in relation to specific external forcing and internal system dynamics. An excellent example of this latter strategy is the evaluation of the role of the insolation forcing as a driver of glacial/interglacial cycles by the SPECMAP group (Imbrie et al. 1992, Imbrie et al. 1993). Until recently there has been relatively little knowledge of past climatic variability on decadal to millennial time scales, a major problem considering that these are time scales on which climate is greatly affected by energy and material exchanges between the atmosphere, ocean, cryosphere, and biosphere.

The field has evolved rapidly in recent years. Some of the main areas of progress include:

- A much improved knowledge of the decadal variability of the surface ocean, ice and continental system over the last millennium.
- The development of an interhemispheric ice-core stratigraphy covering more than a full glacial cycle with a temporal resolution of about 100 to 1000 years, based on both ice parameters (δD, $\delta^{18}O$, dust) and concentrations and isotopic compositions of trapped atmospheric gases (see Chapter 2).
- The acquisition of oceanic time series with sufficient temporal resolution to allow the study of century-scale variability in surface and deep water properties over the last few glacial cycles. Of special importance in this regard have been the discovery and study of oceanic records that capture the millennial scale oscillations known from ice-core records as Dansgaard-Oeschger cycles (Dansgaard et al. 1984, Broecker et al. 1992, Dansgaard et al. 1993), and the ice-collapse episodes known as Heinrich events (after Heinrich 1988, Bond et al. 1992, Bond et al. 1993). Progress is being made on

a common stratigraphy for these events, but it is not yet global.

- The improved calibration of calendar time scales using ^{14}C and U/Th radiochronological methods, and comparison with incremental dating approaches such as annual layer counting. This enables the establishment of precise relationships between external forcing and climatic response, and direct comparisons between ice, ocean and continental paleoclimate records (Stuiver and Braziunas 1993, Sarnthein et al. 1994, Wang et al. 2001).
- The identification of linked chronostratigraphic markers in marine, terrestrial and ice records. The classic approach was based on the detection of specific volcanic ash layers. The newly developed high resolution reconstruction of the earth dipolar magnetic moment, NAPIS 75 (Laj et al. 2000) and the associated evolution of cosmogenic nucleides recorded in the ice (Baumgartner et al. 1998) offer the first possibility of truly global correlation at millennial scale resolution.
- The development of models of intermediate complexity. These can be integrated for thousands of years and facilitate numerical experiments and data-model comparisons that can help to identify key processes involved in past climatic changes (Berger et al. 1994, Rahmstorf 1995, Stocker 2000, Ganopolski and Rahmstorf 2001).

The first section of this chapter discusses climate dynamics on orbital time scales: sea level and glacial/interglacial cycles, monsoon variability, interhemispheric connections, low versus high latitude insolation forcing and the Last Glacial Maximum. The second covers millennial scale variability and associated climatic processes. The final section focuses on interannual to decadal variability. We make no attempt here to review the progress made on proxy development and quantification of local climatic and environmental responses to climate changes. Much of that work may be found in the special issue of Quaternary Science Reviews based on the first PAGES Open Science conference (Alverson and Oldfield 2000).

3.2 Climate change under orbital forcing

The seasonal and latitudinal distribution of energy received from the Sun is modulated by oscillations of the earth's orbital parameters. The major changes derive from precession of the equinoxes (at 19 and 23 ka/cycle) and changes in the eccentricity of the earth's orbit (main periodicities around 400 and 101 ka/cycle). High latitude summer insolation and mean annual insolation are also particularly sensitive to the changes in the Sun's elevation above the horizon (obliquity) which varies with a periodicity of 41 ka/cycle (Berger 1977, Laskar 1990). The record of past climate change thus provides key information about the sensitivity of the earth system to energy balance changes.

3.2.1 Developing a chronology of past climatic change

The study of the sensitivity of the earth's climate to insolation forcing requires a reliable chronology. The first timescale for Pleistocene glacial cycles was established by joint application of magnetostratigraphy and changes in the $\delta^{18}O$ in fossil *Foraminifera* in ocean sediments, a proxy for ice volume (Broecker and Donk 1970, Shackleton and Opdyke 1973). This early chronology linked reversals of the earth's magnetic field recorded by ocean sediments to those recorded in dated volcanic rocks, and showed that the main periods of orbital oscillation are apparent in $\delta^{18}O$ records. Direct links between high northern-latitude summer insolation and $\delta^{18}O$ based records of high sea stands of the last interglacial were dated by Broeker et al. (1968) using $^{234,238}U/^{230}Th$ analysis of Barbados corals. That study, among others, led Imbrie et al. (1984) to propose a revised chronology of the last 800 ka, obtained by tuning paleorecords to orbital frequencies (the so-called SPECMAP method of Imbrie et al. (1989) and Martinson et al. (1987)). The success of this method has been demonstrated by the reevaluation of the K/Ar dating of the last several reversals of the earth's dipolar magnetic field (Shackleton et al. 1990) and further refinements of the chronostratigraphy of the Pleistocene (Bassinot et al. 1994). Such orbital tuning of isotopic stratigraphy has a significant drawback for climatic studies. It presupposes that the interactions between the main components of the climatic system that react with response times similar to the orbital periods (thousand of years or longer) operate with constant phase lags or leads with respect to insolation (Imbrie et al. 1992). It is probable that this supposition will have to be relaxed in order to obtain a better understanding of the interactions that may occur between processes operating on different time scales, such as the influence of ice sheet extent on greenhouse gases and thermohaline circulation.

Uncertainties in interpretation, lack of precision in the reference series, and the presence of higher frequency variability generate intrinsic errors in orbitally tuned chronologies of about 1/4 of the precession period, or ±5 ka. Despite this relatively low resolution, the method has generated considerable progress in our understanding of long term climatic processes. Spectral and cross spectral analysis of orbitally tuned paleorecords have helped to improve the astronomical theory of climate and to elucidate the main interactions between slowly reacting climatic components (Imbrie et al. 1992, Imbrie et al. 1993). Indeed, the orbitally tuned oceanic time scale and its associated marine isotopic stages (MIS) remains the best chronostratigraphic time scale available for the last several million years. It is the reference for all long marine paleorecords and, by extension, for late Quaternary ice and continental records.

Numerous recent and ongoing studies have been developed to improve the absolute chronology of sea level changes, using $^{238,234}U/^{230}Th$ dating of coral reefs (Figure 3.1). Both benthic *Foraminifera* $\delta^{18}O$ records and analyses of other proxies from the same sediment cores may be linked to these chronological markers, because the growth and decay of continental ice sheets change the $\delta^{18}O$ of seawater and of the *Foraminifera* which grow in it. However, local changes in water temperature and salinity account for part of the foraminiferal $\delta^{18}O$ changes, and these effects are not currently independently estimated with sufficient precision (Adkins and Schrag 2001, Duplessy et al. 2001). Another cause of uncertainty derives from the U-series dating. Uranium can diffuse in and out of coral aragonite, a mineral that is also prone to dissolution and recrystallization if exposed to fresh water, which is usually the case given that most sampling is conducted above sea level. The geochemical community has yet to reach a consensus on the correct interpretation of U/Th ages when the $^{234}U/^{238}U$ differs significantly from the modern mean sea water value of 1.149. Such a discrepancy is evident for most coral samples older than 130,000 years. A final major uncertainty derives from local tectonic activity and isostatic readjustment to sea level changes. There are independent estimates for these motions for the last 10 to 20 ka (e.g. Bard et al. 1996, Lambeck and Chappell 2001) but they are much more difficult to constrain for older sea level records derived from raised coral deposits. The problem is especially crucial for the reconstruction of sea levels from glacial periods since most records that span such time intervals are obtained from areas with rapid vertical tectonic motions, such as the Huon peninsula (Chappell et al. 1996).

Fossil remains less than about 40,000 years old may also be directly dated by accelerator mass spectrometry of ^{14}C. Ka calibrations to calendar scales have been developed to correct for the changes with time in the initial amount of ^{14}C. With the help of tree ring measurements, an optimum resolution of 20 years has been obtained for the last 10 ka (Stuiver and Reimer 1993). Beyond this, in the absence of long tree-ring sequences, floating calibrations have been proposed based on radiocarbon dating of macrofossils in annually laminated lacustrine sediments (Goslar et al. 1995, Kitagawa and Plicht 1998) and marine sediments from the Cariaco trench (Hughen et al. 1998, 2000). Lower resolution marine calibration curves have been proposed for the Late Glacial period by comparing ^{14}C and U/Th dating of coral samples (Bard et al. 1990a, Bard et al. 1990b, Bard et al. 1993), or by applying model corrections derived from recorded changes in the earth's dipolar moment (Laj et al. 1996). The uncertainty is roughly 1 ka or slightly more for the period before 18 ka BP. It may exceed 2 to 3 ka around 35 ka BP, a time when the earth's dipolar magnetic moment was significantly smaller than at present (Laj et al. 2000). An additional uncertainty, which complicates precise comparison between oceanic and continental records, comes from the $^{14}C/^{12}C$ isotopic disequilibrium between the atmosphere and ocean surface water. Because of the size of the ocean carbon reservoir and its relatively slow rate of ventilation, this difference is presently a few hundred years to one thousand years. This is the so-called ventilation age, or reservoir effect of surface waters. Estimates of this age have been proposed for specific periods such as the Younger Dryas (Bard et al. 1994) and the last glacial period and deglaciation (Sikes et al. 2000, Siani et al. 2001) using as time markers volcanic ash layers which have been bracketed by ^{14}C AMS dates on continental organic matter and on *Foraminifera* from oceanic sediments. These comparisons indicate a much larger variability on the ventilation age in the past, between 300 and 2000 years, than has been generally considered.

Dating of Greenland ice is performed either by combining ice flow and accumulation models (Johnsen et al. 1992) or by counting of annual layers (Greenland Summit 1997). The GRIP and GISP2 timescales agree very well (within 200 years or so) back to the Bølling/Allerød transition. But this precision deteriorates rapidly for older parts of the records, due to uncertainties in the counting methods for GISP-2 and GRIP, in the accumulation rate model for GRIP, and irregularities in deposition between the two sites. The absolute precision of the Greenland ice chronostratigraphy is not better than a few ka during the period from 30 to 100 ka BP. Other uncertainties are introduced in the comparison of the ice and gas proxies, because the gas from a given level in the ice core is younger than the ice by several hundred years to a few ka, depending on accumulation rate (see Chapter 2).

The dating uncertainties discussed above are particularly important for the study of millennial scale global climate variability. Any detailed comparison of climatic significance must rely on independent stratigraphic and dating tools. Such tools do exist within a regional framework, and include detailed *Foraminifera* $\delta^{18}O$ records and AMS ^{14}C dating for ocean sediment records, identification of common regional climatic patterns in oceanic and continental (or ice) records, and characteristic stratigraphic markers of known ages (ash layers, magnetic anomalies). One good example of such a technique is the strategy developed for the comparison of Antarctic and Greenland ice records, using comparison of the changes in trapped air $\delta^{18}O$ (Bender et al. 1994) and CH_4 concentration from both cores (Blunier et al. 1998, Blunier and Brook 2001).

Nevertheless, much progress is still required in order to generate a common absolute time scale and improve on the current uncertainties of about ±100 yr for the last 10 ka, ±0.5-1 ka until 15 ka, ±1-2 ka until 30 ka and about ± 5 ka (inherent uncertainty within the SPECMAP time scale), for the more distant past.

3.2.2 Understanding glacial cycles

Glacial cycles are recorded in a wide range of oceanic, cryospheric and continental paleorecords (Figure 3.2). They present a high level of similarity and are, in most cases, significantly correlated with the $\delta^{18}O$ record of ice volume changes (Figure 3.1) and its large amplitude, approximately 100 ka period, over the last 400-800 ka. On this long time scale, global climate variability appears to be forced by high latitude Northern Hemisphere summer insolation with a major climatic feedback associated with the waxing and waning of the northern continental ice sheets as predicted 70 years ago by Milankovitch (1930) (Imbrie et al. 1992). Prior to 800 ka ago, 100 ka periodicity was not a dominant mode of variability. Rather 41 ka periodicity was predominant, with climate apparently responding linearly to insolation changes associated with variations in obliquity (Tiedemann et al. 1994). Climate modulation by the precession of the equinoxes (about 20 ka/cycle) is well recorded at low latitudes, in particular by proxies linked to the evolution of the monsoon and trade winds (Rossignol-Strick 1985, Prell and Kutzbach 1987, Imbrie et al. 1989, Bassinot et al. 1994, McIntyre and Molfino 1996, Beaufort et al. 1997). There is thus little doubt that the modulation of insolation by precession and obliquity plays a major role in climatic changes. The interrelations between low and high latitude processes in both hemispheres on these time scales is however still a matter of debate. One approach which has been recently developed is to study, with the same high resolution as for the recent past, older periods which are known to have had very different insolation characteristics. The period around 400-500 ka is a particularly interesting target, since the earth's orbital eccentricity was small, and the only significant modulation of insolation was through the oscillations of obliquity. Isotope Stage 11, at the end of that period, is a major interglacial interval and is prominent in most of the paleorecords (e.g. Rossignol-Strick et al. 1998, Droxler 2000).

Understanding the predominance of a 100 ka periodicity when most of the insolation forcing occurs in the precession and obliquity bands has been an important challenge of the last 10 years. It is generally accepted now that several processes may lead to this low frequency oscillation, either together or independently. These include non-linear threshold responses to insolation forcing from the continental ice sheet and other from climate components, including atmospheric greenhouse gases (Imbrie et al. 1993, Berger et al. 1994, Beaufort et al. 1997, Paillard 1998, Shackleton 2000). The strong non linearity of ice sheet dynamics during growth and decay is probably also involved in the large and rapid amplitude climatic oscillations associated with glacial terminations. These have been described in detail for the last deglaciation, but not for those before. Several independent studies based on coral terraces (Gallup et al. 1994) and ocean sediments (Henderson and Slowey 2000) suggest that the sea level rise associated with the penultimate deglaciation actually preceded its presumed Northern Hemisphere summer insolation forcing maximum by some 15 ka. However, these records are not dated with sufficiently high resolution to permit unambiguous interpretation. The majority of published results support the idea that during glacial terminations, sea level increased rapidly just after, or synchronously with the increase in Northern Hemisphere summer insolation.

For the penultimate termination, the mid-transition is about 5-7 ka prior to the insolation maximum, at 132-135 ka (Gallup et al. 1994, Stirling et al. 1995), compared to 129 ±5 ka in the SPECMAP age scale (Imbrie et al. 1984). There is also a significant lead at that time between the low latitude Devil's Hole (Nevada) vein calcite isotopic record and the global oceanic proxies (Winograd et al. 1988, Coplen et al. 1994). However, ongoing studies suggest an explanation for this lead. Kreitz et al. (2000) show that the proximal south-west California coast became warmer, with a slowing down (or an interruption) of the California current and associated coastal upwelling, before the end of each glacial cycle, prior to the glacial maximum.

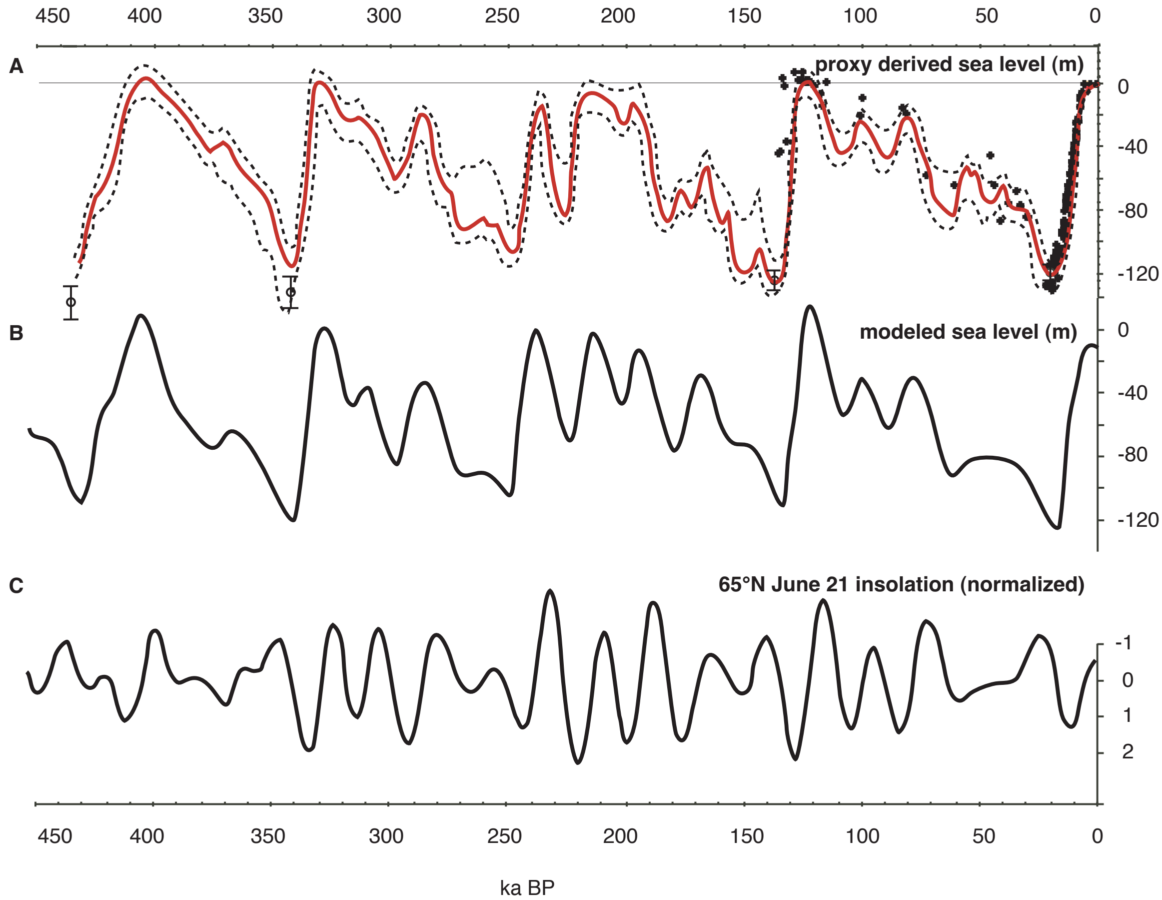

Fig. 3.1. Sea level change over four glacial cycles **A**. Relative Sea Level (RSL) over the last 450 kyr and U/Th dated estimates retrieved from coral terraces and ^{14}C-dated sediments from continental shelves (Bard et al. 1990, Stein et al. 1993, Zhu et al. 1993, Gallup et al. 1994, Stirling et al. 1995, Bard et al. 1996, Chappell et al. 1996, Stirling et al. 1998, Hanebuth et al. 2000, Yokohama et al. 2000, Yokohama et al. 2001). Open circles: RSL low stands estimated by (Rohling et al. 1998). Bold grey line and associated thin lines: composite RSL curve of (Waelbroeck et al. 2001) obtained after correction of the effect of deep water temperature changes on the benthic *Foraminifera* δ^{18}O records of marine-sediment core sites ODP 780 (North Atlantic, McManus et al. 1999) and MD94-101 (Southern Ocean, Gif data base), and a stack of ODP Site 677 and V19-30 for the Pacific Ocean (Shackleton et al. 1983, Shackleton et al. 1990) **B**. Sea level changes over the same time period, as estimated using a simple threshold function of the insolation (Paillard 1998). **C.** June 21 insolation at 65°N (Berger 1977) expressed as deviation to the mean insolation and scaled proportionally to the mean deviation**.**

Such changes may have resulted from a large southern expansion of the Alaskan Gyre during glacial maxima, associated with southerly winds along the California borderland, in agreement with the COHMAP simulation of the Last Glacial Maximum (COHMAP 1988). Such climatology would sufficiently affect the atmospheric hydrological cycle, at least on the regional scale, to explain the isotopic changes observed in the Devil's hole calcite prior to the termination.

Time dependent modeling of the glacial cycle has to date been undertaken primarily with relatively simplified models capable of long integration with relatively minimal computational requirements (Saltzman et al. 1984, Gallée et al. 1991, Gallée et al. 1992, Paillard 1998). One result common to many of these models is that prescribed carbon dioxide variability (Tarasov and Peltier 1997) or destabilization of large continental ice sheets, either through albedo changes (Gallée et al. 1992) or other ice sheet processes, is required to model a full glacial cycle. Improvements in the development of coupled models of intermediate complexity has led to a situation where modeling the full glacial cycle with somewhat more complex models, perhaps even with prognostic atmospheric CO_2, may become possible. Such studies will no doubt soon be published. General circulation modeling on these timescales has been, and continues to be, beyond the limits of computing power although asynchronous coupling schemes and other model simplifications will perhaps allow modeling of the glacial cycle with modified general circulation models in the next decade.

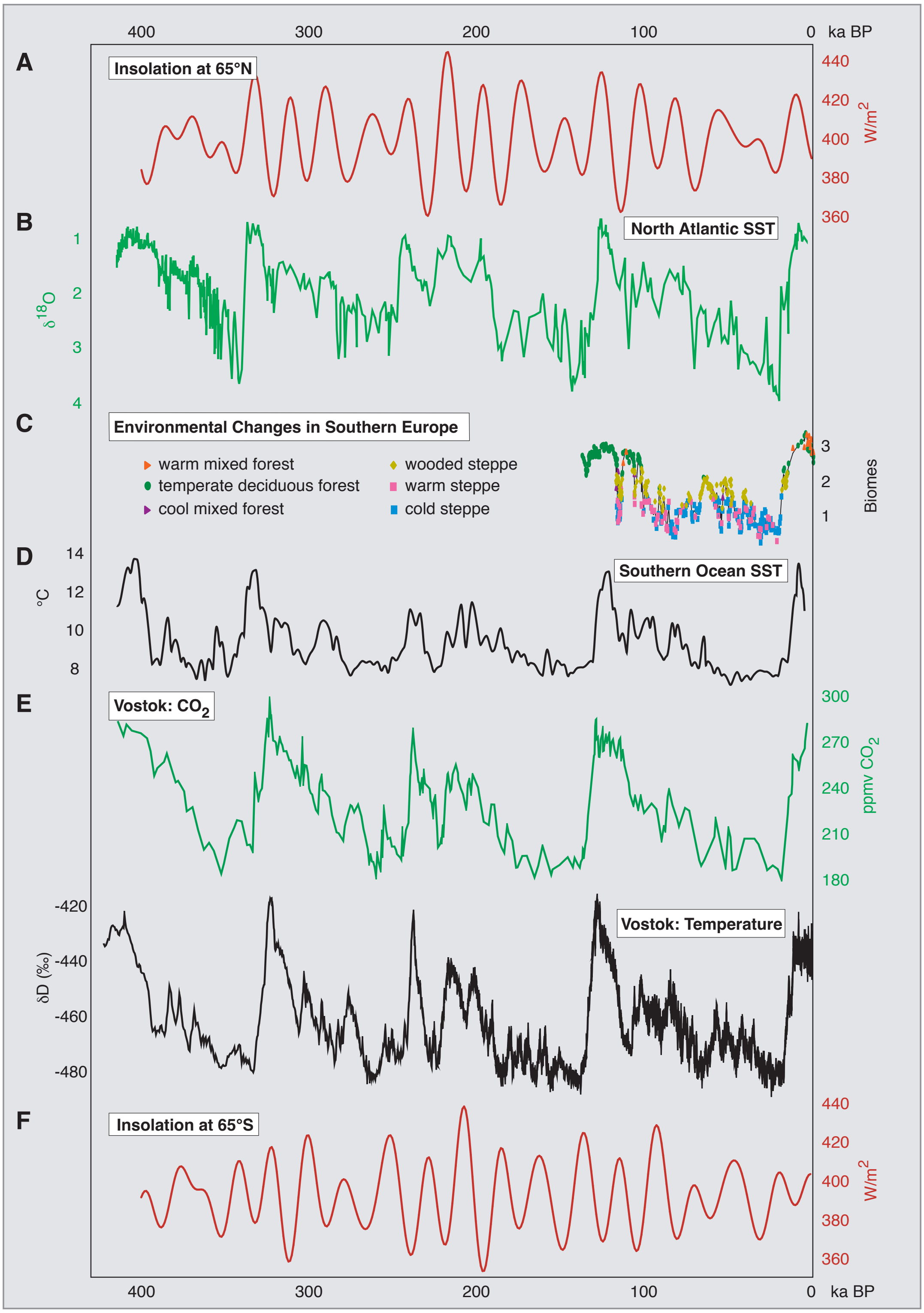

Fig. 3.2. Climatic proxies from both hemispheres over four glacial cycles. The last four glacial cycles as recorded in a variety of paleoclimate records. **A**. Summer insolation (21 June) at 65°N (Berger 1977). **B**. North Atlantic Sea Surface Temperature (SST) record of ODP site 980 (McManus et al. 1999). **C**. Allen et al. (1999) biomes record from lake Monticchio, southern Italy, over the last 100 ka. Biomes are statistical representations of vegetation derived from continental pollen records. **D**. South Indian ocean SST record from core MD 94-101 (Salvignac 1998). Temperatures are obtained by foraminiferal transfer functions after Imbrie and Kipp (1971). **E**. Vostok CO_2 and δD (expressed as temperature) records (Petit et al. 1999). **F**. Summer insolation (21 dec.) at 65°S (Berger 1977).

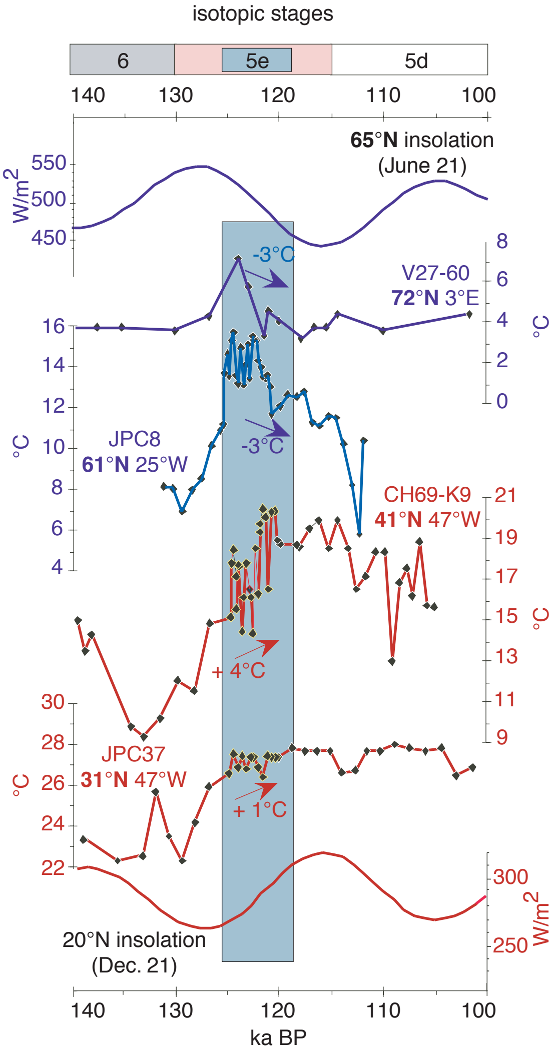

Fig. 3.3. The Last Interglacial (135 to 110 ka BP) in the North Atlantic and Norwegian Sea. High latitude June insolation and summer sea surface temperature versus absolute age for four ocean sediment cores located between 31°N and 72°N (Cortijo et al. 1999). Temperatures are obtained using the modern analog method (Prell 1985) with the North Atlantic data base from Pflaumann et al. (1996). Sea surface temperature and salinity began to decrease at high northern latitudes (72°N) simultaneously with decreases in high latitude summer insolation around 128 ka BP. However between 62 and 55°N, SST stayed high until about 118 ka BP, coincident with initiation of a major sea-level drop. During that period (and until the sea level rise associated with the next increase in insolation at 113 ka BP), lower latitude surface temperature (at 30-40°N) either did not change or increased slightly, suggesting a possible increase in meridional heat flux resulting from the increase in low latitude winter insolation (bottom curve).

3.2.3 Glacial inception

Uncertainties in absolute time scales do not preclude detailed studies of the chain of events. which follow insolation changes, as long as such studies rely on sufficiently precise relative dating based on global stratigraphies. Benthic *Foraminifera* $\delta^{18}O$ in ocean sediments (Martinson et al. 1987), $\delta^{18}O$ of the atmospheric oxygen extracted from ice cores (Sowers et al. 1993, Bender et al. 1994, Shackleton 2000) and methane content in the ice cores (Blunier et al. 1997, Blunier and Brook 2001) have proven their value in this respect.

Let us first consider the inception of the last glaciation, following the decrease in summer insolation at high Northern latitudes after 129 ka BP. Ruddiman (1979) was the first to propose that the pervasive warmth of the surface North Atlantic during periods of ice sheet growth provides a strong positive feedback to glacial growth by inducing precipitation over the ice sheets. Despite some controversy (Gallup et al. 1994, Henderson and Slowey 2000), the end of the high sea level stand is relatively well dated now at 118 ± 2 ka BP (Figure 3.2 and references therein), which constrains the chrono-stratigraphy of the associated changes in ocean $\delta^{18}O$. The suite of North Atlantic and Norwegian sea cores studied by Cortijo et al. (1997) has recently been completed and extended to a common temporal scale by Balbon (2000) (Figure 3.3). These cores show that high-latitude (Norwegian Sea) sea surface temperature (SST) decreased in parallel with summer insolation around 120 ka BP; lower latitude SST stayed warm during that period. These results provide evidence that high-latitude summer insolation controls the development of ice sheets. However, they also highlight the importance of latent heat transport to high latitudes during periods of maximum winter insolation at Northern Hemisphere low and mid latitudes. These periods correspond to the maximum latitudinal gradients in insolation, which should enhance transport of heat and moisture to high latitudes (Young and Bradley 1984, Rind 2000).

Khodri et al. (2001) recently published a 100 year long simulation in a fully coupled ocean-atmosphere general circulation model (GCM) forced with the 115 ka insolation, which supports these hypotheses. Their model produces a build-up of perennial snow cover for ice sheet growth in response to the orbital forcing, with a strong high latitude ocean sea-ice feedback and an increase in atmospheric moisture transport from the low latitudes. De Noblet et al. (1996) have shown, using an atmospheric GCM coupled to a global biome model, that changes in vegetation at high latitudes (forest to tundra) which accompany cooling may provide an additional important feedback enhancing snow accumulation.

It is interesting to compare the last interglacial results to the mean evolution of climate during the Holocene, during which high latitude northern summer insolation decreases and low latitude winter insolation increases. The amplitude of these changes is smaller than during the last interglacial, however, because the earth's orbital eccentricity is smaller now than it was 120 ka ago. Since 6 ka, high northern latitude temperatures have decreased (Johnsen et al. 1995, Koç et al. 1996, Bauch et al. in press) but no significant change has occurred in the tropical Atlantic (Rühlemann et al. 1999).

3.2.4 The Last Glacial Maximum

Describing the climatic state of surface oceans, continents and ice sheets during the Last Glacial Maximum (LGM) was the first large scale coordinated objective of the paleoclimatic community, as exemplified by the international CLIMAP project (Climate: Long-Range Investigation, Mapping, and Prediction (CLIMAP 1976, CLIMAP 1981). Sea surface temperature was derived from the statistical analysis of micro-fossil species distribution in sediments (Imbrie and Kipp 1971). The project also focused on chronostratigraphy to ensure reasonable temporal resolution of the Glacial Maximum (around 18 ^{14}C ka BP, or 21 calendar ka). The reconstructions were published as global maps of the distribution of sea surface temperature, sea ice, continental ice sheets and albedo (August and December) (CLIMAP 1981). One of the major conclusions was that most of the cooling during LGM occurred at high latitudes, with only small changes over the tropical oceans. Although since reevaluated in detail (Mix et al. 2002), the CLIMAP data set is still the basic reference for the understanding of ice-age climate. CLIMAP results have been used for both forcing and validating the recent intercomparison study of atmospheric GCM's and data, by the Paleoclimate Model Intercomparison Project (PMIP) (PMIP 2000). Two types of simulations have been carried out in these studies, the first forced with the CLIMAP global sea surface reconstruction (CLIMAP 1976, CLIMAP 1981) and the second using a coupled slab-ocean (Broccoli 2000). Significant differences exist between the two sets of simulations (Pinot et al. 1999, PMIP 2000). The coupled simulations predict surface ocean temperature and sea-ice distributions in relatively good agreement with recent reevaluations of the CLIMAP results. CLIMAP largely overestimated high- latitude summer sea ice coverage, and underestimated surface ocean temperature in the Northern Atlantic and Norwegian Sea (Sarnthein et al. 1994, 1995, De Vernal and Hillaire-Marcel 2000), as well as in the Southern Ocean (Crosta et al. 1998). Low-latitude sea surface temperatures may also have been over-estimated in the CLIMAP reconstructions. This is particularly apparent in the equatorial Atlantic and Pacific oceans, where recent SST hindcasts suggest LGM temperatures 2 to 5°C cooler than modern at the eastern side of the ocean basins, where equatorial currents interact with boundary currents (Hostetler and Mix 1999, Mix et al. 1999). For the subtropical gyres, however, recent results support the original CLIMAP inference of relatively unchanged SSTs compared to modern.

New proxies for reconstruction of sea surface temperature, such as ketone unsaturation ratios (Uk^{37}) (Brassell et al. 1986, Prahl and Wakeham 1987, Müller et al. 1998), or Mg/Ca ratios in foraminiferal shells (Elderfield and Ganssen 2000, Lea et al. 2000, Nürnberg et al. 2000) have not changed the overall picture significantly. The large amplitude (6°C) LGM cooling derived for Barbados from changes in coral Sr/Ca ratio (Guilderson et al. 1994) is not supported by these more recent studies.

Comparison of PMIP simulations with continental records for the LGM period are in progress (Kageyama et al. 1999, Joussaume and Taylor 2000, PMIP 2000, Kageyama et al. 2001). Often, however, the comparison between ocean and continental records is made difficult due to a lack of common chrono-stratigraphy. High resolution studies in ocean sediments show that the LGM was immediately preceded and followed by strong cold events (Heinrich events H-2 around 23 ka BP and H-1 around 17 ka BP). Similar variability is observed for lake levels in tropical Africa (Gasse 2000) and for the Indian (Leuschner and Sirocko 2000) and the East Asian monsoons (Wang et al. 2001). In low resolution records, an analogous series of events cannot be distinguished from a mean LGM climate. For this reason, a precise definition of the LGM period is important. It has been recently proposed by the EPILOG IMAGES Working Group that the LGM is best defined at 21±2 calendar ka BP (calendar scale) (Mix et al. 2002). This definition is in agreement with the sea level record of Yokohama et al. (2000) from Western Australia showing a –125 m relative sea level minimum between 22 and 19 ka. Because of a paucity of accurate dating or resolution, it remains difficult to integrate many records within the context of the high amplitude millennial scale climate variability surrounding the LGM period. These include observations such as the glacial lowering of snow lines (Klein et al. 1999, Porter 2001, Seltzer 2001) and ground temperature estimates derived from noble gas concentration in aquifers. The noble gas thermometry implies a mean annual air temperature decrease in northern Brazil of 5°C (Stute et al. 1995) at ~16-19 calendar ka, and 6.5°C in Oman (Weyhenmeyer et al. 2000), averaged over the period 16 to 27 ka BP. Both periods are dated with several ka uncertainty. Nevertheless, a general consensus exists that the LGM climate was not only much colder, but also, for the most part, more arid than the present day. Overall, currently available evidence suggests that intertropical areas probably cooled 1 to 3 °C in the surface ocean, and about 4 to 6°C at moderate altitude on the continents. These results are in agreement with recent fully coupled atmosphere-ocean simulations of the LGM climate. A drier continental LGM also explains the significantly lower atmospheric CH_4 concentrations observed in the Greenland and Antarctic ice records for that period (Chappellaz et al. 1993 and Chapter 2).

If we consider the glacial climate in more detail, using coupled GCM simulations, the system appears more complex and is harder to understand. Ganopolski et al. (1998) utilized a coupled ocean-atmosphere GCM of intermediate complexity (dynamical three-dimensional atmosphere of low spatial resolution coupled to a zonally averaged multi-basin ocean), and predicted a mean LGM cooling of about 2°C for the intertropical oceans (4.6°C for the continents). This was driven by sea-ice and a southward shift in deep-water formation, with no significant drop in the overall thermohaline transport. Weaver et al. (1998) using an oceanic GCM in equilibrium with an energy-moisture balance model for the atmosphere, also estimated a mean tropical ocean temperature LGM decrease by about 2.2°C which is consistent with a low to medium climate sensitivity to radiative perturbations. The large cooling over North America and the northern Atlantic is linked in this work to a large drop in the rate of North Atlantic deep water formation. Bush and Philander (1998) used a fully-coupled three–dimensional GCM configured for the LGM and found, in contradiction to the results of Ganopolski and Weaver, a large amplitude cooling of the tropical Pacific Ocean (6°C for the western Pacific) driven by enhanced trade winds, equatorial upwelling and equatorward flow of cold water in the thermocline. However, Bush and Philander (1998) limited their run to 15 years, thus the ocean was not in equilibrium with the atmosphere, except for low latitude surface waters.

Atmospheric circulation was strongly affected by glacial ice sheets over northern mid-latitudes. Model results suggest that an anticyclonic circulation developed over the ice sheets and that planetary waves were enhanced. Baroclinicity increased as a result of stronger meridional temperature gradients and storm tracks experienced an eastward shift especially over the North Atlantic (Kageyama et al. 1999). Such circulation changes played a key role in determining regional climate change patterns. For example, along the west coast of the Americas, an equatorward shift in the westerlies in both hemispheres during the LGM lead to relatively wetter conditions in regions that are quite dry in today's climate (Bradbury et al. 2001). The equatorial shift of the westerlies in the Southern Hemisphere is more controversial than that in the North. If it did occur, it may have been related to a northward shift of the winter sea ice belt (Crosta et al. 1998) and a coastal expansion of the Antarctic ice sheets over the continental margin, since additional forcing by large continental ice sheets must be excluded there. Models are able to reproduce the main regional trends over Eurasia except over western Europe where they suggest that conditions were relatively warm and wet. Reconstructions deduced from pollen data (Peyron et al. 1998, Kageyama et al. 2001) indicate cooler conditions suggesting that the models underestimate the meridional temperature gradient over Africa-Europe during the LGM.

In Greenland, it is generally accepted (since the work of Cuffey et al. (1995) and Dahl-Jensen et al. (1998) using bore-hole temperatures as an independent temperature proxy) that the LGM air temperature reconstruction derived from $\delta^{18}O$ on the basis of Dansgaard's (1964) spatial calibration underestimated the LGM cooling by about 10-15°C. The reasons for this are multiple, but for the most part are linked to changes in the sources, transport and seasonality of snow precipitation over the northern latitudes. These effects are probably smaller in Antarctica (Jouzel et al. 2000).

Over the last several decades, significant strides have been made in understanding the changes in ocean circulation during the last glacial period, using proxies that reflect the timescale of deep ocean ventilation and nutrient content. It has been known for more than 10 years (after the work by Boyle and Keigwin 1982, Curry and Lohmann 1983, Boyle and Keigwin 1985, Boyle and Keigwin 1987, Oppo and Fairbanks 1987, Curry et al. 1988, Duplessy et al. 1988) that the thermohaline circulation was significantly altered during the LGM. The tropical thermocline is thought to have been shallower during the Last Glacial Maximum, at least in the Western Atlantic (Slowey and Curry 1987), which helps to explain the large SST cooling observed in low latitude upwelling areas by Mix et al. (1999). Glacial North Atlantic Deep Water was located above, and not below, a deep-water equivalent of Modern Antarctic Intermediate Waters. A similar inversion of water masses with ventilated nutrient-poor waters above 2000 m, and generally poorly ventilated waters below was also present in both the Indian (Kallel et al. 1988) and Pacific oceans (Duplessy et al. 1988). Deep water temperatures, estimated by comparison of benthic *Foraminifera* $\delta^{18}O$ records from the Norwegian sea, North Atlantic and Pacific Oceans, were about 2 to 4°C colder (Labeyrie et al. 1987, Labeyrie et al. 1992). A similar cooling has been estimated, using the Mg/Ca ratio in ostracodes (Cronin et al. 2000). Benthic $\delta^{18}O$ records also provide constraints on the Glacial-Holocene changes in deep water $\delta^{18}O$ following the melting of continental ice and sea level increase (0.9‰ for the Atlantic and 1.1‰ for the Pacific oceans, Waelbroeck et al. 2001). These values are not significantly different from those derived by Schrag et al. (1996) and Adkins and Schrag (2001) from pore water $\delta^{18}O$ and chlorinity. However, LGM deep water salinity may be more difficult to extract from changes in water $\delta^{18}O$ than previously thought, as sea-ice formation (which does not fractionate $H_2{}^{18}O$ vs. $H_2{}^{16}O$) was probably a large source of salt for deep water production (Dokken and Jansen 1999, Adkins and Schrag 2001).

Several outstanding questions exist regarding the dynamics of ocean circulation at the LGM. For example some data imply a rate of meridional overturning at LGM similar to that of the present day (Yu et al. 1996). In addition, although proxy data inform us about the location of water masses at the LGM, it is difficult to quantify them in terms of the volume of meridional overturning or the quantum of meridional heat transfer (LeGrand and Wunsch 1995). In the modern ocean the wind driven circulation carries enormous amounts of heat into the subpolar latitudes, and in a generally more windy glacial period with a larger equator to pole temperature gradient, this heat transfer would have been enhanced irrespective of what was happening to the thermohaline circulation. The proxy that would best constrain rates of deep water ventilation is clearly ^{14}C, because of the inherent "clock" in its radioactive decay. Initial efforts to estimate overturning rates at the LGM from the ^{14}C difference between planktonic and benthic *Foraminifera* (Broecker et al. 1988) were hampered by relatively large uncertainties. ^{14}C measurements in deep sea corals (Adkins et al. 1998) are a relatively new proxy development and one which may provide the ability to resolve millennial scale changes in deep ocean ventilation rates (Boyle 2000), but few such data are yet available for the LGM period. Despite great progress in mapping the location of water masses at the LGM, there remains much uncertainty in quantifying the rate of deep water ventilation and meridional heat flux in the oceans during this period.

3.2.5 Glacial termination

Rapid terminations of glacial periods have attracted much attention: melting $54X10^6$ km^3 of continental ice (Yokohama et al. 2000) in about 10,000 years is not a small thing. Ice-core records, which include information on both atmospheric composition and temperature at high latitudes (Chapter 2) provide strong constraints on the possible mechanisms responsible for deglaciation. The increases in atmospheric greenhouse gases, CO_2 and CH_4, and of ice δD (a proxy for air temperature over Antarctica), occur quasi synchronously (within the available resolution) during each of the last four terminations in the Vostok ice record (Blunier et al. 1997, Petit et al. 1999, Pépin 2000, Pépin et al. in press). Long-term trends are difficult to separate from millennial variability, even across the last termination where in the Byrd record the timescale is more precise (Blunier et al. 1997, 1998). Thus, we will first present the general picture of the deglaciation, before discussing in more detail millenial variability and associated climate dynamics.

Warming started both in Greenland and Antarctica at about 23 ka BP (Blunier et al. 1997, Alley and Clark 1999, Alley 2000, Blunier and Brook 2001), in phase with the increase in Northern insolation (Figure 3.4), but prior to the LGM. The warming accentuated after 19 ka BP when the Byrd $\delta^{18}O$ signal drifted out of its glacial range of variability. The rise in sea level due to the melting of northern ice sheets started at 19 ka BP (Yokohama et al. 2000), but moved out of its glacial range only at 15 to 14 ka BP, as is the case for the Greenland ice $\delta^{18}O$ signal. Severinghaus and Brook (1999) precisely dated the corresponding warming at 14.7 ka in the GISP ice core, using as proxies the changes in nitrogen and argon isotopic ratios (see Chapter 2, Section 2.1). They were able to show that the CH_4 increase in the same air samples in fact lagged temperature by about 50 years. If CH_4 is interpreted as a proxy for warmer or wetter tropics, such regions could not have been the trigger for initial warming. Within ^{14}C ventilation age uncertainties, deglacial warming in the surface waters of the low-latitude Cariaco Basin off northern Venezuela has been shown to be synchronous with the 14.7 ka rise in temperature in Greenland (Hughen et al. 1996), implying a close coupling between climate change in the tropics and high latitudes of the Atlantic.

The period of most rapid sea level rise, meltwater pulse 1A, follows by a few hundred years the initiation of the rapid warming over Greenland (Fairbanks 1989, Bard et al. 1996). This meltwater pulse, which raised sea level by about 20 m between 14.2 and 13.8 ka BP, corresponds in fact to a

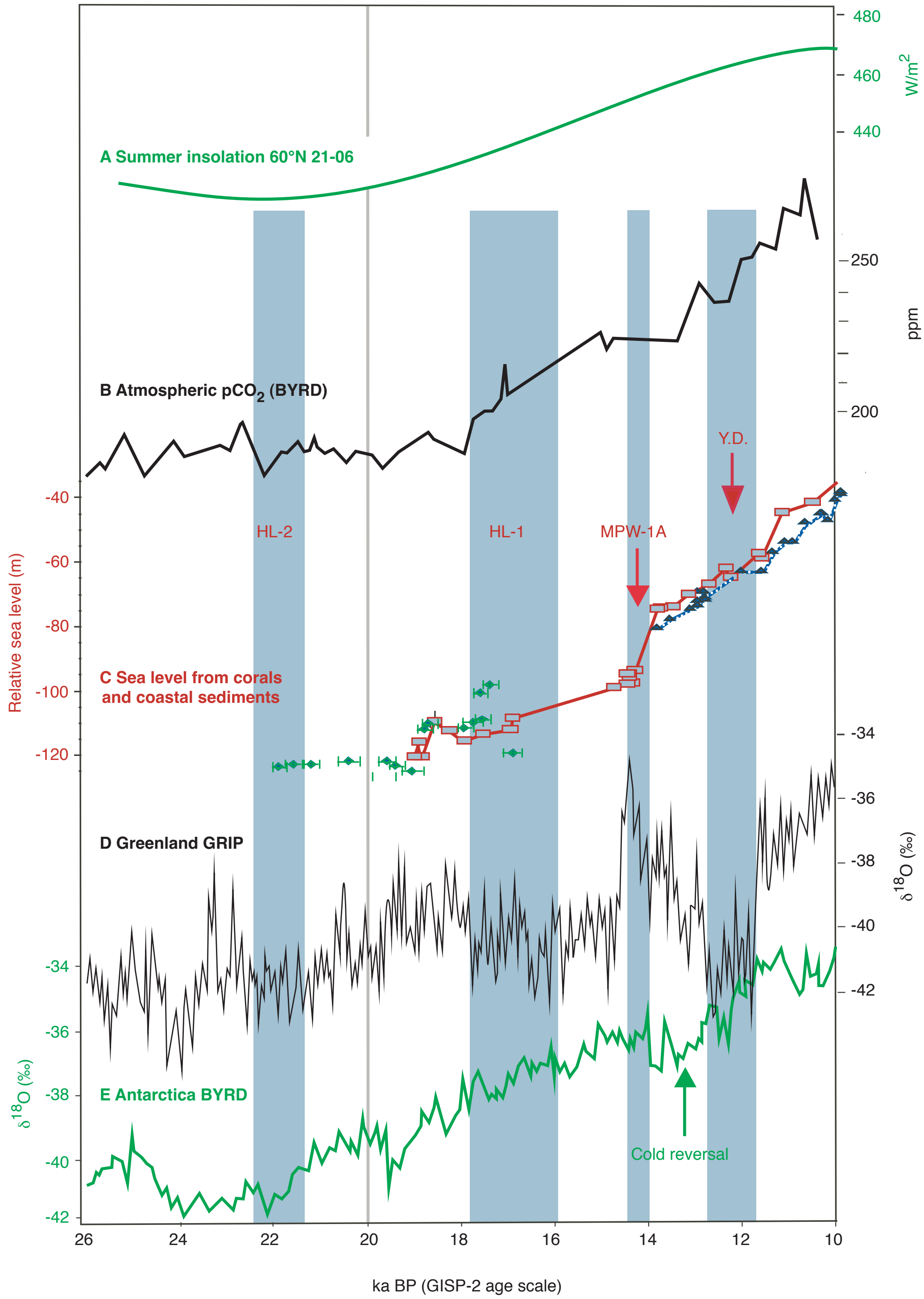

Fig. 3.4. Last deglaciation (25-10 ka BP) through various proxies. **A** Northern summer insolation (Berger 1977). **B** Atmospheric pCO_2 from the BYRD ice core (Blunier et al. 1998). **C** Relative sea level from Barbados coral (boxes) (Fairbanks 1989), Tahiti coral (triangles) (Bard et al. 1996), and sedimentary facies (diamonds) from North-West Australian shelf (Yokohama et al. 2001) and Sunda shelf (Hanebuth et al. 2000). **D** GRIP ice $\delta^{18}O$ (Johnsen et al. 1992). **E** Byrd ice $\delta^{18}O$ (Blunier et al. 1997). The GRIP and BYRD records are shown here on the GISP-2 time scale based on annual layer counting (Grootes et al. 1993, Meese et al. 1994, Stuiver et al. 1995, Grootes and Stuiver 1997). Correlation of the GISP-2 and GRIP $\delta^{18}O$ signals was done with Analyseries (Paillard et al. 1996), and transferred to BYRD using the GRIP-BYRD CH_4 stratigraphy of Blunier et al. (1997, 1998).

short cooling phase in Greenland (known as the older Dryas in Europe, Mangerud et al. 1974).

Temperatures from the Byrd record do not show significant inflexions during meltwater pulse 1A, but decrease later (the Antarctic Cold Reversal), during a slightly warmer phase at GRIP. The cold reversal terminated during the Younger Dryas, at about 12 ka BP. These events illustrate the typical asynchronous (or anti-correlated) North/South warm/cold modulations that appear to have occurred on the millennial scale over a large part of the of the last glacial period and deglaciation (Blunier and Brook 2001). Warming was for the most part complete by about 11.5 ka in Byrd. In the Northern Hemisphere, warming continued into the Holocene (10 ka or later). The northern Arctic seas, in particular the northern Norwegian (Koç et al. 1996) and Barents seas (Duplessy et al. 2001), became ice-free in summer, with progressive warming, only after 11-10 ka.

In North Atlantic surface temperature records, the drastic cooling that culminated in Heinrich event 1 (about 18 to 15.5 ka BP) was a dramatic interruption of the overall warming trend. The Younger Dryas (13 to 11 ka BP) had a similar effect. Available data and models suggest that these cooling phases were caused by catastrophic input of iceberg-derived meltwater over the northern Atlantic. This meltwater flux (about 0.5 to 1 Sv) induced a large decrease in thermohaline heat transport (Broecker et al. 1992, Paillard and Labeyrie 1994, Sarnthein et al. 1995) or a change in the zones of deep convection (Rahmstorf 1995). Active "modern" North Atlantic deep water production appears to have been initiated as late as about 15 ka BP (Bølling/Allerød) and persisted for about 1 ka before diminishing again during the Younger Dryas only to recommence at about 10 ka BP (Sarnthein et al. 1994, Marchitto et al. 1998). Thus, North Atlantic thermohaline circulation may have acted as a direct positive climate feedback only during the main Northern Hemisphere warming phases at ~15 and 10 ka BP.

The European continent reacted directly to these changes (e.g. Björk et al. 1998, von Grafenstein et al. 1998). In contrast, the loess series of China suggests that conditions actually became more humid there during the Younger Dryas interval, though temperatures remained low (Zhou et al. 1996). This probably resulted dynamically from an increased summer monsoon associated with the summer insolation maximum, but a more active polar vortex, with frequent cold northerly winds and active loess formation and transport (An and Porter 1997).

In the Atlantic and Indian sectors of the Southern ocean, surface temperature increases lead the benthic $\delta^{18}O$ record of sea level by 2 to 4 ka during the whole of the deglaciation (Labracherie et al. 1989, Labeyrie et al. 1996, Lemoine 1998). Poor constraints on the ventilation age introduce uncertainties of 0.5 to 1 ka in this estimate. Early warming is also indicated by continental southern hemisphere records. Pendall et al. (2001) recently published a well-dated peat-bog record (δD and pollen) from Patagonia (55°S) that shows climate changes similar to those seen in the Taylor Dome and Byrd records, including a well-defined early warming at about 17 ka. Southern Ocean deep water ventilation, as inferred from benthic foraminiferal $\delta^{13}C$ started very early during the deglaciation process (less than 1 kyr after the initial warming), thus providing evidence for an early renewal of convection and reestablishment of the active exchange between the deep Southern Ocean and the global ocean (Charles and Fairbanks 1992, Lemoine 1998).

Atmospheric pCO_2 presents characteristics similar to the $\delta^{18}O$ signal at Byrd Station, with a progressive increase after 19 ka, and a plateau between 15 and 13 ka BP, during the cold reversal. Such correlations support the idea that the Southern Ocean played a key role in atmospheric pCO_2 changes during the deglaciation (Blunier and Brook 2001 see Chapter 4).

3.3 Interaction among climate system components on millennial time scales

3.3.1 Millennial scale variability in proxy data: high latitude signals

When the rapid and large amplitude temperature oscillations recorded during the glacial period in the Camp Century Greenland ice core were first published (Dansgaard et al. 1984), they did not attract much interest in the climatological community. The delayed reaction (Broecker et al. 1992) occurred a few years after Heinrich's (1988) publication of his interpretation of the succession of sandy layers observed in a sediment core from the northeast Atlantic Ocean. Broecker and his colleagues interpreted these so-called Heinrich events as layers rich in ice-rafted detritus (IRD) that resulted from the catastrophic collapse of ice sheets into the North Atlantic via fast ice streams. These surges would have perturbed the hydrology of the North Atlantic, stopping the thermohaline conveyor belt and cooling regional climate. Simultaneously, Dansgaard et al. (1993) and Grootes et al. (1993) confirmed the presence of about 21 large-amplitude changes in air temperature, now called Dansgaard-Oeschger events (D-O), in the new GRIP and GISP2 Greenland ice cores. Bond et al. (1993) observed that each of the six Heinrich events identified from 60 to 15 ka BP occur at the end of a several kyr long cooling phase and appear to be simultaneous with the coldest of the Greenland stadial events. The (logical) hypothesis of synchronous cooling and warming over northern Atlantic and Greenland has been strengthened by the discovery of two peaks of cosmogenic ^{36}Cl within the GRIP ice core, one located between interstadials 10 and 8 and the other prior to interstadial stadial 6 (Figure 3.6). These correspond, in agreement with the proposed correlation, to the two periods of low-level paleomagnetic field intensity (the Laschamp and Mono lake events) recorded in North Atlantic sediments cores on each side of Heinrich event H-4 (Laj et al. 2000, Wagner et al. 2000). The presence of detrital carbonate (Bond et al. 1992) and very old (over 1 Gyr) detrital silicates (Huon and Jantschik 1993) within the IRD point to the Laurentide ice sheet as the major contributor of these sedimentary deposits. Geochemical studies confirm these results (Gwiazda et al. 1996, Revel et al. 1996) for all the Heinrich events of the last glacial period, save H-3 (around 30 ka BP). Grousset et al. (1993) and Gwiazda et al. (1996) have shown that this particular event probably derived from European or Greenland sources.

Labeyrie and his colleagues (Labeyrie et al. 1995, Cortijo et al. 1997, 2000) mapped the surface sea water $\delta^{18}O$ anomaly that resulted from melting icebergs during Heinrich events. From this they derived a rough figure for the ice volume change, which correspond to a meltwater flux of about 0.5 Sv, probably discontinuous over several hundred years, and an integrated sea level change of about 5 m. Available data from the Huon Peninsula (Lambeck and Chappell 2001) indicate that the sea level change may have been even larger, as much as 10-15 m during some of the Heinrich events. Such shifts correspond to both catastrophic collapses of the Laurentide ice sheet and additional input from grounded ice shelves destabilized by the initial sea level increase. North Atlantic deep-water ventilation significantly decreased during these events (Vidal et al. 1997). Ventilation of intermediate water may have increased in parallel (Marchitto et al. 1998).

Several authors have described the existence of a similar millenial variability at low latitudes in both the Northern (Little et al. 1997, Hendy and Kennett 1999, Sachs and Lehman 1999, Vidal et al. 1999, Peterson et al. 2000) and the Southern hemispheres (Charles et al. 1996, Kanfoush et al. 2000). Southern Hemisphere cooling episodes, possibly triggered by increased trade wind intensity (Little et al. 1997), are marked by an increased flux of ice rafted detritus between 41 and 53°S (Labeyrie et al. 1986, Kanfoush et al. 2000). They appear to have been approximately in phase with periods of warmth and active NADW formation in the Northern Hemisphere. In addition, Southern Hemisphere surface temperatures may have been warmer at the time the Northern Hemisphere was cold prior to and during Heinrich events. This asynchronous climatic behavior as recorded in ocean sediments is similar to that described in the Greenland and Antarctic ice cores (Blunier et al. 1998, Blunier and Brook 2001). However, unlike the ice records which are narrowly tied by their CH_4 signals, most of the millenial-scale oceanic and continental records currently available are not sufficiently well correlated from region to region nor with ice records to define precise interrelationships.

The cause of the ice sheet collapse associated with Heinrich events is still a matter of active debate between the proponents of internal ice-sheet dynamics (the binge-purge hypothesis of MacAyeal 1993), and those who favor an external origin (see discussion in Clarke et al. 1999). Since the interval between events decreases from about 10 kyr (between H5 and H4), to 5 kyr between H2 and H1, they clearly do not occur in direct response to insolation forcing, as originally hypothesized by Heinrich (1988) and McIntyre and Molfino (1996). However, taking into account the massive disruptions in the atmospheric and oceanic circulation linked to these events, it is evident that multiple positive and negative feedbacks must have been operating on a range of different time scales.

The Younger Dryas (YD), as discussed in the previous section, corresponds to the period at about 12 ka BP when Northern Atlantic temperatures returned to glacial levels for more than 1 kyr, despite the fact that Northern summer insolation was at its maximum. Broecker et al. (1989) suggested that the YD event signaled a major rerouting of the Laurentide meltwater from the Mississippi Delta, through which it was flowing until about 13 ka BP (Kennett and Shackleton 1975), to the St Lawrence estuary. The addition of meltwater near the sources of deep water formation would have directly affected the thermohaline circulation. However, De Vernal and her colleagues (1996) presented evidence that during the Younger Dryas, the St-Lawrence estuary was sea-ice covered most of the time, with very limited output of fresh water, and no indication of a significant meltwater spike. Broecker's hypothesis is therefore not clearly supported. Available data indicate, in fact, that continental ice melting decreased significantly during the YD (Fairbanks 1989). Andrews et al. (1995) attributed the YD cooling to a Heinrich-like event (H0). Interestingly, the characteristic sedimentary signature of Heinrich events (with detrital carbonate) is limited to the Labrador Sea at this time, and is not associated with a significant meltwater anomaly. It is therefore not at all proven that the YD corresponds to an ice sheet instability. However, as shown by Fairbanks (1989) and confirmed by Bard et al. (1996), the major meltwater period of the deglaciation occurred at 14 ka BP (Figure 3.4), at the beginning of the cooling that culminated in the YD proper, 1500 yr later. We may therefore imagine that a decoupling occurred, with an initial high latitude cooling linked to a decrease in thermohaline heat transfer, accumulation of snow, ice sheet regrowth, and subsequent collapse. The low latitude western Atlantic Ocean was warmer between 13 and 11 ka BP (which straddles the YD) than between 15 and 13 ka BP (the Bølling-Allerød) (Rühlemann et al. 1999). This accumulation of heat at low latitudes (also observed by Rühlemann and his colleagues between 16 and 15 ka BP) would be a consequence of lower thermohaline heat transport to the Northern Atlantic. But it would also be expected to promote enhanced evaporation and atmospheric water transport to the Laurentide and Fenno-Scandian ice sheets (Labeyrie 2000).

The origin of the relatively rapid D-O cycles is even less well understood than Heinrich events. Patterns of temporal variability with similar frequencies and durations (a few hundred to a few thousand years) have been recorded by numerous paleo-proxies over the Northern Hemisphere and low latitudes from both hemispheres (Grimm et al. 1993, Guiot et al. 1993, Chen et al. 1997, Hatté et al. 1998, Hatté et al. 2001, Wang et al. 2001) as well as in the oceans (Rasmussen et al. 1996a, 1996b, Curry and Oppo 1997, Moros et al. 1997, Kissel et al. 1998, Schulz et al. 1998, Wang and Oba 1998, Cacho et al. 1999, Hendy and Kennett 1999, Sachs and Lehman 1999, Tada et al. 1999, Peterson et al. 2000, Shackleton 2000, van Krevelt et al. 2000 among others). The signal is also clearly apparent in methane records from ice cores (Chapter 2). The characteristic signature of the Greenland ice $\delta^{18}O$ records with 21 large amplitude oscillations over the Last Glacial, each composed of a rapid shift to warm temperature (in few decades), and a slower cooling (several centuries) may help to identify the climatic processes that are directly linked to the D-O episodes. The rapid warming phases are especially significant, because any climatic mechanism, which operates with a rate constant longer than few decades, such as the dynamics of intermediate or deep ocean and ice sheets would have smoothed this signal. All the events and their characteristic temporal evolution are found in proxies which record some aspects of the Northern Hemisphere atmospheric circulation (Mayewski et al. 1994) and wind-driven surface ocean circulation (Peterson et al. 2000, Shackleton et al. 2000, van Krevelt et al. 2000). The new high resolution records of atmospheric CH_4 content (Blunier and Brook 2001) have the same signature, indicating that CH_4 is modulated by processes occurring on the millenial timescale on the Northern Hemisphere continents and possibly along continental margins (Kennett et al. 2000).

Thus, available data would point to large North-South oscillations of the North Atlantic Polar Front, of the westerly wind belt and maybe of the ITCZ and associated trade winds as direct modulators of the temperature over Greenland and the northern Atlantic (Peterson et al. 2000). The Fennoscandian and maybe other Arctic ice sheets were also affected. In sediment cores from the northern North Atlantic, Southern Norwegian and Greenland seas, the amount of ice rafted detritus (IRD) from Scandinavian sources increased (Blamart et al. 1999) and the foraminiferal $\delta^{18}O$ decreased (acting as a tracer of continental ice meltwater (Labeyrie et al. 1995) in apparent phase with each cold stadial (Bond and Lotti 1995, Rasmussen et al. 1996, Rasmussen et al. 1996, Elliot et al. 1998, Vidal et al. 1998, Dokken and Jansen 1999, Grousset et al. 2000, van Krevelt et al. 2000).

We suggest from these observations a simple cause and effect relationship whereby D-O oscillations result from a direct coupling between atmospheric circulation, coastal ice sheets and ice shelves. During interstadials, when relatively warm waters (10-12°C summer sea surface temperature) reached as far as 60°N (Manthé 1998, van Krevelt et al. 2000), heavy snow would have fallen over Iceland, Scandinavia and Southern Greenland, causing fast growth of the ice sheet periphery, development of ice shelves, and expansion of the polar vortex. This would correspond to the progressive cooling phase towards a stadial, and to the southern shift of polar waters to about 45°N (Shackleton et al. 2000b). In parallel, snowfall over the ice sheets would have decreased, while active ice streams would have eroded coastal ice sheets, in turn leading to the disappearance of the ice shelves, the decay of the polar vortex, and the start of a new cycle with the rapid warming of the high latitude ocean.

Recent results may help to establish possible links between Heinrich events and D-O oscillations. IRD and isotope records from the Norwegian Sea (Rasmussen et al. 1996a, 1996b, Dokken and Jansen 1999) and northern North Atlantic (Elliot et al. 1998) show a typical succession of meltwater events with peaks in IRD from the Arctic ice sheets (Blamart et al. 1999) that follow the rhythm of the

D-O events. Magnetic susceptibility measurements on these cores present a temporal succession which is very similar to the Greenland $\delta^{18}O$ record. This signal is interpreted as resulting from the formation and transport of nepheloid layers formed by resuspended magnetic microparticles along the Norwegian Sea volcanic ridges during periods of active deep water convection and thermohaline circulation (Kissel et al. 1999). There is no direct indication of the influence of Heinrich events in these records. By contrast, in the mid-latitude North Atlantic Ocean, both Heinrich and D-O oscillations appear superposed, each with their own typical rhythm and morphology (Bond et al. 1992, Grousset et al. 1993, Weeks et al. 1995, Grousset et al. 2000), as if the longer scale oscillations of the Laurentide and the shorter scale oscillations of the Fennoscandia and other Arctic ice sheets were operating, at least in part, independently. Yet, there is evidence for interrelationships. A major connection may be found by comparing the millennial variability signals, recorded in the Greenland and Antarctic ice cores. A correlation between the GISP-2 and Byrd ice records is now available (Blunier and Brook 2001), using higher resolution atmospheric CH_4 records than those published in Blunier et al. (1998). During each of the major Northern Hemisphere D-O coolings (those corresponding to the Heinrich events), air temperature over Antarctica gradually warmed, peaking exactly at the end of the Greenland cold stadials. The following cooling reached a minimum approximately in phase with the beginning of the next large stadial in Greenland, after which Antarctic air warmed again as part of the next cycle. Such oppposite temperature trends in the Northern and Southern Hemisphere records is also seen for D-O 20 and 21, at the transition between MIS 5.1 and 4 (at about 70-80 ka BP) (Blunier and Brook 2001), as well as for the YD. These contrasting trends may result from a shift between a thermohaline circulation regime similar to the modern one, with active NADW formation and transfer of heat from the Southern to the Northern Hemisphere, and a regime with enhanced deep water formation in the high southern latitudes. This is the so-called bi-polar seesaw hypothesis of Broecker (1998) and Stocker (1998). Meltwater excess at high northern latitudes (in particular during Heinrich events) would slow down NADW formation and activate deep water transport from the south (Figure 3.5). At the end of the Heinrich event, thermohaline circulation would start again in the north, bringing warm waters near ice-sheets, thus facilitating snow accumulation and rapid ice sheet growth until the next Heinrich event. This hypothesis is also supported by the results of Marchitto et al. (1998) who observed in a set of North Atlantic cores from different water depths that during deglaciation and the YD, the nutrient contents of intermediate and deep waters evolved in opposition. The YD was associated with a shallow-nutrient-poor "NADW-like" intermediate water, and nutrient rich "AABW-like" deep waters. The reverse is true for the warm Bølling/Allerød period.

Interestingly, Shackleton and colleagues (2000) recently published a study of an IMAGES high resolution core from about 2000 m depth on the Iberian margin (MD95-2042), which shows in its planktic *Foraminifera* $\delta^{18}O$ record the typical signature of the D-O events, but in its benthic *Foraminifera* $\delta^{18}O$ record precisely the same signal that is recorded in the Antarctic ice (Figure 3.6). The planktic vs benthic foraminiferal oxygen isotope profiles from that sediment core present the same phase relationship as the Greenland vs Antarctic ice records discussed by Blunier et al. (1998). The millenial variability of the Vostok $\delta^{18}O$ is thus apparently a signal of global significance. This provides a strong argument in support of the hypothesis that the millenial variability in $\delta^{18}O$ in Vostok (or Byrd) ice is linked to oscillations in interhemispheric thermohaline heat transport.

3.3.2 Millenial variability of climate at low latitudes

The fundamental direct response of the Asian and African monsoons to changing seasonality of insolation forcing has been well known for nearly two decades. When perihelion coincides with summer, as it did in the Northern Hemisphere 11,000 years ago, the seasonal insolation contrast increases, and monsoonal circulation intensifies. Across North Africa and monsoon Asia, reconstructed lake levels, vegetation, and lake chemistry support the inference of a much-intensified monsoon in the early Holocene (e.g. Street and Grove 1979, Gasse and Van Campo 1994, Lamb et al. 1995, Bradbury in press); see summaries in Winkler and Wang (1993), Overpeck et al. (1996) and Gasse (2000). Paleoceanographic records from the northern Indian Ocean and eastern Mediterranean complement this picture with evidence of increased monsoon-related upwelling and enhanced riverine deposition of terrigenous material and diagnostic pollen types during this period (Rossignol-Strick 1983, Prell 1984, Prell and Campo 1986, Overpeck et al. 1996). The driving force behind these large changes (increased seasonality of solar radiation), has been confirmed with a series of GCM studies (Kutzbach 1981, Kutzbach and Otto-Bliesner 1982, Kutzbach and Street-Perrott 1985). Although the response of the monsoon to orbital forcing was first characterized

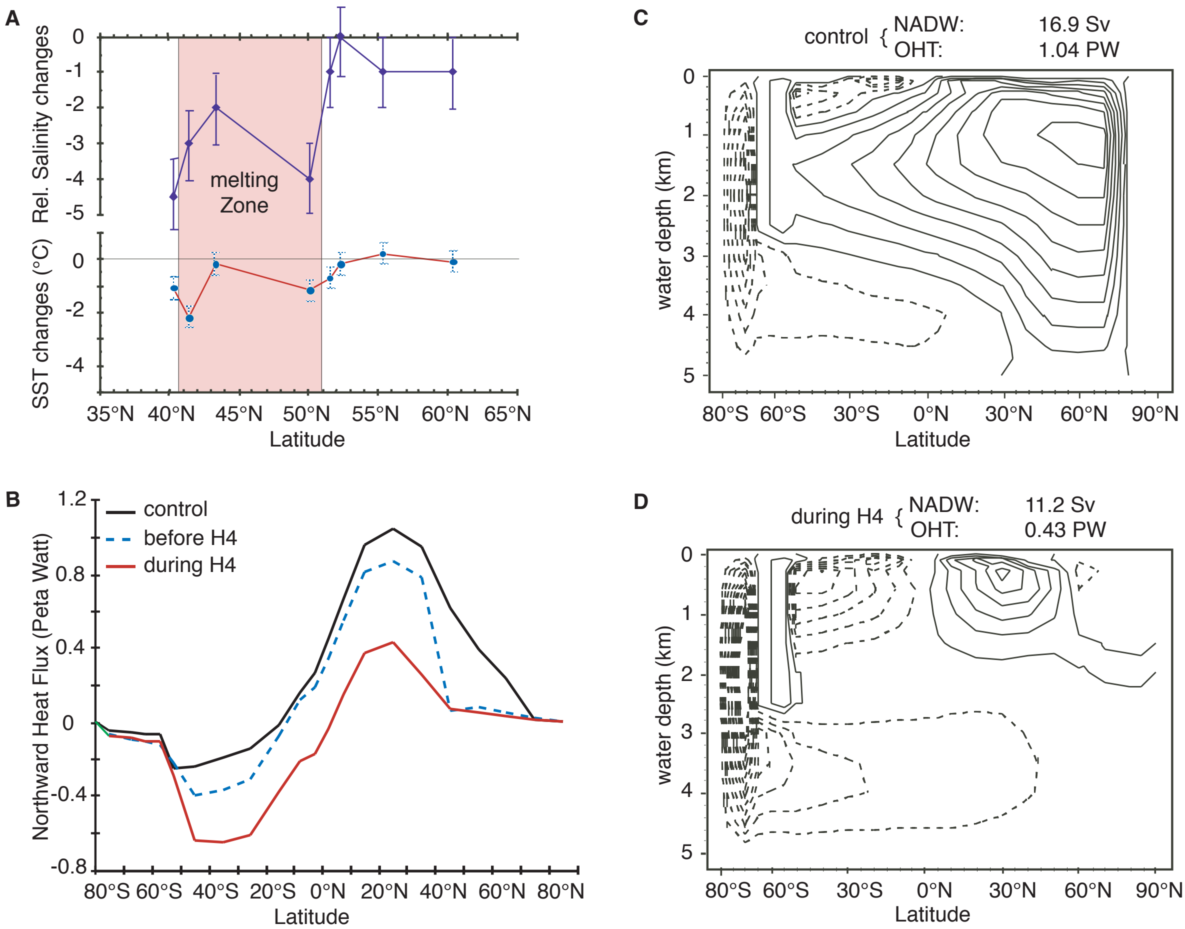

Fig. 3.5. 2-D zonal model simulation of the effect of Heinrich event H4 on thermohaline circulation adapted from (Paillard and Cortijo 1999). **A**. Zonal mean of the anomaly (H4 - beforeH4) for Sea Surface Salinity and Sea Surface Temperature, versus latitude. These data are used as input for the model. **B.** Changes in modeled northward oceanic heat transport for the control run, before H4 and during H4. **C.** Vertical distribution of the ocean heat fluxes for the control run, versus latitude. Flux of North Atlantic deep water (NADW) at its maximum is 16.9 Sv and oceanic heat transport (OHT) 1.04 PW. **D.** During H4: The model predicts a lower northern heat flux (by more that 50%) with a smaller NADW flux (11.2 Sv) and much shallower thermohaline circulation. Low latitude heat is preferentially transported to the Southern Hemisphere.

as linear (Prell and Kutzbach 1987), records from terrestrial and marine systems that have greater resolution, better chronological control, and improved spatial coverage demonstrate that the monsoon does not respond gradually to gradual insolation forcing. The warm and wet monsoon maximum in N. Africa and West Asia (about 13 to 6 ka BP) is interrupted at least twice by dry and/or cool spells (Younger Dryas 12-10 ka BP and 8.5-8 ka BP). The Mediterranean Sea hydrology, also influenced by the evolution of the African monsoon, similarly presents a large deficit in evaporation (and/or excess in precipitation) peaking at 9 ± 0.5 ka BP and 7.5 ±0.5 ka BP, and a drier interval between 8.5 to 8 k yr BP (Mercone et al. 2000). The precipitation excess may have disrupted deep-water ventilation of the Mediterranean Sea, and contributed to the development of anoxic bottom waters (Sapropel S1) (Rossignol-Strick 1985, Fontugne et al. 1994, Kallel et al. 1997, Calvert and Fontugne 2001).

At about the same time, an abrupt cooling affected a large part of the Northern Hemisphere, the so-called "8.2 ka event". This is the only notable event in the Greenland ice isotopic record for that period (Alley et al. 1997), and is recorded as a 1.5‰ negative shift in ice $\delta^{18}O$ (equivalent to a 4-8°C drop in air temperature) which lasted for about 200 yr. A cold event of similar timing is recorded in the Lake Ammersee (Germany) record of von Grafenstein et al. (1998), and as a color shift (interpreted as drier and cold climate with stronger trade winds) in the Cariaco Basin record of Hughen et al. (1996). An increase in IRD deposits occurred at that time

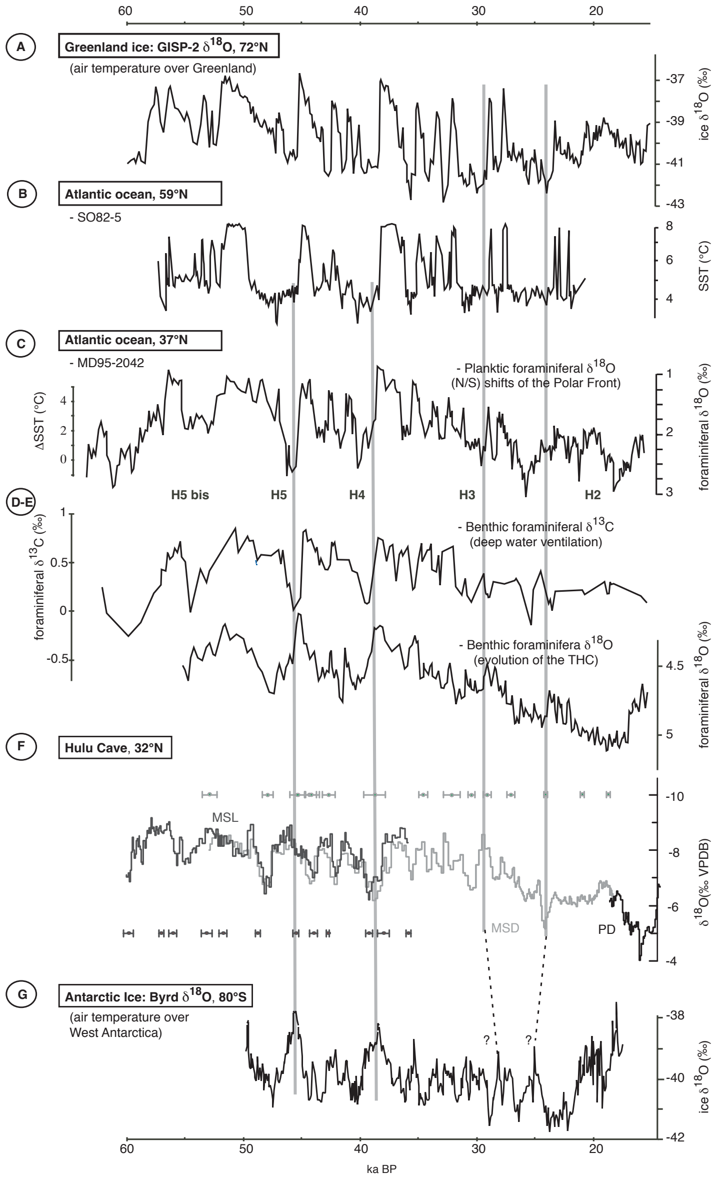

Fig. 3.6. Millenial variability during the Last Glacial period (15 to 60 ka BP). Ocean sediment time scales (calendar ka) are derived from AMS ^{14}C dating and correlation with the GISP-2 $\delta^{18}O$ record. Timing of the large meltwater Heinrich events is indicated by the labels H2 to H5, while the vertical grey lines mark the initiation of each subsequent warming in the North Atlantic system. **A.** GISP ice $\delta^{18}O$ record, a proxy for air temperature over Greenland (Grootes and Stuiver 1997). **B.** North Atlantic SST record from SO82-5 (59°31°W) (van Krevelt et al. 2000), derived from planktonic foraminiferal counts by the SIMMAX modern analog method (Pflaumann et al. 1996). **C.** $\delta^{18}O$ of planktic *Foraminifera* from IMAGES core MD95-2042 (37°48'N 10°10'W 3100 m depth) along the Iberian margin (Shackleton et al. 2000). The corresponding surface water temperature changes are plotted relative to the left axis. Signals **B** and **C** record the North-South oscillations of the surface hydrological system (in particular the polar front) **D.** $\delta^{13}C$ (7 pt Gaussian smoothing) of the benthic *Foraminifera* (C. wuellerstorfi) from the same core, a proxy for deep water ventilation (Shackleton et al. 2000). This signal reproduces, with more noise, the $\delta^{18}O$ of the planktonic *Foraminifera*, thus linking deepwater ventilation and surface hydrology in the North Atlantic. **E.** $\delta^{18}O$ of the benthic *Foraminifera* in the same core (Shackleton et al. 2000), which tracks the temperature and salinity changes of deep water at the core location. **F.** speleotheme $\delta^{18}O$ record from Hulu Cave (East China, 32°N), a proxy for the relative proportion of precipitation derived from winter vs. summer monsoon (Wang et al. 2001). The record is dated by $^{234}U/^{230}Th$ disequilibrium directly in calendar years. **G.** Byrd ice core $\delta^{18}O$, a proxy for air temperature over West Antarctica (Blunier et al. 1998, Blunier and Brook 2001),placed in the GISP-2 time cale by correlation of the CH_4 records. The good analogy between the Chinese monsoon record F and the North Atlantic records A to C supports the idea of a general northern hemisphere reaction to the events occuring around North Atlantic. The so-called North/South <anti-phasing> of the climatic response during the main Heinrich events, which is apparent between the Greenland and Antarctic ice records **A** and **G** is also well apparent within Atlantic core MD95-2042 between the planktic **C** and benthic **E** $\delta^{18}O$ signals. The strong analogy between records **E** and **G** suggests a common causal relationship for the deep ocean and Antarctic records, linked to oscillations of the THC (see sections 2.1 and 2.2) and initiated, about 1 ka before, by the changes in North Atlantic surface hydrology (records **A** to **C**).

in the Northern Atlantic Ocean (South of Iceland to the Newfoundland margin) (Bond et al. 1997, Labeyrie et al. 1999), which probably derived from increased transport of icebergs from northern Greenland and other remnant ice sheets from the Arctic periphery. The atmosphere-surface ocean system was thus affected over much of the North Atlantic Ocean and surrounding continents in a similar way (although with a lower amplitude), than during the Younger Dryas period. There is no direct indication for an associated meltwater spike from the ocean sediment records of the Northern Atlantic. Yet, Barber et al. (1999) have shown evidence supporting a catastrophic drainage of glacial lakes Agassiz and Ojibway at about 8.47 ka BP, just prior to the cold event. That result supports the hypothesis that, once more, a freshwater anomaly could have caused a major breakdown of the North Atlantic thermohaline circulation and a major cold spell that had global effects. The 8.2 ka event appears in the available northern Atlantic records as one of several periods during the Holocene when low-salinity polar water and the iceberg melting zone penetrated southward (Duplessy et al. 1993, Bond et al. 1997, Labeyrie et al. 1999).

These results support the idea that low latitude phenomena such as monsoons are sensitive to the interaction between gradually changing insolation and high-latitude changes (Overpeck et al. 1996, Sirocko et al. 1996, Gasse 2000) and surface feedbacks such as vegetation (De Menocal et al. 2000) and SST (Gasse 2000). A recent synthesis of well-dated, high-resolution monsoon records (Morrill et al. submitted) identifies statistically significant monsoon transitions at 1300, 4500-5000, 11,500, and 13,500 years before present. The earliest two shifts likely relate to North Atlantic changes during deglaciation. The mid-Holocene shift would relate more directly to the non-linear interaction between surface ocean, the hydrological cycle and continental albedo (with its vegetation feedback) to changes in insolation (De Menocal et al. 2000). The more recent shift remains unexplained. Climate model studies of varying complexity support the idea that surface ocean and vegetation feedbacks can add nonlinearity to the response of the monsoon to insolation forcing (De Menocal and Rind 1993, Overpeck et al. 1996, Kutzbach and Liu 1997, Claussen et al. 1999). The abrupt shifts in monsoon intensity that result from these nonlinearities can occur within centuries or even decades. In some cases, paleoclimate records of the terrestrial hydrological cycle may themselves reflect threshold effects. For example, dust records in deep-sea cores may reflect the migration of source areas or the changing location of a particularly effective delivery vector (river plume or wind belt), rather than regional climate patterns. Lake sediment records may likewise show threshold effects, as a lake nears desiccation or reaches a depth at which its chemistry, biota, or sedimentation regime can change abruptly. Despite potential uncertainties in individual records, regional changes in moisture balance can be clearly seen when many individual records are pooled. There are clear examples of regional changes in moisture balance that do not correspond to any obvious forcings, yet are clearly expressed. For example, widespread dry events occur in Asia/Africa around 7ky and 4.5ky BP (summarized in Gasse 2000).

Despite a plethora of new results (only partially presented here), we still lack the temporal resolution and chronological constraints necessary for a full understanding of the processes involved in these oscillations and abrupt climatic changes. Our understanding will increase dramatically with the availability of better continental and marine records, as illustrated by the exceptional isotopic record compiled from speleothems from Hulu cave (near Shanghai, China) recently published by Wang et al. (2001). The record comprises long annually-banded series and spans the last 10-70 ka period with century-scale resolution, with time control provided by more than 50 U/Th dates (Wang et al. 2001), The overall morphology of this record mimics perfectly the GISP, GRIP or the North Atlantic MD95-2042 records of Shackleton et al. (2000). The relative intensity of the summer and winter East Asian monsoons appear to have oscillated over the whole period in parallel with the changes in North Atlantic climate, with each cold North Atlantic event being associated with a proportional reinforcement of the winter Asian monsoon. But what are the climatic links that explain such direct connections? It is evident that the repeated succession of large amplitude, rapid climatic fluctuations in comparative studies of this type can provide a wealth of information relevant to understanding and quantifying the inter-relationships between climatic and environmental changes. This field is in its infancy, and will no doubt offer many surprises in the coming years.

3.3.3 Modeling millennial scale climate variability

The rapidly expanding suite of high-resolution paleoclimatic records that exhibit millennial-scale variability discussed above has provided an enormous stimulus to numerical modeling efforts over the last decade. While model simulations exist that contain variability on this time scale, it should be noted that no model has yet been shown capable of generating millennial variability with the characteristic temporal structure seen in records such as the isotopic temperature from the Greenland ice cores or high-resolution marine sediments. One reason for this is that the time scale typical for the recurrence of D-O events (about 1-2 thousand years) or Heinrich events (5-7 thousand years) is longer than the characteristic time scales of the ocean-atmosphere system. Although the renewal time associated with the deep circulation in the global ocean, about 1500 years, is often called upon as a mechanism of millennial scale variability, none of the models of the thermohaline circulation (2D, 3D, coupled A/O GCM) exhibit natural cycles on these time scales. Rather, the models suggest that oscillations in which the thermohaline circulation is involved have time scales of 3-5 hundred years at most (Mikolajewicz and Maier-Reimer 1990, Pierce et al. 1995, Aeberhardt et al. 2000). This time scale is the typical renewal time of the modern Atlantic basin only and is consistent with the fact that the other ocean basins have so far not been identified as centers of action for this variability.

The apparent absence of a natural time scale in the atmosphere/ocean system strongly suggests that the cryosphere plays a central role in pacing or exciting this variability. Indeed, as discussed above, paleoclimatic records do show, for each of the D-O events, traces of ice rafted debris which have been interpreted as a sign of rapid discharge from circum-Atlantic ice sheets (Bond and Lotti 1995). A minimal model of millennial-scale variability should therefore include the ocean, atmosphere and ice sheets. Entire sequences of millennial-scale changes involving the thermohaline circulation (THC) have been simulated only in simplified models which employ ad-hoc parameterizations of ice sheet discharge (Paillard and Labeyrie 1994, Stocker and Wright 1998, Ganopolski and Rahmstorf 2001). In these models, changes occur on a very regular time scale and are characterized by self-sustained oscillations. In contrast, the paleoclimatic record exhibits millennial variability in a band of time scales.

Recent dynamical modeling of the Laurentide ice sheet (Clarke et al. 1999) confirms the plausibility of earlier suggestions (MacAyeal 1993) of successive destabilizations of ice streams. While the timing of these events in the model (about 1 in 4500 years) is roughly consistent with the paleorecord, the amount of freshwater associated with the discharge of for example, the Hudson Bay ice stream, is rather small (order 0.01 Sv). Based on most ocean model results, such a freshwater perturbation is too small to induce significant changes of the Atlantic thermohaline circulation.

As noted earlier, these ice-sheet modeling fresh water fluxes do not agree with values derived from high-resolution dating of the Huon peninsula sea level record (figure 3.1). The reconstructions suggest sea level changes of the order of 10 m, which corresponds roughly to about 1 Sv sustained for 100 years or more. This implies a globally integrated freshwater flux several orders of magnitude larger than the regional values simulated by an ice sheet model (Clarke et al. 1999). Hence, there must be additional mechanisms in the climate system that multiply the effect of individual ice stream discharges. One possibility would be sea level rise itself, which would act as a synchronizer for discharge from various sites around the North Atlantic and Antarctica. Grounded marine ice sheets are very sensitive to sea level rise and can be effectively destabilized in this manner.

Perturbations of the thermohaline circulation with freshwater discharge have been tested with numerous coupled climate models of different complexity. The degree of collapse depends on the amount and location of freshwater discharge, and is sensitive to the model parameters. The model experiments indicate that perturbations in the order of 0.1 Sv or more are generally needed to induce significant changes of the THC (Manabe and Stouffer 1993, Wright and Stocker 1993, Mikolajewicz and Maier-Reimer 1994, 1997). Indeed, the above estimate of 1 Sv for 100 years, were it to occur in the northern North Atlantic, would be largely sufficient to collapse the Atlantic thermohaline circulation and induce strong cooling in the North Atlantic region (with amplitudes in near surface air temperature of up to 15 ˚C). In compensation, a warming is seen in the south, as proposed by Crowley (1992). Model simulations suggest that in that case, a bipolar seesaw is in action, with northern cooling associated with southern warming (Broecker 1998, Stocker 1998). For partial collapse, changes are more limited to the North Atlantic with little influence on the Southern Hemisphere. However, model simulations suggest that changes in the convection patterns of the Southern Ocean can strongly influence regional temperature response. As noted above (Section 3.1), although the concept of the bipolar seesaw is straightforward, its verification in paleoclimatic records remains controversial.

Coupled climate models therefore appear to contain sufficiently different circulation modes in the ocean to explain hemispheric to global scale reorganisation suggested by the paleoclimatic records. It is evident that rapid warming, as well as rapid cooling, can in principle, be realized by these models. The timescale of decades for these changes (as suggested by some paleorecords) are not inconsistent with simulations. Models make predictions regarding the spatial expression of such changes which are not always the same. Nonetheless, they do generally agree that climate changes are strongest in the North Atlantic region and that they are transmitted to other locations of the Northern Hemisphere via the atmosphere and through oceanic teleconnections (Schiller et al. 1997).

Of course, the goal would be to simulate entire climate cycles such as a series of Dansgaard-Oeschger events. If the climate system, for some reason, happens to be marginally stable, any small perturbation could trigger a mode change. In principle it is possible to construct a climate model that is very close to a bifurcation point at which very small freshwater perturbations can induce mode switches of the thermohaline circulation. The Climber-2 intermediate complexity model (Petoukhov et al. 2000, Ganopolski and Rahmstorf 2001), for example, presents a large hysteresis loop for the response of NADW formation to changes in freshwater flux over Northern Atlantic high latitudes. The "width" of that loop (along the freshwater flux axis) is directly proportional to the meridional oceanic transport of heat. The buoyancy gained by release of heat to the atmosphere has to exceed the buoyancy loss by freshwater input to sustain convection (Ganopolski and Rahmstorf 2001). In this model, the modern 10^{15} W meridional heat flux at 20°N corresponds to a width of the hysteresis loop of 0.22 Sv of fresh water, and all convection occurs north of Iceland. This defines the stability of the modern "warm mode" thermohaline conveyor to freshwater perturbations. During the last glacial period, the main site of convection occurs south of Iceland, and may shift progressively south during periods of ice-sheet extension. There is no clear bifurcation. However, during the intermediate climatic stage of MIS 3, the system appears particularly unstable, with deep convection moving north or south of Iceland under minimal changes of the fresh water flux (0.01 to 0.02 Sv.) (Paillard 2001). It is currently impossible to determine how close the system is to possible bifurcation points during the glacial, because the conditions at the atmosphere-ocean interface and in the ocean interior are not well known. At present, there is no easy way out of this dilemma.

Another question that has been treated only in a cursory fashion is the possible role of oceanic tidal cycles. Tides provide more than half of the total power for vertical mixing in the oceans (Munk and Wunsch 1998). Indeed, Keeling and Whorf (2000) propose that the 1-2 kyr oscillation observed on Last Glacial and Holocene northern Atlantic Ocean paleorecords by Bond et al. (1997, 1999) derives from the 1800 year cycle of the gradually shifting lunar declination from one episode of maximum tidal forcing on the centennial time scale to the next.

A multitude of parameters is recorded in ice, marine sediments and terrestrial records. Among these are greenhouse gases and their isotopic composition, isotopes and trace elements in marine organisms, assemblages and higher organic compounds. Although they are often hard to tie directly to physically relevant parameters, they contain invaluable information about how the climate system components reacted to millennial climate change. The goal of modeling is therefore not only to faithfully simulate certain phases of past records, but also to provide quantitative and dynamically consistent interpretations of these records.

3.4 Climate modes on interannual to centennial scales

Paleoclimate studies of the past few millennia, and in particular the past few centuries, have seen a tremendous expansion in the past decade and have made substantial new contributions to our understanding of natural climate variability. One approach to studying the climate of the past millennium, the synthesis of temperature-sensitive data to produce regional-global temperature histories, is discussed in detail in Chapter 6. Here we describe how paleoclimate reconstructions have contributed to an expanded view of climate dynamics on interannual to century time scales, with a particular focus on how modern patterns or modes of variability have changed through time.

Observational and modeling studies confirm that a substantial portion of modern climate variability can be described in terms of modes with distinctive temporal scales and spatial patterns. In the broadest sense, these modes result from the interactions of the ocean and atmosphere over a spatially heterogeneous surface boundary. The spatial scales are set by fundamental features of the earth's surface (such as ocean/continent geometry and gradients in insolation receipt), and they derive their time scales from the scales of the forcings and responses within the system and the degree to which the slower components of climate (ice, deep ocean, and vegetation) become entrained along with the more rapid variations of the surface ocean and atmosphere. They appear to represent fundamental physical aspects of the modern circulation. To the extent that climate varies by perturbations of these modes, they offer a framework for interpreting the past and possibly anticipating the future.

It is important to distinguish between the fundamental aspects of these modes, which are relatively well described, and their impacts, which occur outside the region of well-described physics. The El Niño/Southern Oscillation (ENSO), for example, is a well-described physical mode in the equatorial Pacific, with characteristic impacts outside this narrow region that are less consistent (Kumar and Hoerling 1997, Trenberth 1998). Changes both within and beyond the tropical Pacific may significantly alter how ENSO influences remote regions; changed teleconnections do not necessarily mean changed ENSO physics in the equatorial Pacific. Both paleoclimatic and model-based studies show that the teleconnections of ENSO are not always stable as background climate changes (Meehl and Branstator 1992, Cole and Cook 1998, Gershunov and Barnett 1998, Otto-Bliesner 1999, Moore et al. 2001). Attempting to reconstruct ENSO based on teleconnected patterns may lead to erroneous conclusions; this caution also holds true for other modes (and probably other timescales).

Both paleoclimate and instrumentally based climate studies have focused on physical modes of climate variability and their impacts as targets for sampling, modeling, and greater understanding. The following sections focus on recent results that bear on the variability and sensitivity of specific modes.

3.4.1 The tropical Pacific: El Niño/Southern Oscillation

ENSO in recent centuries

The well-known interannual variations of the ENSO system arise due to coupled interactions among the atmosphere and the surface and thermocline waters of the equatorial Pacific. Bjerknes (1969) first described the positive feedback mechanisms at the heart of ENSO, and these have subsequently been refined, described and summarized by others, including Philander (1990), Battisti and Sarachik (1995), and Wallace et al. (1998). In the case of El Niño warm phases, weaker trade winds lead to a drop in the normal sea level pressure gradient across the Pacific and a consequent deepening of the thermocline on the eastern side. This leads to weaker upwelling, warm SSTs, a further reduction in the zonal SST gradient, and further weakening of the trades. The opposite set of feedbacks acts to maintain cold La Niña events: stronger trades enhance and shallow the eastern equatorial Pacific thermocline, driving colder and more intense upwelling, strengthening the zonal SST gradient and consequently the trades. Warm SST anomalies in the east are initiated by the Kelvin wave response of the thermocline to seasonal wind anomalies in the western Pacific, and are terminated by slow adjustment processes that act between the ocean surface and the thermocline. These carry the oscillation into a cool phase. The positive feedbacks that maintain anomaly states may not be restricted to the interannual time scale, but could prolong anomalies for longer periods and even operate as a mode of response to external forcing (Clement and Cane 1999).

Interannual variations between extreme states of the ENSO system in the tropical Pacific orchestrate year-to-year climate variability in many parts of the world (Kiladis and Diaz 1989, Trenberth et al. 1998). Teleconnections throughout the tropical oceans, in the western Americas, and in the Indian and African monsoon regions are among the many well-documented impacts of the modern ENSO system. ENSO also influences high-latitude processes, including Antarctic sea ice extent and ice chemistry (Simmonds and Jacka 1995, White et al. 1999), as well as aspects of Atlantic hydrography relevant to thermohaline circulation, e.g. North Atlantic SST and the subtropical freshwater balance (Schmittner et al. 2000, Latif 2001). Interannual changes in the rates of atmospheric pCO_2 increase are attributable to ENSO variability, through changes in surface ocean temperature and upwelling patterns (Chavez et al. 1999, Feely et al. 1999) together with the response of terrestrial productivity to changes in water balance and nutrient feedbacks (Keeling et al. 1995, Braswell et al. 1997, Tian et al. 1998, Rayner et al. 1999, Asner et al. 2000). ENSO-related rainfall changes strongly influence fire frequency in many regions, both naturally (Swetnam and Betancourt 1998) and by abetting anthropogenic burning (Nepstad et al. 1999). ENSO variability is thus imprinted in an extensive suite of parameters relating to modern and past physical climate, biogeochemical cycles, and ecosystem dynamics; changes in ENSO have the potential for diverse and substantial global impacts.

Although modern instrumental studies of ENSO characterize its dominant frequency as interannual, a recent shift in 1976 to warmer and wetter conditions in the tropical Pacific has drawn attention to decadal variability in ENSO (e.g. Zhang et al. 1997). Extratropical decadal variability is discussed in section 3.3. A growing body of paleoclimatic evidence indicates that in the 19th century, variance in the tropical Pacific was weaker on interannual and stronger on decadal timescales, relative to the 20th century. This pattern is seen both in the impact of ENSO inferred from records largely outside the tropical Pacific (Mann et al. 1998, Stahle et al. 1998) and in the characteristic frequency of anomalies within the central equatorial Pacific (Dunbar et al. 1994, Urban et al. 2000 Figure 3.7). This and other changes in the dominant frequency of ENSO appear to be related to the background climate of the tropical Pacific; in the central Pacific, decadal ENSO variability is strongest when the central Pacific is relatively cool/dry (Urban et al. 2000). This decadal variance was felt beyond the equatorial Pacific regions. At least one of the prolonged La Niñas of the 19th century (1855-62) likely played a role in an extended drought in the central US, along the lines of what would be expected from modern climate relationships (Cole et al. submitted).

Another "style" of decadal ENSO variability consists of modulating the frequency of extreme events; some decades have stronger interannual variability than others (Trenberth and Shea 1987). Studies based on geochemical records from long-lived Pacific corals clearly show decadal modulation of the frequency of interannual ENSO extremes (Cole et al. 1993, Dunbar et al. 1994, Tudhope et al. 1995, Urban et al. 2000). The teleconnections of ENSO experience substantial decadal modulation as well. The well-known link between ENSO and the Indian monsoon has virtually disappeared since 1976 (Kumar et al. 1999b, Kumar et al. 1999a).

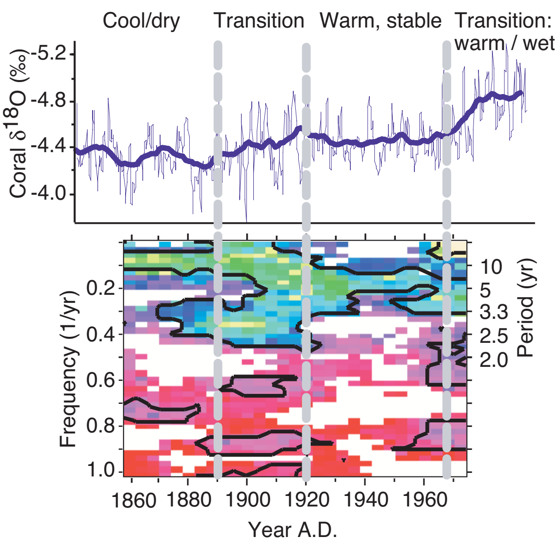

Fig. 3.7. Coral $\delta^{18}O$ data from Maiana Atoll (central Pacific) and evolutionary spectrum of the data plotted on the same horizontal axis (Urban et al. 2000). The top panel shows bimonthly values with a superimposed 21-yr running mean. The bottom panel maps the changing concentrations of variance revealed by evolutionary spectral analysis, in which 40-year segments were analyzed, offset by 4 years. Colored regions are significant above the median (50%) level, and the dark line encloses variance significantly different from a red noise background spectrum at 90%. Changes in the mean of the time series correspond to changes in the frequency domain characteristics of the record, particularly in the correspondence of strong decadal variance and weak interannual variance to cooler/drier background conditions in the 19th century. Vertical lines separate intervals where apparently different background conditions prevailed locally; these coincide with transitions in the spectrum.

Patterns of North American drought that correlate with ENSO variability changed during the twentieth century, a consequence of both changing ENSO strength and interactions with midlatitude systems (Cole and Cook 1998, Figure 3.8). Correlations among paleoclimate records sensitive to ENSO wax and wane (Michaelsen and Thompson 1992), suggesting nonstationarity in spatial patterns and/or intensity of teleconnections. Ice-core based precipitation records on Mt Logan in the Pacific Northwest change the sign of their correlation with ENSO on decadal scales (Moore et al. 2001).

Several candidate mechanisms exist for decadal modulation of ENSO variability (Kleeman and Power 2000). First, the physics responsible for interannual variability may be invoked for longer-period variability (Clement et al. 1999); the positive feedbacks that maintain La Niña and El Niño anomalies could become more persistent than today, due to changes in the background state (e.g. in the Pacific thermocline). Second, decadal changes in subtropical latitudes can propagate along isopycnal surfaces into the tropical thermocline, thereby influencing the temperature of upwelled water (Gu and Philander 1997). Subsurface temperature observations support this idea (Zhang et al. 1998). However, geochemical data from equatorial Pacific corals indicate that the 1976 shift cannot have been caused solely by changes in the thermocline source waters in the northern subtropics (Guilderson and Schrag 1998). Additionally, decadal variations in ENSO may result from physical processes not yet identified; they may originate from external forcing, or they may be stochastic. Cane and Evans (2000) argue that decadal variability in the tropical Pacific is not necessarily created by a single mechanism specific to that time scale, and that it may well result from multiple processes acting over a range of time scales.

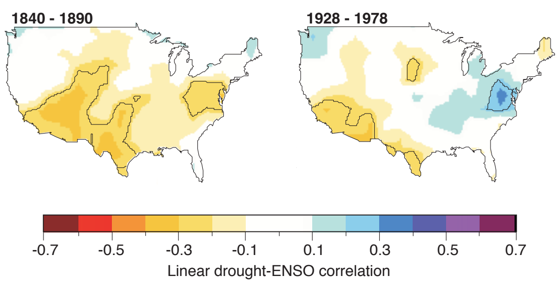

Fig. 3.8. Maps of drought-ENSO correlations calculated for 1840-90 and 1928-78 using tree-ring reconstructed drought indices (Cook et al. 1999) and a coral record of ENSO (Urban et al. 2000). Contours outline regions where correlations are significant at 90%.

ENSO in the late Quaternary

To assess the sensitivity of ENSO to changes in background climate, paleoclimatic studies of ENSO have begun to focus on more distant periods of recent earth history. Perhaps the most complete look at past ENSO variability to date uses a suite of coral records from northern New Guinea, where modern corals and instrumental data show a strong influence of ENSO variability (Tudhope et al. 2001). Tudhope et al. analyzed, at near-monthly resolution, multidecadal sections of coral in well-dated time windows over the past 130,000 years (Figure 3.9), and the results show clearly how the strength of interannual variability at this site has varied. They identify two aspects of long-term climate variability that appear to influence ENSO strength, precessional forcing and glacial background climates. Both act to dampen ENSO variability in past periods; modern samples (even those that predate substantial anthropogenic greenhouse forcing) show the strongest ENSO variability of any interval sampled.

The precessional influence on ENSO has been physically described using climate models of varying complexity. In a simple model of the equatorial Pacific Ocean and atmosphere, seasonal insolation changes associated with the precession of the earth's equinoxes influence the seasonal strength of the trade winds. When perihelion falls in the boreal summer/fall, stronger trades in that season inhibit the development of warm El Niño anomalies. This response is sufficient to generate significant changes in ENSO frequency and recurrence over the late Quaternary (Clement et al. 1999). In a global coupled ocean-atmosphere model, the intensified Asian monsoon at this phase of the precession cycle further enhances Pacific summertime trade winds, cooling the equatorial Pacific and reducing interannual variability (Liu et al. 2000).

The mechanisms for glacial ENSO attenuation are not nearly as well described. Possibilities include weaker ocean-atmosphere interactions in a cooler Pacific and intensified trade winds resulting from a stronger temperature gradient across the Pacific. A lower sea level exposing shallow continental shelves in the western Pacific may also stabilize variability by anchoring the Indonesian Low convection system. Possible mechanisms for strengthening ENSO in a glacial world also exist. For example, a shallower, steeper thermocline in the eastern Pacific could allow for greater interannual variability. The NCAR coupled climate model simulates stronger ENSO variability during the last glacial maximum (Otto-Bliesner 1999). The inference of weaker glacial ENSO from the coral data does not necessarily conflict with this simulation, however. The "glacial" intervals in the Tudhope et al. (2001) study are from less cold periods than the last glacial maximum, when precessional forcing also differs. Additional data will be needed to resolve which of the potentially competing influences ultimately determines ENSO strength, and when.

ENSO in the mid-Holocene

It is becoming clear that ENSO operated very differently prior to the mid-Holocene. Tudhope et al. (2001) document the weakest interannual variability of any time in their record at 6500 yr BP. Interannual variability associated with ENSO along the Great Barrier Reef was absent in a coral record from 5300 yr BP (Gagan et al. 1998). Debris flow deposits in an Ecuador lake, which today occur during El Niño rains, occur with a period of approximately 15 years before about 7ky BP, and show the establishment of modern ENSO periodicities (2-8 yrs) only around 5ky BP (Rodbell et al. 1999). Sediment profiles from archaeological sites along the Peru coast indicate a lack of strong flood events (interpreted as El Niño's) between 8900-5700 years ago at Quebrada Tacahuay (Keefer et al. 1998) and between 8900-3380 years ago at nearby Quebrada de los Burros (Fontugne et al. 1999). Pollen records from both South America and New Zealand/Australia indicate that early Holocene vegetation did not include types adapted to periodic droughts that occur today, associated with interannual ENSO. Such vegetation types became established in these areas only in the late Holocene (McGlone et al. 1992, Shulmeister and Lees 1995). These and other lines of evidence suggest that the global imprint of ENSO was very different before the mid-Holocene (Markgraf and Diaz 2001). Numerical modeling studies provide insight into mechanisms of such changes. Several coupled GCM simulations bear on the question of ENSO in the mid-Holocene. Using the NCAR coupled Climate System Model, Otto-Bliesner (1999) found a cooler equatorial Pacific and no substantial change in ENSO variability. She also noted that teleconnections significant in the modern system (modeled and observed) are absent in the 6K simulation, due to the stronger influence of regional climate changes. Liu et al. (2000) used the Fast Ocean-Atmosphere Model coupled with a low-resolution atmospheric GCM and forced with early and mid-Holocene insolation (6ky BP and 11ky BP). They found a weaker ENSO system in both cases, with a tendency for slightly stronger La Niña events and even weaker El Niños, a consequence of two mechanisms. First, stronger trade winds

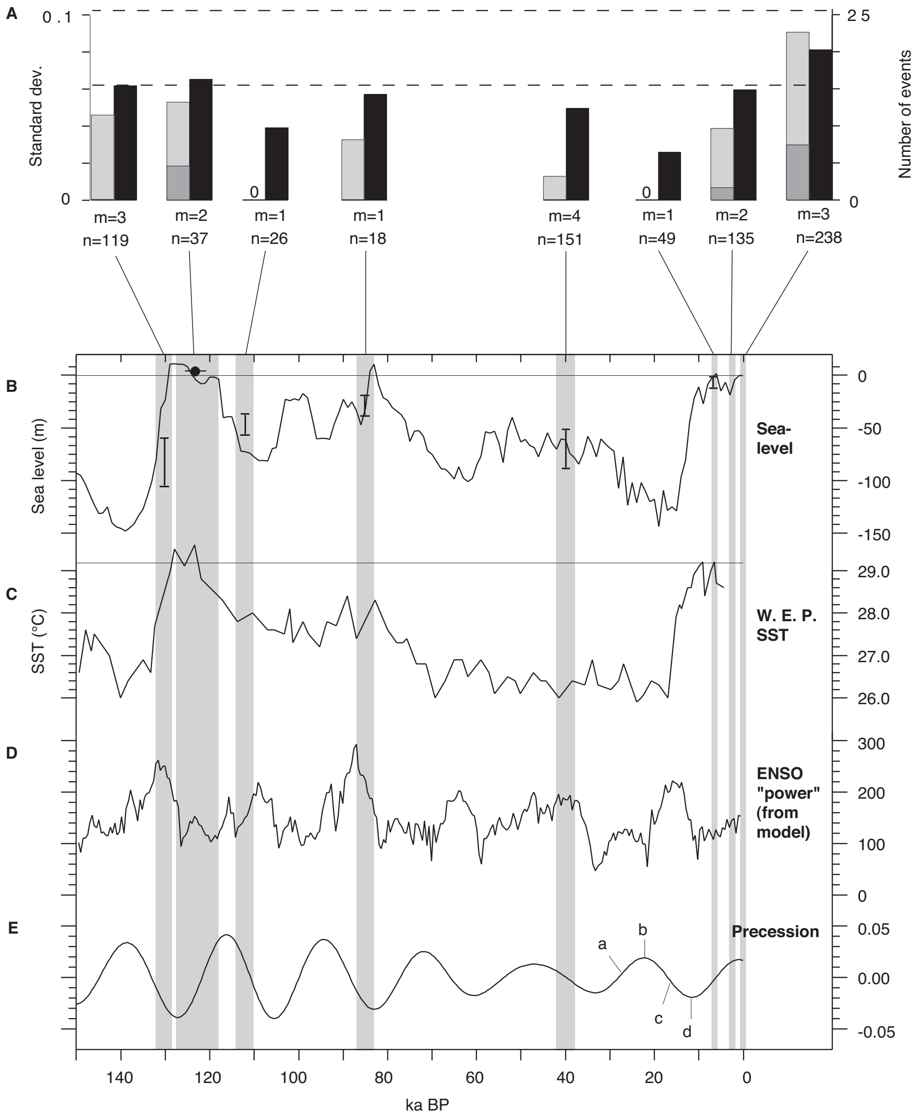

Fig. 3.9. Interannual ENSO variability over the past 130,000 years. **A.** Strength of ENSO variability in coral $\delta^{18}O$ records for eight time periods. Solid black bars show the standard deviation (scaled on the left axis) of the 2.5 to 7 year bandpass-filtered coral $\delta^{18}O$ records from each time period. Shaded bars denote high-amplitude events for each period. The darker bars indicate the percentage of the data in the ENSO bandpass-filtered data that exceeds 0.15 absolute amplitude; lighter bars indicate the percentage of data that exceeds 0.10 amplitude (both scaled relative to the left vertical axis). The number of corals for each group is given by m, and the total number of years represented by all corals in each group is given by n. The horizontal dashed lines indicate the maximum and minimum standard deviation for sliding 30-year increments of modern coral $\delta^{18}O$ 2.5- to 7-year bandpass-filtered time series. **B.** Estimate of global sea level (plotted as meters below present sea level) derived from benthic foraminiferal $\delta^{18}O$ (Shackleton 2000). Bars indicate paleo-sea level estimated from the elevation, age, and uplift rate of corals analyzed in this study. These bars include uncertainty in the water depth in which the corals grew. Estimates of uplift rate are based on an assumed sea level of +5 m at 123 ka (circle and bar). **C.** Sea surface temperature record for the western equatorial Pacific (ODP Hole 806B, 159°22'E, 0°19'N, 2520-m water depth) based on Mg/Ca composition of planktonic *Foraminifera* (Lea et al. 2000). The horizontal line indicates modern SST. **D**. ENSO variability estimated from application of the Zebiak-Cane coupled ocean-atmosphere model forced only by changing orbital parameters (Clement et al. 1999). Shown here is power in the 2- to 7-year (ENSO) band from multitaper spectral analysis of nonoverlapping 512-year segments of the modeled NINO3 SST index. Power is approximately equal to 100x variance. Although there is considerable variation at sub-orbital wavelengths (2s of power estimates ~±71 based on a control run with no change in orbital parameters), the main precessionally related features, including the trend of increasing ENSO amplitude and frequency through the Holocene, are statistically significant (Clement et al. 1999, Clement et al. 2000). **E.** The precessional component of orbital forcing (Berger 1978). For one cycle, the timing of perihelion is indicated as follows: a, boreal autumn; b, boreal winter; c, boreal spring; d, boreal summer.

(resulting from enhanced deep convection in the Asian monsoon region) force a shallower thermocline in the eastern Pacific. Second, a warm anomaly in the equatorial thermocline weakens the vertical temperature gradient and hence the El Niño-La Niña contrast. Bush (1999), using the GFDL atmopheric GCM coupled with a primitive equation ocean model, found an intensified equatorial Pacific cool tongue at 6000 years ago, with greater seasonality (changes in interannual variability were not discussed).

Mid-Holocene model results and paleoclimatic observation are beginning to yield a semi-consistent picture (Clement et al. 2000, Cole 2001). A cooler (more La Niña-like) eastern and central Pacific is shown by all coupled models (Bush 1999, Otto-Bliesner 1999, Liu et al. 2000). Regional warming suggested in the westernmost Pacific (Gagan et al. 1998) supports a picture of a more La Niña-like average state. Reduced interannual variability in northern Australia (McGlone et al. 1992, Shulmeister and Lees 1995, Gagan et al. 1998) and New Guinea (Tudhope et al. 2001) implies a weaker ENSO overall and attenuated teleconnections. Weaker interannual variance around a cooler background mean would be consistent with fewer flood events in an Ecuador lake record (Rodbell et al. 1999), as warm anomalies large enough to generate intense rainfall would be rarer.

The presence of warm water molluscs of mid-Holocene age on the Peru coast (Sandweiss et al. 1996) disagrees with this picture, although local geomorphological complications (de Vries et al. 1997) or regional oceanic influences (Liu et al. 2000) may explain those observations. The inference of comparable ENSO variance at 6ky BP in one modeling study (Otto-Bliesner 1999) does not correspond to the generally observed picture of reduced variance at that time. Well-calibrated data from the regions of strongest ENSO influence are still sparse, and models have yet to simulate modern ENSO variance perfectly, so there is room for improvement on both sides. Records from outside the tropical Pacific may confuse this analysis, as both data and models increasingly reveal the nonstationarity of ENSO teleconnection patterns, particularly as background climates change (Meehl and Branstator 1992, Cole and Cook 1998, Kumar et al. 1999, Kumar et al. 1999, Otto-Bliesner 1999, Moore et al. 2001)

3.4.2 Decadal variability in the extratropical Pacific

Instrumental data offer clear evidence of coherent decadal variability over much of the Pacific Ocean. The spatial pattern associated with this variance is latitudinally broader than ENSO, and the time scale of variability is longer, but the impacts on climate variability around the Pacific are often similar. Many instrumental and modeling studies have described this pattern (e.g. Latif and Barnett 1996, Mantua et al. 1997, Minobe 1997, Zhang et al. 1997, Garreaud and Battisti 1999). Pacific decadal variability clearly has a strong influence on many aspects of natural systems in the Pacific (Ebbesmeyer et al. 1991) including salmon fisheries and glacial mass balance in the northwestern US (Mantua et al. 1997, Bitz and Battisti 1999), the strength and pattern of ENSO teleconnections in North America (Cole and Cook 1998, Gershunov and Barnett 1998 Figure 3.10), and the predictability of Australian rainfall (Power et al. 1999).

Pacific decadal variability is typically described in terms of SST variations in the north Pacific. Trenberth and Hurrell (1994) use a North Pacific Index (NPI) of monthly SLP anomalies over the domain 30-65°N, 160°E-140°W; other studies refer to this index as the North Pacific Oscillation (NPO Gershunov and Barnett 1998). Mantua et al. (1997) describe the Pacific Decadal Oscillation as the time series of the first EOF of Pacific SST north of 20°N; others have redubbed this the Pacific Interdecadal Oscillation (retaining the acronym PDO). Power et al. (1999) refer to the phenomenon as the Interdecadal Pacific Oscillation (IPO) and include both Northern and Southern Hemisphere SST in its definition. The spatial pattern of a positive PDO or IPO is similar to that of El Niño, but with a greater latitudinal spread of warming centered on the equator and a stronger cooling in the western mid-latitudes. When the PDO is positive, low pressure anomalies (a deepened Aleutian Low) exist over the north central Pacific and the NPI is negative.

Time series of the NPI and the PDO calculated from various datasets show a multidecadal time scale since the 1920's, with prominent transitions around 1976, 1946, and 1924 (e.g. Mantua et al. 1997, Minobe 1997, Chao et al. 2000). Before 1925, however, instrumental records of Pacific decadal variability tend to disagree, as data become increasingly sparse (Figure 3-10a). With so few iterations of Pacific decadal variability available in the instrumental record, numerous questions remain open about its fundamental nature. Given the long time scale of this mode relative to most instrumental records, paleoclimatic efforts to reconstruct this phenomenon in the preindustrial era (e.g. from subtropical corals and tree-ring data along the Pacific rim) have important contributions to make in understanding its persistence, stability, and role in regional-hemispheric anomalies.

Tree ring-based reconstructions of Pacific decadal variability typically proceed by either averaging over multiple sites or extracting patterns of common variance from networks of sites in the western Americas. Because this phenomenon influences both hydrologic balance in the subtropical latitudes

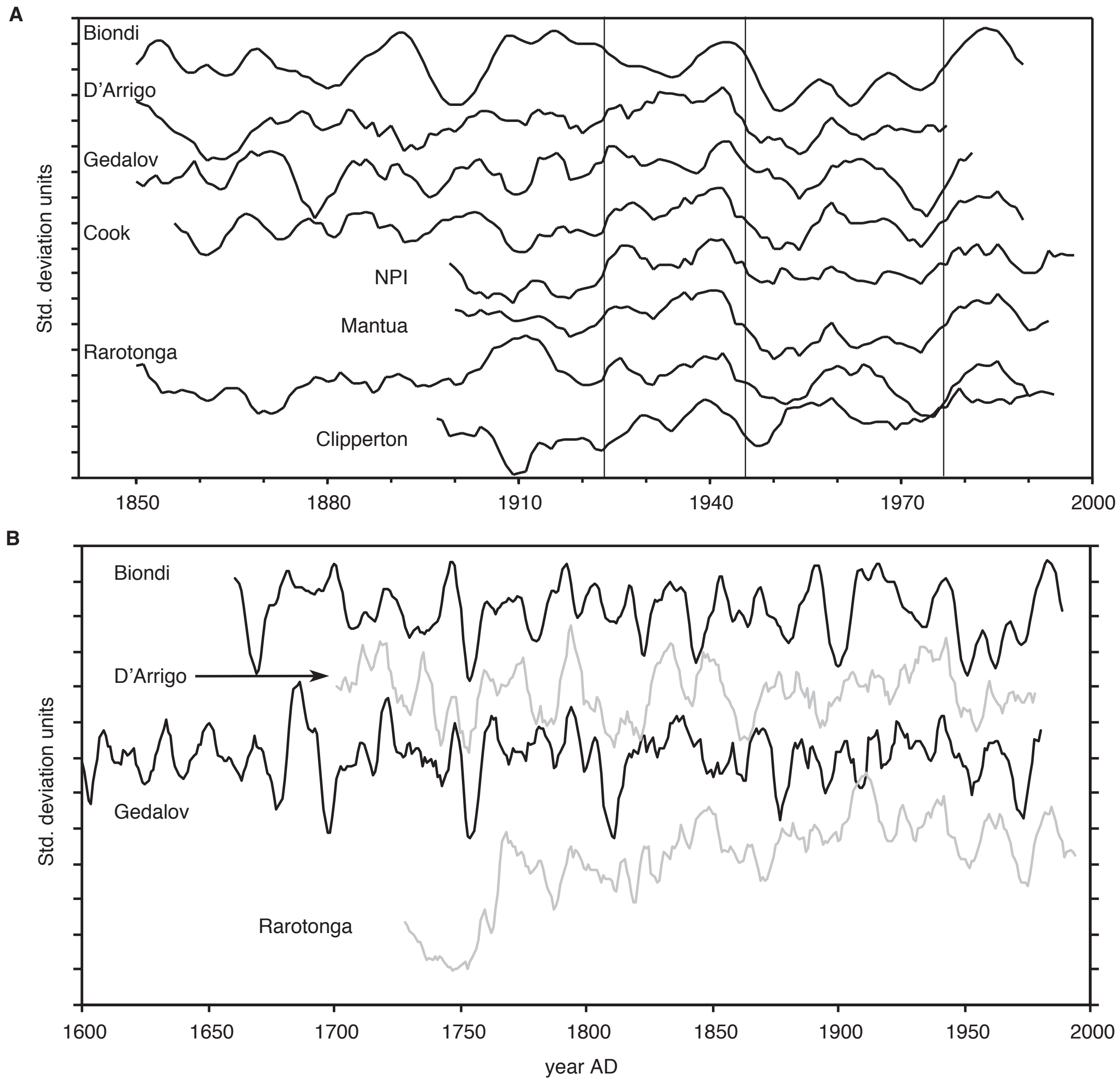

Fig. 3.10. Time series of Pacific decadal variability from instrumental and paleoclimatic sources. Contours outline regions where correlations are significant at 90%. The data sources are, from top: (Trenberth and Hurrell 1994, Mantua et al. 1997, Linsley et al. 2000a, Linsley et al. 2000b, Biondi et al. 2001, Gedalov and Smith 2001, D'Arrigo et al. submitted). Records have been normalized to a standard deviation of 1 in the 1900-1979 common interval and smoothed using a 5-point running mean, to enhance the low-frequency variance. Vertical axes are marked every 1 standard deviation. The NPI (North Pacific Index of Trenberth and Hurrell (1994)) and Rarotonga (Linsley et al. 2000b) SST reconstructions are plotted inversely to facilitate comparison. The Clipperton (Linsley et al. 2000) record indicates the alternation between relatively warm/fresh (upwards) and cool/salty (downwards) conditions; isotopic data do not distinguish temperature and salinity. **A.** Instrumental and paleoclimatic records from 1850 onwards show transitions in the mid-1920's, mid-1940's, and mid-1970's. Before about 1925, however, the records agree less well – even the two instrumental SST-based PDO indices differ, due to sparse observational coverage and different methods for infilling missing values. **B.** Pre-instrumental records of Pacific decadal variability from tree-ring and Rarotonga coral data appear to move in and out of phase, suggesting a variable footprint (and perhaps variable mechanisms) for Pacific decadal variability.

and temperatures at higher latitudes, sites in one or both of these regions are commonly used. Biondi et al. (2001) used a network of precipitation-sensitive sites in southern and Baja California to develop a PDO reconstruction since AD 1661. Gedalov and Smith (2001) reconstructed north Pacific decadal variability since 1599 using six tree-ring chronologies from Oregon to Alaska. Another tree ring-based PDO reconstruction dating to 1700 (D'Arrigo et al. submitted) used sites in coastal Alaska, the Pacific northwest, and northern Mexico to capture both the temperature and hydrologic aspects of this variability. Other studies identified a broad pattern of Pacific decadal variability in tree-ring data from regions near the Pacific coast of North America (Ware and Thomson 2000) or in reconstructions of temperature and precipitation in North and South America (Villalba et al. 2001). The latter study strongly supports the extension of this mode to the Southern Hemisphere.

Although many coral records from the Pacific exhibit a decadal time scale, few describe potential links to the PDO. A coral oxygen isotope record from Clipperton Atoll (10°N, 109°W) shows decadal variations that may reflect changes in the strength of the low-salinity equatorial countercurrent; the correspondence between this coral record and the PDO implies a stronger ECC and northward-displaced ITCZ during positive PDO phases (Linsley et al. 2000). Recent oscillations in the PDO are also present in an SST reconstruction based on coral Sr/Ca records from Rarotonga (21.5°S, 159.5°W) which extend to 1726 (Linsley et al. 2000b). This study provides additional evidence for the symmetry of decadal temperature responses across the hemispheres, at least during some intervals.

These records tend to confirm the oscillations seen in the mid-late 20th century instrumental data, but typically show much less agreement in earlier times (Figure 3.10b). The disagreement likely stems from several sources. First, if taken at face value, these records imply that the Pacific decadal variability in the 20th century is anomalously coherent across a broad latitudinal range, compared to previous centuries. The disagreement in earlier periods may suggest that Pacific decadal variability is not a single fundamental phenomenon, as it appears from the 20th century, but instead involves multiple independent influences, which are more clearly seen when records are extended. Second, the sensitivity of the sites used may not be ideal for representing the PDO. As initially defined, the PDO core region lies in the extratropical North Pacific; all terrestrial data are essentially teleconnections, which may change through time. Finally, the records are located in widely varying locations, and each represents the superposition of local variability on top of large-scale patterns. Local variability may be more of an issue for coral data, which are single-site geochemical records, than for the tree-ring data, which are combinations of records that span larger regions.

We can potentially learn much from comparing these records, despite their dissimilarities. Comparison of the Rarotonga coral record with the Northern Hemisphere data, for example yields strong agreement for the 20th century and certain intervals of the 18th century, but weak agreement in between. Tropical forcing may be most important during those times when the southern and Northern Hemispheres agree (e.g. the 20th century), and during other times, the hemispheres may behave more independently. Also notable is the tendency for the hydrology-sensitive and temperature-sensitive reconstructions to behave differently at times; the PDO has apparently not always caused these anomalies to move synchronously, which calls into question its use in prediction. Finally, there are intervals where most long records do agree (e.g. the mid-late 18th century), which points to promising intervals for further study of the PDO phenomenon. Periods when the greatest covariation is seen among multiple records may provide a useful focus for more in-depth analysis (e.g. mapping of annual anomalies). An event-based analysis may prove useful in identifying how consistent the PDO's impacts are over time; more data from the Southern Hemisphere (e.g. Villalba et al. 2001) are sorely needed for such work.

Paleoclimatic records of Pacific decadal variability also indicate diverse frequency-domain characteristics. Typical time scales include one around a decadal period (identified variously as 12, 14-15, and 10-20 yrs) and one in the multidecadal range (identified as 23, 25-50 and 30-70 yrs). The tree-ring studies address the stationarity of these components to varying extents, with no clear consensus: the 23-yr period of Biondi et al. (2001) strengthens post-1850; the 12-yr period in D'Arrigo et al. (submitted) weakens post-1850, and Gedalov and Smith (2001) found that the multidecadal (30-70) period is concentrated before 1840, while their 10-20 year period is strong throughout. Spectral analysis of instrumental indices of Pacific decadal variability also shows decadal and multidecadal periodic components whose strengths depend on the time interval analyzed (Cole, unpublished analysis). Either there is no clear frequency-domain signal of this "decadal" phenomenon, or the available records have yet to identify it unambiguously.

Although paleoclimatic studies of the PDO have not yet answered the questions raised by instrumental records, we note that this approach is still very new (the reconstructions shown here were all published in 2000 and 2001). Reconciling discrepancies among existing reconstructions, and developing better ones is a focus of substantial ongoing effort in the tree-ring and coral paleoclimate communities. Basic questions on the spatial and temporal pattern and coherency of Pacific decadal variability should soon be answerable with much greater confidence.

3.4.3 North Atlantic Oscillation

Interannual and decadal variations in North Atlantic climate are closely tied to the state of the North Atlantic Oscillation (NAO), broadly defined as the meridional pressure gradient between the high pressure system that lies around the Azores and the low pressure system that covers a broad swath of the Atlantic Arctic centered around Iceland (Hurrell 1995). The state of the NAO governs the strength of the westerly flow across Europe and thus the regional moisture balance and the tendency for severe winter storms.. The NAO is typically monitored by indices of the sea level pressure difference between stations in Iceland and stations near the Mediterranean (e.g. Lisbon, the Azores, or Gibraltar). However, NAO physics have not been well described, and the oscillation itself may reflect the combination of semi-independent processes that influence the strength and extent of the Icelandic Low and the Azores High. The NAO also influences the formation of North Atlantic Deep Water (Dickson et al. 1996, Curry et al. 1998), giving it a potential role in broader patterns and longer time scales of climate change.

The late 20th century experienced exceptional variability in the NAO, including an unprecedented series of high-NAO-index winters between 1988-1995, which led to unusual warmth and storminess over Europe. Paleoclimatic reconstructions have attempted to address whether the recent variability is unusual in the context of longer records. Several reconstructions have been proposed, using various combinations of tree-ring, ice core, and long instrumental data (D'Arrigo et al. 1993, D'Arrigo and Jacoby 1993, Appenzeller et al. 1998, Cook et al. 1998, Luterbacher et al. 1999, Cullen et al. 2001, Glueck and Stockton in press, Mann in press). Recent comparisons of NAO reconstructions (Schmutz et al. 2000, Cullen et al. 2001) suggest that the various reconstructions have different sensitivities and tend to correlate only moderately among themselves. The best-calibrated reconstructions include long instrumental data from European sites, and extend the NAO record back to 1700-1750; the late 20th century appears anomalous in that long-term context (Cullen et al. 2001). The best calibrated reconstructions may not necessarily be the best reflection of the NAO, however. Instrumental SLP records against which paleo data are compared may not be ideally situated to capture variance in the NAO variance (Deser 2000), while the paleoclimatic records have the potential to include a spatially broader NAO signal (Cullen et al. 2001).

Analyses of high-latitude climate variability have identified a mode that is potentially of broader significance (Thompson et al. 1998). The Arctic Oscillation (AO) describes an oscillation that spans the troposphere and lower stratosphere, and raises pressures alternately at the polar cap and along a zonal ring at about 55°N. In its positive phase, the pressure along 55°N is high, which strengthens westerly winds there and steers oceanic storms along a more northerly path. The NAO may be a regional manifestation of the AO; they correlate well, and the AO explains a greater fraction of variance in European climate than the NAO. Alternatively, Deser (2000) has noted that the NAO is strongly linked to Arctic variability, but Pacific-Arctic links are weaker and the AO is dominated by the NAO. Debate about the AO centers in part on whether climate patterns along the full circumference of the southern extent of the AO are actually well correlated (Kerr 1997), a question that paleodata could address, but has not yet, at least directly. Arctic data appear relatively coherent (Jacoby and D'Arrigo 1989, D'Arrigo et al. 1993, Overpeck 1997), but a direct comparison of Pacific and Atlantic regions along the southern extent of the AO is needed.

3.4.4 Tropical Atlantic: the dipole and extratropical links

The main mode of decadal SST variability in the tropical Atlantic has been described as a dipole across the equator (Servain 1991). The northern and southern tropical SSTs are generally uncorrelated (Houghton and Tourre 1992, Rajagopalan et al. 1997, among many others); however the cross-equatorial SST gradient has an important influence on rainfall on adjacent continents (Hastenrath and Greischar 1993). Analyses of instrumental data show strong coherence between this gradient and the North Atlantic Oscillation (Rajagopalan et al. 1998). Rajagopalan et al. speculate that the North Atlantic may respond to tropical heating anomalies and that the cross-equatorial gradient is a reasonable indicator of extratropical linkages. The mechanisms for a dipole in tropical Atlantic SST have been described in a modeling study (Chang et al. 1997) that identified a characteristic dipole time scale of 13 years.

Paleoclimatic records support these inferences over longer periods. An 800-year varved sediment record of trade wind-driven upwelling from the Cariaco Basin, Venezuela, exhibits significant variance at a 13-year period (Black et al. 1999), although the period is nonstationary. Longer-period variance in this record suggests an upwelling response to solar forcing (see Chapter 6, Section 7 for discussion). The strongest correlation with Cariaco Basin upwelling is found with North Atlantic SST. Stronger trade winds and upwelling are associated with colder North Atlantic conditions. This relationship holds true over longer time scales as well, with multidecadal North Atlantic anomalies clearly visible in laminated sections of the Cariaco record during deglaciation (Hughen et al. 1996).

The influence of the dipole on Sahel rainfall may hold a clue to explaining a mystery of past climate - the "Green Sahara" of the early-mid Holocene. Climate models forced with appropriate insolation values fail to explain the magnitude of increased moisture in north Africa reconstructed from lake level and other paleoclimatic data (Joussaume et al. 1999). Even when albedo and moisture recycling feedbacks associated with expanded wetlands and vegetation cover are included, climate models fail to simulate the degree of monsoon intensification observed (Doherty et al. 2000). But when ocean temperatures are calculated to be in equilibrium with the radiative forcings and incorporated into the model, the African monsoon response is significantly enhanced (Kutzbach and Liu 1997).

3.4.5 Global teleconnectivity

As research proceeds on the dynamics of modern climate systems – ENSO, monsoons, the high and low latitude Atlantic – couplings among them, often spanning time scales, are becoming clearer. For example, ENSO influences the freshwater balance in the tropical Atlantic; by changing the salinity of the source waters that eventually feed convection in the North Atlantic, long-term changes in the frequency of ENSO extremes can feed back on thermohaline circulation (Schmittner et al. 2000, Latif 2001). A lagged correlation between ENSO and the NAO supports this connection and implies predictability for the NAO based on earlier ENSO behavior (Latif 2001). Alternatively, recent unusual behavior in the NAO has been attributed to Pacific changes as a direct (immediate) atmospheric response to Indo-Pacific warming (Hoerling et al. 2001). North Atlantic variability influences the hydrologic balance over the Asian landmass, with consequences for the monsoons of the subsequent year. As noted earlier, the well-known linkage between ENSO and the Asian monsoon appears to have broken down as the Eurasian landmass has warmed in the 20th century (Kumar et al. 1999). These and other studies of global teleconnectivity offer targets for paleoclimatic investigation, yet caution us against assuming that such linkages should be permanent in the face of changing background climate.

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