NEWS 5-VERSION A

STATE OF THE STAGE 3 PROJECT
AT THE START OF ITS THIRD PHASE

A SUMMARY OF THE DELIBERATIONS AND DECISIONS
OF THE PHASE TWO WORKSHOP

Cambridge 29-30 June 1998

WITH PLANS, SCHEDULES AND COMMITMENTS FOR
PHASE THREE

This report is issued in two versions:
5-A: for participants in the Phase Two
Workshop of June 28-30; this version
lacks Appendices 2 & 5 distributed at
the Workshop.
5-B: for all other members; because of
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the two most important colour plates
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1. Northern European Stage 3 ice sheet and shoreline reconstructions: Preliminary results.

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Inputs into the Stage 3 climate models include the extent and height of the ice sheets and the geometry of the shorelines for the corresponding epoch. For the recent past, in areas of relative tectonic stability, the two inputs are closely related through the process of glacial isostasy. As ice-sheets grow, water is removed from the oceans and a global lowering of sea-level occurs, but under the influence of the changing surface loads of ice and water, the crust responds by subsidence or rebound and the change in relative sea-level, the change with respect to the land which itself is going up or down, exhibits a quite complex spatial pattern. Near the centres of ice accumulation, for example, the crust is increasingly depressed and the sea encroaches onto the land, even when the overall volume of the ocean is decreasing and globally sea-level is falling. Further from the ice sheets the changes are more subtle but can, nevertheless, be important.

Numerical models with high resolution have been developed over recent years that give realistic representations of the spatial variability of the sea-level change and shoreline evolution if the evolution of the ice sheets through time are known. One unknown in these models is the rheological response of the Earth to the surface loading. Models for the time following upon the Last Glacial Maximum are generally satisfactory and consistent with a broad observational data base from which it has become possible to estimate a set of parameters that describe this response to loading (Lambeck et al. 1996; 1998). Back in time the models become increasingly less reliable because the history of the ice sheets become more uncertain and the observational data base for testing the models becomes more restricted. Nevertheless, it should be possible to arrive at representative models for the epochs in question using the results for the past 20,000 years as guide.

In so far as the Stage 3 climate models are restricted to the European region which has been influenced by the northern glaciations, the models will require very detailed information on the Scandinavian ice-sheets through time in order to include the consequences of crustal rebound, but the more distant ice models must also be considered, although in less detail, in order to include the changes in global ocean volume. Also, because the Earth's response to loading is not instantaneous but contains a memory of earlier loading events, the rebound model will have to start at some time earlier than Stage 3.
Input parameters.

Earth model parameters. Earth model parameters previously found to give an adequate description of the rebound for the last deglaciation and post-glacial stages for northern Europe are used here. The response is described by a three-layered mantle model comprising an elastic lithosphere of thickness 65 km, a viscoelastic upper mantle with a viscosity of $4 \times 10^{20}$ Pa s, and a viscoelastic lower mantle with a viscosity of $10^{22}$ Pa s (Lambeck et al. 1998). Realistic density and elastic moduli profiles are used for the mantle and core.

The eustatic sea-level function. The oxygen isotopic signal of deep-sea cores has provided a useful proxy for eustatic sea-level change through successive ice cycles and the result of Shackleton (1987), scaled using the Huon terrace data, is adopted as a starting point. Figure 1 illustrates the function. A calendric time scale is used throughout, including the last 20,000 years. At the time of the Last Glacial Maximum (OIS-2) sea-levels are assumed to have been globally lower than at present by about 125 m. During the penultimate glacial maximum (OIS-6) at about 145-140 ka, the eustatic level is assumed to have been about 10 m lower than this. The earlier cycle, from the penultimate interglacial to OIS-5 is included in these models, although predictions for intervals OIS-3 and OIS-4 are not strongly dependent on this earlier cycle. The ice volumes locked up in the totality of the land-based ice sheets will have to be consistent with this function.

Figure 1. Eustatic sea-level function for the past 250,000 years based on the $\delta^{18}$O record of Shackleton (1987), and contributions to this function from (i) northern Europe, (ii) North America, (iii) Antarctica and other distant ice sheets.
The Fennoscandian ice-sheet. The limits of the Fennoscandian ice-sheets between the last interglacial stage OIS-5e and OIS-2 have been extracted from a variety of sources, including Donner (1995), papers by Mangerud and his colleagues, and the work of Houmark-Nielsen for the southwestern region. The resulting models are purely empirical, containing no glaciological-model inputs. Several different scenarios have been experimented with for OIS-3.

In the model the ice-free conditions corresponding to the Eemian stage (OIS-5e) came to an end at about 119ka by which time a small ice-sheet formed over the high ground of Norway and Sweden and which reached its maximum extent at about 110ka corresponding to the OIS-5d (see Figure 3b). The maximum ice limits for this period were somewhat smaller than those experienced much later during the Younger Dryas. The region subsequently was ice free (Brorup warm stage, OIS-5c) but again became ice covered during the OIS-5b stage when the ice grew to limits similar to those of the previous cold stage. During both these cold periods the contributions of these ice sheets to fluctuations in eustatic sea-level amount to only a few meters. During the subsequent 'warm' Odderade interstadial at about 82ka (OIS-5a), the region is again assumed to have been ice-free. After 80ka glaciation was again initiated and the region is now assumed to retain an ice cap with time-fluctuating margins until about 9000ka, the end of the Late Weichselian glaciation. The OIS-4 stage is represented by a cold phase at about 65ka during which the ice limits were similar to those that occurred subsequently at about 12 000 BP. The subsequent OIS-3 stage extends nominally from 59 to 24ka. Initially conditions in this interval were warmer leading to the Moershoofd interstadial at about 57ka. After that conditions became colder but sea-level oscillations indicate that this was not a uniform process because relative warm periods occurred at circa 40 and 30ka that correspond to the Hengelo and Denekamp interstadials respectively; they were separated by colder intervals at about 43 and 35 ka (Figure 1).

Two configurations for the Moershoofd interstadial are considered (see Andersen & Mangerud 1989); (i) a maximum ice-limit model in which the ice retreated by only relatively short distances from the OIS-4 limits so as to leave ice sheets that were similar to those of the earlier cold periods, OIS-5d and OIS-5b, and (ii) a minimum ice-limit model in which the ice sheet retreated considerably farther so as to remain mainly over the high ground of Norway and Sweden (cf. Figure 4c below). Ice limit fluctuations during the remainder of OIS-3 are assumed to have remained rather small, with the overall trend being one of an expanding ice front. The maximum ice volume, with limits near to where they again stood at about 12,000ka, occurred at about 35ka. Soon after the Denekamp interstadial the intensive glaciation phase leading up to the OIS-2 Late Weichselian glacial maximum was initiated. Only at this time does a large ice sheet begin to form in the model over the Barents Sea and its island groups; at earlier times the ice-sheet there is limited mainly to island mountain glaciers. The evolution of these ice margins is assumed to be in phase with that of Scandinavia and follows the pattern established for the past 20,000 years by Lambeck (1996). These postulated ice limits rest on few reliable data points but they are believed to be representative of conditions experienced in northern Europe during the Weichselian period when conditions became progressively colder. Even less certain will be the estimates of the ice volumes for this interval. The procedure adopted here is to assume that if at any time T before the Last Glacial Maximum (LGM) the ice limits were similar to those at a time T after the LGM, then
Figure 2. Predictions of sea-level change since the onset of OIS-5e at 129,000 yrs ago at locations in Scandinavia with contributions from (a) the distant ice-sheets of North America & Antarctica, (b) Scandinavia and (c) total predicted sea-level change.
ice thickness and volumes were also similar. For this extrapolation, ice model SCAN-1 discussed in Lambeck et al. (1998) is used as reference for the LGM and post-LGM period. For OIS-6 the large ice limits corresponding to the Saalian glaciation have not been used; instead, the ice sheet at this epoch corresponds to that of the Late Weichselian maximum limits except that the ice thickness has been scaled upwards such that the ice model contributes about 10m more to eustatic sea-level than it does during the LGM (see below). This simplification is of little consequence for the predictions for the OIS-3 and OIS-4 and has been introduced here for a different purpose, namely the examination of sea-levels during the OIS-5e, or Eemian stage.

Figure 1 illustrates the contribution of the Fennoscandian ice to the global eustatic sea-level function. Only at the times of maximum glaciation does this contribution become significant and for most of the remaining time it is no more than the uncertainty of the eustatic estimate and most of the latter must originate from fluctuations in the other major ice sheets, over North America and Antarctica.

The more distant ice sheets. Much less is known about the fluctuations in the North American Wisconsin ice sheet than about the European ice for the same period and even fewer observations exist from which conclusions can be drawn about the changing ice balance. Perhaps the only robust observation is that at no time in the past 160 000 years has the ice sheet extended significantly beyond the margins occupied during Late Wisconsin time. Thus North American ice volumes at the time of OIS-6 are unlikely to have exceeded those at OIS-2 and if the eustatic sea-level at OIS-6 exceeded that of the LGM contribution, then ice volumes of the European OIS-6 ice sheet must have been correspondingly greater than at the time of OIS-2. Hence the increased ice heights introduced above for the Saalian ice model.

Even less is known about pre-LGM Antarctic ice-volume fluctuations. Questions remain about the magnitude of the ice-volume of Antarctica at the time of the last glacial maximum but indirect inferences, based on the observed eustatic sea-level change and on the rebound models for the northern hemisphere lead to the conclusion that there must have been more ice at the time of the LGM than can be accounted for in the known northern hemisphere ice-sheets (Nakada & Lambeck 1988). Hence this missing ice has been placed in Antarctica, with a model in which the ice-volume increases linearly with time from the end of the last interglacial to OIS-2, and then decreases relatively rapidly during the late- and post-glacial stages. The maximum ice volume contributes 35m to efl. This function, together with the Scandinavian function, is subtracted from the 'observed' eustatic function to produce an estimate of the North American ice volumes (Figure 1). The ice-sheet geometry is then inferred from the assumption that when ice volumes at pre-LGM times T are equal to those of post-LGM times, then the ice models were also equal. For as long as the predictions are not made for regions near Antarctica or North America, and provided that the total contributions match the observed eustatic function, these choices have only minimal consequence for the sea-level predictions for the European region.

Model predictions.

The predictions for relative sea-level change across northern Europe will be the sum of the contributions from the distant ice sheets, including the effects of the time-variable water loading of the crust.
Figure 3. Reconstructions of ice sheets, shorelines and topography, and shoreline isobases for selected epochs: (a) onset of OIS-5e (Eemian) at 129 ka, (b) OIS-5d at about 110 ka, (c) OIS-5c, the Brorup interstadial at about 100 ka.
4a. 35 ka. OIS-3. (cold phase between Hengelo and Denekamp interstadials)

4b. 57 ka. OIS-3. Moerhoofd interstadial (Maximum ice limits)

4c. 57 ka. OIS-3. Moerhoofd interstadial (Minimum ice limits)

Figure 4. Same as Figure 3 but for the intervals during OIS-3. (a) a cold phase within OIS-3 at about 43 ka, (b, c) a warm phase early within OIS-3 corresponding to the Moerhoofd interstadial. Two alternatives are illustrated for this latter interval; a configuration in which the retreat of the ice margins since the preceding cold stage has been small (b), and a configuration in which the retreat has been more substantial (c).
and the crustal adjustment of the crust to the waxing and waning of the Scandinavian ice load. The former part is spatially nearly constant across the region (Figure 2) and closely resembles the eustatic function. But the Scandinavian contribution is much more variable: sea-levels from areas near the centre of the ice load (Bothnia) have remained above present sea-level for most of the time since the last interglacial whereas sea-levels from sites near the ice margin (e.g. Öresund) have been below present level for much of this interval.

Of greater interest for present purposes are the shoreline locations for the various epochs. These are illustrated in Figures 3 and 4, together with ice limits, ice thickness contours, topography, and the isobases of the rebound (the elevation at which a shoreline that formed at the epoch in question would occur today). At the start of the Eemian period (129ka) Scandinavia had come out of a very major glaciation that depressed the crust to such an extent that by the time the ice first vanished a major part of the rebound remained and relaxed only during the next ten thousand or so years, similar to the Holocene changes. The estimates given here (Figure 3a) are probably in excess because the adopted Saalian ice-sheet is unrealistically thick (a consequence of increasing the volume without expanding the ice margins beyond those of the Late Weichselian period), but they illustrate how it is possible to have well-elevated remains of Eemian shorelines in Scandinavia without this implying that globally sea-levels were higher at this time. A related consequence is that during the early part of the Eemian the Baltic is open to the Arctic ocean, in this case through central Finland and northern Karelia. (Shifting the ice over Russia as appears to have been the case at the time of the Saalian, may result in a southwards shift of the connection between the Baltic and the Arctic via the lakes of Onega and Ladoga.) At the OIS-5d cold stage global sea-levels have fallen sufficiently for the Baltic to have become isolated from the Atlantic (Figure 3b). However, when this ice has vanished over the region (OIS-5c) this separation remains because the eustatic level is below the threshold level of the Baltic which would remain as a freshwater lake. Shorelines from this epoch above present level should not be unexpected from the northern regions of the Gulf of Bothnia (if the evidence could have survived the subsequent cycles of glaciation).

The OIS-3 results are illustrated in Figure 4: (a) is for a cold phase within the stage, representative of either a period between the Moershoofd and Hengelo interstadials at about 43 ka or a period at about 35 ka between the Hengelo and Denekamp interstadials; (b) is for a relative warm phase within the OIS-3, representative of the Moershoofd interstadial but in which the ice limits correspond to a maximum ice-volume configuration, with the ice margins having retreated only over a relative short distance; (c) is the same as (b) but with the interstadial ice limits corresponding to a minimum ice-volume configuration in which the ice has largely retreated to the high ground of Norway and Sweden. Being the result of ice retreat, thin and stagnant ice is assumed to have remained behind over northern Bothnia, similar to the final stages of the last ice retreat at about 9ka. Because the global eustatic sea-level is the same for both the minimum and maximum models, the decrease in ice volume in the Scandinavian ice being compensated for by an equivalent increase in the ice volume of North America, the shoreline predictions for the two cases are very similar outside the immediate area of glaciation and the primary difference lies in dimensions of the Baltic Lake.
In Figure 4 the Baltic ice lake defined by the blue area corresponds to a lake level that is the same as coeval sea-level whereas the sill separating this lake from the Atlantic lies at several tens of meters above this level. The exact location of the sill depends on details of the ice sheet since part of the barrier, as in the case illustrated in Figure 4b, is formed by the ice itself in southern Sweden. The analogous situation during the last late-glacial stage is one where the sill alternates between the Danish Straits and southern-central Sweden with maximum sill elevations of 25 to 30 m. Thus in the three cases illustrated in Figure 4 the maximum limits of the Baltic lake will have been close to the first height contour (at 25 m elevation) of the topography. Other substantial periglacial lakes will also have formed along other sections of the southern and eastern ice margin, such as in Finland and Russia.

References


