Shoreline displacements in southern-central Sweden and the evolution of the Baltic Sea since the last maximum glaciation

KURT LAMBECK

Research School of Earth Sciences, The Australian National University, Canberra ACT 0200, Australia
(e-mail: Kurt.Lambeck@anu.edu.au)

Abstract: During late- and post-glacial times the Baltic basin has been periodically isolated from the Atlantic and freshwater and saline conditions have alternated. This is a consequence of the interactions between the spatially variable glacial rebound of the region and the simultaneous eustatic sea-level rise in the adjacent Atlantic. Observations of the timing of these isolations and of the location of the barriers therefore provide constraints on the rebound models. High resolution numerical models of rebound of southern Sweden, the Danish straits, and the Baltic have been developed and tested against observations of lake-level change as well as against the observed elevations of the Last Baltic Ice Lake shoreline which formed before about 10.3 ka BP. Predictions of sea- and lake-levels are used to test two alternative hypotheses about the ice thickness over Scandinavia and to estimate parameters that describe the Earth’s response. Optimum values for the latter are: lithospheric thickness of 60 ± 20 km, upper-mantle viscosity of 3.3 × 10^{20} Pa s, and lower-mantle viscosity of 10^{27} Pa s. The rapid lake-level fall at the end of the Baltic Ice Lake stage is estimated to be 30 ± 5 m. The optimum ice model for Scandinavia is one in which the ice thickness over southeastern and southern Scandinavia was relatively thin compared with the ice thickness over Sweden and Norway: the ice thickness in the southeast and south changes slowly with distance towards the ice margin whereas over Sweden and Norway the ice thickness profiles are considerably steeper. The earth and ice models have been combined with high-resolution digital terrain data to develop a comprehensive model for the evolution of the Baltic basin since the time the area last became ice free.

The reconstructions predict the occurrence of barriers at certain times which impede free flow between the Baltic and Atlantic and confirm that the Baltic lake levels are controlled by rebound at four localities: Degerfors in the Närke region of southern-central Sweden, the Värne river outlet through the Göta Alv, Öresund, and the Dass-Langelands-Store belts in Denmark.

Keywords: Scandinavia, Baltic Ice Lake, Yoldia. Ancylus, glacial rebound.

The complexity of the temporal and spatial variability of the Baltic Sea surface since the time that the region last emerged from the Late Weichselian ice sheet has long been recognized and has most recently been reviewed by Björck (1995). Qualitatively, this complexity has been understood in terms of the development and decay of ice and land barriers across southern-central Sweden and the Danish straits or belts that intermittently inhibited penetration of marine waters into the Baltic basin. This paper aims to develop a numerical model, consistent with the geological evidence, of the evolution of the southern Swedish shorelines and of the Baltic Sea since the time of the formation of the first Baltic Ice Lake at about 12 ka BP. This development follows from the high-resolution rebound model by Lambeck et al. (1998b) for Scandinavia and northwestern Europe as a whole but differs from it in that the latter excluded all Baltic lake information and could not therefore be used to predict the various lake stages. The resulting Baltic basin model provides not only a useful interpolation device between scattered observations of the positions and elevations of past shorelines but the comparison of the predicted with the observed shorelines leads to new constraints on the physical parameters that define the model.

The complex sequence of Late Weichselian and Holocene Baltic events has been described in detail by Björck, (1995) (see also Svensson 1989 and Donner 1995) and the following is a brief summary of the principal elements. All ages refer to the conventional radiocarbon time scale. Provided the same time-scale is used consistently throughout the analysis and the observational error estimates reflect the ambiguities or nonlinearities in the 13C time scale for the Late Weichselian and early Holocene, this choice is not important (Lambeck 1998).

During the early stages of ice retreat, starting at about 12.5 ka BP, the emerging lake, the Baltic Ice Lake, was a freshwater domain, separated from the Atlantic Ocean by a land barrier through the Danish straits and by a land-ice barrier through southern-central Sweden. Contact between this lake and the Arctic Ocean was prevented by the ice cover over Finland.

At about 10.3 ka BP, a drastic, nearly instantaneous, lowering of the lake level has been inferred from the shoreline data and has led to the suggestion that it was an ice barrier that had dammed the lake until this time. The fall in lake level appears to have been of the order of 25-30 m and in many localities it left behind a well-developed elevated shoreline, the Last Baltic Ice Lake shoreline, undisturbed by subsequent oscillations in lake levels.

After about 10.3 ka BP, marine influence is detected in the Baltic basin for the first time since the region became ice free. This is the Yoldia Sea phase of the Baltic and lasted until about 9.5 ka BP. The marine or brackishwater condition indicates the concurrence of several events: (i) sea level in the adjacent Skagerrak and Kattegat had risen substantially, (ii) the crust near the ice margin was still sufficiently depressed under the ice load, and (iii) the ice had retreated from the low-lying areas of southern-central Sweden. As a consequence, the sea was able to enter the Baltic basin and any shorelines that formed during the Yoldia Sea phase correspond to coeval sea level.
marine fauna actually occurs some hundreds of years after the ice dam is breached because saltwater could not compete against the outflow of fresh water through the initially narrow passages.)

After 9.5 ka BP the marine fauna was gradually replaced by freshwater flora and the basin was again isolated from the Atlantic Ocean. This forms the Ancylus Lake phase that persisted until about 8.5–8.3 ka BP. The barrier separating this lake from the sea appears to have been the result of the crust in southern-central Sweden rebounding at a rate that was faster than the rise in sea level in the Skagerrak. At the same time the land levels in Denmark and the Danish straits were not keeping up with this rise and the location of the overflow of the Ancylus lake was controlled by the relative rates of rebound at the two localities and by the rate of sea-level rise in the adjacent Skagerrak and Kattegat. Initially the Ancylus lake levels were rising, suggesting that the land barrier in central Sweden controlled the height of the barrier, but after about 9.3 ka BP the lake levels were again falling. A shoreline defining the maximum levels reached during the transgressive phase has been identified at numerous localities throughout the Baltic basin. This is the Ancylus maximum transgression shoreline. The return to falling lake levels is indicative of a shift in the overflow from southern-central Sweden to the Danish belts where the land levels are subsiding to the point that the barrier elevation lies below the Swedish barrier and eventually below sea level. This latter event may have occurred as early as 8.5 ka BP and the subsequent lake stage, containing renewed signs of brackishwater conditions, is known as the Mastagloia Lake, a substage of the subsequent Litorina Sea phase that is characterized by well-developed brackishwater and marine fauna. The boundary between these last two phases is not well defined. In some localities the Litorina Sea is characterized by a period of rising levels, culminating in a maximum Litorina shoreline at about 6.5–7 ka BP, after which the levels gradually fall to the present level. This evolution is a consequence of the interplay between the eustatic sea-level change and the isostatic rebound of the crust rather than to changes in the reference levels of the lake surface.

The shorelines that formed during these various lake stages have been well preserved and investigated in southern Sweden (Svensson 1989, 1991). (Fig. 1 gives the locations of the various sites discussed in the text and Fig. 2 illustrates a result for the Oskarshamn area of Småland in southern Sweden.) A feature common to records such as in Fig. 2 is the rapid drop in lake level at about 10.3 ka BP corresponding to the end of the Baltic Ice Lake. After this event the shoreline elevations continue to fall until about 9.5 ka BP (this corresponds to the Yoldia stage). A period of rising lake levels, the Ancylus transgression, follows until about 9.3 ka BP, to be succeeded by falling shoreline elevations until the present time, possibly with a short duration rise at about 6.5–7 ka BP.

Some of the shorelines are ubiquitous events, evident throughout the central Baltic basin. Of these the Last Baltic Ice Lake shoreline is particularly well developed and has been mapped by Svensson (1991). This result is illustrated in Fig. 3 in which the contours represent the elevations, above present sea level, of the Last Baltic Ice Lake shoreline, interpolated between observed values at the localities marked by the black circles. In most instances the evidence is based on morphological criteria but at several locations, such as in Småland, it has been possible to verify the identification using lake-isolation data. The main feature of this shoreline, expressed as its elevation above present sea level, is a decrease in elevation with increasing distance from the ice margin at 10.3 ka BP; from a maximum greater than 140 m in southern Finland and southern-central Sweden to about 40–80 m in Gotland and vanishing below present sea level in the southern Baltic Sea (Fig. 3). This shoreline is understood to correspond to the highest level reached by the Baltic Lake before the ice barrier was breached and is believed to be a synchronous event. Other shorelines that have been mapped for the region include the maximum level reached during the Ancylus phase and the maximum level attained in the early Litorina Phase (e.g. Eronen 1983).

This previously unused shoreline information will be used here to establish the validity of the rebound model of Lambeck
et al. (1998a). This is particularly important because this earlier work led to the conclusion that the ice over the Baltic basin at the time of maximum glaciation was considerably thinner than has sometimes been inferred. Ice models in which ice heights increase with distance inwards from the ice margin according to a parabolic function result in excessive rebound over the region, such that the Baltic is predicted to have been open to marine influence for much of the Lateglacial period. The inversion of the shoreline data, excluding the pre-Litorina Baltic information, led to the conclusion that whilst such parabolic models were appropriate over northern and western Scandinavia they overestimated the ice thickness over the southeastern and southern region. However, by precluding these data, the observations available over these latter regions are restricted to the postglacial stage and much potential information is not used. Thus this additional data is used here to provide an independent check of the rebound model, particularly during the early stages of the ice retreat, to test whether the conclusions previously drawn are valid, and to examine whether these models can be further refined. The resulting rebound model is then used to predict the locations of possible barriers to Atlantic incursions into the Baltic basin as well as the timing of the isolation and marine incursion events. Finally the various sources of observational information are combined to provide a comprehensive model of the spatial and temporal evolution of the southern Swedish shorelines since the time of the last retreat of the ice from the area.

The glacial rebound model

The behaviour of the lake levels can be attributed, at least in a first approximation, to the consequences of the adjustment of the crust to the deglaciation of Scandinavia and to the global increase in ocean volumes resulting from the concomitant melting of this and the other major ice sheets. Models of isostatic rebound at position \( \varphi \) predict the sea-level change \( \Delta z(\varphi, t) \) with respect to the present sea level, as well as the topography \( h(\varphi, t) \) defined relative to the sea level at time \( t \). The former quantity can be related to various rebound parameters as (cf. Lambeck 1995)

\[
\Delta z(\varphi, t) = \Delta z_0(\varphi, t) + \Delta z_F^E(\varphi, t) + \Delta z_I^H(\varphi, t) \quad (1a)
\]

where \( \Delta z_0 \) is the eustatic sea-level change, \( \Delta z_F^E \) is the contribution from the isostatic rebound of northern Europe, and \( \Delta z_I^H \) is the contribution from the isostatic rebound of the other major far-field ice sheets: Laurentia, Greenland and Antarctica. The isostatic terms contain the vertical displacements of the crust and the deformation of the sea surface (the geoid) caused by the changing gravity field of the Earth as the ocean ice mass is redistributed over the surface. They include, therefore, both the glacio- and hydro-isostatic contributions. An iterative scheme is used to evaluate rigorously the hydro-isostatic contributions. The isostatic terms are functions of the parameters defining the Earth’s response to surface loading on time scales of \( 10^4 \) to \( 10^5 \) years as well as of the geometry of the load and, to a lesser degree, of the geometry of the ocean basins through time. The earth-model parameters considered here represent a three layered model comprising (i) a lithosphere of effective elastic thickness \( H_e \), (ii) an upper mantle with an average, effective, viscosity \( \eta_m \) from the base of the lithosphere down to the \( 670 \) km seismic discontinuity, and (iii) a lower mantle with an effective viscosity of \( \eta_m \). Within each of these layers the density and elastic moduli are depth dependent and consistent with values derived from seismic models. The mantle is, therefore, treated as a compressible medium.

Once the sea-level change is evaluated from (1a) for a particular set of earth- and ice-model parameters, the topography at any time \( t \) is defined by

\[
h(\varphi, t) = h(\varphi, t_0) - \Delta z(\varphi, t) \quad (1b)
\]

where \( h(\varphi, t_0) \) is the present topography at \( \varphi \) and \( h(\varphi, t) \) represents the topography at time \( t \) measured with respect to coeval sea level.

The eustatic function \( \Delta z_0 \) in (1a) is constrained from observations far from former ice margins where the isostatic contributions are relatively small and can be predicted with some degree of certainty (Nakada & Lambeck 1988; Fleming et al. 1998). Likewise, the contribution from the distant ice sheets is relatively small and can be adequately predicted because it is not sensitive to the details of the ice distribution within these particular ice caps. For the Scandinavian region the term of greatest sensitivity to the choice of input parameters is \( \Delta z_F^E \). The isostatic terms are expressed as expansions of surface spherical harmonic functions with the summation carried out to degree 256.

Detailed models for the glacial rebound of Scandinavia have been developed and compared with observational data to estimate the optimum combination of earth-model and ice-model parameters (Lambeck et al. 1998a). The observational data base used in these analyses included more than 1000 observations of the heights and ages of shorelines throughout the Scandinavian, Baltic and North Sea region, all referenced to mean sea level. Within the confines of the Baltic Sea, including the Gulfs of Bothnia, southern Sweden and Finland, data before \( 8.1 \) ka bp were not used in Lambeck et al. (1998a) because of the possibility that the levels corresponded to one or other of the dammed-up Baltic lake levels rather than to sea level, as is predicted by the relation (1a).

For the ice models the positions of the margins are assumed known for the period of ice sheet growth and the subsequent decay, but the thickness is considered to be only approximately known. The models are defined spatially on a \( 25 \times 25 \) km grid and temporally at \( 500 \) year intervals. The preliminary model (denoted by SCAN-0) is based on the ice retreat contours of
Andersen (1981) and Pedersen (1995) and on the ice thickness estimates of Denton & Hughes (1981). The thickness profiles are described as quasi-parabolic functions (cf. Paterson 1994), based on the assumption that the ice sheet is frozen to its base. The maximum ice thickness in SCAN-0 is about 3400 m centered over the Gulf of Bothnia and northern Finland. This thickness is ice-model dependent but it is also one of the parameters that can be inferred from the comparison of model predictions with observations of shoreline age-height relations. Earlier studies have indicated that the maximum value of 3400 m is excessive (Lambeck et al. 1990) and the average scaling value of 0.62, inferred from the comparisons with the geological data (Lambeck et al. 1998a), is adopted here. This scaled ice model is referred to as SCAN-1.

An evaluation in Lambeck et al. 1998a of the discrepancies between the observed shoreline elevations and the corresponding predictions based on the SCAN-1 ice-model revealed a consistent pattern whereby the predicted values were systematically greater than the observed values in the east and south and the reverse occurs in the west and north. This led to the conclusion that the parabolic profile assumption for the ice sheet is inappropriate for the eastern and southern regions; that the ice heights there increased much more slowly with distance in from the ice margin than is assumed to be the case when the parabolic profiles are adopted. This led to the second, empirically derived, model (SCAN-2) in which the ice load was adjusted for a regionally variable scaling parameter. This ice model may not be wholly decoupled from the solution for the earth-model parameters and one aim of this paper is to establish whether the SCAN-2 model is also consistent with the pre-Litorina data from the Baltic region and whether or not there has been an effective separation of the earth- and ice-model parameters in the solution discussed in Lambeck et al. (1998a).

A preliminary comparison of model predictions with observations

The Gotland and Småland Baltic-level time series

A comprehensive discussion of the observed age-elevation relationship of shorelines in Småland and Gotland, south-eastern Sweden, is given by Svensson (1989, 1991) and this data set forms the basis for the initial comparisons between observations and predictions. The data include the heights and timing of basin isolation and ingress events, the elevations of the Last Baltic Ice Lake shoreline, and the height of the maximum Ancylus shoreline. Each of these shorelines are usually assumed to be synchronous events, occurring at nominal times of 10.3 and 9.3 ka bp respectively. The basin isolation events establish the timing of the final separation of small lakes or basins from the appropriate Baltic lake stage. The observations correspond to the periods of falling lake levels during the Baltic Ice Lake and Yoldia Sea phases and again during the post-Ancylus maximum stage. A number of the isolations appear to have occurred during or soon after the drainage of the Baltic Ice Lake and they provide a control on the timing of this event. The ingressions define the timing of flooding of the small basins by the Baltic lake and correspond to periods of rising lake levels from the time of the onset of the Ancylus stage until the time of the maximum Ancylus transgression. A second interval of ingressions may have occurred in the lead up to the Litorina maximum shoreline but this falls outside the time interval considered both by Svensson and the present analysis.

The data set of Svensson contains a number of references to the first occurrence of a brackish-water fauna or sediments but these bear no relationship to the threshold elevation of the basins within which they occur and do not provide a measure of sea or lake level. Some of the observations refer to the timing of the first occurrence of the Ancylus flora. If the basin isolation occurred well within the Ancylus phase then such observations provides a measure of the timing of the end of the isolation. But because it takes time for the Ancylus flora to be widely established within the Baltic basin, the actual time of first isolation could be earlier. Two of the ingress points tabulated by Svensson are stated to be close to an ingestion only and these are not used in the comparisons. Likewise, observations identified as being close to the isolation event are not used. For the majority of the basins sampled the time scale is determined by pollen zones calibrated locally to the conventional radiocarbon time scale (Svensson 1991). Thus all observations refer to the radiocarbon time scale, consistent with the time scale used for the ice-sheet reconstruction and for all the other geological data used in Lambeck et al. (1998a).

Figure 4a compares the predicted and observed sea levels for a subset of the data for eastern Småland encompassing information from sites between Oskarshamn in the north, Nybro in the west and Öland in the east and southeast (see Fig. 1). The prediction shown in this figure is for the average position of the selected sites and no attempt has been made here to reduce the observed heights for the spatial variability in rebound that can be expected over a region of this size. The predictions are based on the SCAN-1 ice model and on the nominal earth-model parameters (earth-model E-1, Table 1) found to be representative of the response of the Earth to surface loading on glacial time scales (Lambeck et al. 1998a, 1998b). These predictions assume that the lake level is, at all times, equivalent to mean sea level (curve i). Other than not modelling the large drop in level at 10.3 ka bp, the other feature of these predictions is that they lie well above the observed levels for the entire period under consideration. Figure 5a, b illustrates the difference between observed and predicted shoreline elevation for Småland and Gotland. (The predictions of the height of mean sea level above the present level are now for the individual localities and depths of each observation.) Rejecting the improbable interpretation that the lake levels lie below open sea level throughout the interval, the immediate result is that these differences are strongly negative throughout the time interval in question due to the predicted levels being too great. Introducing into the predictions the 30 m fall in the Baltic Ice Lake level at 10.3 ka bp, in accordance with the offset seen in the observed values illustrated in Fig. 5a, only exacerbates the discrepancy between the observed and predicted levels (curve ii, Fig. 4a).

The rapid change of about 30 m observed at both the Småland and Gotland localities at the end of the Baltic Ice Lake phase contains no information on the mantle rheology and is everywhere the same, but the amplitude of the subsequent transgression during the early part of the Ancylus phase will be a function of this response and observations of this rise could place constraints on the model parameters. Another feature of these comparisons is that while the differences for the two localities in Fig. 5 exhibit fluctuations at 10.3 ka bp of very similar magnitude, the amplitudes of sea-level change differ by amounts that are model dependent.
predictions for both localities are close to zero, consistent with the Yoldia Sea being at sea level. The subsequent observations point to a rise in the Ancylus Lake level to a maximum of about 13 m above the coeval sea level. Again, the pattern of the differences for the two localities, Småland and Gotland, are consistent with each other. However a small offset occurs between the result for the two regions, with the Gotland residuals being about 10 m greater than those from Småland (Fig. 5c, d), suggesting that there is scope for some further model improvement. Figure 4b illustrates the predicted lake heights based on this earth ice-model combination and in which the Baltic Ice Lake barrier is held constant at 30 m above sea level (curve ii). The Ancylus barrier is allowed to develop to a maximum elevation of about 13 m above sea level between 9.5 and 9.3 ka BP before returning back to mean sea level by about 8.5 ka BP. Other than for a systematic height offset of about 15 m, these predictions represent well the general characteristics of the Gotland–Småland observations.

The elevation of the Last Baltic Ice Lake shoreline

The equation relating the predicted and observed levels of the Last Baltic Ice Lake shoreline at a location \( \varphi \) for the \( n \)th data point is (cf. Lambeck et al. 1998a)

\[
\Delta \zeta_{\text{ancylus}}(\varphi) + e_\sigma(\varphi) = \Delta \zeta_{\text{ancylus}}(\varphi, E_{\text{ancylus}}, \eta) + \delta \zeta.
\]

The \( \Delta \zeta_{\text{ancylus}}(\varphi, E_{\text{ancylus}}, \eta) \) represents the predicted height, relative to present sea level, at position \( \varphi \) corresponding to the \( n \)th observation \( \Delta \zeta_{\text{ancylus}} \) of the lake level, for a particular earth (\( E_{\text{ancylus}} \)) and ice model (\( \eta \)). The corrections to the observations are denoted by \( e_\sigma(\varphi) \) and the standard deviations of the latter are \( \sigma_\varphi \). The \( \delta \zeta \) represents the quasi-instantaneous drop from the Baltic Ice Lake level to the Yoldia sea level at 10.3 ka BP (a positive value means a fall in lake level). If the lake-lowering event occurred in a very short period of time so that the overall ice volumes did not change in the interval, the lake-rise fall can be assumed to have been spatially uniform. The estimate of the lake elevation above mean sea level after the fall at 10.3 ka BP follows from (2) as

\[
\delta \zeta = \frac{\sum_{n=1}^{N} [ (\Delta \zeta_{\text{ancylus}}^{\text{ancylus}}(\varphi) - \Delta \zeta_{\text{ancylus}}^{\text{ancylus}}(\varphi)) / \sigma_\varphi^2 ]}{\sum_{n=1}^{N} \sigma_\varphi^2}
\]

and for observations of equal variance this reduces to

\[
\delta \zeta = (\Delta \zeta_{\text{ancylus}} - \Delta \zeta_{\text{ancylus}}^{\text{ancylus}}).
\]

If the model is satisfactory then the residuals for the individual data points

\[
\sigma_\varphi(\varphi) = (\Delta \zeta_{\text{ancylus}} - \Delta \zeta_{\text{ancylus}}^{\text{ancylus}}) + \delta \zeta
\]

would be normally distributed with amplitudes similar to the observational accuracies. The intrinsic measurement error of the shoreline elevation is about 1-2 m but a more conservative estimate of 3 m has been adopted here to allow for uncertainties in the identification of the lake level and to allow for the uncertainty in the formation age attributed to the shoreline. Comparisons of the observed \( \Delta \zeta_{\varphi} \) and predicted \( \Delta \zeta_{\text{ancylus}} \) shoreline elevations, with the latter based on the SCAN-1 ice model and the nominal earth model E-1, are quite unsatisfactory, yielding \( \phi_{\varphi} = -38 \) m, whereas the expected value, based on the
earlier comparisons for Småland and Gotland, is of the order +30 to +35 m. If an offset \( \delta = 35 \) m is assumed (cf. Fig. 5) then the residuals \( r_n \), defined by (3c), are everywhere strongly positive except for the northern Vättern region (the dashed contours in Fig. 6a). Thus if the earth-model parameters are correct, a major modification of the SCAN-1 ice sheet is in order in which the ice thickness over the southern and southwestern areas must be significantly reduced. An approximate estimate of such an ice-thickness scaling factor \( \beta \) is given by (cf. Lambeck et al. 1998a)

\[
\beta = \frac{(\Delta \zeta_n - \Delta \zeta_p)(\Delta \zeta_p - \Delta \zeta_e)}{(\Delta \zeta_e - \Delta \zeta_n)}
\]

where \( \Delta \zeta_n \) is the eustatic sea level corresponding to the far-field ice sheets of North America and Antarctica. Figure 6a (the solid contours) illustrates this scaling function for the region and indicates that a substantial reduction in ice thickness is appropriate over southern Finland, particularly over the eastern region, as well as over the southern Baltic region and southern-most Sweden. This scaling factor applies to the SCAN-1 ice model and the corresponding scaling of the original ice model SCAN-0 would be \( 0.62 \times \beta \). Hence the overall reduction of the ice thickness for these regions is by as much as 60-70% of the thickness estimates in the original Denton & Hughes (1981) model. The spatial variation of this scaling factor is in fact very similar to that inferred independently in Lambeck et al. (1998a) from the Scandinavia-wide shore-line data.

A similar comparison of observations with predictions based on the SCAN-2 ice model and the relations (3) yields much improved results. Now \( \delta = 30 \) m, of the expected sign and magnitude, the residuals \( r_n \), are much reduced and exhibit a less systematic spatial pattern than before, and the ice-scaling parameter \( \beta \) is near unity in most areas (Fig. 6b).

**The Baltic lake barrier locations and lake levels**

If no erosion or deposition of sediments occurred since the retreat of the ice, the topography \( h(t) \) at any time \( t \), expressed relative to sea level at that time, follows from (1b). For the preliminary modelling the present-day topography is taken from the GTOPO30 data base (USGS 1996) which has a resolution of 30" or about 1000 m in latitude by 500 m in longitude. The data base for the bathymetry of the Baltic Sea is from the Danish Hydraulic Institute complemented with digitized nautical charts for the Danish straits. The bathymetry for the large lakes Vänern and Vättern, not included in these data bases and unimportant in the present context, has been set to zero. Once the sites for potential barriers to the exchange of Atlantic and Baltic waters have been identified their exact locations and heights are established from the appropriate
1:50 000 topographic maps (Lantmäteriverket 1997; Statens kartverk 1995). In some localities the potential barriers consist of narrow and deep valleys where the effective threshold level will exceed the level of the rock barrier. For example, the present relationship between input and discharge of the Vänern basin, with the latter occurring through the incised Götaland, is such as to keep the lake level a few meters above the rock threshold level. At times of higher input into the lake this difference between the rock threshold and the water level will be correspondingly higher. These cases will be discussed as they arise and the preliminary predictions are all based on the present water or rock threshold levels.

Figure 7 illustrates a typical reconstruction of the topography for southern Sweden, the western Baltic, and the Danish straits at 10.5 ka BP, about the time of the Younger Dryas. This is based on equations (1), the SCAN-2 ice model and the nominal earth model E-1. The ice limit illustrated reflects the relatively coarse definition (25 km x 25 km) of the ice model used in the rebound calculations. At this epoch the Baltic is separated from the Atlantic by the elevated terrain of eastern Denmark and by the ice barrier to the west of Vättern. The resulting lake level, corresponding to the Last Baltic Ice Lake, could therefore be higher than illustrated, with the maximum elevation above sea level depending on the height of the barrier in the Danish straits, or about 25 m for this particular model.

For this earth and ice model, the topography has been generated at successive time steps from the time of the Baltic Ice Lake until the onset of the Litorina Sea. At each epoch the lake level is raised until it overflows into the Atlantic and this establishes both the location and minimum elevation of the barrier for that epoch. Figures such as 7 confirm that flow between the Atlantic and Baltic was impeded in one or both of two regions: within the Danish straits, including the Öresund and the Store and Langelands belts, as well as the 'Darss sill' between Darss in Germany and Falster in Denmark (Fig. 1), and across southern-central Sweden between latitudes of about 57.5° and 59° north, a region that is presently characterized by large lakes, incised river valleys and low-elevation plains.

More detailed reconstructions are shown in Fig. 8 for the Danish belts for two different epochs (10.3 and 8 ka BP). Shown are the present coastline, the predicted open-ocean
does the sill elevation evolve through time but the position of the sill also shifts because of the differential vertical movement of the crust across the region. Thus at around 10.3 ka BP the Öresund represents a broad saddle of near constant elevation of about 25 m that controls the lake level. But at 8 ka BP it is the Langelands Belt that determines this level. The evolution of the sill elevation is further illustrated in Fig. 9a for several localities. Initially the lake level is controlled by the Öresund topography with the sill elevation increasing from the time the area first became ice free to a maximum at about 11 ka BP. Subsequently the sill height is lowered and from about 9.2 ka BP onwards the control has shifted to the Langelands Belt and Dars’s sill. The first predicted marine incursion through the Danish straits for this particular earth-ice model combination is after 8 ka BP. The evolution of the heights of these three sills are based on the assumption of zero sedimentation or erosion but curve (ii) corresponds to the Langelands sill on the assumption that there has been 5 m of sedimentation across it after about 7 ka BP when the Öresund barrier was open to the Atlantic. (A similar modification of the Dars’s sill would be required to open up the Baltic to the Atlantic before 8 ka BP.) This has the effect of advancing both the time at which this sill first becomes the outflow to the Baltic and the time of the onset of the Litorina phase. Curve (iii) illustrates the consequence of a hypothetical rapid erosion event within the Öresund during an early phase of the Baltic Ice Lake: the lake level before this event (illustrated here to occur at 11.2 ka BP with a lowering of 7 m) would be higher than otherwise predicted leaving behind a well-defined shoreline in the deglaciated parts of the Baltic basin.

Fig. 8. Shoreline reconstructions for the Danish straits at two epochs: (a) 10.3 ka BP before the rupture of the ice barrier at Mt Billingen in southern-central Sweden and (b) at 8 ka BP. Shown are: (i) the present shoreline (the boundary between the darker and paler grey shades), (ii) the marine shoreline (the boundary delimiting the white zone) and (iii) the Baltic shoreline (the boundary delimiting the dashed zone). The lake limits at 10.3 ka BP correspond to the Baltic Ice Lake; those at 8 ka BP to the late stage of the Ancylus Lake. The prediction for 8 ka BP shows that the Langelands and Store belts are close to sea level and tidal fluctuations could result in the first marine incursion occurring before this time.

shoreline at the stated epoch, and the predicted maximum lake shoreline, the last on the assumption that no lower sill elevations occur outside of the belt region. Of note is that not only

From the sequence of predictions, of which Fig. 10 forms a subset, potential sill locations controlling the flow across southern-central Sweden, with the sill heights estimated from
the 1:50,000 topographic maps for Sweden and Norway in parentheses, are:

(a) east of Vättern, near Motala (89 m) (location 3, Fig. 10a), controlling the flow from the Baltic into Vättern;
(b) between Vättern and Vänern through Viken (91 m) (location 2, Fig. 10a), controlling the flow between these two lakes;
(c) south of Degerfors in the Närke area (location 4, Fig. 10b) where the connection between the Baltic and the Vänern basin occurs via Mälaren and Hjälmaren to the west of Stockholm (105 m);
(d) the outlets of Vänern, via the Gota Älv (44 m, present lake level above mean sea level) and Uddevalla (53 m) (location 5, Fig. 10a);
(e) a northern outlet of Vänern via a series of northerly trending lakes and valleys, with a sill occurring at Otteid (location 6, Fig. 10a) in southeastern Norway (112 m), separating Store Le and Oymarksjöen, and a second sill to the east of Halden in the narrow Stenselva gorge (105 m, present water level above mean sea level).

All of these locations have previously been identified as possible barriers that may have controlled flow between the various water bodies (e.g., Björck 1995). Figure 9b illustrates the predicted evolution of these potential sill elevations for the nominal earth- and ice-model parameters. During Lateglacial times all the barriers are below sea level and if it were not for the ice cover, the region as a whole would provide a broad passage between the Baltic Sea and Atlantic Ocean. Immediately after the breach of the ‘Billingen ice barrier’ at about 10.3 ka BP, all of the sills are still below sea level but, because the northern passage through Närke remains closed by ice, the initial contact between the Atlantic and Baltic is predicted to occur via Vättern through the Motala and Viken gaps (cf. Fig. 10a). As the ice retreats northwards the depressed Närke and Degerfors passage becomes ice free and the flow between the two bodies of water now occurs primarily here. With time, the isostatic rebound causes these passages to become increasingly constricted. Thus, at about 9.4 ka BP both the Viken and Motala sills are lifted above sea level and the connection between Vänern and Vättern via Viken is closed. However, the connection between the two large lakes is maintained, via the more northerly connection through Askersund (location 8, Fig. 10b), until such a time as the Degerfors sill is lifted above sea level. This is not until 8.85 ka BP for this particular set of model parameters (Fig. 9b), and the predicted isolation of the Baltic from Vänern only begins at this time. The outlets of Vänern to the Atlantic are also evolving and the first barrier between the two water bodies occurs in the northern passage via Otteid and Stenselva, at about 9.1 ka BP, followed by the Uddevalla barrier at about 8.6 ka BP and finally the Gota Älv.

The predictions illustrated in Fig. 9b express the evolution of the elevation of the Baltic lakes in the absence of any control from barriers outside of southern-central Sweden so that these results need to be combined with the parallel evolution of the

Fig. 9. Evolution of the barrier elevations through time at (a) the three controlling locations within the Danish straits; the Öresund, the Darss sill and the Langelands Belt based on earth-model E-1 (Table 1) and the SCAN-2 ice model. The prediction (i) for the Langelands Belt is based on the present bathymetry of the belt and prediction (ii) is based on the assumption that there has been 5 m of sedimentation in the belt during Mid- to Late Holocene time (c. 3 ka BP) once the Öresund passage is open. The prediction (iii) for the Öresund is based on the assumption of a rapid erosion of tills and other sediments in this locality so as to reduce the barrier level by 7 m at about 11.2 ka BP. (b) Evolution of the barrier elevations through time at the controlling locations within southern-central Sweden: near Motala on the western side of Vättern where the topography controls the connection between this lake and the Baltic; near Viken, a low topographic passage between Vänern and Vättern; near Degerfors in Närke where the topography controls the flow between the Baltic and Vänern; the upper Gota Älv near Vänersborg for (i) present water threshold level of 43 m and (ii) a threshold level 5 m higher; at Otteid, northwest of Vänern. The Stenselva sill gives a similar result to that for Otteid. (c) The combination of the significant sills from the two localities. The heavy dashed line indicates the predicted evolution of the elevation of the Baltic Lake for the E-1 and SCAN-2 models.
level of about 10 m is predicted to occur at about 8.5 ka BP. After this time the falling barrier in the Danish straits lies below the southern-central Sweden barriers and the former now controls the Baltic level, with the next marine incursion predicted to occur at about 7.6 ka BP, initially through the Langelands and Store bælt and the Darss passage, marking the onset of the Mastagloua sub-stage.

The predicted times for these events are generally later than the observed times and point to a need to modify some aspects of the model. The sequence of predictions illustrated in Figs 9 and 10 rests on a number of model assumptions which will be addressed below. One assumption is that the earth-model parameters E-1, based on the solution in Lambeck et al. (1998a), is consistent with the Baltic shoreline data and this is examined first in the following section.

**Baltic Ice Lake test of earth and ice models**

The Baltic Ice Lake shoreline information was not used in the modelling of the glacial rebound in Lambeck et al. (1998a) and it therefore provides an independent test of the model; one that is particularly important because the observations are from the southern Baltic shore where few data are otherwise available. A similar procedure to that used in Lambeck et al. (1998a) is adopted here to systematically explore the earth-model space in order to establish a set of parameters that result in an optimum description of the rebound. The criterion adopted for a measure of the agreement between observations and predictions based on an earth-model (E) is

$$\Psi_k = \frac{1}{N} \sum_{n=1}^{N} \frac{[\langle \Delta h_{n,0} \rangle - \langle \Delta h_{n,0}^* \rangle]}{\sigma_n^2}$$

where the summation is over all \( N \) observations of the Baltic ice lake elevation given by Svensson (1989). The \( \Delta h_{n,0} \) refer to the predicted sea-level change corresponding to the observation \( n \) for the earth-model \( E \). If the predictive model is satisfactory and the estimates of \( \sigma_n \) are reasonable, then the expected value of \( \Psi_k \) is unity. The earth model that leads to the least variance is denoted by \( E_k^* \).

An estimate of the accuracy of the earth-model parameters for this least variance model follows from the quantity (cf. Lambeck et al. 1998a)

$$\Phi_k = \frac{1}{N} \sum_{n} \left( \frac{\Delta h_{n,0}^k - \Delta h_{n,0}^*}{\sigma_n} \right)^2$$

where \( \Delta h_{n,0}^k \) is the predicted shoreline elevation for the optimum earth model \( E_k^* \) at observation site \( n \), and \( \Delta h_{n,0}^* \) is the lake-level offset corresponding to \( E_k^* \). Earth models for which \( \Phi_k \leq r \) differ from the best-fitting model \( E_k^* \) by an amount equal to or less than \( r \) times the average observational error of the data.

The limits of the parameter space searched for the minimum value of \( \Psi_k \) are

$$\begin{align*}
30 \leq H_k & \leq 150 \text{ km} \\
2 \times 10^{10} \leq \eta_m & \leq 2 \times 10^{11} \text{ Pa s} \\
10^{21} \leq \eta_m & \leq 3 \times 10^{22} \text{ Pa s}
\end{align*}$$

which bracket values found in most previous studies of Scandinavian glacial rebound data (Lambeck et al. 1990; Mitrovica & Peltier 1993; Fjeldskaar 1994). Partial results based on the ice model SCAN-1 are illustrated in Fig. 11 for

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Fig. 11. Variance factor defined by (5) as a function of (a) upper- and lower-mantle viscosity for $H = 65$ km and SCAN-1 and (b) lithospheric thickness and upper-mantle viscosity for $\eta_u = 10^{22}$ and SCAN-1. The dashed lines represent contours of constant $\delta \pi$ defined by (3a). (c) Same as (a) but for SCAN-2, and (d) same as (b) but for SCAN-2. The shaded regions in (c) and (d) correspond to the range of models defined by the model-reliability criterion $\Psi_k = \leq 2.25$ (eqn. 6). (e), (f) Same as c, d but in which the offset $\delta \pi$ has been set to 30 m rather than included as an unknown in the solution.

The model subspace defined by $\eta_u - \eta_l$ with $H = 65$ km (Fig. 11a) and by $\eta_u - H_l$ with $\eta_l = 10^{22}$ Pa s (Fig. 11b). The dashed lines correspond to the estimate of $\delta \pi$ for each of the corresponding earth models $E_k$. Throughout the model space (7) the solutions are unsatisfactory, with large values for $\Psi_k$ (80) and negative values for $\delta \pi$, confirming earlier conclusions (Lambeck et al. 1998a) that, irrespective of the choice of earth-model within this parameter range, the ice model SCAN-1 provides a poor description of the glacial rebound of Scandinavia.

Similar analysis for the SCAN-2 ice model yield a much better agreement between predictions and observations of these shoreline elevations (Fig. 11c & d). The minimum $\Psi_k$ is now much reduced and the corresponding $\delta \pi$ estimate is consistent with the observations from localities such as Småland and Gotland. The optimum solution $E_k^*$ within the
entire $H_l$-$\eta_m$-$\eta_m$ model space occurs where $\Psi^2_{\Phi}$ is a minimum and the corresponding accuracy estimates, based on (6) are indicated by the shaded parts in the two model subspaces illustrated in Fig. 11c & d. In these examples $\Phi^2_{\Phi} \leq 2.25$ has been adopted such that models that lead to predictions that agree with the corresponding observed values to within 1.5 $\sigma_\alpha$ with $\sigma_\alpha = 3$ m, are assumed to be acceptable. (Because the value of 3 m is greater than the original accuracy estimates of 1–2 m, this corresponds approximately to the 90–99% confidence limits.)

The optimum solution of the model parameters is then defined as

$$H_l = 60 \pm 30 \text{ km}$$
$$\eta_m = (3.3 \pm 1.5) \times 10^{20} \text{ Pa s}$$
$$\delta_\zeta = 32 \pm 5 \text{ m}.$$  

(8)

The upper-mantle viscosity is well constrained at about $(3-4) \times 10^{20}$ Pa s but the lower-mantle viscosity $\eta_m$ is largely unconstrained within the bounds set in (6). The error bars for $\delta_\zeta$ reflect the range of values that occur within the earth-model parameter space defined by $\Phi^2_{\Phi} \leq 2.25$. These solutions point to a strong trade-off occurring between the $\eta_m$ and $\delta_\zeta$ as can be seen in Fig. 11c, particularly when $\eta_m \geq 4 \times 10^{20}$ Pa s. Thus models with $\eta_m$ near the lower and upper limits of the explored range yield values for $\delta_\zeta$ that are either significantly less than or greater than the observed values from Småland and Gotland. In so far as the lake-level fall has been estimated independently at these localities, the observed estimate $\delta_\zeta$ could be used as a further constraint in the above models. Repeating the search through the model space (6) with ($\Lambda_{\eta_m} - \Lambda_{\eta_m}$)=$\delta_\zeta$, in equation 5 yields a much improved solution for the lower-mantle viscosity, as is illustrated in Fig. 11e and f for $\delta_\zeta = 30$ m, while the estimates for the other two earth-model parameters remain essentially unchanged. Assuming an accuracy estimate for $\delta_\zeta$ of 5 m and propagating this into the estimates of the uncertainty of the earth-model parameters, now yields the solution

$$H_l = 60 \pm 20 \text{ km}$$
$$\eta_m = (3.5 \pm 1.0) \times 10^{20} \text{ Pa s}$$
$$\eta_m = 1(\pm 0.5) \times 10^{22} \text{ Pa s}$$

(9)

with

$$\delta_\zeta = 30 \pm 5 \text{ m}.$$  

(9b)

This earth-model parameter solution is similar to that previously found from the shoreline elevation-age data for the Scandinavian region which excluded the pre-Litorina Baltic data (cf. (9) with the model $E-1$ defined in Table 1). It is also consistent with the results for earth- and ice-model parameters inferred from the analysis of the more recent mareograph data for the region (Lambeck et al. 1998b) (compare the model results E-1, E-2 with E-6 in Table 1 where E-6 represents the mareograph results).

While these model parameters yield a satisfactory prediction of the elevations of the Baltic ice lake, some of the residuals, defined by (3c), remain larger than expected from observational accuracy considerations alone. The spatial pattern of these residuals, as well as the resulting estimate of the scaling parameter $\beta$ defined by (4), are very similar to that found previously for the nominal earth-model $E-1$ (Fig. 6b). The pattern of these residuals is of shorter wavelength than would be expected if they were the result of erroneous earth-model

<table>
<thead>
<tr>
<th>Earth model</th>
<th>Lithospheric thickness (km)</th>
<th>Upper-mantle viscosity ($\times 10^{20}$ Pa s)</th>
<th>Lower-mantle viscosity ($\times 10^{22}$ Pa s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>E-1</td>
<td>75 $\pm$ 10</td>
<td>3.6 $\pm$ 0.5</td>
<td>0.8 $\pm$ 0.5</td>
</tr>
<tr>
<td>E-2</td>
<td>60 $\pm$ 20</td>
<td>3.5 $\pm$ 1.0</td>
<td>1.0 $\pm$ 1.0</td>
</tr>
<tr>
<td>E-3</td>
<td>65</td>
<td>3</td>
<td>1.0</td>
</tr>
<tr>
<td>E-4</td>
<td>100</td>
<td>3</td>
<td>1.0</td>
</tr>
<tr>
<td>E-5</td>
<td>65</td>
<td>5</td>
<td>1.0</td>
</tr>
<tr>
<td>E-6</td>
<td>100 $\pm$ 35</td>
<td>$5^{(+3)}_{(-5)}$</td>
<td>$2^{(+3)}_{(-5)}$</td>
</tr>
<tr>
<td>E-7</td>
<td>80</td>
<td>3.5</td>
<td>1.0</td>
</tr>
</tbody>
</table>

Table 1. Summary of earth-model parameters for the three-layer models

parameters (compare Fig. 6b with the model errors illustrated in fig. 13 of Lambeck et al. 1998a) and a more plausible explanation is that they reflect local inadequacies of the ice model. Thus a reduction in the SCAN-2 ice thickness by about 10% over southern Sweden, south of Vättern, removes much of the discrepancy there. Likewise a comparable reduction in the ice heights over the southwestern part of the ice sheet removes the discrepancy over south-eastern Finland. The negative values for the residuals over western Finland indicate that a small (about 10%) increase in the SCAN-2 ice thickness over the Gulf of Bothnia is appropriate. Such further adjustments are small when compared with the initial scaling of the SCAN-0 model and do not vitiate the overall conclusions drawn about the ice volumes and asymmetry in the ice-thickness profiles between the western part of the ice sheet and the southern and southeastern parts.

Discussion

While the analysis of the different data sets lead to consistent estimates for the earth-model parameters as well as to broadly consistent conclusions about the ice thickness estimates, the uncertainties in these parameters remain substantial (cf. models E-1, E-2, E-6 in Table 1). This can be expected to introduce uncertainties into the predictions of the timing and elevations of the sill elevations. Figure 12a illustrates some examples of this dependence for models summarized in Table 1. An increase in lithospheric thickness results in the onset of both the Ancylus and Litorina stages being brought forward in time: increasing the lithospheric thickness from 65 km to 100 km, for example, results in an earlier onset of the Ancylus transgressive phase by about 400 years (compare the results for E-3 and E-4 in Fig. 12a). Similar dependencies exist for the upper-mantle viscosity: an increase in this parameter reduces the amplitude of the pre-Holocene lake level, increases the duration of the Ancylus transgression and delays the onset of the Ancylus and Litorina stages (compare the results E-3 and E-5 in Fig. 12a).

The dependence of the sill elevation predictions on ice thickness is significant, as is illustrated in Fig. 12b in which the sill evolution is given for the earth model E-3 and in which the ice thickness of the SCAN-2 model has been uniformly
modified by a factor $\beta$. A decrease of ice thickness over the entire region ($\beta=0.8$) leads to a reduction of the duration of the Yoldia stage and to an earlier peak in the Ancylus transgression as well as to an earlier onset of the Litorina stage. An increase in overall ice thickness ($\beta=1.2$) has the opposite effects: the duration of the pre-Ancylus opening of the Baltic is lengthened, the timing of the onset as well as the end of the Ancylus isolation is brought forwards, and the amplitudes of the lake levels are lowered.

What the above model results indicate is that predictions of the principal stages of evolution of the Baltic lakes are sensitive to some of the model-parameter assumptions and that differences between these predictions and observations may provide further constraints on the model. These results also indicate that some trade-offs between the various model parameters can be expected. For example, increasing the lithospheric thickness reduces the predicted duration of the Yoldia stage (compare results for models E-3 and E-4 in Fig. 12a) but this can be countered by an increase in the ice thickness (compare the results for $\beta=1.0$ and $\beta=1$ in Fig. 12b).

The model predictions are based on the assumption that the topography has not been modified by erosion and sedimentation over time and that estimates used for the present elevation are realistic. With the 30$^\circ$ area-averages used to reconstruct the palaeotopography, the minimum elevation of the valley floor and mountain saddles defining the overflows tend to be overestimated and the predictions for the timing of the isolations may occur earlier than would be the case for lower threshold elevations and for this reason the high resolution topographic maps have been used for the potential sill localities. For example, for the Göta Älv the maps indicate that the present water threshold level occurs at between 43 and 45 m above sea level whereas the 30$^\circ$ resolution topographic data puts the sill elevation at about 55 m to the south of Vänersborg. Most of the potential sill sites through river valleys and saddles are close to or at bedrock and erosion since the region became ice free may not have been important. One exception is the Langelands Belt and the Dariö localities where, as appears to be the case from an examination of a number of cores in the area, there has been sedimentation since the time of the opening of the Öresund (e.g. Bennike & Jensen 1998).

The model elevations of the sill may therefore be underestimated by the corresponding amount and the flooding of the sill would occur earlier than predicted. This is illustrated in Fig. 9a for 5 m of sedimentation in the Langelands Belt. In this case the estimate of the onset of the marine influence of the Baltic is advanced by nearly 500 years (cf. curves i and ii).

The predictions illustrated in Fig. 9 assume that for the constricted passages such as the Göta Älv the present relationship between the rock level and water level threshold is representative of past conditions but this need not have been the case when there were still residual ice sheets and the water input into the various basins was greater than today. Björck (1995) has examined this relationship in some detail and concludes that the past water level relative to bedrock at Göta Älv may have been as much as 5 m higher than at present, yielding a critical level of about 48–49 m. Because the rate of rebound is relatively slow by 8 ka BP an increase in the threshold level by 5 m advances the isolation of Vänern from the sea by about 400 years (Fig. 9b). Likewise, Björck places the effective elevation of the Vänern-Lapland barrier at 20 m above the present water surface, yielding a critical level of 125 m that is above the Ötteen sill.

The sill-elevation predictions are based on the nominal eustatic sea-level function in which ocean volumes have been constant for the past 6000 years. Other studies have led to the conclusion that this may not have been the case and that eustatic sea levels at 6 ka BP may have been a few metres lower than today (e.g. Nakada & Lambeck 1988; Fleming et al. 1998). If earlier eustatic sea levels have also been underestimated in the nominal model, then this would lead to higher elevations for the Danish barriers.

In the following discussion the predictions of shoreline elevations and locations are based on the solution 9 (solution E-2, Table 1) and the SCAN-2 ice model. These predictions are first compared with the observations from Småland and Gotland discussed above and then compared with the predicted Last Baltic Ice Lake shoreline elevations throughout the region. Next, the sill locations, their timing and heights are predicted for this model and compared with some of the observational evidence. Finally, based on the outcomes of these comparisons, a sequence of reconstructions of the Baltic is proposed. As before, all ages correspond to conventional radiocarbon ages.

The model predictions for the elevation of the Last Baltic ice Lake shoreline based on the solution E-2 leads to residuals between observations and predictions that are similar to those illustrated in Fig. 6b for the nominal earth model E-1 (Table 1). The majority of the residuals lie within the
range ± 10 m and exhibit a spatial pattern that can be attributed to inadequacies in the ice thickness values of the SCAN-2 model of within ± 15%. In particular, from a series of forward-model calculations, the residuals point to a need for a small decrease (c. 5%) in ice thickness over southern Sweden and for a similar magnitude increase over the Baltic Sea and Lithuania. Such changes are wholly within the error limits of the SCAN-2 model and much smaller than the overall differences between the SCAN-1 and SCAN-2 ice-thickness estimates.

Figure 13 illustrates the discrepancies between the predictions and observations for Småland and Gotland based on the above earth- and ice-model combination, with the predictions based on the assumption that the lake level corresponds to mean sea level at all times. If the predictive model is correct then these differences for the two localities should be identical and provide a measure of the departures of the Baltic lake level from mean sea level. As for the nominal earth-model results (Fig. 5c, d), the amplitudes of the fluctuations for the two localities are similar but there is a small systematic offset between the two (Fig. 13a). Within the range of permitted earth-model parameters (Table 1), it is the dependence of this offset on lithospheric thickness that is most marked; an increase from 60 to 80 km removing about half of the discrepancy. The predictions are also sensitive to minor adjustments in the load model, as is illustrated in Fig. 13b in which the predictions for both localities are based on the same mantle viscosity model as before (solution E-2) but on a lithosphere of 80 km thickness and in which the ice thickness over southern Sweden, but not over Gotland and the western Baltic, has been reduced by 5%. Both modifications are within the noise level of the parameter values. Despite the agreement between the results for the two localities, a small discrepancy remains in that the predicted lake levels for the Yoldia stage lie above sea level (Fig. 13b). Thus a further constraint on the models could be imposed by requiring that lake levels during the Yoldia stage lie at sea level. This elevation is most critically dependent on differential changes in the ice heights for the two localities and can be largely removed by a decrease of about 5% in the SCAN-2 ice thickness over southern Sweden and an increase in ice thickness of about 10% over the Baltic east of Småland (Fig. 13c). Such modifications are consistent with the inferences drawn previously from the Baltic Ice Lake shoreline information.

**Elevation of the Baltic Ice Lake**

Figure 14 illustrates the evolution of the sill elevations for the combination of the earth-model E-2 and the ice-model SCAN-2. These predictions confirm that the level of the Baltic Ice Lake is controlled by the topography of the Oresund because the other barriers within the Danish straits are at higher elevations at this time. From the time this region became ice free, soon after 13 ka BP, the lake level is predicted to have risen initially and then to have remained at a nearly constant height of about 28 m between about 11.5 ka BP and 11 ka BP after which it fell slowly up to the time of its drainage (Fig. 14a), a behaviour that is predicted for a range of other plausible earth models that encompass the range of parameters expressed by models E-1, E-2 and E-6. At the time of drainage the lake is predicted to have been at 25-26 m above sea level. This is somewhat lower than indicated by the Småland and Gotland observations (Fig. 13c) or by the analysis of the shoreline elevations across the Baltic as a whole (the solution 8) but values of 20-28 m have been inferred from the evidence in Finland (Donner 1995). (The predicted level is of the topographic barrier; the lake level may be somewhat higher if flow out of the Baltic is restricted.) However, relatively modest
adjustments of some of the model parameters do lead to higher elevations than the values indicated in Fig. 14a. For example, a small reduction in ice thickness over southern Sweden or over the southwestern Baltic region enhances the sill elevation as is illustrated in Fig. 12b. Alternatively, a decrease in upper mantle viscosity leads to an increase in this elevation, as does a decrease in lithospheric thickness (see Fig. 12a).

Björck & Möller (1987), from field evidence in Blekinge, identified an earlier fall in the Baltic Ice Lake, of about 10 m in amplitude, at 11.2 ka BP. Svensson (1989) identified the same feature in Småland but with an amplitude of only about 5 m. The model predictions for the Danish straits do not point to any change in the sill location before the principal drainage event because the other potential barriers lie above the Öresund threshold throughout the Last Baltic Ice Lake phase (Fig. 14a). Thus the simplest interpretation of an earlier drainage event is that the sill at Öresund was deepened at this time, either by bed-rock erosion or by erosion of the sediments left by the retreating ice across eastern Denmark, as are now found in the southeastern part of the Kattegat and which could have formed a temporary barrier to the Baltic ice Lake. As the build up of water behind this barrier increased, the sediments would have been eroded until the bedrock, forming the present-day sill, was reached. At this time the new rock barrier is still above sea level so that no marine incursion into the Baltic is predicted at the time of this drainage event. The alternative hypothesis of Björck & Möller (1987) is that this lowering of lake level may have been associated with an early deglaciation phase around Mt Billingen followed by a readvance until the final ice retreat at 10.3 ka BP from the area. However, the crustal depression in this area is substantial at this time and any ice retreat would have resulted in a lowering of the lake level to sea level, analogous to the 10.3 ka event.

**Duration of the Yoldia stage**

The onset of the Yoldia stage is determined by the assumed date of the major drainage event north of Mt Billingen and the termination of this stage is determined by the timing of the formation of a new barrier to Atlantic flow into the Baltic basin. For the model predictions illustrated in Fig. 14 the termination is controlled by the Degerfors sill because the lowest of the potential sills to Vänern, Göta Alv, remain below sea level until much later (Fig. 14a, b). The predicted location for this barrier is south of Degerfors and corresponds to a bare bedrock locality, sometimes referred to as the 'Svea River' and already identified by H. Munthe early this century. However, the termination of the Yoldia stage is predicted by this model to have occurred at about 9 ka BP (Fig. 14a, b), later than the nominal termination date at about 9.5 ka BP. An initially higher sill, eroded during the subsequent overflow of the Ancylus Lake, would advance this date but this is improbable as the sill elevation does not appear to have been subjected to significant postglacial water erosion (Björck 1995).

One possibility for advancing the prediction of the termination of the Yoldia stage is to assume that the thresholds to the west have been modified through time such that it was these that controlled the actual lake level and that the Vänern basin remained part of the Baltic lake for much of the subsequent Ancylus stage, as suggested by Björck (1995). For the model results illustrated in Fig. 14a, b, the three sill localities of Otteid, Uddevalla and Göta Alv, as well as the Stenselva barrier, are predicted to be well below sea level until

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*Fig. 14.* (a) Predicted evolution of sill elevations for earth model E-2 and ice model SCAN-2. Curve (i) corresponds to the Öresund sill in which erosion of sediments occurred at about 11.2 ka BP. Curve (ii) corresponds to the Göta Alv sill for which the water level threshold has been raised 5 m above present level. Curve (iii) corresponds to the sill threshold that is required at the Göta Alv for the Vänern level to be at the same height as the Baltic lake. (b) Expansion of (a) in the time interval 9.5 to 7.5 ka BP. (c) Predictions of the Göta Alv sill elevation for: (i) earth model E-2 and SCAN-2, (ii) the same model parameters as in (i) except for an increase in lithospheric thickness to 80 km, (iii) same as (i) but with a reduction in ice thickness over southwestern Sweden of 10%; (iv) same as (ii) but with a 15% reduction in ice thickness, (v) predictions for the Degerfors sill with the same parameters as (ii), (vi) the same as (v) but for an increase in ice thickness over the Närke region of 5%. The Langelands prediction is based on the same parameters as in (ii) but with a reduction in ice thickness of 5% over southern Sweden.
some time after 9.5 ka BP. The Stenselva and Otteid passages are the first to be closed, at about 9.2 ka BP, followed by the closure of the Uddéavalla passage some 200 years later and by the Göta Älv passage, assuming the 48 m level adopted by Björck (1995), at 8.7 ka BP (curve ii). Thus some substantial sill modifications are required if the isolation of Vänern and the Baltic from the Atlantic occurred as early as 9.5 ka BP or even 9 ka BP. As for the other sills, an effective way of advancing the timing of this event is to modify those model parameters that reduce the amount of crustal depression in glacial and lateglacial time; increasing the lithospheric thickness or decreasing the ice load. Alternatively, the introduction of a higher effective threshold level at the three localities also advances the timing of the termination. Erosion since deglaciation may have been significant in the case of the Göta Älv sill where the present river bed is incised into the surrounding bedrock by up to 12 m (see Fig. 9 of Björck 1995) but presumably this incision is a cumulative result of successive periods during the Weichselian and earlier when Lake Vänern overflowed, and not just of the latest episode. The threshold level for the Otteid sill may also have been higher since the rate of outflow at this locality is likely to have been controlled not by the Otteid topography itself but by the narrow passages downstream in the Stenselva between Aspern and Femsjön (Björck 1995). Likewise, the threshold level within the Uddéavalla passage will have been higher than the present rock level because of the constructive channels that the waters have to navigate about midway between Vänersborg and Uddéavalla (Lantmäteriverket 1997, map 8B NO Vänersborg). However, if the adjustments of the threshold elevation were the only parameters to be modified, the Göta Älv threshold would have had to be some 20 m above the present lake level (curve iii, Fig. 14a, b) in order for the termination of the Yoldia stage to be advanced to 9.5 ka BP. This is an unlikely high value as it exceeds the depth of the downstream gorge below the surrounding terrain.

The Ancylus transgression

If, once the Baltic has been isolated from the Atlantic, the lake level is controlled by the evolution of the sill south of Degerfors, the model predictions indicate that the transgressive phase would have lasted about 350 years by which time the Danish strait barrier had been sufficiently lowered to form the subsequent outlet for the lake (Fig. 14a). This compares with an observed duration of the transgression of about 200-300 years (see also Fig. 13c), although the observational constraint on this interval is not well defined (Björck 1995; Svensson 1989). The predicted elevation of the maximum transgression is about 12 m (Fig. 14a) and this compares well with the observed amplitude at the Smaland and Gotland localities (Fig. 13c).

If, on the other hand, the Baltic lake level is assumed to have been controlled by the Göta Älv barrier, then the Ancylus transgression is predicted to have occurred over a much longer time interval, of about 600 years (curve iii, Fig. 14a, b), because the rate of crustal rebound in this region was substantially less than for the Närke-Degerfors region. The observation of the duration of the transgression, therefore, does not favour the hypothesis that it was the Göta Älv barrier that controlled the level of the Ancylus lake. Also, the minimum increase in the level of this threshold required to keep Vänern at the same level as the Baltic to the east of Degerfors is a still improbably large 15 m, if all other model parameters remain unchanged. Such an increase does bring the onset of the Ancylus stage forward to about 9.3 ka BP, in better agreement with the observed time of onset, but the duration of the transgression remains considerably longer than the observed value.

Other combinations of parameters can also bring the onset of the Ancylus forward and keep the Vänern and Baltic water bodies at the same level. Increasing the lithospheric thickness to 80 km, within the range of acceptable earth models as defined in Table 1, and reducing the ice thickness over southwestern Sweden by about 60%, for example, yields sufficient uplift for the Göta Älv to act as a barrier by 9.5 ka BP, provided that the lake threshold was about 5 m above the present water level (curve iv, Fig. 14c). But by increasing the lithospheric thickness, the Degerfors sill-elevation prediction is also raised such that the Baltic level will be higher than the Vänern level during the late stage of the transgression (curve v, Fig. 14c) unless the ice height over the Närke region is increased by about 5% (curve vi, Fig. 14c). However, this does delay the timing of the maximum transgression to about 8.9 ka BP.

Clearly there is an inconsistency between the three observations relating to the Ancylus transgressive phase that cannot be resolved by any modification of the model predictions. A rapid transgression in the early Ancylus stage, within about 300 years, is consistent only with the Degerfors sill acting as the barrier. But if the critical observation is the absence of a height difference across this sill this means that the Göta Älv barrier controlled the height of the Baltic, implying that a substantial (c. 15 m) down cutting of the barrier occurred in lateglacial time. However, if the latter sill controlled the Ancylus lake level then the duration of the transgression is increased to about 700 years. One possibility is that this lengthening is a consequence of the use of the radiocarbon time scale which is not strictly linear in the time interval covered by the Ancylus stage (e.g., Stuiver & Reimer 1993), but departures from linearity are too small to explain these discrepancies between the model predictions and observations of the duration of the transgressive phase. An alternative explanation, that the rebound of the Vänern outlets was faster than predicted by the model, is inconsistent with the evidence for rebound from other localities in the region. At Hanneberg, close to the Göta Älv sill (Björck & Digerfeldt 1982), downstream from this sill in the areas of Rissveden (Svedhuse 1985) and Sandsjöbacka (Passe 1987), as well as for the Kroppeljäll area north of the Uddéavalla passage (Björck & Digerfeldt 1991), the rebound model based on the parameters (9) and the SCAN-2 ice model give an adequate representation of the observed changes for the interval from about 9-12 ka BP when the localities were all open to marine conditions (Fig. 15); the small differences between the predictions based on this model and the observations can be readily attributed to small changes in either the earth- or ice-model parameters.

End of the Ancylus stage

The end of the Ancylus stage is marked by the first appearance of brackish-water diatoms, the first signs of which occur in the Blickinge area of southern Sweden at about 8.5 ka BP. This suggests that the influx of marine water through the nearby Danish straits occurred at this time, but because the early salinity levels were very low, the barrier may only have been
Fig. 15. Comparison of observed (with error bars) and predicted relative sea-level changes in southwestern Sweden based on the earth model E-2 and the ice model SCAN-2. The observed values are from (a) Björck & Digerfeldt (1982), (b) Björck & Digerfeldt (1991), (c) Svedhage (1985), (d) Pässe (1987), (e) Björck & Digerfeldt (1986).
breached during times of very high ocean tides. Full marine conditions are believed to have been established by 7.8-7.5 ka BP. The inference from the Småland-Gotland observational evidence (Fig. 13c) is that the Baltic level reached mean sea level before about 8 ka BP, whereas the model predictions place this event a few hundred years later, at about 7.7 ka BP in the case of Fig. 14a. A small reduction in ice thickness does bring this event forward in time, as does an increase in lithospheric thickness (Fig. 12) but this model dependence is generally inadequate to explain the entire discrepancy. Any sedimentation within the belters subsequent to the diversion of the main Baltic flow through the Öresund, predicted to occur just before 7 ka BP (Fig. 14a), will also lead to an earlier marine incursion into the Ancylus lake but some 7-8 m of sediments are required in the Langelands Belt or across the Darsö sill to bring this event back to about 8.3 ka BP (cf. Fig. 9a).

Conclusion

The solutions for earth-model parameters E-2, based on the observed elevations of the Last Baltic Ice Lake shoreline, are consistent with values estimated from a more comprehensive data base of shoreline age-height relationships for Scandinavia as a whole. However, this latter analysis excluded all pre-Litorina Baltic data because of the unknown height differences between the various lake levels and mean sea level for the earlier periods. In consequence data coverage for Late Glacial times was limited for the southern regions of the Scandinavian ice sheet and the Baltic Ice Lake data provides a complementary and independent data base for testing the rebound models. That the two solutions for the earth-model parameters are in good agreement, provides therefore, a satisfactory test of the results. Further verification of the model parameter inferences comes from the analysis of the mareograph records for both sea-level and lake-level change across Scandinavia (Lambeck et al. 1998a) (model E-6, Table 1). However, the comparisons of predictions of the Baltic lake-level oscillations based on the earth model E-2 with the observations from Småland and Gotland suggests that some modifications of this model may be warranted, in particular, that the lithospheric thickness should be increased to about 80 km (cf. Fig. 13). Likewise the observational constraints on the timing of the various lake changes also favour a small increase in this parameter. This modest increase in the lithospheric thickness compared to the value from solution E-2 (model E-1 Table 1) remains within the stated range of uncertainty of this parameter (cf. models E-1 and E-2, Table 1). Together, the Scandinavian data provide good constraints on effective rheological parameters that define the mantle response to surface loading on time scales of glacial cycles, constraints that are consistent with those obtained from comparable analysis for the British Isles, for example (Lambeck et al. 1996).

The rebound analysis based on the Ice Lake elevations also confirms that the SCAN-2 ice sheet leads to a very considerable improvement in rebound modelling when compared with the ice model SCAN-1. Not only does the analysis confirm that thick Late Weichselian ice sheets of some 3500 m maximum thickness at the time of the Last Glacial Maximum are ruled out, but it supports models in which the ice thicknesses over the Baltic region are proportionally thinner than for the northern and western parts of the ice sheet. But, while the SCAN-2 ice model yields a good first-order solution to the rebound of the region, discrepancies between model predictions and observations do point to further modifications that can be made. Modifications that are also consistent with those inferred from the analysis of the mareograph data (Lambeck et al. 1998b). These include a small decrease in ice thickness over southern Sweden, the Danish straits region, and the eastern part of the Gulf of Finland, and a small increase in ice thickness over the western part of the Baltic Sea, Lithuania, and over the Gulf of Bothnia. With these modifications, the model predictions of the various stages of the evolution of the Baltic (Fig. 14) are in good agreement with the observational evidence, the only substantial discrepancy that cannot be reconciled being the question of whether the Degerflors or Götä Alv sill was the controlling barrier for the Baltic Lake. All the modelling evidence points to it being the former, but in this case a significant height difference develops between this lake level and the Vänern basin, something that appears to be inconsistent with the observational evidence (Björck 1995).

The agreement between observations and predictions is sufficiently satisfactory to attempt to use the model as a predictive tool for lake level and shoreline evolution across the Baltic region and for lake-level changes at specific localities. Such predictions are given in Figs 16 and 17 primarily to illustrate the complex spatial and temporal pattern of this change across the region and to identify some of the characteristics of the changes, further observations of which should contribute to subsequent improvements in the models.

Figure 16 illustrates the reconstructions of the region for four epochs. These results are based on the earth-model parameters E-7 (defined in Table 1). This model differs from E-2 only in that the lithospheric thickness has been increased to 80 km, consistent with values that provide overall best agreement with the Baltic Ice Lake shoreline evolution as well as with the Småland-Gotland data. The reconstruction for the Baltic Ice Lake Shoreline corresponds to the epoch just prior to the drainage event at 10.3 ka BP. The Baltic is at about 28 m above sea level and overflows through the Öresund. It extends into Vättern while Vänern is open to the Atlantic via a number of channels, including the Götä Alv. The Yoldia Sea reconstruction corresponds to an epoch of about 10 ka BP with the Baltic open to marine influence via Vänern and Vättern, the more northerly connection via the Närke region being closed by the ice sheet. The opening between Vänern and the Atlantic is via narrow channels, Götä Alv and the Udevalla passage, and a major marine influence in the Baltic is improbable in view of the considerable meltwater originating from the retreating ice.

The maximum Ancylus transgression reconstruction corresponds to the epoch at about 9.3 ka BP and the lake is predicted to be at about 12 m above sea level. The barriers to marine input into the Baltic are at Degerflors and in the Danish straits. Vänern is seen to be isolated because this reconstruction is based on the digital data base for the present topography which gives higher average elevations for the incised and narrow river valley. As discussed above, the more detailed topography indicates that the lake is still open to the Atlantic at this time. Vättern is isolated from both the Baltic and Vänern, with the sills at Motala, Askersund and Viken all being above the height of the other water bodies. The Danish straits represent a barrier with the lowest elevations occurring in the Darsö and Langelands areas. The final reconstruction corresponds to the early Litorina stage when the Baltic is isolated from the Atlantic through southern-central Sweden but is open to marine influence, initially though the Langelands Belt and the Darsö area.
Figure 17 illustrates the predicted lake shoreline elevations based on the same earth-ice-model combination used to construct Fig. 16. The selected localities are identified in Fig. 16. Illustrated are the predicted shoreline-height relationships based on the assumption that all shorelines correspond to mean sea level (curve i) or that the shorelines correspond to lake levels that may at times lie above mean sea level (curve ii).

The model-derived lake level function used in this calculation is also shown (curve iii).

The predictions for the localities of Norrköping and Stockholm are representative of areas of rapid rebound. For the early part of the record (13 ka BP) sea-level (curve i) remains relatively constant because: (i) the major phase of deglaciation starts only at about 15 ka BP; (ii) of the nature in
Fig. 17. Predicted evolution of shoreline height-age relationships at localities in southern-central Sweden. The curves marked (i) correspond to predictions of the shoreline elevations in the absence of ice or land barriers between the Baltic and Atlantic. Those marked (ii) predict the lake-level elevations on the assumption that the Baltic has been periodically isolated from the sea. The elevation of the lake level with respect to coeval sea level is indicated by the curve (iii).
which the various terms contributing to the sea-level change combine. (iii) the delaying nature of the rebound response to the changes in surface loading creates the maximum rebound after the time of maximum glaciation. These northern localities are, however, ice-covered until about 11 ka BP and early Baltic shorelines will not have formed there. Further south (such as at Karlskrona or Öresund), this early part of the sea-level curve is characterized by a falling level.

For the northerly sites a small cusp (A) in the sea-level curve (i) is predicted to occur at about 10.5–10 ka BP because of the stationarity of the ice front during the Younger Dryas. A second cusp (B) is predicted to occur at about 6 ka BP due to the completion of melting of the ice by this time, a feature that becomes more pronounced to the south such as in Gotland.

Also illustrated in these figures are the predicted lake-level curves (ii) for the various localities. The evidence for the major drainage event at 10.3 ka BP is predicted to lie below sea-level south of about Blekinge and at Bornholm where elevated lake shorelines are predicted to occur between about 12.5 and 6.2 ka BP. No raised shorelines are predicted for any time after the region became ice-free for localities within and to the west and south of the Langelands and Store bæts.

During the Ancylus transgression the rate of rebound for all sites is less than the rate of uplift of the Baltic sill at Degerfors, with consequence that this lake stage is marked by a rising level throughout the region bounded by Fig. 16. In these models the lake surface again coincides with sea level at about 8.1 ka BP after which shoreline elevations north of about Oskarshamn (e.g. Stockholm) continue to fall. In contrast, at Oskarshamn, Gotland and further south, shoreline elevations increase towards a maximum elevation corresponding to the Litorina transgression when, in these models, occurs at 6 ka BP when much of the melting of the major ice sheets has ceased.

Whilst the results illustrated in Figs 16 and 17 are consistent with much of the observational evidence for the Baltic lake phases since the time of the last glaciation, several areas can be identified where further improvements are desirable.

(i) The incorporation of other lake shorelines such as the maximum Ancylus shoreline and the Litorina maximum, and the expansion of the Baltic data base to include other localities.

(ii) The development of improved ice models that have the essential characteristics of the empirical model SCAN-2 but that are consistent with glaciological principles and geomorphological indicators of base surdness.

(iii) Use of the calendar time scale, both for the shoreline age information and ice-margin retreat, so as to remove uncertainties that may arise from non-linearities in the radiocarbon time scale, such as the plateau effects in the Last Weichselian and early Holocene. This will, however, require a major new dating and re-interpretation effort for much of the shoreline evidence.

(iv) Explore rheological models with a greater degree of viscosity stratification than the three-layer models considered here (cf. Lambbeck et al. 1996).

(v) Use higher resolution topography and bathymetry information for reconstructing high resolution shoreline positions that could be used as a guide to both the interpretation of the shoreline and lake-sediment evidence and to any archaeologically interpretations of the relationships between human land use and the Baltic shore (e.g. Åkerlund 1996).

(vi) Extend the model back to Early and Mid-Weichselian time.

C. Smither provided valuable assistance with the numerical modelling and, together with C. Krayshek digitised the bathymetric charts for the Danish straits. M. Ekman of the National Land Survey of Sweden and B. G. Harson of the Norwegian Mapping Authority provided the 1:50 000 scale maps for the sill regions. I am grateful to N.-O. Svensson of Lund University for valuable discussions on the lake-level data for Småland and Göteborg, and to S. Börck from the University of Copenhagen for helpful discussions on the evolution of the Baltic. Helpful comments on the manuscript were also received from G. Hannon and B. Berglund from Lund University. This research was partly funded by an International Science and Technology grant from the Commonwealth of Australia, Department of Industry, Science and Technology.

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Received 24 March 1998; revised typescript accepted 22 September 1998.

Scientific editing by Mike Humbley.