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Environmental processes of the ice age: land, oceans, glaciers (EPILOG)

Alan C. Mix[a,*], Edouard Bard[b], Ralph Schneider[c]

[a]College of Oceanic and Atmospheric Sciences, Oregon State University, Corvallis, OR 97331-5503, USA
[b]CEREGE, Aix-en-Provence, France
[c]University of Bremen, Bremen, Germany

Abstract

Knowledge of the state of the earth at the Last Glacial Maximum (LGM, an interval around 21,000 years ago) is an important benchmark for understanding the sensitivity of global environmental systems to change. Much progress in understanding climates of the LGM has occurred in the ~ 20 years since the end of the CLIMAP project of the 1970s (Climate Long-range Investigation, Mapping and Prediction). Here we review this progress, based on presentations and discussion at a first open science meeting of the EPILOG project (Environmental Processes of the Ice age: Land, Oceans, Glaciers) held in Delmenhorst, Germany, May 1999. We outline key controversies and document protocols for EPILOG contributions, so that a new synthesis of the LGM Earth can emerge as an open project of the world's community of scientists.

1. Introduction

A long-standing goal of paleoclimatologists is to understand climate changes of the last ice age, the largest manifestation of natural climate change that remains relatively well preserved in the geological record. This interest is more than just academic curiosity. Current debate about the extent of future climate changes, whether natural or manmade, demands that we understand the extent, rate, and frequency of natural climate changes of the past.

Assessing the likelihood of future changes requires the use of computer models that link components to simulate the atmosphere, ocean, land surface, ice, and biota. In spite of growing sophistication of such efforts, models inevitably simplify reality and must be tuned to simulate both the present system and its sensitivity to change. This limitation of models underscores the importance of reconstructing and understanding the state of the Earth during the Last Glacial Maximum (LGM). If models are able to simulate the full range of dynamic climate changes actually observed, then we gain confidence in their ability to predict the future. If models fail in their simulations of the past, we learn from the failure about overlooked or poorly represented processes that must be understood and incorporated into the next generation of models. At present, confidence in model predictions of future climate change is limited, especially because many of the models incorporate oceanic processes in a limited way.

If the geologic record is to provide a useful test of climate models, we must first have confidence that we know what really happened during the last ice age. How cold was it? Was cooling relative to modern climate uniform over the globe, or do geographic patterns of change reveal the processes responsible for change? What mechanisms set the sensitivity of Earth's climate system to large-scale changes?

2. The CLIMAP reconstruction

The challenge of reconstructing the surface of the Earth at the LGM was taken up nearly 30 years ago by the CLIMAP (Climate Long-range Investigation, Mapping, And Prediction) project, which built on a method for quantitative estimation of past environments developed by Imbrie and Kipp (1971). During the 1970s, the CLIMAP group pioneered a first global reconstruction of the ice-age climate. Along the way they revolutionized the study of global environments in the geologic record as they developed strategies for quantitative stratigraphy

* Corresponding author. Tel.: + 1-541-737-5212.
E-mail address: mix@oce.orst.edu (A.C. Mix).

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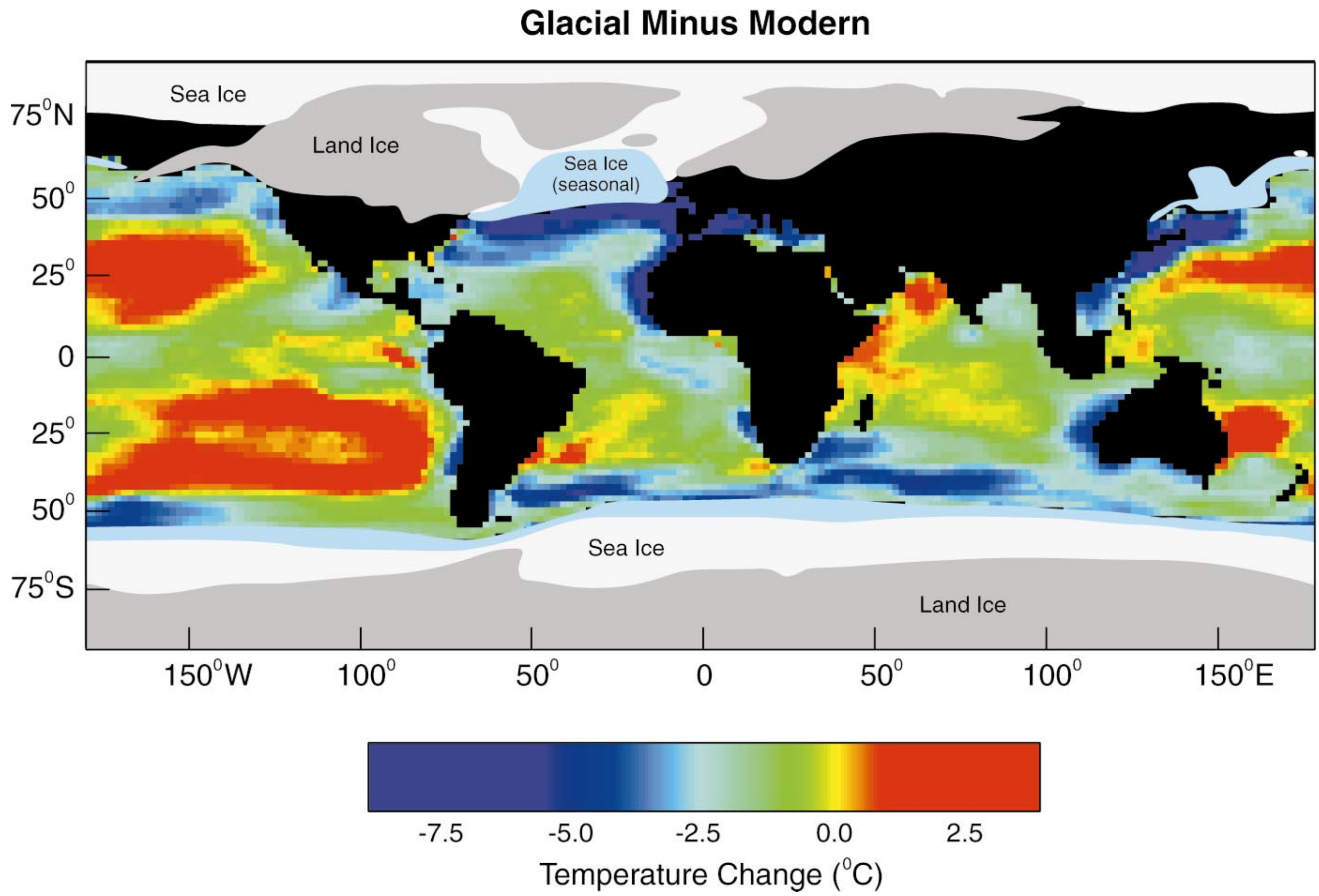

Fig. 1. Changes in the annual average sea surface temperatures (LGM-modern), with geographic distributions of LGM seasonal and perennial sea ice, and LGM land ice, based on seasonal estimates of CLIMAP (1981). A major conclusion of CLIMAP, surprising at the time, was that cooling associated with ice ages was regional, rather than global.

on a global scale, established a preliminary chronology for global climate change, and advanced methods to estimate quantitatively seasonal temperatures based on distributions of fossil species (e.g., Cline and Hays, 1976).

The CLIMAP project culminated in a series of maps that depicted the state of the Earth's surface at the LGM (CLIMAP, 1981). Although these maps emphasized seasonal sea-surface temperatures, they also included the distribution of ice both on land and in the ocean, as well as vegetation and albedo (reflectivity) of the land surface.

A primary result of CLIMAP, surprising at the time, was that the magnitude and even direction of temperature change was very different between regions rather than uniform globally (Fig. 1). The largest inferred changes were in the mid-to-high latitudes, especially in the North Atlantic region bordered by large ice sheets. In the CLIMAP reconstruction, sea ice advanced in the Greenland and Norwegian Seas and persisted through the summer. Seasonal meltback of sea ice was also diminished in the Southern Ocean. Upwelling systems associated with eastern boundary currents (such as off NW Africa) cooled significantly. Lesser cooling was inferred near the equator in all oceans. The centers of the subtropical gyres and the western Pacific warm pool yielded the largest surprise; based on CLIMAP's species-based methods temperatures in these areas stayed essentially the same as (and in some cases even warmed relative to) modern conditions. Seasonal contrast increased near the equator (implying stronger seasonal trade or monsoon winds), but decreased at high latitudes (due to perennial ice cover on land and in the sea). Overall, CLIMAP (1981) inferred that Earth cooled surprisingly little during the ice age, except at high latitudes. This concept, often referred to as "polar amplification" of climate change, implies a feedback mechanism driven by increased albedo associated with expanded snow and ice cover.

On land, forests shrunk, deserts expanded, and albedo was adjusted accordingly (Peterson et al., 1979). Debate within the glaciological community left uncertainty about the extent and thickness of LGM ice sheets. CLIMAP proposed two options, both based on equilibrium profiles along presumed glacier flowlines (Hughes et al., 1981). A "maximum" model assumed areas of ice sheets as large as possible based on the greatest extent of moraines or other evidence (in some cases undated) and on an interpretation of ice flowlines that yielded a relatively high Laurentide Ice Sheet dominated by a central dome. Significantly expanded marine ice included an Arctic ice shelf that helped to buttress thick ice on land. A "minimum" model was based on much more restricted ice limits in many places, particularly around shallow marine basins, and on flowlines inferred from a different interpretation of the glacial geology that in some cases yielded thinner ice sheets. Hughes et al. (1981) acknowledged that real ice sheets, which respond slowly to climate forcing, may not have achieved full equilibrium profiles during the LGM, but given the data available at the time there was no reliable basis for reconstructing transient behavior of the global ice sheet system. Indeed, Hughes et al. (1981) stated, "Our most important conclusion is that the distribution of late Wisconsin-Weichselian ice sheets is not well known". This remains an important consideration, because climate feedback processes associated with ice depend on the regional details of ice extent and height (Ramstein and Joussaume, 1995; Broccoli, 2000).

3. Two decades since CLIMAP

Over the past 20 years, several elements of the CLIMAP (1981) reconstruction have been called into question. Ice sheets based on regional sea-level data are thinner and have about 35% less change relative to modern ice sheets than those of the CLIMAP maximum ice sheet and 10% less than the minimum ice sheet reconstruction (Fleming et al., 1998; Peltier, 1994, 1998a, b; Lambeck et al., 2000), although recent data and isostatic modeling suggest a total volume change similar to the CLIMAP minimum ice sheet (Yokoyama et al., 2000). In the high-latitude oceans, diatom floras appear to conflict with CLIMAP's (1981) inference of perennial sea ice (Burckle et al., 1982; Crosta et al., 1998a). In the tropics, many climate models cannot reconcile CLIMAP's relatively small changes in tropical sea-surface temperature with evidence for advances of mountain glaciers, modified pollen assemblages (Rind and Peteet, 1985; Thompson et al., 1995) or rare gases in lowland groundwater (Stute et al., 1995). New methods for reconstructing ocean temperature and other properties are being developed, but in some cases these conflict with each other. For example, Sr/Ca measurements in coral reefs suggest much greater cooling than that inferred by CLIMAP (Guilderson et al., 1994), but an organic geochemical thermometer based on unsaturation of long-chain alkenones, $U^{k'}_{37}$, suggest tropical temperature changes of about 2°C that are to first approximation in agreement with CLIMAP's results (e.g., Bard et al., 1997).

As yet, no large-scale re-analysis of the LGM interval exists based on recent estimates of ocean temperature, ice sheets, and other environmental properties. Recognizing the need for such a synthesis, 57 scientists met in May 1999 at the Hanse Institute for Advanced Study in Delmenhorst, Germany, to initiate the EPILOG program. Research programs from nine nations (Australia, France, Germany, Netherlands, Norway, Russia, Spain, UK, US) were represented, and the effort was sponsored by IGBP/PAGES and IMAGES. This first public gathering of EPILOG set as its goals: (1) to assess advances made in the 20 years since CLIMAP, finding points of consensus where possible, (2) to outline strategies that will help to resolve remaining conflicts, (3) to set standards for international contributions to a new community assessment of the LGM, and (4) to outline a plan of action leading to a new synthesis. The meeting emphasized current understanding of sea-surface and continental temperatures at the LGM.

Here we report on the views presented at this first EPILOG meeting. We focus on the rationale for a new synthesis of LGM environments. We review substantial advances in chronology that help to refine research strategy and note key chronology issues that remain unresolved. We assess strengths and weaknesses of various geologic proxies for surface temperature and other information needed for understanding ice-age climate, and note emerging regional syntheses based on new data. Our primary motivation is to highlight key controversies that need resolution, to outline opportunities for research to fill gaps in knowledge, and to document protocols for EPILOG contributions so that a new synthesis can emerge as an open project of the world's community of scientists interested in the evolution of past and future climate.

4. Why the Last Glacial Maximum (LGM)?

The LGM is an important topic for paleoclimatic study today, as it was 30 years ago at the start of CLIMAP, for many of the same reasons:

- The LGM represents a global climate state dramatically different from that of today, and thus provides a useful test of climate models' sensitivity to change. Although other extreme states of climate may also exist, the clearly documented influence of global ice sheets on climate (Clark et al., 1999) justifies a thorough understanding of the climate state dominated by ice.
- The LGM is reasonably close to an equilibrium state of climate (to the extent that equilibrium is ever obtained in the recent geologic past); short-term variability appears to be smaller during the glacial maximum than in older intervals such as marine oxygen isotope

stage 3, which is dominated by numerous short-term climate oscillations.

- Primary boundary conditions for the LGM climate, such as continental geography, orbital configurations, and atmospheric pCO_2 are extremely well known. Others, such as sea level, ice area, and ice-sheet heights are reasonably well known, and refinements are imminent.
- The LGM interval is within the range of both ^{14}C and U-Th dating, and additional methods such as surface exposure dating with cosmogenic nuclides offer the potential for unified chronologies on land and in the sea. Major advances in this regard have been made since the CLIMAP study, including calibration of the radiocarbon timescale to so-called calendar ages, and the advent of accelerator measurements of ^{14}C and other cosmogenic isotopes in extremely small samples.
- An extensive set of LGM geologic data and model runs based on both CLIMAP and post-CLIMAP studies already exists. Refinements to such work will be efficient and cost effective.
- Tremendous growth in computing technologies has advanced the state-of-the-art in climate modeling, particularly with the development of coupled atmosphere-ocean models with a range of resolution and complexity.

The relatively few numerical climate models that existed at the time of CLIMAP were limited to simulations of "perpetual" seasons (Gates, 1976; Manabe and Hahn, 1977; Kutzbach and Guetter, 1986). This modeling limitation helped to motivate CLIMAP's experimental strategy of estimating boundary conditions of "February" and "August" sea-surface temperatures, in spite of questions about the statistical independence of such estimates. In contrast, today's models allow simulation of full seasonal cycles that include more realistic feedback processes such as moisture and vegetation feedbacks at the land surface, more precise depictions of topography of the land and the sea floor, and elements of an interactive ocean. Some models can now be run through thousands of years of simulation, allowing exploration of multiple stable states of the climate system and better tests for the statistical significance of relatively small changes. With this growing sophistication of modeling, an opportunity arises to use paleoclimate data in a new way for model verification, rather than as fixed boundary conditions.

4.1. Models and mechanisms

Recently, an array of 16 different atmospheric general circulation models (some with mixed-layer ocean components) was tested using CLIMAP LGM conditions (SST as either a boundary condition or as a verification data set) as part of the Paleoclimate Modeling Intercomparison Project (PMIP). A summary of these model runs (Pinot et al., 1999) supports earlier inferences that CLIMAP LGM SSTs may have underestimated ice-age cooling in the tropics, but also reveals that LGM cooling is clearly more pronounced over land than over the oceans for all models. CLIMAP (1981) estimated that average tropical SSTs cooled by 1.4°C at the LGM. In the eight PMIP models that used these SSTs as a boundary condition, atmospheric cooling at the sea surface was 0.8 ± 0.1°C in the tropics, but over land the average cooling was 2.4 ± 0.4°C. In the eight models that simulated SSTs with simple "mixed-layer" oceans, average cooling over the sea surface was 2.1 ± 1.0°C, while over the land surface it was 3.4 ± 1.2°C. However, not all the models agreed about the overall magnitude of temperature changes. For example, LGM cooling over the tropical continents varied between 1.8 and 5.4°C in different models, implying that GCMs could be further improved. A key element that all of these models lack is an interactive ocean that includes processes such as ocean currents and upwelling of cool, subsurface waters.

Experiments using the current generation of coupled or partially coupled atmosphere–ocean models underscore the importance of re-assessing the state of the LGM world, including the circulation and properties of the deep sea. Here we highlight three of these efforts, along with the questions that arise from their results.

Ganopolski et al. (1998) considered a model with a dynamical three-dimensional atmosphere of low spatial resolution (10° latitude by 51° longitude grid spacing, 16 vertical levels) coupled synchronously to a zonally averaged multi-basin ocean (Atlantic, Pacific, and Indian Oceans). Ocean currents were parameterized based on latitudinal vorticity balance and meridional component of Ekman transport. The model calculated sea ice thermodynamically as the fraction of ice cover in each grid cell and transported it using advection and diffusion. On land, vegetation was prescribed, and runoff was simulated by assigning excess moisture accumulating at each land grid point to an appropriate point in one of the two-dimensional ocean basins. The simplicity of this model allowed for long model runs (5000 years) approaching equilibrium conditions even in the deep sea. Model output consisted of an average of the last 100 years of each simulation.

The LGM model run of Ganopolski et al. (1998) produced large atmospheric cooling at high latitudes, especially in winter in the northern hemisphere (Fig. 2a). The tropical oceans cooled less, 2.3°C on average (Fig. 2b), while the average land surface in the tropics cooled by 4.6°C. Near the equator the greatest atmospheric cooling, 4–5°C, occurred over the Atlantic Ocean and the easternmost Pacific Ocean (although the model ocean is two dimensional and thus does not really have east-west structure). In the deep sea (Fig. 2c and d), the total rate of thermohaline overturn stayed about the same, but in the

LGM run the site of deepwater formation in the North Atlantic shifted southwards by about 20° of latitude (two grid boxes in the model). This southward shift of oceanic heat transport, accompanied by advance of sea ice in the North Atlantic, accounted for about half of the large climate response in the north (CO_2 and land ice accounted for the rest). In the Pacific Ocean, limited modern formation of North Pacific Intermediate Water expanded to reach 2-km depth and this water mass crossed the equator to ventilate parts of the South Pacific. This expansion was driven primarily by cooling of the atmosphere at high northern latitudes, linked to the North Atlantic region via hemispheric winds. During the LGM the total rate of deepwater formation in this model was similar to the present rate in the North Atlantic and greater than at present in the North Pacific; thus net northward heat transport increased. As a result of this northward heat transport, significant cooling also occurred in the Southern Hemisphere, especially near 50–60°S.

Weaver et al. (1998) performed a similar experiment, but at a higher resolution (1.86° latitude by 3.75° longitude grid spacing), using a more complete three-dimensional ocean model. The 19-level primitive-equation ocean included components for thermodynamic sea ice and ice-albedo feedback, and was coupled to a simple energy and moisture balance model of the atmosphere (i.e., heat and moisture are transported via diffusion). In this model atmosphere, winds that drive ocean currents were approximated from the energy balance using a statistical scheme. The effect of atmospheric $p\mathrm{CO}_2$ was imposed by assuming that LGM reduction of $p\mathrm{CO}_2$ was equivalent to a $3.2\,\mathrm{W\,m^{-2}}$ reduction in radiative forcing. Precipitation on land (which had no topography) returned to the ocean via 21 prescribed drainage basins. Length of runs was not specified, but the authors stated that all runs reached equilibrium.

In the Weaver et al. (1998) study, northern hemisphere cooling was strongest over North America and in the North Atlantic (Fig. 3a and b). This result is similar to that of Ganopolski et al. (1998) but was caused by a completely different mechanism in which a nearly 50% reduction in the rate (but not the location) of North Atlantic Deep Water formation (Fig. 3c and d) caused a net

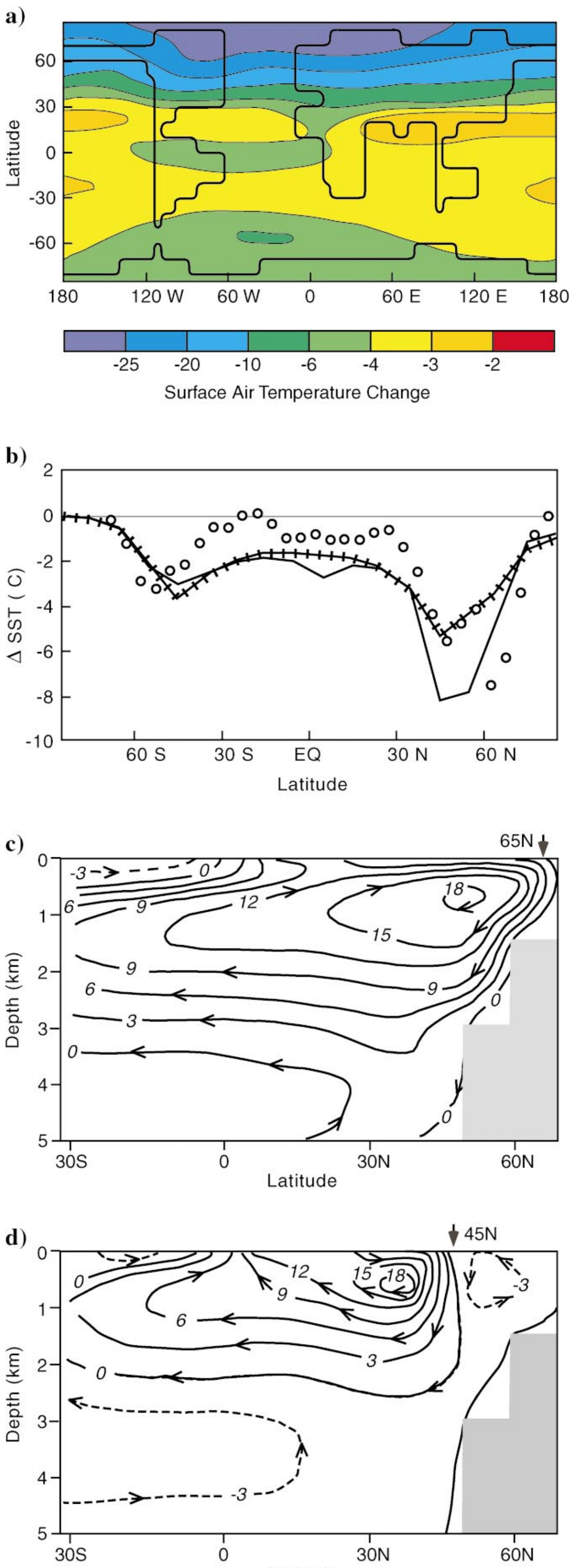

Fig. 2. Results from coupled atmosphere–ocean models of Ganopolski et al. (1998); (a) Surface air cooling (LGM — modern control simulation). (b) Sea-surface temperature change (LGM — modern), zonal averages of the two-dimensional ocean basins. Hatched line is a slab-ocean model. Solid line is the coupled atmosphere–ocean model, and the dots are the CLIMAP reconstruction. Note that the coupled model produces substantial cooling (> 8°C) in the northern hemisphere. Both models reveal somewhat more tropical cooling than CLIMAP (1981). (c) Meridional overturning streamfunction, modern Atlantic basin, (d) as in c, but for LGM. Note the southward movement of the deepwater source area to $\sim 45°$N, and total Atlantic deep-water formation at LGM similar to today (~ 18 Sv).

reduction of northward heat transport (opposite the sense of change in the model of Ganopolski et al., 1998). As a result, oceanic heat built up in the South Atlantic and Southeast Pacific sector of the Southern Ocean, mitigating regional cooling caused by lower greenhouse gases, and resulting in little or no net change in sea surface temperature in the Southern Hemisphere. Oceanic cooling in the model did occur near Australia, and in the North Pacific (based on diffusion in the energy balance atmosphere from the Atlantic and North American cold spot). Cooling in the tropical ocean was small (about 2°C) and asymmetrical with latitude (greater cooling in the northern hemisphere) and longitude (greater cooling in the eastern reaches of each ocean basin). Climate feedback mechanisms were analyzed by sequentially turning each one off. Weaver et al. (1998) found that the most important mechanism for total cooling was radiative forcing associated with greenhouse gases, followed by ice-sheet albedo feedback. The effect of changing ocean circulation was primarily to change the spatial pattern of temperature changes.

Bush and Philander (1998) used a fully coupled three-dimensional dynamical atmosphere–ocean model (2.25° latitude by 3.75° longitude, 14 atmosphere levels, and 15 ocean levels). The computational cost of this model limited the control run to 40 years, and the LGM run to 15 years, with the last six years used for output. This model run was too short to reach a full-equilibrium solution for the ocean interior, so results were only documented for the tropics and subtropics, where the dominant ocean circulation is shallow and the approach to equilibrium is rapid. The LGM Ocean is significantly different in the model of Bush and Philander (1998) that in those of Weaver et al. (1998) or Ganopolski et al. (1998). Cooling of the atmosphere at high latitudes compressed westerly wind belts equatorward in both hemispheres, and thereby intensified both the westerly jets and the trade winds. As a result of greater subduction of cold water into the high-latitude thermocline and stronger equatorial upwelling, the tropical oceans cooled dramatically in the Bush and Philander (1998) model, for example by more than 6°C in the western Pacific.

The similarities and differences between these three model runs are instructive both for the difficulties involved in understanding the LGM climate and for the opportunities that lie ahead. Although forced with

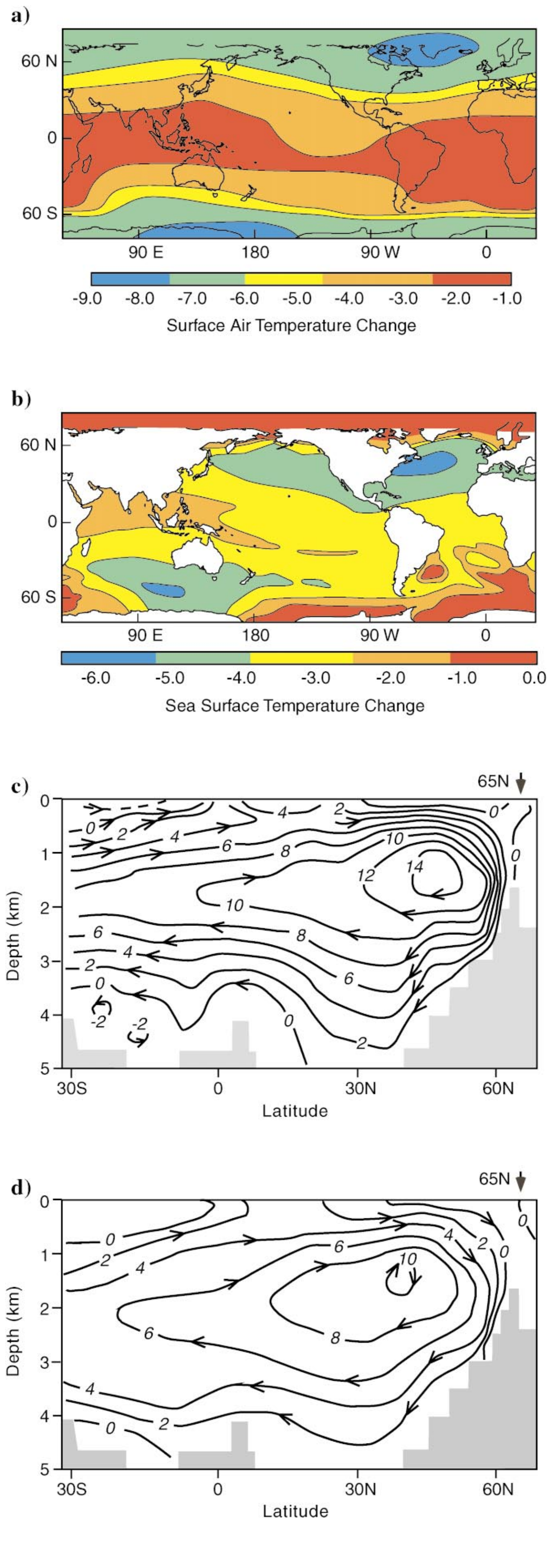

Fig. 3. Results from coupled atmosphere-ocean model Weaver et al. (1998): (a) Surface air cooling (LGM — modern control simulation). (b) Sea-surface temperature change (LGM-modern); note limited cooling in the southern ocean Atlantic and eastern Pacific sectors, a result of lower northward heat transport in the LGM. (c) Meridional overturning streamfunction, modern Atlantic basin, (d) as in (c), but for LGM. Note the stable location of the deepwater source area near 65°N, with a reduction of Atlantic deep-water formation at LGM (~ 10 Sv) relative to today (~ 14 Sv).

similar boundary conditions (modified radiation, reduced greenhouse gases, continental ice sheets), the three models give significantly different results and emphasize completely different mechanisms. The mechanism noted by Ganopolski et al. (1998) is primarily oceanic (sea ice and a shift in deep-water formation), whereas that of Weaver et al. (1998) is mostly atmospheric (CO_2 and ice-sheet albedo). It is interesting to note that in each of these models the primary mechanism that amplifies change occurs in the subsystem modeled in least detail. In the Bush and Philander (1998) study, the strongest mechanism is interaction of surface winds with the shallow ocean — a process that the other models do not include in detail. All three models highlight the importance of understanding three-dimensional circulation of the ocean interior, because of the role deep waters play in controlling large-scale heat transports from low to high latitudes, or between hemispheres. Other primary processes are also possible in this highly linked system, such as radiative feedback associated with water vapor (Webb et al., 1997), and changes in the transports of fresh-water vapor between ocean basins or from low to high latitudes (Zaucker et al., 1994; Rahmstorf, 1995; Hostetler and Mix, 1999).

Conflicting results from models and differences in key feedback mechanisms that set the sensitivity of each model to change beg the question, "what really happened during the LGM?" Before further progress can be made in modeling and understanding the mechanisms of ice-age climate change, a new comprehensive reconstruction of LGM climate is needed. This reconstruction can then be used to test whether mechanisms suggested by the models are reasonable or not.

5. When was the Last Glacial Maximum (LGM)?

The LGM can be defined as the most recent interval when global ice sheets reached their maximum integrated volume during the last glaciation. When did this occur? Based on stratigraphy available in the early 1970s, CLIMAP (1976) selected the celebrated benchmark of 18,000 ^{14}C-yr BP and assumed climate stability for the interval between 14,000 and 24,000 ^{14}C-yr BP. Not all records considered by CLIMAP included radiocarbon dates. In many cases, CLIMAP cores were dated by $\delta^{18}O$ stratigraphy, in which the LGM was defined as the midpoint of oxygen isotope stage 2 (Shackleton, 1977). In some cases $\delta^{18}O$ data were not available and the stage 2 boundaries were inferred based on some other regional lithostratigraphy or biostratigraphy, such as $\%CaCO_3$ (McIntyre et al., 1976) or the percentage of indicator species such as the radiolarian *Cycladophora davisiana* (Hays et al., 1976). Within that broad interval, a representative estimate was chosen based on an average of values within the LGM interval, or a single value at the depth of the highest $\delta^{18}O$ value, or in some cases the coldest inferred value with the general LGM interval.

Significant progress in geochronology and paleoclimatology made during the 20 years following the CLIMAP project allow us to revisit this issue. During the first EPILOG meeting participants, assessed new evidence and reached an agreement on a practical time window for an LGM reconstruction. Three separate issues arose: (1) establishing a clear definition of the LGM, (2) assessing the precision and accuracy of LGM dates, and (3) determining the timing and duration of LGM conditions on a global basis.

5.1. *Defining the LGM based on proxy data*

We continue to define the LGM as the time of the most recent maximum in globally integrated ice volume. Within this interval, however, we seek a practical definition of the largest chronologic interval within which global climate was reasonably stable, while avoiding major climatic shifts or transients. Much work done within and since the CLIMAP project has confirmed that regional temperature records do not necessarily reach minimum values at the same time, so a definition based on the coldest observed temperature or some other local extreme conditions would be inappropriate. Similarly, it is unlikely that all ice sheets and glaciers on the earth reached a synchronous maximum. The challenge is thus to characterize unambiguously the maximum of continental ice volume in paleoclimatic archives that integrate global effects, to establish the timing of such effects, and to find ways of linking various local or regional climate records into this global framework.

Two independent methods are commonly used to assess globally integrated ice volume. First, the oxygen isotopic composition of the mean ocean water, $\delta^{18}O_{mow}$, changes in response to storage of ^{18}O-depleted ice on land (Emiliani, 1955; Shackleton, 1967). $\delta^{18}O$ is 1000 $(^{18}O/^{16}O_{sample}/^{18}O/^{16}O_{reference} - 1)$ and reference is Standard Mean Ocean Water (SMOW) for water or atmospheric oxygen samples, or Pee Dee Belemnite (PDB) for calcite samples. Second, sea level changes relative to the modern baseline as water is removed from the ocean to build ice sheets. This portion of sea level change is called equivalent sea level (ESL), which is inferred from various local records of relative sea level (RSL) by modeling the gravitational effects of loading by ice on land or water in the ocean basins.

Changes of $\delta^{18}O_{mow}$ have been inferred from different archives such as planktonic and benthic foraminifera (Mix and Ruddiman, 1985; Shackleton, 1987; Labeyrie et al., 1987; Mix, 1987), deep-sea sediment pore waters (Schrag and DePaolo, 1993; Schrag et al., 1996) and molecular oxygen from atmospheric bubbles trapped in polar ice ($\delta^{18}O_{atm}$, Sowers and Bender, 1995). To a first approximation the three techniques provide the same basic information, i.e., $\delta^{18}O_{mow}$ was significantly higher during the LGM than at present. However, ever since the pioneering work of Emiliani (1955) and Shackleton (1967), uncertainties about the exact magnitude of changes in $\delta^{18}O_{mow}$ have been debated. This is because $\delta^{18}O$ from each archive combines changes in $\delta^{18}O_{mow}$ with other sources of isotopic variability.

Planktonic foraminifera $\delta^{18}O$ curves are affected by local changes in ambient temperature, fluctuations of the regional balance of evaporation and precipitation, and transport of fresh water via atmospheric vapor between oceans or from low to high latitudes (Broecker, 1986; Mix, 1987). Near coastal sites, river runoff with $\delta^{18}O$ values lower than local seawater adds additional complexity (Pastouret et al., 1978). Local or regional anomalies that decouple planktonic $\delta^{18}O$ values from global ice volume effects also include glacial meltwater that enters the ocean (Kennett and Shackleton, 1975; Cortijo et al., 1997). Given all of these regional complications, as much as one-third or even half of the total variability in planktonic foraminiferal $\delta^{18}O$ curves may reflect local environmental changes, which do not necessarily mimic those of ice volume.

Benthic foraminifera inhabit the deep sea, which is generally more stable and uniform in its properties than are surface waters. Nevertheless, it is now recognized that most benthic foraminiferal $\delta^{18}O$ curves, with total ranges between 1.5 and 1.9‰, are also affected by significant deep-sea temperature changes. Estimates of abyssal temperature changes vary from 2°C (Shackleton, 1987), to 4.5°C (Dwyer et al., 1993), which is equivalent to a $\delta^{18}O$ range in calcite samples of 0.5–1.1‰, respectively. Considering likely changes in temperature and other local effects, foraminiferal $\delta^{18}O$ data suggest a glacial-interglacial change due to ice volume of 1.1–1.4‰. This is roughly equivalent to excess LGM ice volume of $48–59 \times 10^6\,km^3$. Here we assume modern ice volume of $32 \times 10^6\,km^3$, a mean $\delta^{18}O$ of modern glacier ice (including Antarctica) of -47‰ relative to SMOW, and mean $\delta^{18}O$ of LGM ice (including Antarctica) of -44‰ relative to SMOW (-37‰ for ice outside of Antarctica). The total ice volume implied by these assumptions is smaller than the maximum ice sheet reconstruction of Hughes et al. (1981), but near that of the minimum model ($52 \times 10^6\,km^3$ more ice than at present). For comparison, recent excess LGM ice volume estimates based on modeling of sealevel data range from $40 \times 10^6\,km^3$ (Peltier, 1994) to $52 \times 10^6\,km^3$ (Yokoyama et al., 2000).

Changes in the isotopic composition of glacier ice through time suggest the possibility that changes in $\delta^{18}O_{mow}$ may lag changes in ice volume by as much as a few thousand years during major transitions (Mix and Ruddiman, 1984). In addition, the deep ocean was probably not isotopically homogenous. The influence of ^{18}O-depleted meltwater during the deglaciation probably first reached the Atlantic Basin, which was closest to the former ice sheets. Using sediment data and box models, several authors suggested that the $\delta^{18}O$ record of the deep Pacific could have lagged that of the Atlantic significantly, depending on the deep-sea circulation rate that prevailed in the past, although such transient effects are most prevalent in the Atlantic ocean where most of glacial meltwater entered the ocean (Broecker et al., 1988; Bard et al., 1991; Duplessy et al., 1991). A finding that the $\delta^{18}O$ shift following LGM conditions in the deep Pacific appears to precede rapid changes in sea level (Mix et al., 1999a) suggests that deep-sea warming may have preceded major deglaciation (meltwater pulse 1A as dated by Bard et al., 1990a). An alternate scenario, based on a new finding of rapid sea-level rise near 19,000 cal-yr BP, may suggest that isotopic change in the deep sea lags behind the termination of the LGM interval (Clark and Mix, 2000; Yokoyama et al., 2000).

Another isotopic method of assessing isotope budget changes, based on the oxygen isotopic composition of molecular oxygen in air bubbles preserved in ice cores (here, $\delta^{18}O_{atm}$) also does not record the pure oceanic component of $\delta^{18}O_{mow}$ variations. Changes in $\delta^{18}O_{atm}$ also record the balance of marine and terrestrial production and consumption of oxygen by photosynthesis and respiration (the "Dole Effect"; Dole and Jenks, 1944). They may also reflect divergence of surface ocean $\delta^{18}O$ from that of the mean ocean, due to changes in the rate of isotopically depleted water vapor from low to high latitudes where it is preferentially transported into the deep sea (Broecker, 1986; Mix, 1992). The observed maximum of $\delta^{18}O_{atm}$ during the last ice age occurred very late (i.e., 15,000 cal-yr BP) in the well-dated GISP2 ice core (Sowers and Bender, 1995) relative to other estimates of the glacial maximum (near 21,000 cal-yr BP, see below) and the extent of change from 15,000 to 8,000 cal-yr BP is quite large (1.5%; Sowers and Bender, 1995). In addition, the $\delta^{18}O_{atm}$ record generated for the four last glacial cycles in the Vostok ice core (Petit et al., 1999) is clearly dominated by a $\sim$22-kyr cycle, which contrasts with the classical 100-kyr cycle thought to dominate Late-Quaternary ice-age oscillations (Imbrie et al., 1993). Thus, it appears likely that regional climatic or biotic effects shift the timing and pattern of changes in $\delta^{18}O_{atm}$ recorded in ice cores relative to that of global ice volume. This limits the use of $\delta^{18}O_{atm}$ in determining the timing and amplitude of the LGM.

Schrag and DePaolo (1993) and Schrag et al. (1996) used direct measurements of $\delta^{18}O$ in deep-sea sediment pore waters to estimate the $\delta^{18}O_{mow}$ during the LGM. This technique, in which the record is deconvolved from data smoothed by diffusion, suggests a much lower range of $\delta^{18}O$ changes (0.8–1.0‰, Schrag et al., 1996) than that based on temperature-corrected foraminifera or corals (1.1–1.4‰, Labeyrie et al., 1987; Mix and Pisias, 1988; Fairbanks, 1989), or that implied by recent sealevel reconstructions (Clark and Mix, 2000; Yokoyama et al., 2000). The cause of this misfit is not presently understood. Future efforts to analyze chlorinity and hydrogen isotope composition of the same pore water samples may help to constrain the interpretation of porewater data (D. Schrag, pers. comm., 2000). Although useful for assessing the amplitude of change in $\delta^{18}O_{mow}$, the porewater method does not provide precise information on the timing of the LGM, and is not useful as a tool to locate LGM sediment samples, because the signal is heavily smoothed and transported from its original stratigraphic position by diffusion of pore waters within the sediment column.

Perhaps the most direct way to detect the glacial maximum is by documenting the low-stand of sealevel caused by the lockup of water on land-based ice sheets. It is possible to document deglacial sea levels by drilling tropical reefs and coastal sediments. To a first approximation all relative sea-level curves track the global sea-level changes that occurred since the LGM. However, second-order fluctuations superimposed on the common eustatic change are linked to regional and local effects (see atlas compiled by Pirazzoli, 1991). This complexity justifies the use of geophysical models which take into account all data on sea level and on the areal retreat of continental ice sheets, including comparison of sea-level records far from the ice sheets (Edwards et al., 1993; Bard et al., 1996) with those closer to the ice sheets (Fairbanks, 1989; Bard et al., 1990b). Results based on this approach range from a total equivalent sea-level change of about 105 m (Peltier, 1994), to 120–130 m (Fleming et al., 1998; Peltier, 1998a and b). However, new data from sediments on the Australian margin suggest a total equivalent sea level lowering of about 135 m (Yokoyama et al., 2000), which implies excess LGM ice volume of $52 \pm 2 \times 10^6\,m^3$. At present, no high-resolution sea-level curves based on dated corals extend through the full range of the LGM. Sea-level reconstructions consider only land ice, as floating ice shelves would displace water and thus would yield little or no change in sea level.

For purposes of defining the LGM interval, the EPILOG participants agreed that local glacial maxima or temperature minima do not provide a reliable stratigraphic link to the globally integrated glacial maximum. The best estimate of such an integrated maximum is based on the equivalent sea level low stand (i.e., after removing local isostatic effects), followed by benthic foraminiferal $\delta^{18}O$ (recognizing the presence of some temperature or watermass effects even in the deep sea). Because various components of the geologic record will not contain either of these measures, however, the most practical definition of the LGM horizon for purposes of mapping is one based on chronology. Links into this definition will inevitably use a variety of stratigraphic and chronologic approaches. This raises the question of accuracy of dates for the LGM.

5.2. Accuracy for dating the LGM interval

Radiometric dates for LGM events are most commonly based on radiocarbon. These radiometric dates can be adjusted to provide an estimate of calendar ages by correcting for changes in ^{14}C production rates and some reservoir effects. Calendar calibrations now extend through the LGM to 20,400 ^{14}C-yr BP (equivalent to 24,000 cal-yr BP in CALIB-4), based on a combination of ^{14}C-dated annually layered materials such as tree rings, as well as marine carbonates from the near-surface ocean dated with both ^{14}C and U-Th (Stuiver et al., 1998; Bard et al., 1998). A software program commonly used for calendar corrections of radiocarbon dates, CALIB, updated recently to version 4.2 (http://depts.washington.edu/qil/calib), is recommended as the standard correction scheme for EPILOG contributions.

CALIB uses a simple atmosphere–ocean box model to propagate inferred changes in atmospheric production of radiocarbon into the surface and deep-sea reservoirs, and is well constrained at accepted calibration points, but it does not explicitly assess changes in ocean circulation that may change the regional distribution of ^{14}C within the oceans. Thus, CALIB corrections are most accurate when applied to pristine samples from land (such as wood) or carbonate shells or corals that grew in the near-surface waters of the low-latitude ocean that are near equilibrium with the atmosphere. Calendar corrections in the deep sea or polar oceans are less certain, due to the large regional (and possibly temporal) variations in reservoir ages (e.g., Shackleton et al., 1988; Bard et al., 1994; Broecker et al., 1998; Sikes et al., 2000). Corrections for the LGM interval are less certain than in the Holocene, due to the limited number of well-constrained calibration points. It is likely that additional details, such as radiocarbon "plateaus" will emerge (e.g., Kitigawa and van der Plicht, 1998), but inclusion of such effects is premature at present.

Direct dating of corals using both ^{14}C and U-Th methods provides the best available estimate of the true age of the LGM based on dating the interval of lowest sealevel. The scarcity of LGM samples recovered from coral reefs and shallow-water sediments, however, is a major obstacle to this approach. Sites lacking major complications of tectonic uplift are now submerged and thus can only be sampled using ocean drilling — an expensive and logistically complicated proposition.

Existing coral data (Fig. 4) reveal rapid sea-level rise near 12,000 ^{14}C-yr BP (14,000 cal-yr BP). This event, commonly referred to as Meltwater Pulse 1A (MWP-1A, Fairbanks, 1989) postdates the beginning of the Bølling warm event in Europe (near 12,600 ^{14}C-yr BP, or 14,800 cal-yr BP). Recent data from organic debris off the Sunda Shelf in Southeast Asia are consistent with rapid sea-level rise, and radiocarbon dates here suggest timing between 14,100 and 14,700 cal-yr BP (Hanebuth et al.,

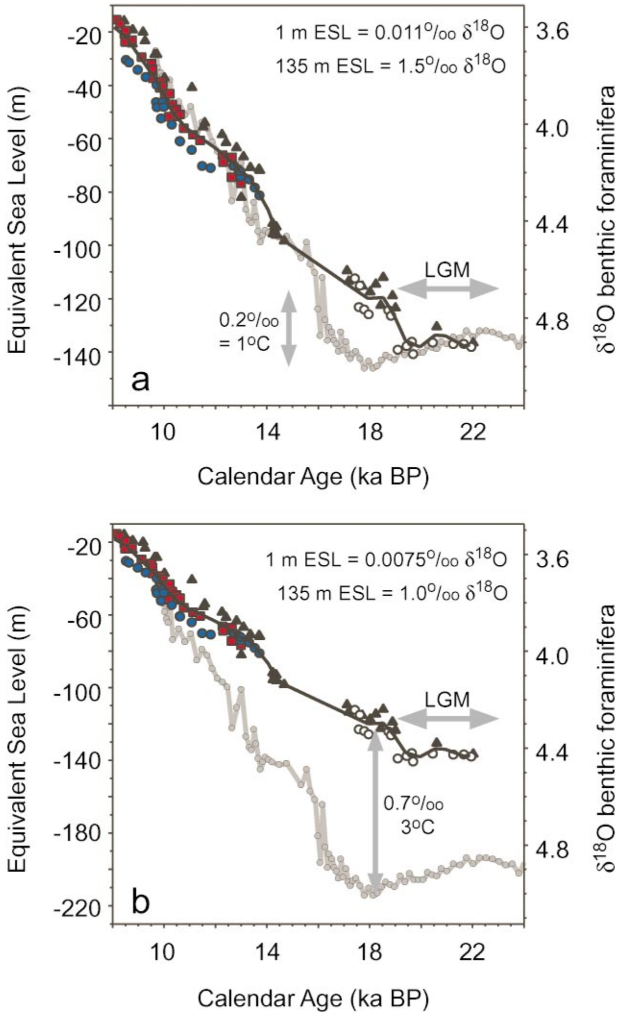

Fig. 4. Equivalent sea level (ESL) estimates based on coral data corrected for global isostatic effects (K. Lambeck, personal communication, 2000) compared to deep Pacific benthic foraminifera $\delta^{18}O$ (Mix et al., 1999a). Black triangles are based on data from Barbados (Fairbanks, 1989; Bard et al., 1990b), red squares are based on data from New Guinea (Edwards et al., 1993), and blue circles are based on data from Tahiti (Bard et al., 1996). Open circles are isostatically corrected sediment facies data from northern Australia (Yokoyama et al., 2000). The solid black line is a 1000-year block average of the ESL data, calculated every 500 years. Note rapid ESL rise near 19,000 cal-yr BP, which terminates the LGM interval. The gray line with small dots is a deep Pacific $\delta^{18}O$ record based on benthic foraminifera. Note rapid $\delta^{18}O$ change near 16,000 cal-yr BP, suggesting abyssal warming that postdates initial sealevel rise. The two panels span a range of possible scales for the relationship of ESL $\delta^{18}O$ in the deep Pacific. In panel (a) the sealevel and isotope records are scaled such that 1 m change of ESL is associated with 0.011‰ change of $\delta^{18}O$ (Fairbanks and Matthews, 1978). This scaling implies a total glacial-interglacial $\delta^{18}O$ change due to ice volume of 1.5‰, limited deep-Pacific cooling of about 1°C early in the deglaciation, and warming to essentially modern conditions near 16,000 cal-yr BP. In panel (b) the sealevel and isotope records are scaled such that 1 m change of ESL is associated with 0.0075‰ change of $\delta^{18}O$. This scaling implies a total glacial-interglacial $\delta^{18}O$ change of 1.0‰ (Schrag et al., 1996), substantial deep-Pacific cooling of about 3°C (relatively to modern bottom water temperature at the side of 1.7°C), and continual warming throughout the deglacial interval in a series of steps near 16,000, 13,000 and 10,000 cal-yr BP.

2000). U–Th dates at Barbados (which require no reservoir or calendar corrections as do the ^{14}C dates) constrain rapid sea-level rise of MWP-1A to younger than 14,300 cal-yr BP (Bard et al., 1990a).

The lowest stand of sea-level occurred prior to 16,000 ^{14}C-yr BP (19,000 cal-yr BP) (Fairbanks, 1989; Bard et al., 1990a, b) and probably around 18,000–19,000 ^{14}C-yr BP, or 21,200–22,400 cal-yr BP (Fleming et al., 1998). Yokoyama et al. (2000) provide new evidence from the tectonically stable Australian margin for a rapid sealevel rise near 19,000 cal-yr BP, which is preceded by a relatively stable 3000–4000 yr interval of an equivalent sealevel low stand $\sim$135 m below present. If confirmed, this early sea-level rise provides a logical younger limit to a well-defined LGM, centered at 21,000 cal-yr BP and extending from 19,000 to at least 23,000 cal-yr BP. This definition of the LGM interval is remarkably close in time to the original definition of CLIMAP (1976, 1981) of 18,000 ^{14}C-yr BP. One recommendation of the EPILOG group is that this LGM sea-level low stand must be understood better, through efforts to drill, sample, and date submerged coral reefs and shallow-water sediments.

5.3. Transient events and regional timing of the LGM conditions

Was the LGM a very brief event or an extended period of stable climate? This question is crucial as it affects the possibility of reconstructing and simulating the LGM as an equilibrium climate state. There is also a practical reason for evaluating the duration and regional timing of LGM conditions; geologic records are generally obtained as discrete data points rather than continuous curves, and it is rarely possible to assign exact dates to each data point. Thus, a useful definition of an LGM interval is needed to allow for reasonable error in dates, and for appropriate averaging of paleoclimatic proxies with different time resolutions. In addition, transitions into and out of the LGM climate state may have occurred at different times in various places on the globe. A final chronologic definition of the LGM should be based on calendar ages, and should be equally applicable (i.e., avoiding regional transient climatic shifts) in as many areas as possible.

Although sea-level data offer the best available integration of the global ice volume, they are less useful in assessing fine-scale variability within the LGM interval, and in many cases cannot assess regional differences in timing of major events. Since the CLIMAP (1976, 1981) studies, many high-resolution records have become available from ice cores and marine sediments that offer guidance about choosing a practical LGM interval. These show that the climate of the last ice age was far more variable than previously thought. In particular, the LGM does not always correspond to the coldest temperatures. For example, in the North Atlantic Ocean and surrounding continents, temperature minima typically occurred during Heinrich Events (HE) (e.g., Bard et al., 2000). These events are defined based on dating of sand layers (so-called Heinrich Layers) in North Atlantic sediments, which are inferred to be caused by ice-rafting (Heinrich, 1988; Bond et al., 1992). The LGM, as defined by sea-level records, probably took place in the interval bracketed between the end of HE-2 (19,500 ^{14}C-yr BP) and the start of HE-1 (15,000 ^{14}C-yr BP), two prominent events which have been dated precisely with accelerator mass spectrometry radiocarbon measurements on foraminifera (Elliot et al., 1998; Thouveny et al., 2000). These dates correspond to calendar ages of about 23,000 and 18,000 cal-yr BP, respectively. Global climate was clearly not at steady state during these events of massive surges and melting of icebergs, which affected many components of the ocean–atmosphere–cryosphere–biosphere system. The consensus of the EPILOG group is that Heinrich Events should be excluded as much as possible from an LGM reconstruction.

Other important regional climatic events may also need to be excluded from the LGM time window. For example, the so-called Dansgaard/Oeschger (D/O) Events, defined in Greenland ice cores, are brief episodes of warming in the North Atlantic Ocean and surrounding area. In particular, D/O-2, a prominent warm event, should probably be excluded from the LGM time window. It remains uncertain to what extent D/O Events are expressed throughout the globe, although it has been suggested that equivalent warm events exist in the North Pacific (Hendy and Kennett, 1999).

It is also important to consider the major regional climatic events that occurred in the Southern Hemisphere. Events of the high southern latitudes, including the end of extreme cold conditions associated with the LGM, are thought to have led those of the northern hemisphere (Imbrie et al., 1989; Charles et al., 1996; Clark et al., 1996). Southern Hemisphere polar events are so far best dated in ice cores from coastal and central Antarctica (i.e., at the sites of Byrd, Vostok and Taylor Dome), which have been correlated to those of Greenland.

Here (Fig. 5) we evaluate the potential for errors or biases in an LGM reconstruction associated with regional climatic changes or transient events, using well-dated high-resolution records of benthic foraminiferal δ^{18}O (North Pacific core W9809A-13PC; Lund and Mix, 1998, Mix et al., 1999a) and $\delta^{18}O_{ice}$ from the GISP2 and GRIP ice cores in Greenland (Grootes et al., 1993; Johnsen et al., 1997) and from the Byrd ice core in Antarctica (Johnsen et al., 1972).

The chronology for the deep-sea record of δ^{18}O from core W8709A-13PC is based on 41 AMS ^{14}C ages (36 within the window 10–30 kyr cal BP) measured separately on planktic and benthic foraminifera (Mix et al., 1999a). Radiocarbon ages were corrected to calendar ages using CALIB-4. In this record, benthic foraminiferal δ^{18}O values are remarkably stable in the interval around the LGM (Fig. 5a). This is probably not an artifact of smoothing caused by bioturbation, as sedimentation rates are high, about 15 cm ka^{-1} in this interval. However, the highest δ^{18}O values occur near 18,000 cal-yr BP and 26,000 cal-yr BP, both prior to and following the LGM interval defined by the lowest stand of sealevel near 21,000 cal-yr BP (Yokoyama et al., 2000).

The rapid change in benthic foraminiferal δ^{18}O near 16,000 cal-yr BP appears to precede the rapid rise in sea-level of MWP-1A near 14,000 cal-yr BP (Fig. 4) but postdates the early sea-level rise near 19,000 cal-yr BP (Yokoyama et al., 2000). If all these records are correct there is some decoupling of benthic foraminiferal δ^{18}O from sea level. The timing differences are larger than transients lags induced by ocean mixing (Duplessy et al., 1991). A solution may involve the timing of deep-sea temperature change (Mix et al., 1999a), and the dynamics of ice shelves (which affect the isotope budget but not sea-level) and land ice (which affect both the isotope budget and sealevel), Clark and Mix (2000). Thus, although these δ^{18}O data do not support rapid transient events of ice volume or deep-sea temperature during the glacial maximum, the apparent decoupling of sea-level and isotope records suggests that not all events are clearly recorded by δ^{18}O. For purposes of defining the LGM interval, however, a conclusion is that the highest value of benthic foraminiferal δ^{18}O is not precisely aligned with the lowest sealevel stand, and may even be offset in time by as much as a few thousand years. These differences, if confirmed, should be considered when employing δ^{18}O stratigraphy in the assembly of LGM data.

The most precisely dated deglacial record of water δ^{18}O in ice (which mainly reflects regional temperature of precipitation) comes from the Byrd core (Johnsen et al., 1972). Three independent techniques have been used to date this record. First, the electrical conductivity method (ECM) provides stratigraphic information on annual bands for layer counting (Hammer et al., 1994). Second, the record of $\delta^{18}O_{atm}$ fluctuations in the Byrd core has been matched to that of the GISP2 ice core in Greenland (Sowers and Bender, 1995; Bender et al., 1999), which was previously dated by layer counting (Alley et al., 1993). Third, methane concentration data from the Byrd core have been correlated to similar fluctuations in the GRIP ice core in Greenland (Blunier et al., 1998).

Correlation of the Byrd ice core to Greenland in the LGM interval (and by inference the relative timing of extreme cold conditions in the high latitudes of the southern and northern hemispheres) remains unclear. The variations of $\delta^{18}O_{atm}$ and CH_4 in this interval are small and not uniquely diagnostic, which leads to potentially significant chronological errors in the LGM time range. Indeed, Sowers and Bender (1995) acknowledge that "between 16,000 and 22,000 cal-yr BP, $\delta^{18}O_{atm}$ varied by less than $\pm$ 0.1 permil. Thus during this period, we

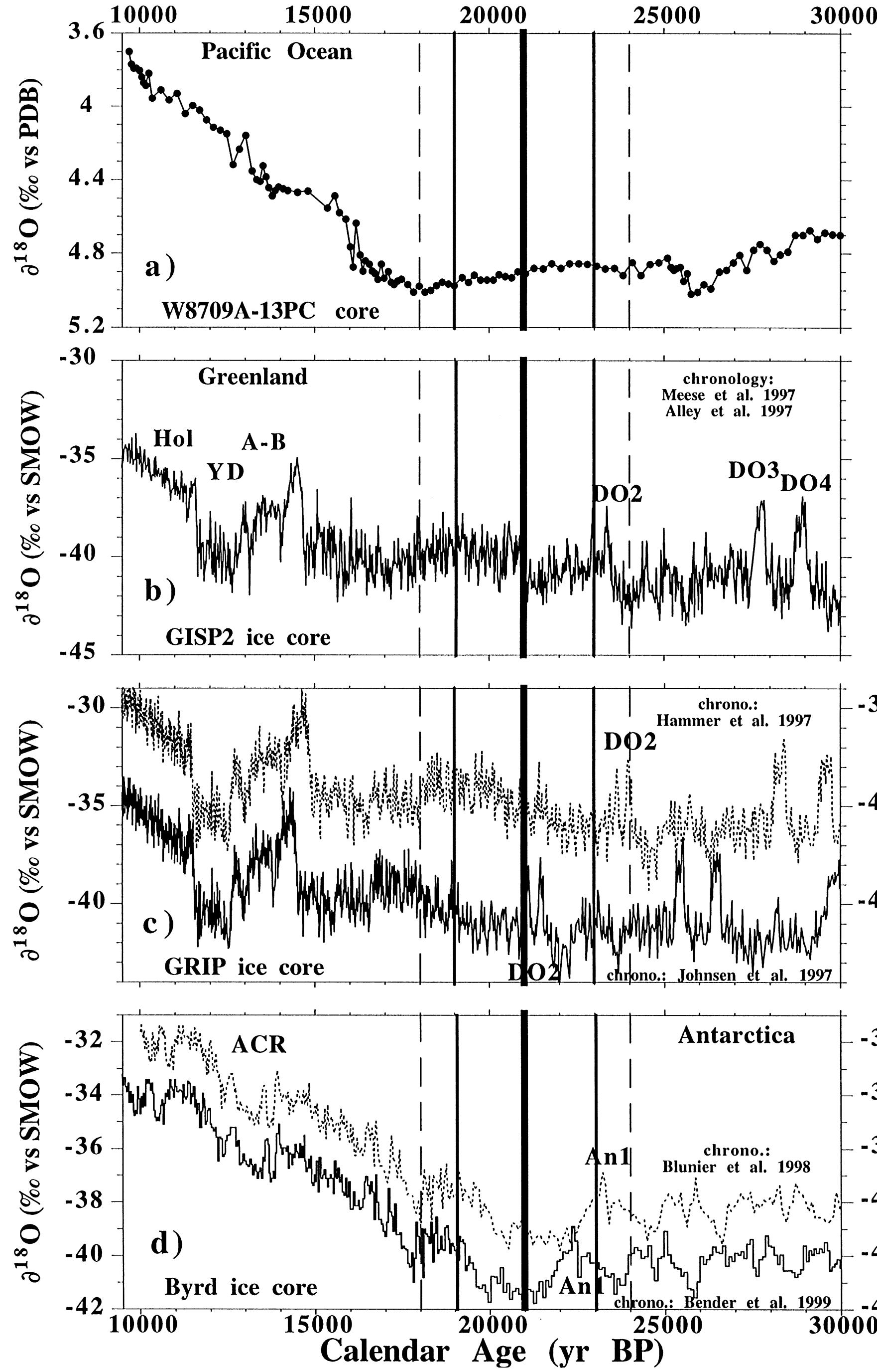

Fig. 5. Oxygen isotope time series for the time period 10–30 cal-kyr BP. Here, Hol is Holocene, YD is Younger-Dryas, A-B is Allerød-Bølling, ACR is the Antarctic Cold Reversal, and DO2 is Dansgaard-Oeschger event 2. An1 refers to the last $\delta^{18}O$ maximum before the LGM in the Byrd core from Antarctica. (a) $\delta^{18}O$ in benthic foraminifera from the Pacific core W8709A-13PC. The time scale is based on 41 AMS ^{14}C ages (36 within the window 10–30 kyr cal BP) measured on planktic and benthic foraminifera (Mix et al., 1999a), corrected to calendar ages using CALIB-4.0. (b) $\delta^{18}O$ in ice from the GISP2 core in Greenland (Stuiver and Grootes, 2000). The time scale is the GISP2 chronology available on the 1997 GRIP-GISP CD ROM (file gispd18o.dat). The main reference for the age scale is Meese et al. (1994) but the CD-ROM file includes the revisions by D.A. Meese as of September 1994, later published by Alley et al. (1997) and Meese et al. (1997). (c) $\delta^{18}O$ in ice from the GRIP core in Greenland. The lower curve with left y-axis is plotted versus the 'conventional' GRIP time-scale based on a flow-model (Dansgaard et al., 1993, Johnsen et al., 1997); i.e., files gripd18o.dat and gripage.dat of the GRIP-GISP CD-ROM). The upper curve with right y-axis shows the same $\delta^{18}O$ data plotted versus the stratigraphic chronology by Hammer et al. (1997); i.e., file gripstrt.dat of the 1997 GRIP-GISP CD-ROM. (d) $\delta^{18}O$ in ice from the Byrd core in Antarctica (Johnsen et al., 1972). The lower curve with left y-axis is plotted with the time-scale by Sowers and Bender (1995) slightly modified by Bender et al. (1999). The upper curve with right y-axis is plotted versus the time-scale of Blunier et al. (1998). Both Byrd chronologies are indirect and based on synchronizing geochemical proxies fluctuations measured in air bubbles ($\delta^{18}O_{atm}$ and CH_4) with the corresponding variations dated in Greenland ice. Consequently, the Byrd chronology by Bender et al. (1999) (lower curve in (d)) is compatible with the GISP2 chronology (b). Similarly, the Byrd chronology by Blunier et al. (1998) (upper curve in (d)) is compatible with the 'conventional' GRIP chronology (lower curve in (c)).

are unable to provide well-constrained time scales for other ice cores using $\delta^{18}O_{atm}$". Similarly, Blunier et al. (1998) state "the time period 17,000–25,000 cal-yr BP was excluded from the synchronization because only low CH_4 variations appear during that period, making the correlation uncertain". Recognizing the uncertainties implied by these difficulties in correlating the Byrd ice core to Greenland, the coldest (lowest ice $\delta^{18}O$, Fig. 5d) interval in west Antarctica appears to occur between about 19,500 and 22,000 cal-yr BP (Bender et al., 1999) or between 20,000 and 22,500 cal-yr BP (Blunier et al., 1998). Ice $\delta^{18}O$ values rise unambiguously above the range of typical glacial values by 17,000 cal-yr BP.

Within the Greenland ice cores, the timing of LGM events based on the "conventional" GRIP and GISP2 chronologies differ by about 2000 yr (compare Fig. 5b and lower curve in 5c). It is thus remarkable that the two Byrd chronologies developed by synchronization to either GRIP (via CH_4, Blunier et al., 1998) or GISP2 (via $\delta^{18}O_{atm}$, Bender et al., 1999) are quite close (Fig. 5d) even though the GRIP and GISP2 records in Greenland disagree by two millennia (Fig. 5b and c). This must be an artifact. Correlating methane and $\delta^{18}O_{atm}$ in the Byrd core to the same two data types in GISP2 would shift the record near the LGM by nearly 3000 years. The problem may lie in details of the records within the LGM interval. Consider the last $\delta^{18}O$ maximum of the LGM interval at Byrd (An1), illustrated in Fig. 5d. Bender et al. (1999) date this event at 22,300 cal-yr BP, while Blunier et al. (1998) place it 900 years earlier, at 23,200 cal-yr BP. For comparison, in the GISP2 ice core DO2 occurs almost 1,000 years earlier than An1 (Bender et al., 1999), while in the GRIP ice core DO2 occurs almost 2 kyr later than An1 (Blunier et al., 1998). Further synchronization of the Byrd stratigraphy to GISP2 based on CH_4 peaks now suggests an age for the ice $\delta^{18}O$ minimum between ca. 21,000 and 23,000 cal-yr BP (Blunier and Brook, personal communication, 2000), slightly older than the ages of Bender et al. (1999). Clearly some puzzles remain to be solved regarding ice-core chronologies near the LGM.

The isotopic record from the Taylor Dome (Steig et al., 1998) was dated by a combination of dating methods, including ^{10}Be fluxes and synchronization of CH_4 and $\delta^{18}O_{atm}$ records to those of Greenland. The δD record (i.e., the deuterium isotopic composition of ice) here is characterized by a rather late and abrupt deglacial warming near 14,800 cal-yr BP. This event appears to be synchronous with the start of the Bølling warm event as dated in Greenland ice. At Taylor Dome, the interval between 20,000 and 15,000 cal-yr BP exhibits cold temperatures with limited variability (note, however, a warm event centered on 16,000 cal-yr BP). Unfortunately the Taylor Dome record does not extend beyond 20,000 cal-yr BP. Thus it is not possible to assess the climate variability at this site over the full range of the LGM. Furthermore, a recent comparison of calcium records of Dome C and Taylor Dome suggests that deglacial warming at Taylor Dome is older than previously thought and synchronous with other sites in Antarctica (Mulvaney et al., 2000).

The Vostok ice core has a lower accumulation rate and thus a larger gas age − ice age difference than the cores from Antarctica coastal sites. These properties lead to larger dating uncertainties associated with correlation to the Greenland records. Indeed, discrepancies on the order of 2000–3000 years appear during the last deglaciation when comparing the Vostok chronology based on correlation to the GISP2 chronology (Sowers and Bender, 1995; Steig et al., 1998) with those based on the "conventional" GRIP chronology (Blunier et al., 1998). For example, the level in the Vostok ice core dated at 16,500 cal-yr BP by Blunier et al. (1997) is inferred to be 19,700 cal-yr BP by Steig et al. (1998). These discrepancies illustrate the overall uncertainties in the dating process (see also the discussion by Fischer et al., 1999). Nevertheless, it is still possible to conclude that the deglacial warming (based on rise of δD) in Central Antarctica started somewhere between 18,500 and 21,500 cal-yr BP. The latter date would imply that transient warming in Central Antarctica started during the LGM interval defined by sea level.

Given differences in the various ice cores, and the deep-sea records, it remains uncertain exactly when deglacial warming began on a global basis. Clear differences in timing and character of the records exist between Greenland and Antarctica, as well as between the various records within Antarctica. Significant dating uncertainties exist in the ice cores for the interval around the glacial maximum. Nevertheless, we can summarize broad regional differences based on available data. In the North Atlantic region, major warming was roughly coeval with the start of the Bølling-Allerød interstade, near 14,600 cal-yr BP, and extreme cooling events appear to be associated with Heinrich Events 1 and 2 (17,000 and 25,000 cal-yr BP, respectively). In Antarctica, significant deglacial warming began by 17,400 cal-yr BP, following initial warming that dates sometime between 19,500 cal-yr BP (Byrd core) and 20,000 ± 1500 cal-yr BP (Vostok core). Based on discussion at the EPILOG meeting, it was agreed that an LGM time slice should minimize the effects of this early warming in Antarctica, and avoid major regional transient events, if possible.

5.4. Choosing an LGM chronozone

Considering the sea-level constraints and the detailed records of regional climate change available from the ice cores, the EPILOG group reached a consensus that a preferred LGM chronozone (referred to here as LGM Chronozone Level 1) can be defined as the interval between 19,000 and 23,000 cal-yr BP (i.e., 16,100–19,500 ^{14}C-yr BP). This 4000-yr time window, centered on

21,000 cal-yr BP, encompasses the center of the LGM event defined previously by CLIMAP (1976, 1981), and is long enough to allow the inclusion of much existing paleoclimatic data in a new synthesis. It is coeval with the lowest stand of sea level (Yokoyama et al., 2000), avoids all known Heinrich Events in the North Atlantic region, and excludes most of Dansgaard-Oeschger climate event D/O-2, as dated in the GISP2 ice core and in the GRIP core (with the chronology of Hammer et al., 1997). The time window avoids the major deglacial warming in Antarctica, although it may include the beginning of the earliest warm events in the south (Blunier et al., 1998; Bender et al., 1999). This issue will require strengthening of ice core dating to resolve the discrepancies among the published studies.

The EPILOG group recognizes that realistic uncertainties in dating must be accommodated in any time-slice reconstruction, and suggests a second option for the LGM chronozone (Level 2) between 18,000 and 24,000 cal-yr BP (15,000–20,400 ^{14}C-yr BP; Bard, 1999). This is slightly broader than the preferred chronozone, and may partly include transient events such as DO2 in the North Atlantic, and the start of deglacial warming in the Antarctic.

Finally, recognizing that some records will have limited chronological control but may still provide useful information we suggest a third, less restrictive, definition of LGM Chronozone Level 3. Our purpose in providing these different levels of certainty is to strike a balance between reconstruction of a precise chronozone representative of the global glacial maximum, and inclusion of as much data and geographic coverage as possible.

- LGM Chronozone Level 1 19,000–23,000 cal-yr BP. Chronologic control based either on annually counted layers extending through the LGM chronozone, or two radiometric dates within the interval, such as U–Th dates or reservoir-corrected ^{14}C-yr dates adjusted to the calendar scale using version 4 of the CALIB software (Stuiver et al., 1998).
- LGM Chronozone Level 2 18,000–24,000 cal-yr BP. Chronologic control based on two bracketing radiometric dates of any kind within the interval 12 and 30 ka (i.e., within marine oxygen-isotope stage 2, or by correlation of non-radiometric data to similar regional records that have been dated to match the level 1 protocol (for example, δ^{18}O stratigraphy).
- LGM Chronozone Level 3 18,000–24,000 cal-yr BP. Chronologic control based on other stratigraphic constraints (for example, a regional lithologic index such as %$CaCO_3$) that are correlated to similar records dated elsewhere to match the level 2 protocol.

A variety of regional transient climate events may occur, and it is unlikely that any time-slice study will be able to avoid all of them. Thus, the EPILOG group explored the possibility of data treatment protocols to maximize the probability that an LGM reconstruction will approximate an equilibrium climate state. Our tests examined the three data time series of ^{18}O in Fig. 5 (one oceanic, W8709A-13PC from the North Pacific and two from ice cores, GISP2 from Greenland and the Byrd core from Antarctica). These examples do not reveal all possible issues with data treatments, but we consider them representative of the typical high-resolution paleoclimatic time series that might be analyzed in different parts of the world (whether based on isotope measurements or other data types). Our goal was to find data treatments that return stable estimates of the "typical" value and variability within the LGM chronozone, i.e., representative values that are insensitive to realistic errors in chronology or differences in resolution likely to be found in geologic records.

As estimates of the typical LGM value within the chronozone we considered (1) the data point closest to 21,000 cal-yr BP, (2) the arithmetic mean of all points within the chronozone, (3) the median, (4) the minimum, (5) the maximum, (6) the midrange (halfway between the minimum and maximum values), and (7) the trimmed mean (Tukey, 1970). For the trimmed mean, the data values in a sample are ordered, the highest 25% and lowest 25% of the values are removed, and the average of the middle 50% of values was taken. This method is useful for removing outliers and reducing skewness in non-Gaussian or noisy data sets. As measures of variability within the LGM chronozone, we considered the range and the standard deviation (without culling outliers).

We tested stability of each of these measures in each core to the time-slice definition (Chronozone Levels 1 and 2), to dating errors of $\pm$ 1000 years, and to data density. Dating errors were simulated by moving each time slice younger or older by 1000 years. The effect of sample density was assessed by removing samples from each timeseries to broaden the sample interval in steps until only a few samples remained within the LGM interval. For example, for the Byrd ice core dataset we resampled the 84 measurements within the Level 2 chronozone to 42, 21, 10, and 5 samples by iteratively culling every second data point.

While each method may work well in specific circumstances, we found that the most stable estimate of the LGM values was the arithmetic mean, followed closely by the trimmed mean. Values of individual data points nearest 21 ka, the minimum value, and the maximum value were all unacceptably sensitive to errors in dates or data densities. The median and midrange estimates were of intermediate stability in the data sets considered here.

For estimates of variability, the standard deviation was stable, even if only a few data values were available within a chronozone. The range of values in the chronozone was not a stable measure of variability because it was highly dependent on the number of data points included.

The differences in estimates of mean value and variability between LGM chronozone levels 1 and 2 were not significant, if the quality of dating was good; for example the difference in LGM mean $\delta^{18}O$ values between the two chronozone definitions are 0.02‰ at both W8709A-13PC and in the GISP2 ice core, within analytical errors of single measurements. In the Byrd ice core, the chronozone definition produced slightly more error, about 0.2‰, but still small relative to the total glacial-interglacial range. In this ice core, the LGM level is centered on a $\delta^{18}O$ minimum (i.e., cold extreme), and is bounded by relatively warm intervals. Dating errors of up to 1 kyr thus would induce small but significant differences (up to 0.4‰ $\delta^{18}O$) in mean LGM values of the Byrd core. Although larger than analytical precision of a single analysis, these errors are small relative to the ~ 10‰ contrast in glacial and interglacial values, and thus acceptable for purposes of producing a global reconstruction of the LGM.

Considering the tests done with these example datasets, we propose that EPILOG LGM data values be presented as the arithmetic mean and standard deviation of all values within the chronozone, noting the level of chronologic control included in the estimate. Although we can imagine cases in which it would be appropriate to cull clearly defined outliers from the LGM dataset, our tests did not reveal any reliable method that could be applied to all data sets. We recommend that individual investigators explore and document methods appropriate to their datasets. In all cases, analytical errors should be presented as standard deviations, and for samples that contain less than three data values within the chronozone the analytical error will be used as an estimate of the error for the mean value, but not as a measure of true variability within the chronozone.

6. New approaches to reconstructing LGM climate

6.1. The oceans

For the marine realm two methods are now commonly applied for basin-wide mapping of surface-ocean paleotemperatures; those based on species abundances of fossil plankton (using a variety of statistical techniques), and those based on organic geochemistry (using different types of alkenone unsaturation indices, typically $U^{k'}_{37}$). Oxygen isotope data based on planktonic foraminifera have been used in support of temperature reconstructions, but rarely as a stand-alone temperature index because of the large (and in some cases unconstrained) influences of ice-volume and local watermass effects. Inorganic chemical methods that have been explored but not yet widely applied include those based on trace metal elemental ratios such as Sr/Ca, U/Ca, and Mg/Ca in aragonitic corals or in calcitic shells of planktic or benthic foraminifera. Other indices such as those based on the shell size of certain species of planktonic foraminifera (Bé and Duplessy, 1976) have the potential to add information in specific areas but are probably not uniquely related to temperature or applicable in widespread areas. The EPILOG group discussed all methods in common use, based on both published work and ongoing developments.

6.1.1. Faunal and floral methods

Statistical prediction methods for estimating water-column properties from species assemblages are largely empirical. The classical transfer function methods (Imbrie and Kipp, 1971) resolve faunal assemblages in modern (core-top) samples, excluding redundant information via a factor analysis scheme. These assemblages are then calibrated in terms of sea-surface temperature or some other property of the upper ocean using multiple regression methods. The primary assumption is that modern spatial patterns of faunal abundance are controlled by the same mechanisms that control changes through time at any site. In theory, estimates of mean annual SST using statistical transfer functions do not require the fauna or flora to live at the sea surface, or in all seasons, or even to be controlled mechanistically by temperature. The method requires, however, that whatever properties actually control species distributions are linearly related to SST. If this chain of relationships varies through time, then estimates based on transfer functions may be biased.

Theory suggests that prediction of multiple environmental properties is possible from rich multi-species datasets (Imbrie and Kipp, 1971) but does not offer definitive guidance about which environmental properties are most important. Debate continues regarding the potential for biases in estimates and the difficulty in uniquely assigning variations in the fauna to one or another environmental property, in part because of limits imposed by intercorrelation among environmental variables in the field (Watkins and Mix, 1998). Nevertheless, most paleoceanographers accept that species assemblages can yield useful quantitative or semi-quantitative indices of upper-ocean temperature, as well as estimates of either biological productivity or pycnocline structure (noting that these two variables tend to be correlated in the modern ocean).

Transfer functions based on the Imbrie and Kipp (1971) method were the primary source of information for the CLIMAP (1976, 1981) reconstructions of the LGM oceans. Several faunal and floral groups were considered, including foraminifera (prevalent in the Atlantic and Indian Oceans; McIntyre et al., 1976; Kipp, 1976), calcareous nannofossils (which received limited use in the Atlantic and Pacific Oceans; McIntyre, 1967; Geitzenauer et al., 1976), Radiolarians (prevalent in the Pacific and parts of the Southern Oceans; Hays et al., 1976; Lozano and Hays, 1976; Moore et al., 1980), and diatom-rich sediments (primarily in the Southern Oceans, Cooke and Hays, 1982). Concordance of transfer functions based on different species groups strengthened the CLIMAP reconstructions in some areas such as the South Atlantic (Molfino et al., 1982) but led to questions and a difficult choice among conflicting estimates in others, such as the tropical Pacific (Moore et al., 1980).

CLIMAP (1981) chose to make seasonal reconstructions of the earth's surface, based on calibration to August and February conditions. Questions arose regarding the statistical significance of such seasonal estimates, because modern calibration values from the two seasons are highly correlated over large areas (Mix et al., 1986a, b; Pisias et al., 1997). CLIMAP chose these seasonal estimates in part because of limitations of computer models of the 1970s; model simulations typically were performed as "perpetual" seasons rather than full seasonal cycles (Gates, 1976). Such limitations no longer apply.

Recognition of possible problems and biases in statistical transfer functions, such as the estimate of seasonal properties and preservation problems (for example, Parker and Berger, 1971; Berger and Gardner, 1975) led eventually to the development of alternate statistical methods. Most widely used is the Modern Analog Technique (MAT), which attempts to match a geologic sample from the past with a set of modern samples containing a similar fauna or flora (Hutson, 1980; Prell, 1985). Environmental estimates are then based on an average, or weighted average, of a number of best analog modern samples. The basic assumption of MAT is similar to that of classical transfer functions; i.e., that modern spatial variability (in core-top samples) serves as a proxy for past temporal variability (down-core at a site).

The MAT differs from transfer functions in providing an opportunity to reconstruct the full ensemble of environmental properties associated faunal or floral populations. For example the MAT method might estimate a full seasonal cycle of upper-ocean temperature, pycnocline depth, and seasonal productivity for a single sample, simply by averaging all the relevant atlas values in the analog sites. It is not clear to what extent all of these estimates are statistically significant. Clearly only the properties that truly control the populations would be reliable, while other estimates would reflect correlation of properties within the calibration data set. Imbrie and Kipp (1971) note that for a sample set containing n statistically independent variables (such as faunal factors), at most $n-1$ environmental properties can be reconstructed.

A variety of schemes have been explored to make an average of the environmental properties in an array of best-analog samples, including a simple arithmetic average (Prell, 1985), weighting by geographic distance (the SIMMAX method of Pflaumann et al., 1996), or gridding and smoothing the core-top data along property gradients (the RAM method of Waelbroeck et al., 1998). In some cases, species percentage data is modified with a log transform, to minimize problems associated with species dominance (Guiot, 1990; deVernal and Hillaire-Marcel, 2000).

Although MAT methods have been applied mostly to planktonic foraminiferal data, they have also been extended to other planktonic microfossil assemblages (Abelmann and Gowing, 1996; Abelmann et al., 1999; Brathauer and Abelmann, 1999; Crosta et al., 1998a). Prell (1985) shows that the MAT and classical transfer-function methods yield roughly similar results in the tropics. Further comparisons between methods using both living (sediment-trap) and fossil (core-top) assemblages suggest somewhat less bias in estimates for the MAT than for transfer functions (Ortiz and Mix, 1997), although a sparse data set for calibration such as exists in the North Pacific may yield imprecise or noisy estimates.

A potential problem for both transfer function and MAT methods occurs when a lack of suitable modern analogs exists for a geologic sample. This is called the "no analog" problem. Such problems have been known for many years, for example in the eastern tropical Pacific and Atlantic Oceans (Moore et al., 1981; Mix et al., 1986a, b; Ravelo et al., 1990) and parts of the tropical Indian Ocean (Hutson, 1977). The primary problem is that faunal assemblages defined based on core-top samples do not capture the full range of faunal variation that occurred during the ice ages. Mix and Morey (1996) and Mix et al. (1999b) address this issue, and develop a revised transfer-function strategy in which down-core samples, including those representing glacial conditions, are used to define more robust faunal assemblages. These assemblages are then applied to core-top faunal counts, and new transfer functions are calibrated in the normal way. The down-core factor method improves the estimates in areas without appropriate modern analogs, but does not remove all potential biases from the transfer function; for example the precision of the estimate appears to be worse than the apparent precision of previous estimates. A primary result of this procedure (Mix et al., 1999b) is an estimate of LGM cooling in the eastern equatorial Pacific and tropical Atlantic oceans more extensive than that proposed by CLIMAP (1981), in better agreement with some geochemical data from oceanic sediments (for example, Hastings et al., 1998) and with information on land (Hostetler and Mix, 1999).

New mathematical procedures based on microfossil assemblages, such as artificial neural networks and response surfaces (Malmgren and Nordlund, 1997) are also under development, with a goal of better isolating the effects of interdependent environmental variables such as sea-surface temperature, mixed-layer depth, primary productivity; all having the potential to affect the composition of microfossil assemblages and induce biased estimates based on the other statistical methods.

Discussion of these methods by the EPILOG group suggested that all of the statistical methods are worth pursuing, with a goal of further assessing which method is most applicable in different situations. For the foraminifera, preparation of samples and taxonomic identification is well established and large surface data sets are readily available, which make the microfossil techniques a very useful approach for global reassessment of LGM temperatures. Other fossil groups are less well developed, and additional efforts will be needed to develop these groups. Further work is needed to compare the various statistical approaches. One important step will be to make faunal datasets available publicly. Although taxonomic concepts are reasonably well established for the major fossil groups, the community would benefit from regular intercalibration of taxonomies by sharing samples and data.

Although most workers accept that faunal and floral methods provide useful estimates of upper ocean properties such as temperature and pycnocline depth, uncertainties remain about the extent to which multiple environmental parameters can be extracted from a multi-species data sets. For example, it is unclear whether faunas in the tropics are most appropriately used to estimate temperature, pycnocline depth, or biological productivity, or whether the various influences result in biased estimates (Mix, 1989; Ravelo et al., 1990; Andreasen and Ravelo, 1997; Watkins and Mix, 1998; Beaufort et al., 1999; Cayre et al., 1999; Cayre and Bard, 1999). It is also unclear if faunal assemblages from the high latitudes make sense when applied to tropical samples, because the abundance of such polar species in equatorial sediments of the LGM have no modern analog and are thus outside the range of statistical calibrations based on surface sediments (Ravelo et al., 1990; Mix et al., 1999b).

Some species groups provide limited information at high latitudes. Foraminifera here have low species diversity that limits the precision of estimates at low temperatures. One possible solution, shown by U. Pflaumann, is to improve precision by counting a large number of specimens (perhaps thousands), so that relatively rare but important species may be identified reliably. Other species groups may also resolve this problem. For example, dinoflagellates are relatively diverse in the high latitudes, and thus may provide useful estimates of multiple water column properties in these regions (deVernal et al., 1994).

A potential problem for all methods that use core-top sediments to calibrate or confirm proxy data in terms of temperature or other upper ocean data lies in the assumption that core-top sediments reflect modern conditions. Many older sediment cores used in calibration studies have not been dated, and it remains unclear whether the core-top sediments are really modern. Bioturbation mixes older material into surface sediments. This is a special problem in areas of low sedimentation rates. For example, at sedimentation rates of less than $1\,\mathrm{cm\,ka^{-1}}$, bioturbation may mix material of 10,000 years or older into surface sediments (Peng et al., 1977). Because bioturbation moves material farther up the sediment column than it does downcore, LGM values are somewhat less sensitive to alternation of their mean by mixing, as illustrated by benthic foraminiferal $\delta^{18}O$ data (Zahn and Mix, 1991).

Following discussion of the EPILOG group, a consensus emerged that transfer function or MAT estimates of annual average and seasonal range (e.g., Mix et al., 1986a, b; Pisias et al., 1997; Pisias and Mix, 1997) would be more appropriate than separate seasonal temperature estimates. Mean and range values are essentially independent of each other in many calibration data sets, and thus more likely to be robust as independent estimates.

Discussion continues on the accuracy and precision of the SST estimates based on different statistical methods; for example no consensus was reached about the efficacy of geographical distance weighting between the site studied and best analogs (the SIMMAX method, Pflaumann et al., 1996) and smoothing or gridding of species data sets along environmental gradients (the RAM method, Waelbroeck et al., 1998) which both appear to improve precision of MAT estimates. Nevertheless, taking into account the errors of estimates (typically ± 0.8–$1.7^{\circ}C$) calculated for the various statistical methods, results are fairly consistent and the choice of one method or another may depend on specific regional issues.

6.1.2. Organic geochemical methods

Ketone unsaturation ratios have gained a lot of attention in the last decade as a new tool for reconstructing past SST. These proxies were discussed at the EPILOG meeting, but firm recommendations were deferred to a meeting focused on alkenones, which occurred in fall of 1999 at Woods Hole Oceanographic Institution (see the forthcoming special issue of G^3, edited by T. Eglinton, J. Hayes, M. Conte, and G. Eglinton; http://146.201.254.53/).

Unsaturation ratios are thought to represent temperatures during production of long-chain alkenones by haptophytes, a group of primary producers widespread in the ocean. Following early studies that identified the promise of alkenones as paleoceanographic proxies (e.g., Marlowe et al., 1984; Brassell et al., 1986; Prahl and Wakeham, 1987), it has been shown that alkenone measurements in modern (core-top) sediments support the concept of using the $U^{k'}_{37}$ index ($[C_{37:2}]/([C_{37:2}] + [C_{37:3}])$ as a reasonable measure of SST (Herbert et al., 1998; Müller et al., 1998). No significant diagenetic effects on the $U^{k'}_{37}$ index have been detected (Prahl et al., 1989; Madureira et al., 1995; Teece et al., 1998). Patterns of temperature change based on the index appear to behave in systematic ways through the last few hundred thousand years (Schneider et al., 1996, 1999; Bard et al., 1997). The alkenones thus provide a powerful tool for estimating LGM temperatures on a global scale, except for those regions where alkenone productivity and contents in sediments are very low, such as the centers of subtropical gyres, and the polar oceans where temperatures are very low.

As with all geologic proxies, care must be taken to evaluate potential biases and errors on a case-by-case basis. Second-order problems include the following: (1) Observations of peak alkenone production at subsurface depths suggest that the $U^{k'}_{37}$ index may not record true sea-surface temperatures in all areas (Prahl et al., 1993; Okhouchi et al., 1999); (2) Observations of seasonal blooms in alkenone production suggest that the $U^{k'}_{37}$ index may not always record annual-mean SST (Prahl et al., 1993); (3) Kinetic effects of cellular growth, observed to affect the calibration of the $U^{k'}_{37}$ index in some batch culture studies, suggest the potential for chemical or ecological biases (Epstein et al., 1998), although this was not observed in steady-state cultures (Popp et al., 1998), which raises questions about which type of culture experiments adequately represent the real ocean; (4) The finding of different relationships of $U^{k'}_{37}$ to temperature in different species or strains of haptophyte algae in laboratory cultures leads to uncertainties in the choice of temperature calibrations applied to the geologic record in some areas (Conte et al., 1998); (5) Erosion and redeposition at the sea floor suggest that alkenone signatures in some areas such as sediment drifts could be contaminated with material that is older or from another geographic area (Weaver et al., 1999; Benthien and Müller, 2000); and (6) Bioturbation in the sediment column, which appears to be strongest for fine particles that carry alkenones, may slightly shift and differentially smooth the stratigraphic record of alkenone indices relative to other proxy data carried on larger particles (Bard, submitted 2000).

The EPILOG meeting included a discussion of an international laboratory intercomparison experiment, coordinated by the European TEMPUS project (Rosell-Melé et al., 2000). In this coordinated effort 26 laboratories worldwide participated in an anonymous intercomparison exercise and analyzed (doing several replicates) five unmarked sediment samples and three alkenone standard mixtures with a wide range of $U^{k'}_{37}$ values and concentration of alkenones. The experiment evaluated whether there are differences amongst the laboratories results for $U^{k'}_{37}$ and alkenone absolute concentration that are significant for climate reconstructions, and if so to what extent this relates to the analytical methodology employed. In relation to$U^{k'}_{37}$, it is remarkable that despite being the first large exercise of its kind, most laboratories have provided data that are significantly similar (95% confidence). However, there is room to improve the methodology in some instances, and not all laboratories are capable of analyzing all the samples with the same proficiency. Quantification of alkenone concentrations yielded larger inter-laboratory variability than the ratio index $U^{k'}_{37}$. Such quantification, although desirable, appears to have little effect on temperature estimates in most cases. The most important aspects of the procedure are the quality of the gas chromatographic analysis and the skill of analyst, rather than the type of methodology. Chemical cleanup of organic extracts is not always required, but may be needed for some samples to obtain reliable alkenone data. A direct outcome of the exercise will be the recommendation of analytical guidelines to carry out reliable $U^{k'}_{37}$ and alkenone abundance determinations (Rosell-Melé et al., 2000).

A potential problem exists in some areas with very cold or very warm waters at both ends of the alkenone temperature calibration (e.g., Sikes and Volkman, 1993; Sikes et al., 1997; Sonzogni et al., 1997). Here it becomes difficult to properly quantify the minor peak on chromatograms ($C_{37:2}$ and $C_{37:3}$ for cold and warm temperatures, respectively) and analysis requires a high amount of alkenones preserved in sediments. A further problem is deviation of the temperature calibration from a constant slope near the cold and warm ends of the calibrated range. Such effects have been confirmed in cultures for some Haptophyte algae strains (Conte et al., 1998). However, in general the error of SST estimates is similar to that in microfossil assemblage techniques and ranges between 1 and 1.5°C, depending on whether a regional basin-wide or the global surface sediment calibration is used (Müller et al., 1998).

Alkenone records are often in agreement with other records in the low- and mid-latitude bands. For the mid-latitude sites studied by Prahl et al. (1995) and Pichon et al. (1998) there is a good agreement between alkenone results and temperature estimates based on radiolarians and diatoms, respectively.

Some relatively large differences have been noted in several instances. For example, to reconcile transfer function, alkenone, and $\delta^{18}O$ records measured in a core located off Mauritania, Chapman et al. (1996) invoked large changes of the seasonality of foraminiferal growth and coccolithophorid growth. However, further work on Mg/Ca in the same core (Elderfield and Ganssen, 2000) suggests a limited cooling of about 2–3°C during the LGM, in agreement with alkenone and $\delta^{18}O$ records. In an Indian Ocean core, temperature changes based on foraminiferal species systematically lead those based on $U^{k'}_{37}$ (Cayre and Bard, 1999), although the total range of temperatures is similar in both proxies.

In the subtropical Atlantic, regional $U^{k'}_{37}$ estimates exhibit less glacial-to-interglacial change than estimates based on Sr/Ca or U/Ca in corals (Guilderson et al., 1994; Min et al., 1995).

In the equatorial Atlantic, Mg/Ca and $U^{k'}_{37}$ records both suggest slight cooling (2–3°C) at the LGM, although the patterns of change through time are somewhat different, and Mg/Ca temperature estimates (measured in the foraminiferal species *G. sacculifer*) are cooler than $U^{k'}_{37}$ estimates in the same samples by 0–3°C (Nürnberg et al., 2000; Schneider et al., 1996). The authors attribute these differences to habitat preferences of *G. sacculifer*, which in this area is most prevalent during cool upwelling events of the austral fall and winter. In contrast, the alkenone producers in this region are thought (based on core-top correlations) to live in near-surface waters throughout the year (Müller et al., 1998). Transfer-function and modern-analog estimates of seasonal SST based on foraminiferal species distributions (Wefer et al., 1996) suggest larger temperature changes than $U^{k'}_{37}$ or Mg/Ca in the same core (Nürnberg et al., 2000). This careful multi-proxy comparison highlights the difficulty in judging which estimate is the most appropriate measure of mean annual sea-surface temperature.

High-latitude regions also display conflicts between indices. For example, in the Southern Ocean, alkenone-based temperatures are warmer by about 2°C than those derived from diatom species abundances (Pichon et al., 1998). For a North Atlantic site at 56°N, Sikes and Keigwin (1994) reported a similar amplitude and timing of deglacial warming with alkenone and foraminiferal species data. However, the SST estimates derived from alkenones are about 4°C higher than those based on foraminifera. Weaver et al. (1999) observed a similar pattern in a core located at 59°N. These authors suggest that both records could be reconciled by using a different calibration that assumes that alkenone data are representative of summer growth conditions rather than mean annual growth conditions.

One common problem encountered in sediments from the northern North Atlantic (Sikes and Keigwin, 1996; Weaver et al., 1999) is the rather low concentration of alkenones, which complicate the standard GC measurements. For example, Weaver et al. (1999) had to measure alkenones by GC-MS-CI in sediments with concentrations ranging between 0.1 and $10\,\mathrm{ng\,g^{-1}}$. Such North Atlantic sediments are probably more susceptible to contamination by reworking. Indeed, Weaver et al. (1999) observed that in the glacial section of their core, "the in-situ flora becomes very rare and the number of reworked Cretaceous and Paleogene specimens increases dramatically".

The EPILOG group consensus is that alkenone indices offer great potential for widespread reconstructions of upper-ocean temperatures at the LGM. More work is needed to sort out the specific biases (especially those of seasonal or subsurface production), but these efforts should not impede the use of alkenones as part of a multi-proxy strategy for reconstructing properties of the ice-age oceans. Alkenones may be especially useful in areas of relatively high biological production (such as near continental margins) and where other fossil groups are poorly preserved (such as in deep water where calcite tends to dissolve).

6.1.3. Trace metal methods

The search for new and independent paleothermometric methods has been accelerated recently by the apparent discrepancy between the classical CLIMAP reconstruction and the Sr/Ca ratios measured in corals from Barbados (Guilderson et al., 1994). This widely publicized study is based on a single site and is often used as evidence that CLIMAP's foraminifera-based SSTs were much too warm in the intertropical zone and that a severe cooling of about 5°C affected the warm pools during the LGM. Comparison of this result to other nearby data based on many techniques, however, suggest that cooling based on the Sr/Ca estimate is at the extreme of all methods. Crowley (2000) points out that cooling of 5°C in the tropics would fall below the survival temperature of most corals and would likely have restricted the LGM coral distribution to perhaps 5% of its present range, an outcome for which there is no evidence.

For corals, several authors have shown that besides temperature the recorded Sr/Ca can be affected by biological control on Sr and Ca uptake (Ip and Krishnaveni, 1991; Ip and Lim, 1991; de Villiers et al., 1995), growth rate effects (Weber, 1973; de Villiers et al., 1994), intra- or inter-species dependence (Weber, 1973; Alibert and McCulloch, 1997; Boiseau et al., 1997; Cardinal et al., 2000), skeleton heterogeneity (Allison, 1996; Hart and Cohen, 1996; Greegor et al., 1997), and bias in calibration due to distortions of seasonal signal (Barnes et al., 1995; Cardinal et al., 2000). The preservation of the Sr/Ca signal may not be always ideal because of diagenetic dissolution (Amiel et al., 1973; Cross and Cross, 1983; Yoshioka et al., 1986). The interpretation of the measured Sr/Ca may be complex because of changes of seawater Sr/Ca (deVilliers et al., 1994; Shen et al., 1996; Stoll and Schrag, 1998; de Villiers, 1999). Recently, Martin et al. (2000) proposed a new reconstruction of the oceanic Sr/Ca ratio that exhibits large secular changes with maxima on the order of 3–4% in phase with glacial maxima. The immediate conclusion derived by these authors is that glacial SST changes derived from Sr/Ca in corals have been overestimated by at least a factor 2.

U/Ca and Mg/Ca are also potentially useful indices of temperature in corals. However, U/Ca could be affected by changes of the carbonate ion concentration (Min et al., 1995; Lea et al., 2000) and/or the salinity (Shen and Dunbar, 1995). The interpretation of the measured U/Ca may be complicated because of intra- or inter-species dependence (Min et al., 1995; Cardinal et al., 2000). Similarly, the interpretation of the measured Mg/Ca in corals may be complicated by growth-rate effects (Swart, 1981). Several authors have observed unidentified scatter and/or discrepancies between calibrations (Schrag, 1999; Sinclair et al., 1998), and heterogeneous records within coral skeletons (Allison, 1996). In addition, the diagenetic dissolution is identified as a potential bias (Amiel et al., 1973; Cross and Cross, 1983). For both U and Mg in corals, more work remains to be done to assess these methods.

Determinations of Mg/Ca ratios in foraminiferal shells provide a promising new tool for estimating not only sea-surface temperature (Rosenthal et al., 2000) but also intermediate or deep-water temperatures when applied also to planktonic deep dwellers or benthic foraminifera and ostracods (Dwyer et al., 1993). One advantage of this method is its combination with measurements of oxygen isotopes at the same material, which offers the potential to estimate paleo-$\delta^{18}O$ of seawater. The combination of isotopes and elemental ratios then may help to constrain past salinity variations. Knowledge about paleo $\delta^{18}O$ water–salinity relationships in different regions is still needed for precise salinity estimates. Application of climate models that incorporate isotope tracers may help to estimate the paleo $\delta^{18}O$ of water vapor and river runoff (Rohling and Bigg, 1998; Schmidt, 1999; Delaygue et al., 2000). Efforts are now underway to build a global database of $\delta^{18}O$ and salinity data, to improve the basis for such modeling efforts (G. Schmidt, pers. comm. 2000; http://www.giss nasa.gov/data/018data).

Besides temperature, the recorded Mg/Ca can be affected by cleaning procedure, pH, and salinity (Lea et al., 1999). Several authors have observed complications such as intra- or inter-species dependence (Nürnberg et al., 1996), unidentified scatter or bias in the calibration, i.e., cultures vs. core tops (Nürnberg et al., 1996; Mashiotta et al., 1999), shell heterogeneities due to secondary growth of gametogenic calcite (Brown and Elderfield, 1996; Nürnberg et al., 1996). The influence of dissolution has been noted (Rosenthal and Boyle, 1993; Brown and Elderfield, 1996; Rosenthal et al., 2000).

Although the elements are simple to measure the method requires proper cleaning techniques of the calcitic shells, which should be standardized and calibrated between different labs. Since this method is fairly new no global or basin-wide surface-sediment calibration against atlas values exist for planktonic foraminifer Mg/Ca ratios. Culture experiments have been carried out so far only for different surface dwellers, e.g., *G. ruber*, *G. sacculifer*, (Nürnberg et al., 1996), and for *G. bulloides* (Lea et al., 1999). Caution is also needed because of potential bias due to calcite dissolution and ecological factors, e.g., seasonality and depth of calcification, as well as mineralogical heterogeneity in the shells (gametogenic calcite). These factors have so far kept this method from general use, but it should be considered an innovative technique that deserves further attention.

6.2. The continents

Building on earlier global compilations of land climates by CLIMAP (Peterson et al., 1979) and COHMAP (1988), Farrera et al. (1999) surveyed land climates of the tropics during the LGM. These studies define the LGM based on radiocarbon ages as 18,000 $\pm$ 2000 ^{14}C-yr BP. This definition is sufficiently close to the EPILOG calendar age definition of LGM Chronozone Level 1 (19,000–23,000 cal-yr BP), for comparison to new data from the ocean. A difference from the EPILOG definition, however, is that the Farrera et al. (1999) included at each site the single data point closest in age to 18,000 ^{14}C-yr BP, rather than an average of values within the LGM time interval.

Data sources on land include pollen, plant macrofossils, lake levels, noble gases in groundwater, and $\delta^{18}O$ from speleothems. The large international group contributing to the Farrera et al. (1999) study assessed which climatic variables were best inferred from their data, and concluded that estimates of mean temperature of the coldest month, mean annual temperature, plant-available moisture, and runoff (i.e., precipitation less evaporation) were most appropriate.

In this compilation, mean annual temperatures in tropical lowlands cooled by 2.5–3°C at the LGM, only about 1.5°C more than tropical ocean cooling inferred by CLIMAP (1981). The results varied by area, however, with LGM cooling estimated to be 5–6°C in Central America and northern South America, 1°C in New Guinea and the Pacific Islands, and 2–3°C in Africa and Southeast Asia. High-altitude sites suggest much greater average LGM cooling, however. For example, sites near 3000 m elevation cooled about 6°C in the Andes (Thompson et al., 1995), New Guinea and Africa (Farrera et al., 1999). This suggests significantly steeper vertical lapse rates (temperature change with altitude) during the LGM, accompanied by a weaker hydrologic cycle. Farrera et al. (1999) suggest that terrestrial lapse rates (defined as the vertical temperature gradient on the earth surface) may have diverged significantly from free-air lapse rates.

As noted above, modeling studies of the Paleoclimate Modeling Intercomparison Project (PMIP), summarized by Pinot et al. (1999), concur with the earlier study of Rind and Peteet (1985) that it is difficult to reconcile the CLIMAP (1981) ocean reconstruction with terrestrial evidence from the tropics. Atmospheric lapse rates varied slightly in the models, but not as much as implied by the data of Farrera et al. (1999) in some areas. Experiments with computed sea-surface temperatures in the models with interactive mixed-layer oceans suggested that tropical sea-surface cooling was slightly greater than that of CLIMAP (1981), and may vary regionally.

That lapse rates do not change substantially in atmospheric GCM's does not prove their stability, however, as these models parameterize (i.e., simplify) sub-grid scale vertical convection in the atmosphere. In a detailed, one-dimensional radiative-convective model, Betts and Ridgeway (1992) found significant changes in free-air lapse rates associated with surface winds (steeper lapse rate with lower winds), or poleward heat transport (steeper lapse rate with greater heat transport). They concluded that observed advances in tropical glaciers are possible even with CLIMAP's (1981) mean tropical cooling of 1.4°C, but would be more plausible with slight additional cooling of the sea surface totaling 2°C or more averaged over the tropics.

Crowley (2000) also assessed the effect of CLIMAP sea-surface temperature on continental climates. He argued that apparent misfits with tropical snowline data may not be as large as previously thought, and that relatively small changes in free-air lapse rate, coupled to modest additional sea surface cooling of about 1°C, would be enough to explain most of the terrestrial data. Hostetler and Mix (1999) find that regional cooling in the tropical Atlantic and eastern Pacific is sufficient to explain high altitude glaciers and changes in lake level in tropical South America and Africa, and especially note that parts of the Andes were wetter as well as cooler, in contrast to tropical aridity in the lowlands. Crowley (2000) and Crowley and Baum (1997) point to further work in understanding the roles of air–sea interaction, tropical convection, changing vegetation, and the effects of atmospheric moisture in setting model sensitivity. Resolving the effects of these variables in controlling LGM climates are important because all have implications for prediction of future climate changes.

Both the modeling studies, and the data compilation of Farrera et al. (1999) support the concept of LGM aridity in many areas of the tropical lowlands, although this concept has been challenged in some areas (Colinvaux et al., 2000). Other challenges for continental LGM reconstructions discussed during the EPILOG meeting included (1) the difficulty in dating many land records, (2) finding adequate methods for inferring regional patterns from sparse local records, and (3) the need for better methods to reconstruct of past lapse rates, including the differences between terrestrial lapse rates inferred by Farrera et al. (1999) and free-air lapse rates that are calculated in GCMs. Land proxies such as pollen or eolian dust in marine sediment cores may help synchronize climate information from land and sea. Analysis of regional climate models offers the potential to understand linkages between large-scale climate patterns in global GCMs and geologic data in specific locations. As with all geologic proxies, further work on intercalibration and coordination of methodologies will be needed for detailed global-scale reconstructions.

7. Toward a regional synthesis of the LGM

Here we assess progress toward a new reconstruction of the Earth's surface during the LGM interval in various regions. Although a full synthesis is beyond our reach, we highlight points of agreement, as well as unresolved controversies. Our primary purpose is to outline strategies that will lead to resolution of remaining conflicts, and to identify key opportunities to fill gaps in our current knowledge of the LGM.

7.1. High latitudes

In the North Atlantic region, several new LGM SST syntheses are currently in progress using transfer function or modern analog techniques applied to assemblages of foraminifera and dinoflagellates, as well as with organic geochemical methods. Preliminary results agree that ice-free surface waters existed during summer season in the eastern North Atlantic and Norwegian Sea, perhaps as far north as 70°N, but differences in estimates of sea-surface temperature have raised controversy.

In the Nordic seas, LGM temperature estimates based on foraminiferal modern analogs (SIMMAX method of Pflaumann et al., 1996) suggest relatively cold sea-surface conditions, near freezing in winter and between 2 and 5°C in summer (Sarnthein et al., 1995; Weinelt et al., 1996). In contrast, temperature estimates based on dinoflagellates (deVernal, 1993, 2000) and on the alkenone index $U^{k'}_{37}$ (Rosell-Melé and Comes, 1999), both suggest seasonally warm sea-surface temperatures, perhaps as high as 13°C — similar to or in some cases even warmer than modern values. This is surprising, given findings from ice cores in Greenland suggesting even greater cooling that previously estimated (Cuffey et al., 1995; Johnsen et al., 1995). Alkenone SST estimates may be questioned due to very low production of alkenone-forming organisms in cold water and the possibility of contamination with allochthonous older material. In addition, the alkenone estimates for very low temperatures may be biased by extraordinary amounts of the $C_{37:4}$ ketone reaching more than 10% of total C_{37} alkenones, a feature which may be related to reduced salinities (Rosell-Melé, 1998). Given these uncertainties, Rosell-Melé and Comes (1999) discount the traditional alkenone temperature index, but develop an alternate index that yields LGM temperatures associated with summer bloom conditions in the Nordic seas near 6°C.

The apparent disagreement between the LGM temperature estimates based on foraminifera and those based on dinoflagellates in the Nordic seas remains. To resolve this disagreement better knowledge is needed about possible differences in the depth habitats and the living seasons of foraminifera and dinoflagellates in the Nordic seas. Perhaps the dinoflagellates respond to environmental parameters other than SST. Perhaps they record temperature, but under no-analog conditions of a water column strongly stratified in summer by fresh-water inputs from melting glaciers. Perhaps the foraminifera record lack the resolution to resolve temperature due to dominance by a single species in cold environments, or perhaps they reflect subsurface conditions (Kohfeld et al., 1996). Discussion at the EPILOG meeting highlighted the importance of resolving the apparent differences between indices by better understanding the physical and biological controls on each biotic group.

The North Pacific remains a poorly known ocean; due to much more limited sampling than has occurred in other ocean basins and difficulties in defining the LGM chronozone due to poor preservation of carbonate. This region is particularly important as an upwind source of heat and moisture to North American ice sheets (Peteet et al., 1997), and will require focused attention. Here, CLIMAP estimates of LGM temperatures were uncertain, and subject to some artificial corrections to adjust for perceived transfer function bias (Moore et al., 1980). Since CLIMAP (1981), it has become clear that substantial changes in upper ocean structure of the NW Pacific were associated with increased intermediate water formation in the Sea of Okhotsk (Keigwin, 1998). However, the Gulf of Alaska appears to have been more stratified at the LGM than at present (Zahn et al., 1991). This stratification, along with strong cooling, appears to have yielded seasonal sea ice cover in the NE Pacific during LGM, based on dynocyst assemblages (deVernal and Pedersen, 1997).

In the northern source areas of the California Current, Ortiz et al. (1997) used foraminiferal and isotopic data to infer LGM cooling of about 4°C, driven by small southward translation of the subpolar front. Such cooling is consistent with other estimates based on radiolarians (Sabin and Pisias, 1996), and alkenone indices (Prahl et al., 1995; Doose et al., 1997). Conflicts appear off Southern California, however, where alkenone and radiolarian indices suggest relatively small LGM cooling of about 2°C (Herbert et al., 1995; Sabin and Pisias, 1996), whereas foraminifera and stable isotopes suggest larger cooling of >5°C in the Santa Barbara basin (Hendy and Kennett, 1999). The EPILOG group noted that the North Pacific will require much more work to define patterns and mechanisms of change during the LGM, and that apparent disagreements between proxies in the region may in some cases reflect true spatial variability of this large and dynamic region.

Progress in the Southern Ocean has come from better understanding of the ecology of diatoms and radiolarians at all latitudes (e.g., Zielinski and Gersonde, 1997; Welling and Pisias, 1998; Abelmann et al., 1999). Such information applied to down-core records has yielded better knowledge of open ocean temperatures (e.g., Pichon et al., 1992; Brathauer and Abelmann, 1999), as well as revisions to reconstructions of maximum extent and seasonal meltback (or fractional coverage of) sea ice (Burckle et al., 1982; Burckle, 1984, Burckle and Mortlock, 1998; Crosta et al., 1998a, b; Armand, 2000). Inferred sea-ice distributions, especially in southern summer, are substantially different than those of CLIMAP (1981), and this deserves further attention due to the importance of sea ice as a high albedo surface, as a thermal insulator decoupling ocean and atmospheric temperatures, and as an effective barrier to exchange of heat, water vapor and gases (such as CO_2) between the ocean and atmosphere (e.g., Ramstein and Joussaume, 1995; Stephens and Keeling, 2000).

Reconstruction of the Southern Ocean remain limited by the difficulty in linking local stratigraphic records into a global framework, due largely to the absence of calcareous fossils for stable isotope stratigraphy or radiocarbon dating in many southern high-latitude sediments. In some cases a lack of sediment cores (especially in the Pacific sector) also limits progress. Development of additional stratigraphic tools, such as isotope and/or metal indices in siliceous sediments, and further field studies, may be helpful in this region.

7.2. Low to mid-latitudes

For the low and mid latitudes new LGM SST compilations are developing based on a variety of faunal transfer functions and other statistical approaches noted above. A consensus appears to be emerging that sea-surface temperatures in the tropics were slightly but not severely cooler than those of CLIMAP (1981), and that temperature reduction was probably not uniform within the tropics (Broecker, 1986; Mix et al., 1986a, b; Broccoli and Marciniak, 1996; Pisias and Mix, 1997; Rosell-Melé et al., 1998; Bard, 1999; Mix et al., 1999a, b; Broccoli, 2000; Crowley, 2000).

CLIMAP's (1981) data coverage was sparse in the subtropics, especially in the Pacific. For example, recent data suggest significant LGM cooling near Hawaii (Lee and Slowey, 1999), where CLIMAP suggested either no change or seasonal warming. Broccoli and Marciniak (1996) note that many apparent conflicts between data and models occurred where CLIMAP interpolated results between sparse or uncertain data points. Significant conflicts between various proxy methods remain, especially between Sr/Ca data from corals and other data in the warm pool areas of the Atlantic and Pacific (Crowley, 2000), and such conflicts need to be resolved. Uncertainties also remain about the spatial pattern and gradients of low-latitude temperature changes, which in climate models have proven important in changing the distribution of precipitation (Hostetler and Mix, 1999; Yin and Battisti, 2000, in press).

CLIMAP (1981) reconstructed relatively little change in SST, about 1°C, in the western Pacific warm pool. More recent studies based on faunal transfer functions or Modern Analog methods have generally confirmed that finding, but have shrunk the region of stability. For example, foraminiferal estimates support stability of the warm pool, but also imply significant cooling off eastern Australia and in the marginal basins off southeast Asia, both areas where CLIMAP had little data (Anderson et al., 1989, Miao et al., 1994; Thunell et al., 1994). Barrows et al. (2000) confirm this result, but also note that the coolest time in this area may not be the LGM. These findings of small SST changes within the oceanic warm pool are supported by $U^{k'}_{37}$ data (Okhouchi et al., 1994), although Sr/Ca in corals from the region suggests substantial temperature changes within the Holocene (Beck et al., 1997), in apparent conflict with regional estimates based on foraminifera.

The extent to which upper ocean processes manifested in pycnocline depth and/or density contrast (Andreasen and Ravelo, 1997; Mulitza et al., 1997; Watkins and Mix, 1998; Ravelo and Andreasen, 1999; Wolff et al., 1999) can be assessed remains uncertain. Progress here offers the potential to reconstruct the strength of ocean currents based on paleogeostrophy calculations.

It was recommended during discussion of the various regional studies of low- and mid-latitudes that annual mean SST and seasonal contrast should be estimated rather than seasonal temperatures. This strategy accomplishes much the same thing as CLIMAP's (1981) estimate of seasonal temperatures, but helps to maintain statistical independence of the calibration data sets.

If the differences between the various proxies of temperature can be resolved and understood, additional insights into climate systems may come from combinations of proxies. For example, the combination of oxygen isotope measurements on foraminifera with independent estimates of temperature offers potential insight into changes in sea-surface salinity related to regional distributions of evaporation and precipitation, which is important in tropical climate systems.

Further progress on all proxies will come from coordinated study of the same samples with multiple proxies. The EPILOG group encourages public distribution of datasets from both modern and LGM samples via established databanks (for example, the National Geologic Data Center, NGDC, in the US, and the PANGEA databank in Europe). For proxies that require calibration with modern atlas data, we encourage all calibrations to be done with one data product of uniform quality. Use of the World Ocean Atlas, WOA98 (National Oceanographic Data Center, 1999) for purposes of proxy calibration is recommended, (http://www.nodc.noaa.gov/OC5/woa98.html).

8. Conclusions and general recommendations

Much progress in understanding climates of the LGM has occurred in the ~ 20 years since the end of the CLIMAP project. New geochemical methods and new statistical approaches to reconstruction of past environments reveal significant modifications are needed in the CLIMAP (1981) reconstruction of the surface of the ice-age Earth, and a new synthesis is warranted. Such a synthesis will provide an important benchmark for understanding the processes that set the sensitivity of the global system to change, and for evaluating climate models that will be used to predict future changes. Here we recommend key protocols and strategies so that the global community of earth scientists can join together in creating this new synthesis, which has come to be known as EPILOG (Environmental Processes of the Ice age: Land, Oceans, Glaciers). Specific findings and recommendations arising from the first open meeting of EPILOG, held in Delmenhorst, Germany in May 1999, include the following:

The EPILOG definition of the LGM is chronologic, based on ages 19–23 cal-kyr BP (Chronozone level 1), or 18–24 cal-kyr BP (Chronozone level 2). A level 3 chronozone is defined with the same ages as level 2, but with less certainty about dating. We recommend conversion of radiocarbon dates to calendar ages using the CALIB-4 correction scheme and that all chronologic data be presented in both uncorrected and corrected forms.

It is desirable to know both the mean value and variability of paleoenvironmental estimates within the LGM chronozone. We evaluated a variety of statistical treatments to assess these properties while excluding climatic transients or outliers, and found that in most cases the arithmetic average and standard deviation of available data points within the chronozone at each site is the most reliable estimate of each property. Our tests were not exhaustive, however. We recommend that individual scientists evaluate their data sets to find the best way of representing their data in a community reconstruction. An ultimate goal is to understand the transitions into and out of the LGM, as well and the mean state and variability within the chronozone.

A major change since the completion of the CLIMAP project is the rise of multi-proxy approaches, with new statistical methods applied to faunal and floral data, and new geochemical indices. With this new work, we note points of emerging consensus, such as that tropical sea-surface temperatures cooled slightly (but not dramatically) more than suggested by CLIMAP (1981), as well as points of controversy, such as seasonal temperatures and sea-ice cover in the high latitudes of both hemispheres. Resolution of these and other conflicts will likely yield deep insights into the proxies and their underlying systems, and into the processes driving global change. To facilitate intercomparison between proxies, and between various workers using similar proxies, EPILOG recommends (1) intercalibration experiments, following the model of TEMPUS (Rosell-Melé et al., 1998), (2) archival of available datasets in public databases, (3) efforts to share samples to the extent possible, so that comparisons among datasets can be made with rigor, and (4) continued efforts to obtain high-quality materials through field programs. For calibrations involving use of atlas data, we recommend use of a uniform dataset — for example in the ocean; the World Ocean Atlas 98 provides a suitable product.

It is clear that geographic coverage of LGM sample sets is insufficient for meaningful global reconstructions in the ocean or on land. Gaps in sample coverage are particularly acute in the Pacific, Antarctic, and Arctic Oceans, and on land areas outside of North America and Europe. However, attention to climatically important gradients in all regions is warranted.

Although the first EPILOG meeting concentrated on reconstruction of sea-surface temperature and related upper ocean properties as well as tropical land temperatures, a full understanding of the LGM will require a much broader perspective, including detailed reconstructions of glaciers and ice sheets, the land surface and its vegetation, a number of sea-surface properties including seasonal and annual temperature, salinity, productivity, and upper ocean structure, and changing character of the deep-sea and thermohaline circulation. Rather than making seasonal reconstructions of the sea or land surface we recommend estimation of annual means and season ranges of ocean properties. This accomplishes much the same thing as seasonal estimates, but helps to maximize statistical independence of real-world calibration data sets based on spatially distributed modern samples.

Desirable products for a new LGM reconstruction include the mean state of the LGM, variability within the LGM interval, and the transitions into and out of the LGM state. Creating a synthesis of such breadth is too large a task an individual or even a small group of scientists. Instead, we envision an open effort in which the global community of earth scientists contributes to a group synthesis. To facilitate this effort, we encourage the organization of regular topical meetings spanning several years. The next EPILOG meeting, to be held in October 2000 (facilitated by P. Clark, E. Bard, A. Mix), will focus on the reconstruction of the global ice sheet system.

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