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Earth and Planetary Science Letters 181 (2000) 513–527

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Global ice volumes at the Last Glacial Maximum and early Lateglacial

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Received 7 February 2000; accepted 12 July 2000

Abstract

Observations of sea levels during the Last Glacial Maximum (LGM) at localities far from the former ice margins constrain the global change in ice volume from the LGM to the present at about 52×10^6 km^3. Regional studies of sea-level change from observations near and within the margins of the former ice sheets constrain ice volumes of the individual ice sheets and have led to an imbalance between the global estimate of ice volume and the sum of the individual ice-sheet volumes. The latter estimates are reliable only when the observational record from localities close to the ice mass extends well into Lateglacial times, which is generally not the case for the major ice sheets. Ice volumes during the LGM and earliest part of the Lateglacial period can therefore be substantially increased without affecting the predictions of Lateglacial and Postglacial sea level in a significant manner, provided that a rapid reduction in ice volume occurred in early Lateglacial time. New far-field data for LGM and Lateglacial sea-level change indicates that a rapid rise in sea level of about 15 m occurred at about 16 500–16 000 ^{14}C (or 19 200–18 700 calibrated) years ago. This leads to the inference that during the LGM the ice sheet volumes of the major ice sheets were greater than inferred from regional rebound analyses and that rapid reductions in volume occurred at the termination of the LGM. The timing of this occurrence does not coincide with any recognised Heinrich event, although pulses of ice-rafted debris originating from both northern ice sheets do occur in some high latitude cores. An Antarctic contribution to the post-LGM melting event also cannot be ruled out, the timing coinciding with evidence for the onset of warming in southern latitudes and the addition of meltwater into the Southern Ocean.

Keywords: sea-level changes; last glacial maximum; glacial extent; eustacy; meltwater

1. Introduction

The Last Glacial Maximum (LGM) was a time when large ice sheets covered the continental land masses of North America and Europe and when sea levels globally stood lower than today. It is an interval in Earth climate history that is distinctly different from present conditions, yet it occurred sufficiently recently for reliable and quantifiable evidence of the climate to be preserved in sediment and ice records. This period becomes, therefore, important for testing climate models under conditions that are quite different from those of today; models that are also used for predicting future trends in climate change (e.g. [1,2]). An important element in understanding the LGM conditions is the timing of this event and of the

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0012-821X/00/$ – see front matter
PII: S0012-821X(00)00223-5

ice distribution and ice volumes at this and Lateglacial times. This is the issue addressed in this paper through an evaluation of sea-level change during the LGM–Lateglacial interval.

Few direct measurements exist of the thickness of ice formerly over the continents. For small ice sheets such as that over the British Isles, it has been possible to quantify the ice thickness by dating the formation of weathering profiles on mountain peaks that stood above the ice at the time of the LGM, while in some localities of Antarctica, it has been possible to measure the change in ice thickness since the LGM by dating remnant marginal moraines. But in North America and Fennoscandia, locations of the largest latest Pleistocene ice cover, no mountain peaks stood out above the ice domes and in these cases, only minimum estimates of ice thickness can be inferred. Instead, to estimate past ice volumes, recourse to glaciological modelling is necessary and this requires a series of assumptions concerning the supply of ice-forming moisture, ablation conditions, and thermal bed-rock and stress states at the base of the ice cover. Independent checks on LGM and Lateglacial ice thickness estimates are clearly desirable and it has long been recognised that observations of crustal rebound and sea-level change provide such constraints: as the ice sheets melted in post-LGM times, the crust underlying the ice rebounded, sea levels rose or fell depending on their locality, and shorelines advanced or retreated concordantly. Observations of sea-level change and shoreline migration, coupled with observations of the ice retreat, provide the basis for estimating not only parameters defining the earth rheology but also ice volumes during the LGM and Lateglacial times. Sea-level data from the formerly glaciated regions (from near-field sites) provide some constraint on the nearby ice volumes and observations far away from the former ice sheets (far-field sites) constrain the total volume of land-based ice over time (e.g. [3–7]).

Far-field observations of sea level during the time of maximum glaciation are indicative of a prolonged period when sea level was nearly constant, from about 19 ka BP ($\times 1000$ years before present) or earlier, to 16.5 ka BP, followed by a short period of very rapid sea-level rise, of about 15 m in less than 500 years [8]. (All ages are in radiocarbon years because much of both the sea-level and ice-retreat data are referred to this time scale and only in the final section is the calibrated time scale used.) This evidence comes from both the coral record from Barbados and from new sedimentological–palaeontological records from stable continental shelves [9–11]. Together, they place a reliable constraint on the total land-based ice volume during this interval.

Analyses of sea-level change from localities near and within the margins of the former ice sheets, for eastern Laurentia [3], Laurentia as a whole [4,5], Scandinavia [6] and the Barents–Kara seas [7], have led to the conclusion that the corresponding ice-sheet volumes at the LGM and/or Lateglacial time are usually less than the volumes assumed for the ice models (e.g. [12–14]) used in the rebound calculations, models that are generally characterised as single-domed, thick and with quasi-parabolic ice-height profiles. A consequence of these analyses is that the sum of the regional land-based ice-volume estimates is less than the total volume inferred from the far-field LGM sea levels and that other ice sheets experienced a substantial reduction in volume so as to make up for the 'missing' melt-water [14–15]. This led to the postulate that Antarctica contributed more than 10^7 km^3 to the global ocean volume [14], an amount equal to the difference in ice volume between the Denton and Hughes LGM model for Antarctica [13] and the volume estimated from the present ice sheet [16]. Direct tests of this hypothesis have not been possible because, while Late Holocene rebound appears to have been quite substantial in several localities along the Antarctic coast, the available sea-level evidence is inadequate to permit a comprehensive inversion of changes in overall Antarctic ice volume to be made [17]. A limitation of the regional rebound inversions for ice volumes is that the problem is not well constrained for Lateglacial and LGM times and the resulting volume estimates for the individual ice sheets for this early period can be quite uncertain unless there is adequate early sea-level information from sites within the former maximum ice limit [18].

It is the new results for the sea levels in the far

field and the poor resolution for the LGM volumes of the individual ice sheets that leads to the following scenario that may not require as large a LGM Antarctic ice volume as suggested above: Ice sheets during the LGM were larger than inferred from the regional rebound analyses, but these volumes were retained for only a relatively short time period before a rapid reduction in volume occurred at about 16.5–16 ka BP. The ice thicknesses of the individual ice sheets deduced from the regional studies of the northern hemisphere ice sheets, constrained by late- and post-glacial sea-level information, are appropriate for Lateglacial times only, rather than for the LGM and earliest Lateglacial. The inference then is that one or more of these ice sheets experienced a substantial reshaping and loss of mass at about 16.5–16 ka BP, raising global sea level at this time by about 15 m. Thus an equivalent amount of extra ice could be stored in one or other of the northern ice sheets during the LGM rather than requiring that it be stored in Antarctica.

This is the primary issue that is explored in this paper, using mainly the Scandinavian data as example with the understanding that the Laurentian ice sheet may have experienced similar changes. Specifically the question to be addressed is: can more ice be placed into the LGM ice sheets without introducing inconsistencies with the inferences drawn from the local rebound evidence? Of the two major northern ice sheets, the Scandinavian record for rebound is the better one for such an evaluation because the observations in several instances extend well into the Lateglacial period and in one instance into the LGM [6], whereas the observational sea-level record for Laurentia, where the ice cover persisted into more recent time, is generally of shorter duration [19]. Any conclusions reached about the limits of the early northern ice sheets will be at least equally valid for inferred ice volumes for Antarctica during the LGM where the rebound record does not extend much beyond 8000 years BP [17].

A second and related issue, addressed briefly, is the source of the meltwater at 16.5–16 ka BP. If it can be established from the far-field analyses that the ice sheets were larger at the LGM than inferred from the regional studies, the absence of good local rebound data for this time does not permit a determination to be made of which ice sheet provided the meltwater source. Ocean sediment records do record the occurrence of influx of land-based ice into the oceans through their ice-rafted debris [20] or oxygen isotope signals [21] but the high northern-latitude sediments do not contain unambiguous records of this early melting event. Possibly the Antarctic ice sheet was responsible, as has been suggested for a later meltwater pulse [22].

2. Total ice volumes since the Last Glacial Maximum

Sea-level change information from far-field sites provides an estimate of changing ocean volumes and hence of the total ice sheet volume with two provisos: that (i) the observation site is free from tectonic movements or that these movements can be corrected for; and (ii) corrections have been made for the glacio-hydro-isostatic contributions to local sea-level change.

The sea-level equation for a tectonically stable area is [23,24]

$$\Delta\zeta_{\mathrm{rsl}} = \Delta\zeta_{\mathrm{e}} + \Delta\zeta_{\mathrm{i}} + \Delta\zeta_{\mathrm{w}} \tag{1}$$

where $\Delta\zeta_{\mathrm{rsl}}(\varphi,t)$ is the height of the palaeo sea surface relative to the present and is a function of position φ and time t. $\Delta\zeta_{\mathrm{e}}(t)$ is the 'ice-volume-equivalent sea-level change' (defined by Eq. 3, below), or simply the equivalent sea-level change, associated with the change in ocean volume resulting from the melting or growth of land-based ice sheets on a deformable earth [23,24]. The $\Delta\zeta_{\mathrm{i}}$ and $\Delta\zeta_{\mathrm{w}}$ are the glacio- and hydro-isostatic contributions to sea-level change from the isostatic crustal displacement and associated planetary gravity-field or geoid change. Both $\Delta\zeta_{\mathrm{i}}$ and $\Delta\zeta_{\mathrm{w}}$ are functions of position and time. The water depth or terrain elevation at time t at any location φ, expressed relative to coeval sea level, is

$$h(\varphi,t) = h(\varphi,t_0) - \Delta\zeta_{\mathrm{rsl}}(\varphi,t) \tag{2}$$

where $h(\varphi,t_0)$ is the present-day (t_0) bathymetry or

topography at φ. The palaeoshorelines occur where $h(\varphi,t)=0$. Both isostatic terms in Eq. 1 are functions of the earth rheology. Also, the glacio-isostatic term is a function of the ice mass through time and the hydro-isostatic term is a function of the spatial and temporal distribution of the water load: of the relative sea-level variation and of the migration of shorelines.

The term $\Delta\zeta_e(t)$ relates to the total change in land-based ice volume V_i according to

$$\Delta\zeta_e(t) = \frac{\rho_1}{\rho_o}\int_t \frac{1}{A_o(t)}\frac{dV_i dt}{dt} \tag{3}$$

where $A_o(t)$ is the ocean surface area and ρ_i, ρ_o are the average densities of ice and ocean water, respectively. The ocean surface area in Eq. 3 is a function of time because of the shifting shorelines as the relative position of land and sea is modified and because of the retreat or advance of grounded ice over shallow continental shelves and seas. The first dependence is a function of earth rheology and ice-load geometry which together determine the local rate of sea-level change. The second time dependence is a function of the location of the ice limits and whether the ice sheets are located above or below coeval sea level. Hence the relationship between ice volume V_i and equivalent sea level $\Delta\zeta_e$ is weakly model dependent. It should be noted that with these definitions, the ice volume in Eq. 3 includes the ice grounded on the shelves and which displaces ocean water. It should also be noted that the ice-volume-equivalent sea level defined by Eq. 3 is not the same as eustatic sea level which is usually defined as the globally averaged change in ocean level. This distinction arises from the changing depth and shape of the ocean basins and, even when melting has ceased and the ocean volume is constant, the ocean basin geometry continues to be modified by the isostatic adjustment of the land and sea floor. In consequence $\Delta\zeta_i$ and $\Delta\zeta_w$ are non-zero when averaged over the oceans at any time t [23–25] and the difference is not inconsequential. Writing

$$\Delta\zeta_i(\varphi,t) = <\Delta\zeta_i(\varphi,t)>_{Ao} + \delta\zeta_i(\varphi,t) \tag{4a}$$

$$\Delta\zeta_w(\varphi,t) = <\Delta\zeta_w(\varphi,t)>_{Ao} + \delta\zeta_w(\varphi,t) \tag{4b}$$

where $<\ >_{Ao}$ denotes the average value over the ocean surface A_o at time t and $\delta\zeta_{i,w}$ denote the spatial dependence part of the two isostatic terms, then the eustatic sea level is

$$\Delta\zeta_{eus}(t) = \Delta\zeta_e(t) + <\Delta\zeta_i(\varphi,t)>_{A_o} + <\Delta\zeta_w(\varphi,t)>_{A_o} \tag{5}$$

In evaluating these ocean-averaged values, the time dependence of the ocean surface area again needs to be taken into consideration. (These last two terms are a more precise definition of the conversion from the sea-level equivalent to eustatic sea level drop used in Denton and Hughes ([13], p. 274).) We use $\Delta\zeta_e(t)$ here rather than $\Delta\zeta_{eus}(t)$ because it is the former that gives a direct measure of ice volume through Eq. 3. (Some inconsistency has crept into the usage of the term eustatic sea level to which the senior author has contributed. Initially the distinction between the $\Delta\zeta_e(t)$ and $\Delta\zeta_{eus}(t)$ was maintained (e.g. [14,24]), but in some more recent papers (e.g. [26]), the ice-volume-equivalent sea level was dropped in favour of eustatic sea level, mainly because of the clumsiness of the former term.)

If both the ice distribution through time and the earth's response to loading are known, then the equivalent sea level, and ice volumes through Eq. 3, follow from observed relative sea levels $\Delta\zeta_{rsl}^{obs}$ according to

$$\Delta\zeta_e = \Delta\zeta_{rsl}^{obs} - (\Delta\zeta_i + \Delta\zeta_w) \tag{6}$$

where the $\Delta\zeta_i$ and $\Delta\zeta_w$ are the model-dependent isostatic corrections from Eq. 1. The accuracy with which $\Delta\zeta_e$ can be established depends on the accuracy of the local sea-level observation and on the accuracy with which the isostatic corrections can be made. For sites far from the ice margins these latter terms represent 10–15% of the sea-level signal for the LGM and Lateglacial times and if the corrective terms can be evaluated with an accuracy of 10% the resulting uncertainty introduced into $\Delta\zeta_e$ is of the order 1–2 m, smaller than most observational uncertainties for the glacial stages.

Valuable data for the evaluation (Eq. 6) of the

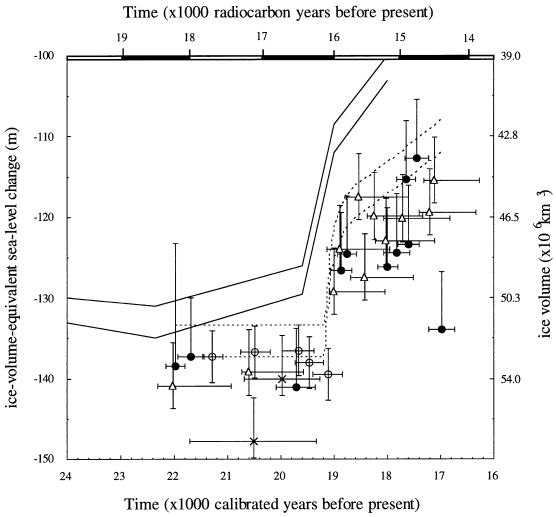

Fig. 1. The upper and lower limits (dashed lines) of the ice-volume-equivalent sea level inferred from observed sea levels from northern Australia and Barbados. The open circles refer to brackish water fauna and the solid circles refer to shallow marine fauna from northern Australia [11]. Two older observations from the same region [28] are shown by the crosses. The Barbados corals are shown as open triangles [9,10]. The right hand axis gives the corresponding volume of land-based ice, including ice grounded below sea level on the shelves.

LGM and Lateglacial equivalent sea level are the depth–age relationship of fossil corals from Barbados [9,10]. This locality has been subject to vertical tectonic movement, but this has been corrected for using the elevation and age of the Last Interglacial (5e) shoreline as a measure of the average rate of vertical movement [27]. Two corals (*Porites* sp.) occur at a corrected depth of around 130 m and have been dated at between 18.5 and 17 ka BP. The main period of early coral growth (*Acropora* sp.) appears to have occurred later, after about 16 ka BP, and these corals now occur in water depths of about 110 m. It is the two older corals, rather than the second group, that we believe mark the LGM and that the younger group is representative of immediate post-LGM conditions (see below). The corals mark a lower limit to sea levels at this locality, the actual sea level having been as much as 5–8 m above the *Acropora* corals at the time of their growth and possibly more for the *Porites* corals.

Another important indicator of past sea levels is the age and depth of sediments or peats that mark the transition from terrestrial or lagoonal conditions to marine conditions as sea level rose from its LGM lowstand to its present level. In the protected palaeo Bonaparte Gulf of northwestern Australia, this evidence has been undisturbed and preserved and has provided detailed evidence of the sea-level change throughout the LGM and Lateglacial [11,28].

Fig. 1 illustrates the LGM and early Lateglacial ice-volume-equivalent sea levels inferred from the various data sources based on the relationship Eq. 6 and nominal earth- and ice-model parameters discussed below. Also shown are the corresponding land-based (including grounded) ice volumes. The observations for the time of maximum glaciation, both the coral data from Barbados and the sediment data from the Bonaparte Gulf, are indicative of a prolonged period when equivalent sea-level was constant, from 19 ka BP or earlier, until 16.5 ka BP. This is followed by a period of very rapid sea-level rise, of about 15 m in less than 500 years after which the sea-level rise was less rapid due to the more gradual and more uniform decay of the ice sheets.

3. Crustal rebound and sea-level change in the formerly glaciated regions

Sites where the sea-level or crustal rebound signal is most sensitive to the thickness of the ice load are those near the centres of former glaciation (near-field sites), but the observational record is limited in time to the ice-free period only. Observations from near the ice margins for the LGM and early part of the Lateglacial period are therefore few. Of the two major northern ice sheets, the Scandinavian one is the better for rebound analysis because the observational record of sea-level change is more complete and in several instances the record extends into the Lateglacial period and in one instance into the LGM.

3.1. Fennoscandia

Earlier studies [4,29] concluded that ice models such as that proposed by Denton and Hughes [13]

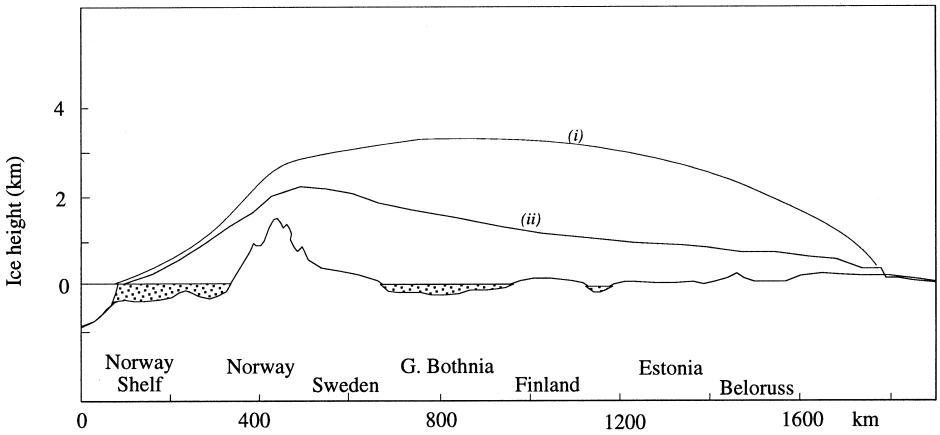

Fig. 2. Ice-height profile across Scandinavia at the Late Glacial Maximum according to: (i) Denton and Hughes [13]; and (ii) the model that is consistent with the inversion of sea-level data from the region [6].

contained more ice than is acceptable by the sea-level data. In this latter, glaciologically argued model, the maximum ice thickness at the LGM exceeds 3000 m and is centred over the Gulf of Bothnia. But for a very wide range of earth-model parameters it leads to an excessive rebound [6] with consequences that the Baltic Sea is predicted to have been open to the Arctic Ocean in Late-glacial times when the ice had retreated from Finland, that much of Finland remained submerged until Mid and Late Holocene time, and that the Baltic remained open to the Atlantic at the times of the fresh-water Baltic Ice Lake and Ancyllus Lake, all inconsistent with observational evidence [30]. The outcome of inversions for both earth-rheology and ice thickness estimates is that during the LGM and Lateglacial time, the ice is relatively thin over eastern and southern Scandinavia and that the ice height increases relatively slowly with distance from the ice margin when compared with the steeply domed models with thick quasi-parabolic profiles of Denton and Hughes [13]. In the north and west, the ice-height profiles appear to be more consistent with such quasi-parabolic models [6,30]. Fig. 2 compares the ice thicknesses for the thin-ice model with the LGM model proposed by Denton and Hughes for a section across Scandinavia. This comparison clearly illustrates the very different ice volumes contained in these two ice models.

In the inversion [6] for the Scandinavian ice volume it has been assumed that ice-height profiles remain similar throughout the deglaciation phase, implying that basal conditions controlling the ice-height profiles have remained unchanged through time. Because few early-Lateglacial observations exist, tests of this assumption are few although the few ice-margin observations that are available from the Norway coast (e.g. from Andøya) indicate that the model predictions underestimate the rebound [6]. Thus a greater ice volume is appropriate in at least northern Scandinavia for this early period and one scenario is that during the LGM, the Scandinavian ice sheet overall contained more ice than is inferred from the rebound inversions: that thicker ice could be supported during the LGM because the conditions at the base of the ice sheet were initially favourable for this, i.e. cold and rigid, but that in early Lateglacial time a change in basal conditions occurred leading to a rapid thinning of the ice and ice-thickness profiles that are more consistent with those inferred from the inversion of the sea-level data (Fig. 2) [31]. The rapid rise in the ice-volume-equivalent sea level noted from the far-field localities (Fig. 1) suggests that this transition could have occurred in the short interval 16.5–16 ka BP.

Fig. 3 illustrates the time dependence of the ice volume (model S0) resulting from the inversion of

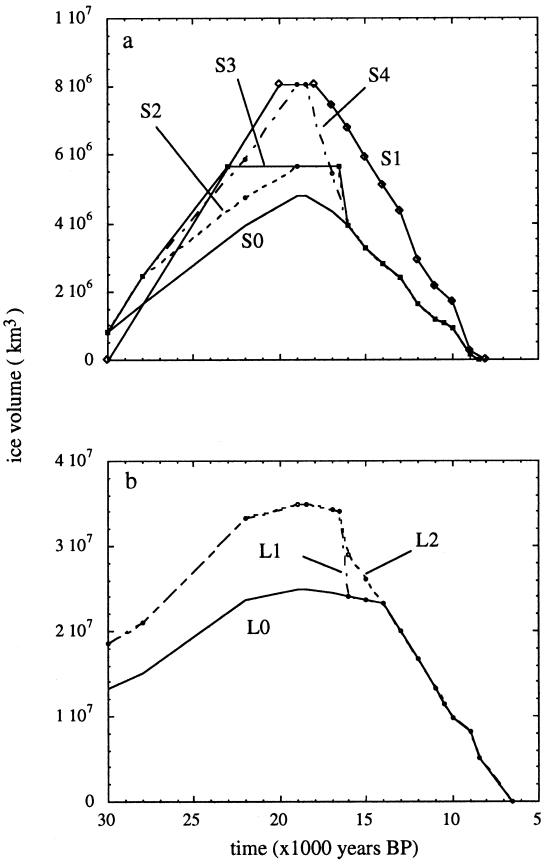

Fig. 3. Ice volumes of (a) Scandinavian and (b) North American ice models since the LGM. In a, S0 corresponds to the model inferred from the inversion of the Scandinavian sea-level data [9] and S1 is an adaptation of the Denton and Hughes ice model [13]. S2–S4 are variants of S0 in which ice volumes are enhanced during the Last Glacial Maximum. In b, L0 is adapted from ICE-1 [34] and L1 and L2 are variants of this model, differing in the amount by which ice volumes have been increased during the Last Glacial Maximum.

the Scandinavian sea-level data [6] as well as that of the preliminary ice model (S1) [9] that was based on the Denton and Hughes LGM model [13] and on published ice retreat reconstructions [32,33]. The three other models (S2–S4) differ from S0 only in their ice content during the LGM and earliest Lateglacial time. These variants are not necessarily intended to have physical meaning, rather, they are introduced to establish the sensitivity of the rebound predictions to the early part of the ice history. In models S2 and S3, the ice at the time of the LGM is assumed to follow the ice thickness profiles of Denton and Hughes, but their amplitudes have been scaled by 0.7 such that the maximum ice thickness is about 2400 m [29]. In S2, the maximum ice persists from 19.0 ka BP until 16.5 ka BP at which point it is reduced everywhere and linearly over the next 500 years to the ice thicknesses of model S0 at 16 ka BP. This model implies that a relatively thick ice sheet did initially develop over the Gulf of Bothnia and Finland, but that at 16.5 ka BP, it rapidly decreased in volume with the centre of loading shifting westwards. Model S3 is the same as S2 except that the duration of the LGM is longer, with maximum ice volumes first attained by 23 ka BP. Model S4 is constructed in a similar way to S2 except that the full Denton and Hughes [3] model is used for a short-lived LGM which then decays to the model S0 in the interval 18.5–16 ka BP. All of these European ice models include a Barents–Kara ice sheet [7]. Other ice sheets used in the calculation include a modified ICE-1 ice sheet [34] for the western half of the northern hemisphere (see below), and a model for Antarctica based on the difference between the Denton and Hughes [13] model and the present ice sheet [16], scaled down so as to contribute 25 m to equivalent sea level at the time of the LGM and whose rate of melting is in phase with that of the northern hemisphere.

The earth model parameters adopted for the preliminary calculations are those discussed in earlier analyses of European sea-level observations and which have been found to give a satisfactory agreement between observations and predictions. The model used here (E0) is characterised by a lithosphere of effective elastic thickness H_l of 65 km, an upper mantle with an average effective viscosity η_{um} of 4×10^{20} Pa s, extending from the base of the lithosphere down to 670 km depth, and a lower mantle with an average effective viscosity η_{lm} of 10^{22} Pa s. (This nominal earth model differs in a small and unimportant for present purposes way from the optimum solution inferred in [6]).

Fig. 4 illustrates the predicted sea levels for the past 20 000 years at four representative sites in Scandinavia. (Fig. 5 illustrates the location of

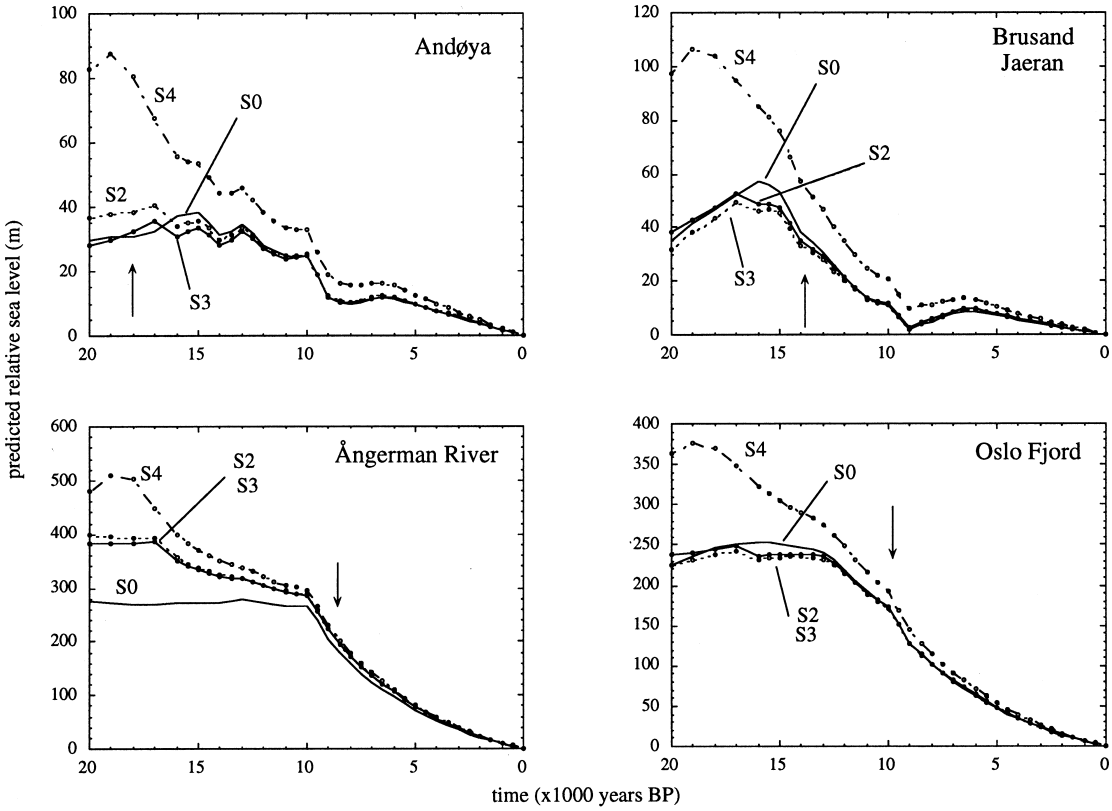

Fig. 4. Predicted sea levels at four locations in Scandinavia (see Fig. 5a for locations) according to different ice models (defined in Fig. 3, and including the standard models for the other ice sheets). The arrows indicate the oldest observations in the region.

these and other sites, along with the limits of the ice sheet at the time of the LGM.) The arrows for the predicted curves at each site indicate the age of the oldest evidence of sea-level change at that site. Models S2 and S3 lead to nearly identical predictions throughout Holocene and Lateglacial times and the differences become notable only for the LGM and earlier. Hence the assumed duration of the LGM is not critical for the predictions of Lateglacial and more recent sea levels. At most of the localities, the differences in the predictions of sea level, based on ice model S0 and S2, remain relatively small for the period of observation and comparable to the uncertainty estimates of the observational data (cf. [6]). Where this is not so is for Æråsvatn (Andøya) where the record extends back to 18 ka BP [35]. (Here there is an inconsistency in the model in that at this time the ice limit is assumed to have stood near the outer edge of the continental shelf [32], whereas the northern end of Andøya appears to have already been ice free by this time. Similar inconsistencies occur in Lateglacial times at some other Norwegian coastal locations (e.g. in Jæren [6]), and this emphasises the need for an improved chronology for the ice margin and its retreat. For present purposes, these inconsistencies are embarrassments rather than being of serious consequence.) Otherwise, for most sites with observational data, little distinguishes the predictions based on models S0 and S2. At both Æråsvatn and Brusand (Jæren), predictions based on model S4 do differ substantially from those based on S0, indicating that there is a limit to the amount of ice that can be introduced into the LGM without perturbing the Lateglacial predictions in a significant manner.

Fig. 6 illustrates the comparisons of observed

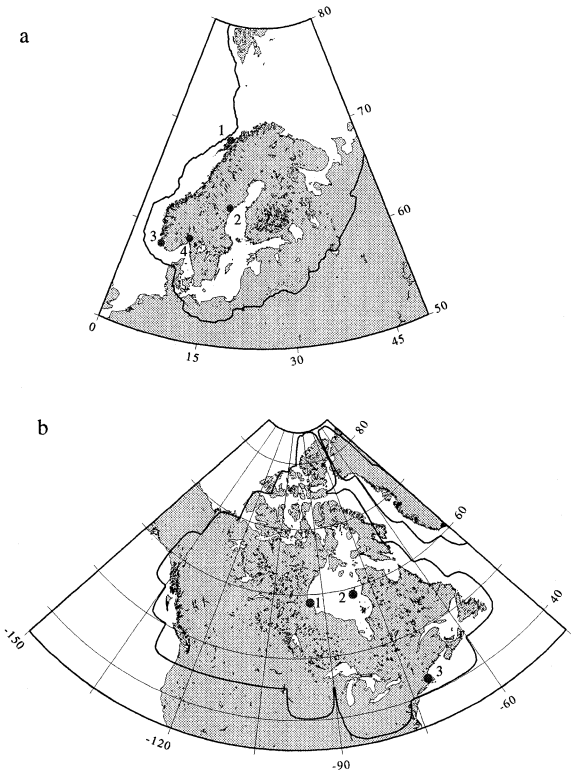

Fig. 5. Locations of sites corresponding to the predictions illustrated in Figs. 4 and 7 and model ice limits at time of Last Glacial Maximum. (a) Scandinavia: 1 = Æråsvatn, Andøya, 2 = Ångermanälven, 3 = Brusand (Jæren), 4 = Oslofjorden. (b) North America: 1 = Churchill, 2 = Ottawa Island, 3 = Boston.

and predicted sea levels for the large Scandinavian observational data base [6] for the three models S0, S2 and S4. The best agreement between observations and predictions occurs for model S0 as is expected because the ice model was the result of the inversion of this sea-level data base and for a earth model similar to that used here. For the other two models, the distributions of the normalised residuals

$$(\Delta\zeta_{\mathrm{rsl}}^{\mathrm{obs}}-\Delta\zeta_{\mathrm{rsl}}^{\mathrm{pred}})/\sigma \qquad (7)$$

where $\Delta\zeta_{\mathrm{rsl}}^{\mathrm{obs}}$ and $\Delta\zeta_{\mathrm{rsl}}^{\mathrm{pred}}$ are vectors of the observed and predicted sea-level values and σ is a vector of standard deviations of the former, become increasingly skewed, indicating that, particularly for S4, the predictions, based on the nominal earth model E0 used, overestimate the rebound. This does raise the question whether the ice models inferred from the rebound analysis are dependent on the earth-model parameters, whether for the model S2, for example, there is a set of three-layered earth-model parameters that leads to a better agreement with the observations than does the combination (E0, S0). This will need further investigation.

3.2. North American ice sheet

Glaciological models of the Laurentian ice sheet at the time of the last peak glaciation suggest that this ice sheet, excluding the Cordilleran and Innuitian contributions, contained enough ice to globally raise sea level by between about 40 and 110 m [12,13,36] and considerable divergence of views remain about the ice thickness through time. The use of sea-level observations to constrain models of the ice load here have generally been less successful than in the case of Scandinavia because the quantitative evidence for rebound is mostly limited to the eastern and northern parts of the former ice margin. Nevertheless, different analyses [3,4,37] have all led to a similar conclusion, that thick, single-domed, quasi-parabolic profiled ice sheets, that persist over North America until early Holocene time, result in an overestimation of the rebound and that substantial reductions in ice volumes are required for Late-glacial times. Nakada and Lambeck [38] concluded that the ice thickness of the Laurentide part of the ICE-1 model [34] needed to be scaled down to about 70% of its initial values in order for the predictions to match reasonably the observed sea-level changes in the Hudson Bay area. Likewise, the more comprehensive analysis by Tushingham and Peltier [4] resulted in an ice model (ICE-3G) whose thickness over the same region at 10 ka BP is significantly less than that contained in ICE-1 at the same epoch, although both models have similar, but not identical, ice thicknesses here for the LGM. However, even more than in the case for Fennoscandia, inversions of the available sea-level data resolve only poorly the early part of the ice record because the

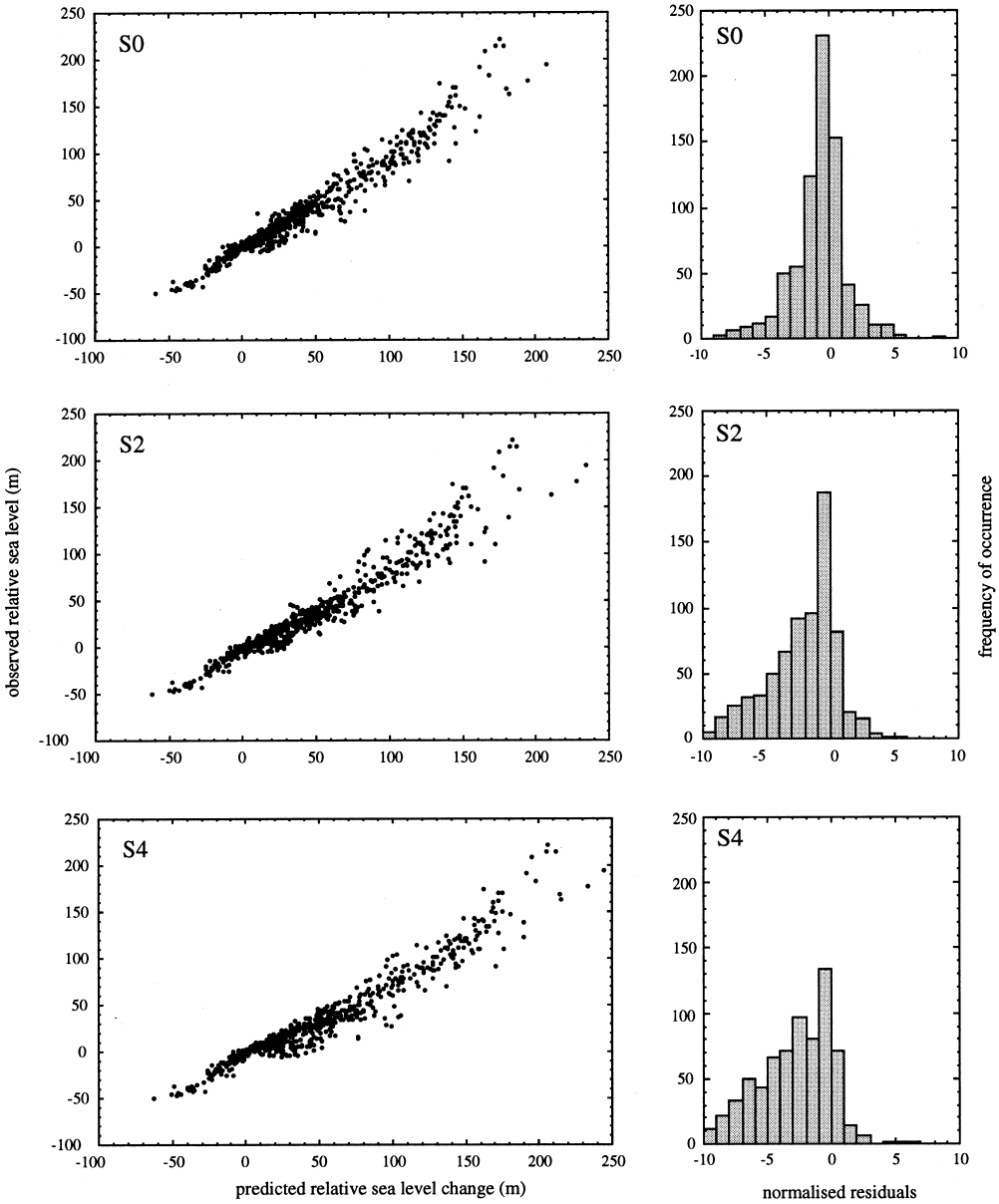

Fig. 6. Observed versus predicted sea levels and histograms of the normalised residuals (defined by Eq. 7) for earth model E0 and three Scandinavian ice models S0, S2, S4.

observational evidence within and near the ice limit does not extend far back in time. Within the Hudson Bay area, for example, the region did not become ice free until about 8000 years BP and the record does not predate this time. Hence, analogously with the Scandinavian example, it is possible to introduce more ice into the early part of the Laurentian LGM ice sheet without this making a significant impact on the comparison of predicted and observed sea-level variations for Holocene time.

Predictions for sea-level change at a number of

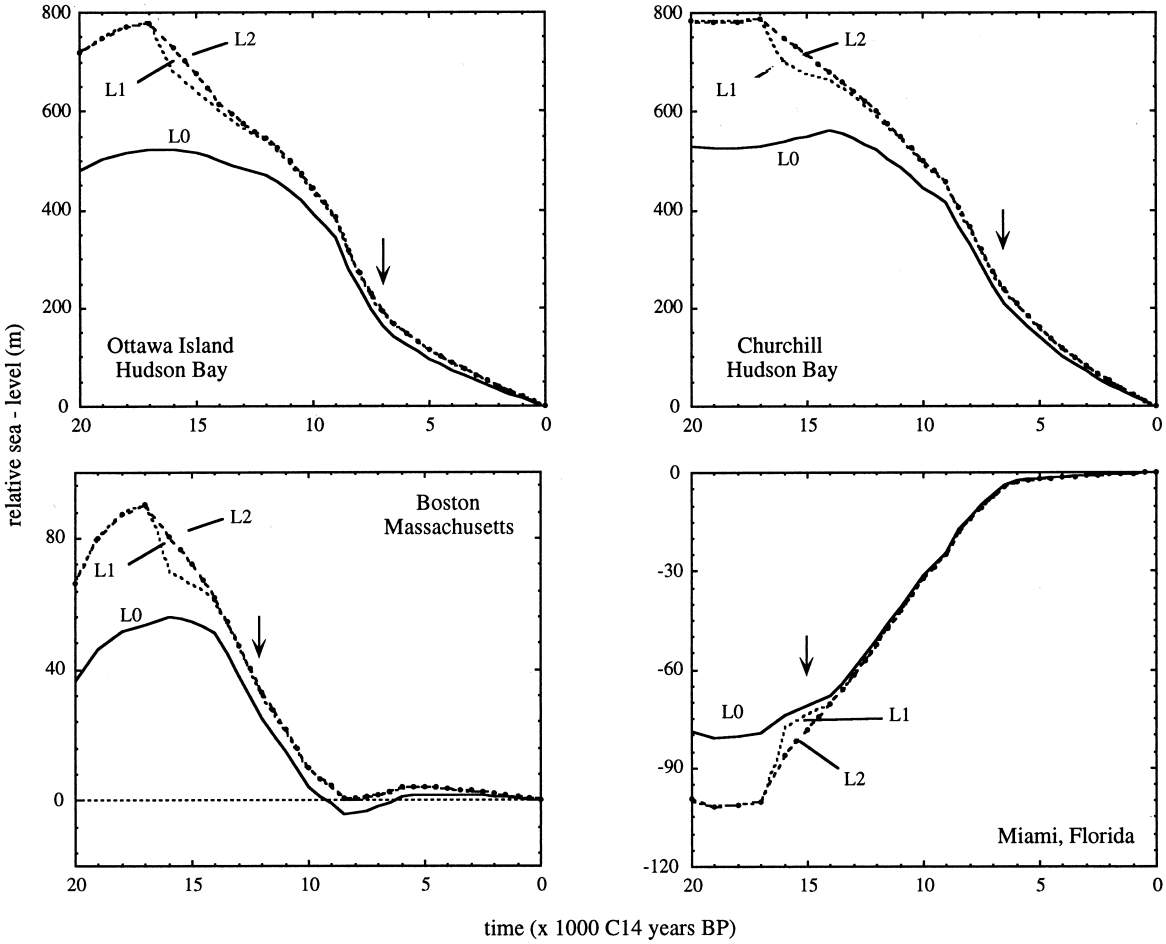

Fig. 7. Predicted sea levels at four North American localities (see Fig. 5 for localities of the Hudson Bay sites of Churchill and Ottawa Island). Predictions are based on earth model E0 and Laurentian ice models L0, L1, L2. Standard ice models are included for the other ice sheets.

representative North American sites are illustrated in Fig. 7 for ice models that differ only in their volumes before early-Lateglacial time. (Fig. 3b illustrates their ice-volumes and Fig. 5b gives the locations of three of the sites for which the predictions are made.) Model L0 is based on ICE-1 [34] smoothed and interpolated onto a 1° latitude by 1° longitude spatial grid and a 1000 year time grid, with a pre-LGM record that follows the pattern of fluctuations inferred from the $\delta^{18}O$ marine record and the Huon reef data [39]. Model L1 is the same as L0 after 16.0 ka BP, but the ice thickness from 16.5 ka BP back into pre-LGM time is everywhere scaled upwards by a factor of 1.4 and with a linear transition between 16.5 and 16 ka BP. Model L2 is another variant of L1 in which the ice thickness from 0 to 14 ka BP is the same as in L0, the same as L1 for 16.5 ka BP and earlier, and with the transition from 16.5 to 14 ka BP occurring in two steps; initially rapid over 500 years starting at 16.5 ka BP and then at a reduced rate from 16 to 14 ka BP. As for the Scandinavian case, these ice models are not based on any particular physical arguments but are introduced only to examine the influence of the early melting history on sea-level predictions in Late- and Postglacial times.

From the comparisons in Fig. 7 it is clear that

the rebound predictions are insensitive to the early melting history, particularly for the last 8000 years for which most of the observational data from within the maximum ice limit exists. In the Hudson Bay area, for example, the data extend back to only about 7 ka BP and all three models lead to very similar sea-level predictions for this interval. For sites close to the maximum ice margin, such as Boston, the record extends back to about 12 ka BP but, here also, the predictions based on the three models differ by amounts that are less than the combined uncertainties arising from the observational data and from inadequate knowledge of the local details of the ice sheet. Southwards, the record extends further back, to about 15 ka BP along the coast of Virginia, but at these distances from the ice sheet the isostatic terms are of second order and the observations have little power to discriminate between the different LGM models (cf. the predicted sea levels for Miami in Fig. 7). Hence, from rebound inversions alone, the LGM ice volumes can be increased by up to at least 40% over that in the initial model without leading to significant discrepancies in sea-level predictions for Postglacial times, provided that this additional ice is removed early in Lateglacial time.

4. Discussion

The above indicative analysis for the Scandinavian and North American rebound illustrates that when evidence for past sea-level positions is absent for LGM and early Lateglacial times for localities close to or within the limits of former maximum glaciation, the ability to estimate maximum glaciation ice volumes for the individual ice sheets is limited (Figs. 4 and 7). This conclusion is even more true for the Antarctic ice sheet with its very limited sea-level data base [17]. In contrast, high resolution data for this period from localities far from the ice margins do provide a good constraint on the total volume of ice and of the rate of melting (Fig. 1). In particular, this evidence indicates that a rapid rise of sea level occurred at about 19 ka BP (calibrated), raising the level by about 15 m in as little as 500 years or less. (All ages in the discussion section now refer to the calibrated time scale [40].) Ice sheets over either Scandinavia and North America in which a rapid and substantial reduction in ice volume occurred at this time, lead to rebound that is consistent with the observational evidence from within the former ice margins and, in fact, is required by data from Andøya. In the case of Scandinavia, the rebound evidence dictates a relatively thin ice sheet over the eastern and southern regions in Lateglacial time and the inference is that an initially thick ice sheet became unstable at about 19 ka BP that, while discharging a substantial volume of ice into the oceans, resulted in the change of ice thickness profiles over the eastern and southern parts of the ice dome (cf. Fig. 2). Physically, this suggests that basal conditions were initially cold and frozen but that basal thawing commenced as early as 19 ka BP (cf. [31,36]). There was never enough ice in the Scandinavian ice sheet to alone explain the 15-m sea-level rise noted for this period (Fig. 1) and this implies that the Laurentide ice sheet, or possibly the Antarctic ice sheet, must have experienced a comparable collapse at about the same time, possibly a destabilising consequence of the rising sea level from one ice sheet adjustment on the other.

Fig. 1 compares observed and predicted ice-volume equivalent sea-level change where the latter is based on the northern hemisphere ice models S2 and L2 and the nominal Antarctic ice model whose volume change since the LGM has been uniformly scaled down from a maximum of 37 m [14] to 25 m of equivalent sea level. That is, only the northern hemisphere ice sheets are assumed here to have participated in the rapid ice-volume reduction event at 19 ka BP. The predictions represent quite well the rapid change at the end of the LGM, although the magnitudes of the predicted sea levels for both the LGM and Lateglacial are less than the observed values, suggesting that the ice volumes of the individual ice sheets could be further increased for the early part of the Lateglacial.

Rapid changes in relative sea level since the time of the LGM have been previously reported. Fairbanks [9], for example, inferred from the Barbados coral record a rise of about 25 m between

14.7 ka BP and 13.7 ka BP, denoted as Meltwater pulse 1A (Mwp-1A) (see also [40]), and Blanchon and Shaw [41] inferred, from the same coral evidence, a rise of 13.5 m in less than 300 years within this interval. Possibly there is a chronology problem with the Lateglacial and LGM data. All ages for the northwest Australian data have been AMS radiocarbon dated [11]. Many of the faunal specimens lived in brackish water conditions so that reservoir corrections should be less than that for full-marine fauna. For the latter a correction of about 1000 years appears appropriate for LGM conditions [42] and 400 years has been adopted for the LGM brackish-water samples. Possibly the radiocarbon time scale is not linear at times of the rapid sea-level rise. This time scale is based on an assumption of equilibrium being established between ^{14}C production in the atmosphere and uptake in the oceans (ignoring other sinks), with a major part of this uptake occurring during the subduction of surface waters in the North Atlantic. This North Atlantic deep water formation is sensitive to changes in the salinity and temperature of the surface waters and the injection of fresh water will affect the rates and sites of the deep water formation and also change the balance between rates of ^{14}C production and uptake. Hence non-linearities in the ^{14}C time scale can be anticipated, as occurred at the time of the Younger Dryas [40], but this is unlikely to lead to errors of the magnitude required if the terminal-LGM meltwater pulse is the same as Mwp-1A, and the two must be seen as distinct events of rapid sea-level rise.

Clark et al. [22] examined possible origins for the Mwp-1A and their arguments are equally valid for the 'terminal-LGM Mwp', reported here, whose flux of freshwater input is $0.3–0.4 \times 10^6$ m^3 s^{-1}, about the same as that inferred for Mwp-1A. If the origin of the meltwater is the Laurentian ice sheet, then the only plausible escape route for the majority of this water into the North Atlantic is by ice calving and transport via the Hudson Strait, there being no likely mechanism that can melt large volumes of ice quickly enough to escape via the Mississippi drainage system [22]. Like Mwp-1A, the large flux of water into the ocean at the end of the LGM does not appear to have left a clear and consistent geochemical signal in the ocean sediments in contrast to the smaller influxes associated with Heinrich events seen in the North Atlantic [20], the major post-LGM event H-1 occurring well after the terminal-LGM meltwater pulse. These events are seen as sharp peaks in the relative abundance of ice-rafted debris as well as in planktonic foraminifera signatures that are indicative of cooling and freshening of the surface waters. However, such peaks are not restricted everywhere to the timing of the established Heinrich events and a similar peak occurs in some sediment cores at the time of the terminal-LGM meltwater pulse. For example, in core SU90-24 [43], the amount of ice-rafted debris deposited at this time is similar to, or greater than, that deposited at times corresponding to some of the recognised Heinrich events. Such a peak is also seen in Nordic seas cores [44], indicating that it may be quasi-synchronous for the two major northern ice sheets.

Clark et al. [22] have suggested that the source of the Mwp-1A may have been the Antarctic ice sheet. They review the limited evidence for, and divergent views on, the post-LGM history of this ice sheet and note, interalia, that Antarctic ice volumes could have been substantially larger at the LGM without this being detectable from observations of rebound of the continent's margin. Recent evidence of raised shorelines from the Vestfold Hills and elsewhere indicates that a substantial reduction in ice volume must have occurred in Lateglacial times [17], but the observational record does not extend sufficiently far back in time to provide a useful constraint on the ice volumes for the earlier period. The $\delta^{18}O$ signal in one deep-sea core from the Southern Ocean [45], as well as in the Vostok ice core [46], indicate that warming in the southern hemisphere may have started quite abruptly at about 20 calibrated ka BP (about 16.7 radiocarbon ka BP) and a partial dispersal of the ocean-bordering ice sheet may have followed soon after. This would be consistent with evidence from the $\delta^{18}O$ signal in diatoms of meltwater input into the Southern Ocean during the last cold stage [47].***[AC]***

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