PERGAMON

Quaternary Science Reviews 21 (2002) 1–7

QSR

Ice sheets and sea level of the Last Glacial Maximum

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Abstract

This paper outlines the general issues regarding ice sheets and sea level of the Last Glacial Maximum (LGM), which formed the basis of an EPILOG Project workshop. Papers in this special issue of *Quaternary Science Reviews* provide a comprehensive assessment of these issues from the perspective of geological reconstructions of ice sheet extent, records of sea-level change, ice sheet modelling, geophysical models of glacial isostatic adjustment, and geochemical proxies of ice volume. This new assessment has substantially narrowed the uncertainties in the total changes in ice sheets and sea level and their proxies, suggesting a net decrease in eustatic sea level at the LGM ranging from 120 to 135 m.

1. Introduction

The Environmental Processes of the Ice-Age: Land, Oceans, Glaciers (EPILOG) program is facilitating development of a comprehensive reconstruction of the state of the earth during (and the transitions into and out of) the Last Glacial Maximum (LGM), defined as an interval centered on 21,000 years ago, and to understand the processes involved in those changes (Mix et al., 2001). EPILOG is an open program that arose in 1999 as a multi-national working group of International Marine Global Change Study (IMAGES), a program of IGBP-PAGES. Through a series of workshops intended to highlight consensus and encourage collaborative work on outstanding issues, EPILOG is revisiting the landmark CLIMAP study of the LGM earth, which created maps of sea-surface temperatures, glacial ice, and albedo (CLIMAP Project Members, 1981). This attention, and particularly the mismatches between an evolving array of data and model predictions, yields insight into the sensitivity of Earth systems to change, the role of external forcing relative to internal feedback in driving natural change, and linkages between various climate subsystems. Further understanding of these issues is important for improving our understanding of climate processes and models that predict future climate changes.

Such a reexamination of the CLIMAP reconstruction is timely given that several of its key components have been questioned, and new methods developed since the CLIMAP project offer the opportunity for a more complete understanding of the ice-age climate system. One outstanding issue, that of ice sheets and sea level of the LGM, was the basis of the second EPILOG workshop, held in October, 2000, at Timberline Lodge, Mt. Hood, Oregon. For the ice sheets, CLIMAP presented two optional reconstructions: a "minimum" model in which ice margins were largely restricted to near continental margins and accounted for 127 m of ice-equivalent sea-level lowering, and a "maximum" model which included significantly expanded marine-based ice sheets and accounted for 163 m of sea-level lowering (Fig. 1; Table 1) (Denton and Hughes, 1981). The CLIMAP maximum reconstruction, which included a high-elevation single-domed Laurentide Ice Sheet, became the standard ice-sheet boundary condition used in many simulations of LGM climate with general circulation models. Nevertheless, significant challenges were mounted against this reconstruction based initially on the interpretation of field evidence for less extensive ice, particularly in the high Arctic (Dyke et al., 2002; Miller et al., 2002), and then on the basis of models of glacial isostatic adjustment that infer the distribution of ice loading by inverting relative sea-level records. These approaches suggested substantially thinner ice sheets

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0277-3791/02/$ - see front matter
PII: S0277-3791(01)00118-4

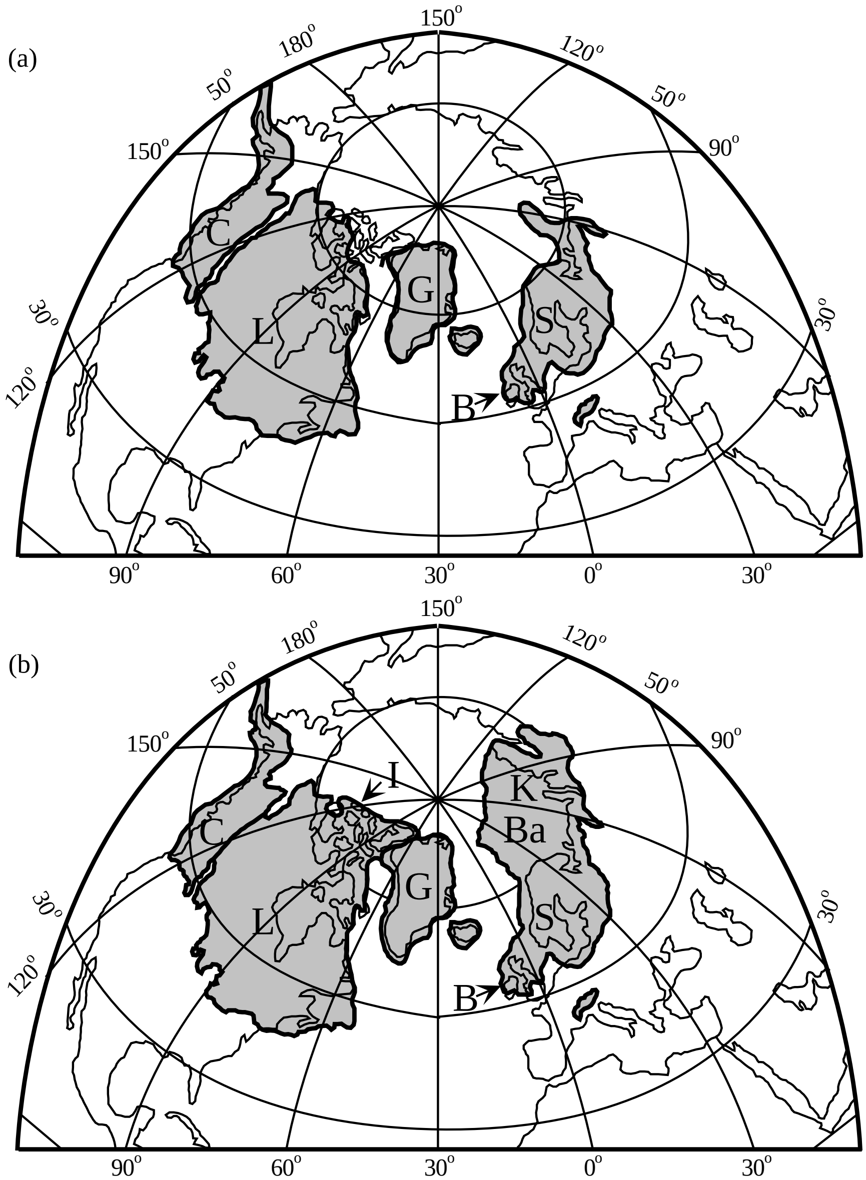

Fig. 1. Two alternatives of the area covered by the large Northern Hemisphere LGM ice sheets as reconstructed by CLIMAP (Denton and Hughes, 1981). (A) The CLIMAP minimum model. (B) The CLIMAP maximum model. Ice sheets are identified by letters: (C) = Cordilleran, L = Laurentide, I = Innuitian, G = Greenland, B = British, S = Scandinavian, Ba = Barents Sea, and K = Kara Sea.

that, by one estimate, accounted for ice-equivalent sea-level lowering of 105 m when including only ''explicit'' ice and 117.8 m when also including ''implicit'' ice (Peltier, 1998).

Other evidence of global ice volume has presented similarly conflicting results. For example, the growth of land-based ice caused lower sea-level and changes in ocean chemistry. Sea level records that come from drilling coral reefs can be dated precisely with U–Th and ^{14}C (with corrections to convert ^{14}C ages not associated with U–Th ages to calendar years to account for variable production of radiocarbon). Coral-based records have yet to sample the full LGM interval in detail, however, and currently only constrain relative sea level at Barbados to have been at least 120 m lower at 19 cal ka BP (Bard et al., 1990). A less direct record of ice volume comes from chemical proxy data—for example, the growth of continental ice sheets causes a change in the δ^{18}O composition of the global ocean. However, other factors (temperature, local salinity) that affect the isotopic signal measured in carbonate fossils partially obscure the ice-volume component.

Papers presented in this special issue of *Quaternary Science Reviews* address several key questions raised at the 2000 EPILOG workshop, including (1) the extent

Table 1
Estimates of excess ice-equivalent sea level for LGM ice sheets

Ice Sheet	CLIMAP Min (m)	CLIMAP Max (m)	Ice sheet Min (m) [a]	Ice sheet Max (m)[a]	Peltier[b] (m)	ANU[c] (m)	Milne et al. [d] (m)
Antarctica	24.5	24.5	14.0	21.0	17.6		
North America[e]	77.0[f]	92.0[g]	82.4	82.4	64.3		
Greenland	1.0	6.5	2.0	3.0	6.0[h]		
Scand/Barents[i]	20.0	34.0	13.8	18.0	25.5		
All others[j]	5.0	6.0	6.0[k]	6.0[k]			
Total[l]	127.5	163	118.2	130.4	113.5	130–135	115–135

[a] Based on ice sheet modeling (see Denton and Hughes, 2002; Huybrechts, 2002; Marshall et al., 2002; Siegert et al., 1999). No corrections made for ice grounded below sea level or changing area of ocean basins.
[b] Peltier (2002).
[c] Australian National University (Lambeck et al., 2002a).
[d] Milne et al. (2002). Allows for changes in area of ocean basins since LGM.
[e] Includes the Laurentide, Cordilleran, and Innuitian ice sheets.
[f] Laurentide contribution is 76 m.
[g] Laurentide contribution is 85 m.
[h] The 6 m number is known to be high. A recent glaciological reconstruction of the Greenland ice sheet (Tarasov and Peltier, submitted) suggests ~3 m.
[i] Includes Scandinavian, Barents, and Kara ice sheets except for the CLIMAP minimum model, which only includes the Scandinavian ice sheet. Ice Sheet Min and Ice Sheet Max from Siegert et al. (1999).
[j] All other LGM ice caps and glaciers.
[k] Value used is based on the CLIMAP maximum model estimate.
[l] Ice-equivalent sea level (S_i) can be converted to ice volume (V_i) by $V_i = (S_i A_o \rho_w)/\rho_i$ where A_o is area of the ocean and p_w and p_i are the average densities of water and ice, respectively. Note that this does not account for differences in ocean area between the LGM and present.

and volume of glaciers and ice sheets during the LGM, (2) the magnitude of global sea-level lowering as a result of growth of LGM ice sheets, (3) the isostatic response to the global redistribution of mass (ice and water) during the LGM, and (4) the relation between LGM ice sheets and global climate change.

2. Geological records of ice sheet extent

The extent of former LGM ice sheets is now reasonably well understood. Geophysical surveys of the Antarctic continental shelf suggest that the LGM extent of the Antarctic Ice Sheet extended to the shelf edge around much of the continent (Anderson et al., 2002), which would imply substantial change in ice volume in Antarctica. The LGM extent of the three North American ice sheets (Laurentide, Cordilleran, and Innuitian) is also well known, and is similar to that of the CLIMAP maximum reconstruction (Fig. 1) (Dyke et al., 2002), although the ice limit in several regions still requires refinement (Miller et al., 2002). Radiocarbon dating indicates that the Laurentide Ice Sheet, which was the largest of the former Northern Hemisphere ice sheets, may have advanced rapidly to its maximum extent as early as 27–29 cal ka BP, well before the LGM, and remained near that limit until ~17 cal ka BP (Dyke et al., 2002). In contrast, the Cordilleran Ice Sheet remained small at 29 cal ka BP and did not reach its maximum extent until 4000 yr after the LGM (Clague and James, 2002).

The outlines and ages of the LGM Scandinavian and British ice sheets are similarly well established (Bowen et al., 2002; Marks, 2002). They advanced to their maximum LGM position >26 cal ka BP along western margins and ~23.5 cal ka BP along southern margins, but were not coalescent at the LGM. Data support the existence of the Barents Ice Sheet during the LGM similar to that depicted in the CLIMAP maximum reconstruction (Landvik et al., 1992) (Fig. 1).

An ongoing debate concerns the extent of the Kara Sea Ice Sheet in northwestern Russia. Geomorphic evidence demonstrates that an extensive ice sheet once covered this region (Grosswald and Hughes, 2002), but most chronological data suggest that this large ice sheet existed early during the last glaciation or before, and that the LGM ice sheet was limited to the region west of Novaya Zemlya (Mangerud et al., 2002). Another debate centered on ice extent in eastern Siberia, where some geomorphic evidence suggests the presence of an ice sheet (Grosswald and Hughes, 2002) but radiocarbon-dated sediment records also question its LGM age (Brigham-Grette et al., 2001). The relative contribution of these Siberian ice sheets to the total extent of global ice is small (perhaps 5% of the total area), however, indicating that LGM ice sheet extent that contributes substantially to the global ice volume is reasonably well known.

3. Sea level: a global integrator of ice volume

Records of changes in global sea level provide the most direct means of determining changes in ice volume, but few of these records sample the LGM in detail. Furthermore, debate continues on how to best extract the isostatic component of these records in order to resolve the ice-equivalent (eustatic) component of sea level change.

Drilling into tropical coral reefs offers an excellent opportunity to develop relatively continuous sea level records. Such records have provided high-precision dates based on U–Th dating that are not subject to the uncertainties of reservoir ages and calendar corrections, as are radiocarbon dates. Although no record of dated corals spans the full LGM interval (Fig. 2), the Barbados record suggests an interval of rapid sea level rise prior to 13.7 ± 0.1 cal ka BP and younger than 14.2 ± 0.1 cal ka BP. This interval is referred to as meltwater pulse 1 A (MWP-1A) (Fairbanks, 1989; Bard et al., 1990).

Records of shallow-water sediment facies are preserved on submerged continental margins, but in some cases the links between facies and sea level may be imprecise. Nevetheless, these records provide valuable information on LGM sea level. Two new records of sediment facies from the Sunda Shelf, off Vietnam (Hanebuth et al., 2000), and Bonaparte Gulf, off northern Australia (Yokoyama et al., 2000), extend the detailed record of sea-level history into the LGM. Bonaparte Gulf and New Guinea coral data suggest a rapid drop to LGM sea level starting ~32 cal ka BP (Lambeck et al., 2002a). LGM sea-level may then have been terminated by a rapid sea-level rise starting at ~19.0 cal ka BP (Fig. 2), although the rate of change within this event is subject to uncertainties in the depth range of sediment facies in the Bonaparte Gulf (Yokoyama et al., 2000). The radiocarbon-dated record from the Sunda Shelf suggests a calendar-corrected age for MWP-1A of 14.6-14.3 cal ka BP (Hanebuth et al., 2000), earlier than suggested by U–Th dates from Barbados corals (Bard et al., 1990). This apparent conflict in the timing of MWP-1A must be resolved. In addition, identifying the source of MWP-1A remains an important issue in understanding the last deglaciation (Clark et al., 1996). The exact timing of transitions and rates of change remain open to debate, but with available data it appears that the low sea level at ~21 cal ka BP was relatively stable (Lambeck et al., 2002a).

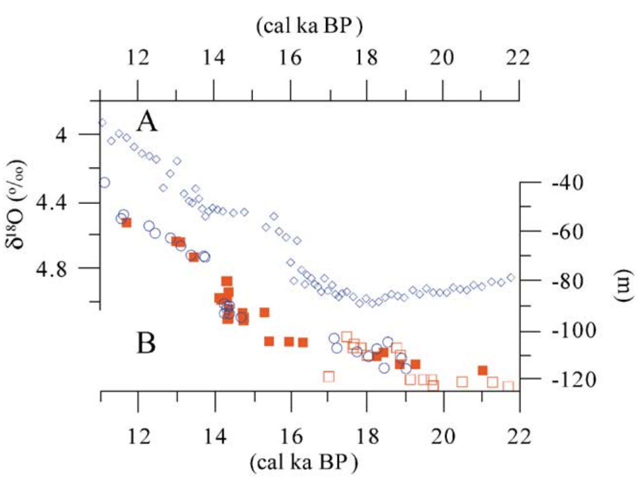

Fig. 2. (a) A benthic marine oxygen isotope record from the Eastern Pacific Ocean (Mix et al., 1999) compared to (b) relative sea level records from three sites: Sunda Shelf record (red filled squares) from Hanebuth et al. (2000); Bonaparte Gulf record (open red squares) from Yokoyama et al. (2000); Barbados record (open blue circles) from Bard et al. (1990).

4. Ice volumes based on ice sheet models

New ice sheet model estimates of ice volume (expressed as ice-equivalent sea level) in the large LGM ice sheets (Antarctica, Greenland, North American, Fennoscandian/Barents) range from 112.2 to 124.4 m relative to present (Table 1) (Denton and Hughes, 2002; Huybrechts, 2002; Marshall et al., 2002). Adding in the CLIMAP estimates of small ice caps and glaciers (cf. Barrows et al., 2002; Hulton et al., 2002; Owen et al., 2002) yields an additional ~6 m, suggesting a global contribution ranging from 118.2 to 130.4 m of ice-equivalent sea-level lowering at the LGM.

Current three-dimensional thermodynamic ice-sheet models have improved significantly since CLIMAP. Further progress in ice sheet modeling will come from better representation of processes such as iceberg calving, sensitivity of ice sheets to climate, ice deformation, and basal boundary conditions (Charbit et al., 2002; Hughes, 2002).

5. Ice volumes based on models of glacial isostatic adjustment

Growth and decay of ice sheets and concomitant changes of sea level caused global isostatic adjustments that continue today, thousands of years following the termination of the last ice age. The primary evidence for this glacial isostatic adjustment (GIA) comes from observations of relative sea-level change. For example, observations near and within formerly glaciated regions (near-field sites) are predominantly affected by the ice-induced deformation of the solid earth. These data can thus be used to infer local ice thickness (Shennan et al., 2002). None of these estimates includes the ice volume contribution from small ice caps and glaciers that CLIMAP estimated at 5–6 m.

Sea level data from sites distant from the former ice sheets (far-field sites) that are less affected by isostatic deformation are more useful for constraining the glacial meltwater signal and thus the total volume of land-based ice (Milne et al., 2002). Results from such studies of far-field sites give an LGM ice-equivalent sea-level lowering ranging from about 120 m (Peltier, 2002) to as much as 130–135 m (Table 1) (Lambeck et al., 2002a, b; Milne et al., 2002). A discussion and reply between Peltier and Lambeck et al. (2002), however, indicates that the solution may be leaning towards a sea level >120 m.

6. Geochemical proxies of ice volume

Geochemical records of oxygen isotopes ($\delta^{18}O$) from deep-sea cores and ice cores provide supporting evidence for the volume of ice at the LGM and for the timing of key changes in the ocean-climate system. Detailed marine records of $\delta^{18}O$ dated by radiocarbon reveal that the last isotopic maximum (LIM, centered near 18 cal ka BP and significant reduction beginning ~ 17 cal ka BP) may be younger than the LGM sea-level lowstand, which ended at ~ 19 cal ka BP (Fig. 2). If both the isotope and sea level records are correct, the differences in the timing of the earliest events of glacial termination imply some combination of (1) deep-sea cooling in the interval between these maxima (to maintain high $\delta^{18}O$ in calcite shells in spite of reduced seawater $\delta^{18}O$), (2) conversion of land ice to floating ice (which would raise sea level without lowering seawater $\delta^{18}O$), (3) a large lag in the propagation of the $\delta^{18}O$ signal of ice melting through the ocean system, or (4) changing isotopic composition of ice sheets during deglaciation (cf. Clarke and Marshall, 2002). The first rapid shift to lower $\delta^{18}O$ values in deep-sea foraminifera began near 16 cal ka BP, earlier than MWP-1A that occurred near 14 cal ka BP. This mismatch with MWP-1A suggests early warming of the deep sea, and implies that deep ocean circulation may have played a key role in the termination of the last ice age (Mix et al., 1999).

Significant debate exists regarding isotopic constraints on the total volume of LGM ice sheets. A growing data set of pore water $\delta^{18}O$ measurements, when deconvolved, suggests a total range of sea-water $\delta^{18}O$ change of about 1‰, with relatively small changes of 0.7–0.8‰ in the deep Atlantic, and slightly larger changes of ~ 1.1‰ in the Southern Ocean (Schrag et al., 2002). This finding seems to require cooling of the deep sea during the LGM in many places to nearly the freezing point. Estimates of the ocean's glacial–interglacial change in $\delta^{18}O$ based on benthic foraminifera (Duplessy et al., 2002; Waelbroeck et al., 2002) allow the possibility of such regional changes (constrained by the freezing point of seawater); however, regional distributions of foraminiferal $\delta^{18}O$ suggest that the deep Atlantic was about 2°C warmer than the Southern Ocean. A possible explanation could be that brine formation during freezing of sea ice pumped a relatively ^{18}O-depleted watermass into the deep North Atlantic during glacial time (Zahn and Mix, 1991).

An implication of this hypothesis that waters of anomalously low $\delta^{18}O$ (relative to the mean ocean) were pumped into the deep sea (yielding relatively low-amplitude glacial–interglacial changes in the deep North Atlantic) is that waters of the warm surface ocean would compensate by becoming anomalously high in $\delta^{18}O$ yielding high-amplitude glacial–interglacial changes (Mix, 1992). Lea et al. (2002) help to constrain such changes in the warm surface ocean by using Mg/Ca ratios to remove the effects of temperature from the $\delta^{18}O$ record in planktonic foraminifera. These results suggest average glacial–interglacial changes in $\delta^{18}O$ of surface waters in the eastern tropical Pacific of 1.2 ± 0.1‰ over the last four glacial terminations. These early results offer the potential for future regional reconstructions of watermass $\delta^{18}O$, which would provide for better understanding of the global ice volume effect, for the isotopic composition of the global ice sheets, and for regional transports of fresh water in the atmosphere related to evaporation and precipitation, as well as helpful constraints on changing upper ocean temperatures.

An alternative constraint on ice history comes from $\delta^{18}O$ in molecular oxygen preserved in polar ice cores; this proxy must in part reflect changes in surface-ocean $\delta^{18}O$, without the temperature effects that complicate the interpretation of marine $\delta^{18}O$ records. A key debate, however, concerns how to remove the imprint of changing Dole effect (the isotopic fractionation during the processes of photosynthesis and respiration). In one view, the Dole Effect is so dominated by cycles near 23,000 yr (similar to orbital precession) that a simple correction for this contribution can be made (Shackleton, 2000). Jouzel et al. (2002) argue that such simple corrections underestimate oceanic $\delta^{18}O$ at some times, implying that the history of the Dole effect is too complex to allow derivation of global oceanic $\delta^{18}O$ histories from the ice-core data.

7. Ice sheets and climate

There is a general consensus that ice sheets influence global climate by affecting the planetary albedo, atmospheric and ocean circulation, and the hydrological cycle (Clark et al., 1999). The increased albedo associated with expanded ice sheets and exposure of continental shelves contributes a substantial fraction of the total radiative forcing at the LGM, with most cooling from ice sheets occurring in the northern hemisphere (Broccoli, 2000). The increased continental elevation in glaciated regions strongly affects the surface and upper elevation wind flow. Lower temperatures and altered atmospheric circulation cause growth of sea ice, thus further increasing albedo and affecting ocean circulation. Millennial-scale changes in ice sheets also have a significant influence on climate. Andrews and Barber (2002) present evidence suggesting that the Hudson Strait ice stream, known to play an important role in the generation of Heinrich events, may also be involved in Dansgaard–Oeschger events. Millennial-scale climate change may also have played an important role in the last deglaciation (Alley et al., 2002; Charbit et al., 2002).

8. Conclusions

LGM ice sheet margins were reasonably close to those postulated in the CLIMAP maximum model, except in the Kara Sea region where the margin was likely smaller. The magnitude and spatial distribution of the ice load, however, was substantially different than in the CLIMAP model. The results from ice sheet and sea level models identify a range of possible solutions for ice volume, expressed as ice-equivalent sea level lowering, from a "minimum" model of ~ 118 m to a "maximum" model of 130–135 m (Table 1). However, the distribution of the mass among the Laurentide and Scandinavian/Barents ice sheets suggested by the ice sheet models differs significantly from that suggested by one GIA model. Nevertheless, both solutions include substantially less ice than the CLIMAP maximum model result of 163 m sea-level lowering, but bound the CLIMAP minimum estimate of 127 m. The geologic data on ice limits require that either of these two ice volume estimates be distributed regionally in the same area as the CLIMAP maximum model, implying that some or all of the LGM ice sheets were thinner than those in the CLIMAP minimum model.

With the papers in this volume, the EPILOG Project has substantially narrowed the uncertainties in the total changes in ice sheets and sea levels and their proxies. Further resolution of LGM ice volume estimates will require additional field data on relative sea level, especially near the centers of ice loading, and from far-field sites that sample the LGM as well as add details to the deglaciation. Additional field data are also needed on ice-margin extent and the spatial distribution of basal processes. We expect that improvements in modeling ice sheets and the solid earth response to the redistribution of mass should converge on the mass distribution and total volume of LGM ice sheets and consequent eustatic decrease in sea level.

Acknowledgements

We thank the Paleoclimate Program of the National Science Foundation, the IMAGES Program, PAGES, INQUA, and the College of Science at Oregon State University for generous financial support of the EPILOG workshop held October 1–5, 2000, at Timberline Lodge, Oregon. We thank all of the participants of the EPILOG workshop for their contributions to the workshop, Jim Rose for his unerring help, and the reviewers of papers in this special issue for their timely and helpful reviews.

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