The amplitude and decay of the glacial forebulge in Fennoscandia

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Forebulge is upward movement of the surface peripheral to glaciated areas above its equilibrium position. The amplitude and decay history of the forebulge will differ for different models of the viscosity of the mantle and rigidity of the lithosphere. For a low-viscosity asthenosphere situated between the rigid lithosphere and mantle mesosphere, the forebulge could be significant. The other extreme, a uniform mantle viscosity, would give no forebulge at all. Tilting of palaeo-shorelines in peripheral areas and the pattern of present uplift and subsidence is best modelled using the following parameters: (1) a lithosphere of flexural rigidity 10^{23} Nm; (2) a mantle of viscosity of 1.0×10^{21} Pa s; and (3) asthenosphere of viscosity 1.3×10^{19} Pa s. This model is used to study the amplitude and decay history of the Scandinavian forebulge. It is shown that the latest Fennoscandian glaciation produces a forebulge of 60 m at 15,000 BP, collapsing smoothly without any migration. The zero uplift isoline in Fennoscandia is relatively stationary over time in late- and post-glacial time, close to the maximum extent of the glacier. The minimum sea level at 15,000 BP is modelled to be 125 m below the present sea level, located 100 km from the former ice front.

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Introduction

The magnitude of post-glacial uplift in regions peripheral to glaciations has been discussed for many years. There is clearly little agreement in the literature regarding the isostatic behaviour in the peripheral areas. It is intuitively expected that mantle material squeezed out from under glaciated areas would tend to produce peripheral bulges. When the glaciers melted, the forebulge would migrate and collapse. The models were first articulated by Barrell (1914), followed by a paper by Nansen (1922) where he assumes the mantle masses to be transferred horizontally below the rigid lithosphere so that a significant forebulge was formed around the glacier. From the very beginning the discussion focused on channel flow ('bulge') models and deep flow ('punching') models. In the deep flow model the peripheral regions will first rise and then subside to isostatic equilibrium, while for the channel model the flow of material will be horizontal, into the central region which must be compensated by peripheral subsidence. The channel model gives first subsidence, then uplift to isostatic equilibrium in peripheral regions. The channel flow and deep flow models were both shown to account equally well for the history of uplift in the central, most rapid uplifting areas of Fennoscandia. When Daly (1934) observed a down drop rather than the expected uplift in peripheral areas, he invented the 'punching' hypothesis. On a theoretical basis Brotchie & Silvester (1969) found that elastic upward bending of the lithosphere above its equilibrium position could produce a forebulge that amounts to 3% of the maximum depression, independent of the mantle rheology. Walcott (1970) found that the distance the forebulge occurs from the ice edge depends on the flexural rigidity, and is independent of the ice load. Artyushkov (1971) and Mörner (1979) concluded that the uplift data suggest channel flow in a low viscosity asthenosphere situation between the rigid lithosphere and mantle mesosphere. Cathles (1980) has shown that a constant viscosity Newtonian mantle gives no peripheral bulges, while a model with a more fluid upper mantle than the lower mantle can give forebulges of significant amplitude. Brevik & Jensen (1992) have estimated the forebulge to within approximately 30 m by modelling the lithosphere as an elastic membrane using non-slip boundary conditions at the lithosphere-mantle interface.

In high resolution studies of the Fennoscandian uplift it has been shown that neither of the extreme models, a uniform mantle or a channel flow model, can explain the pattern of the observed present rate of uplift (Fjeldskaar & Cathles 1991b). It was also shown that a two-layered mantle, with a lower mantle viscosity slightly higher than upper mantle viscosity (Peltier 1987; Peltier & Tushingham 1989; Lambeck et al. 1990) is also not a viable option. The best-fitting model is the one that has a mantle viscosity of 1.0×10^{21} Pa s overlain by a 75 km asthenosphere of viscosity close to 1.3×10^{19} Pa s (Fjeldskaar and Cathles 1991a, b). This paper is focused on the theoretical amplitude and decay of the peripheral bulge of Fennoscandia in late- and post-glacial time using this best fit model. It is shown by a series of illustrations that the zero uplift isoline in Fennoscandia will be relatively stationary with time.

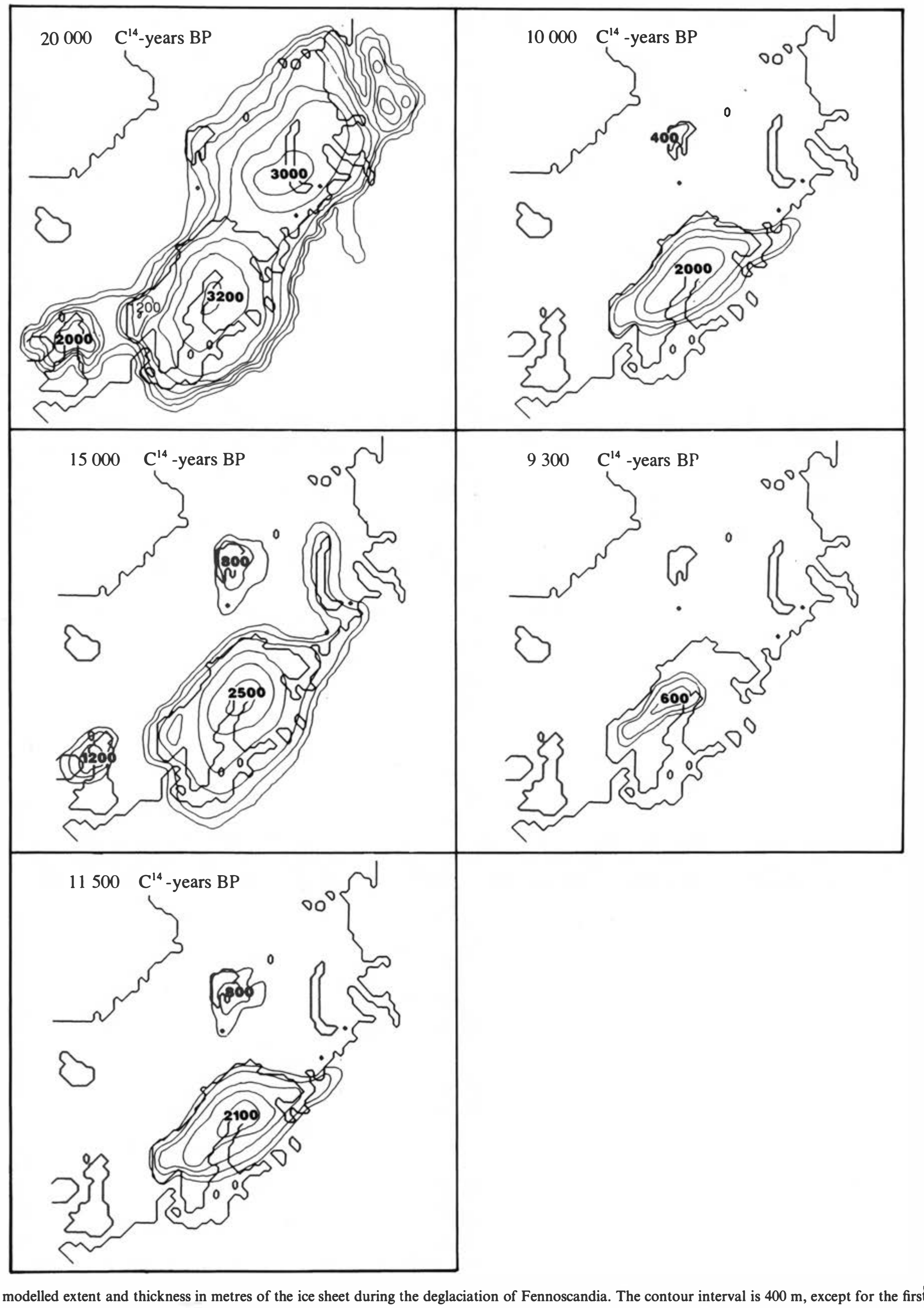


Fig. 1. The modelled extent and thickness in metres of the ice sheet during the deglaciation of Fennoscandia. The contour interval is 400 m, except for the first (800 m). The contour interval for the ice sheet of 9300 BP is 200 m, except for the first (400 m). Partly based on Denton & Hughes (1981).

Deglaciation data

The deglaciation of the last ice age is relatively well established by observations of marginal moraines. The deglaciation history used here (Fig. 1) was compiled by B. G. Andersen (Denton & Hughes 1981). Glacial thicknesses, however, are uncertain based on a paucity of direct geological evidence. The changes from one ice-sheet configuration to the next are assumed linear with time. The area is assumed to have been ice free 8500 BP, and the density of the glacier ice is assumed to be 917 kg m^{-3}.

Model approach

The Earth is modelled by a non-spherical viscous fluid mantle in which the viscosity may vary with depth, overlain by a uniformly thick elastic lithosphere. With this flat Earth model, we are able to treat the isostatic problem analytically, by the Fourier transform technique. The method used here is described in detail in Cathles (1975) and Fjeldskaar & Cathles (1991a).

Eustasy

Eustasy is vertical changes of sea level, and is of three types: (1) glacial eustasy, controlled by variation of the ocean water volume; (2) tectono-eustasy, controlled by variation of the ocean basin volume; and (3) geoidal eustasy. Geoidal eustasy represents changes in the ocean water distribution, caused by variations in the Earth's gravity field. This is an important eustatic factor, and is taken into account in the calculations here (method described in Fjeldskaar 1991). The remaining parts of the eustatic change (glacial- and tectono-eustasy) are here approximated by the eustatic curve of Fig. 2.

Hydro-isostasy

Hydro-isostasy, the isostatic compensation as a result of changes in the water load, is taken into account in an indirect way. The change in the water load is taken care of indirectly by the Fourier transform technique because the technique requires a load redistribution, i.e. the melt-water change equals the ice melting. By appropriately adjusting the computational box, the melting of the ice gives a sea-level curve (Fig. 2) roughly in accordance with some published eustatic curves (Fairbridge 1961; Shepard 1963; Mörner 1969). However, it gives smaller changes than reported by Fairbanks (1989). The melt-water effect of the total global ice redistribution were taken into account in this fashion. The model does not, however, take into account the real land–ocean distribution, and, as such, the technique implies that the melt-

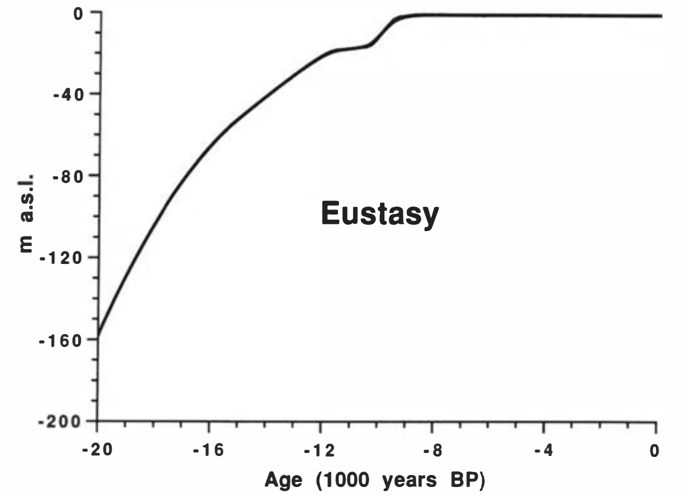


Fig. 2. Eustatic sea-level curve used to calculate the hydro-isostatic effect.

water changes take place outside the former glaciated area.

Present rate of uplift

The observed present rate of uplift in Scandinavia relative to mean sea level increases from 0 mm/yr at the western coast of Norway to 9 mm/yr in the Baltic Sea (Fig. 3). To obtain the uplift of the crust relative to the Earth's centre rather than relative to mean sea level, the uplift rate has to be corrected for eustatic changes. This involves (1) a correction for the gravitational effect of the uplift and (2) a correction for the uniform eustatic sea level change. The uniform eustatic component would, probably, add approximately 1 mm (cf. Lambeck &

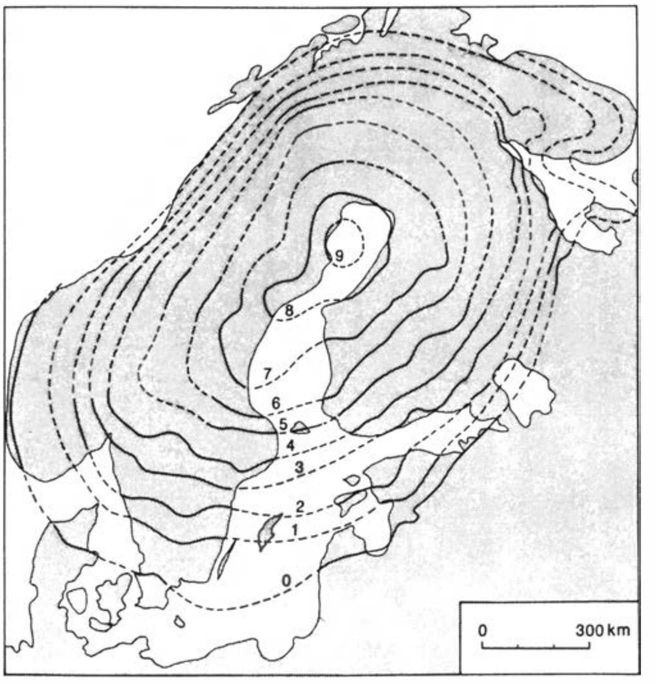


Fig. 3. Observed apparent rate of uplift in mm/year. After Ekman (1989).

Nakigoblu 1984) to the numbers given in Fig. 3. The theoretical gravimetric effect of the present rate of uplift gave a maximum geoidal rise of 0.47 mm/yr in the central Baltic Sea (Fjeldskaar & Cathles 1991b). The uplift of the crust relative to the Earth's centre is thus the sum of present rate of uplift, the uniform eustatic component and the gravimetric effect, adding up to 10.5 mm/year in central Fennoscandia.

Best fitting mantle model

Previous calculations (Fjeldskaar & Cathles 1991b) based on deglaciation models show that the present uplift pattern is mainly determined by the viscosity profile of the mantle. Changes in lithosphere rigidity (at least within the range of $1-100 \times 10^{23}$ Nm) cause only minor adjustments in the pattern. The best-fitting model is the one that has a mantle viscosity of 1.0×10^{21} Pa s overlain by a 75 km asthenosphere of viscosity 1.3×10^{19} Pa s (Fig. 4).

Flexural rigidity

The post-glacial sea-level changes in Fennoscandia have also been mapped by shoreline diagrams, which show the displacement and tilting of palaeo-shorelines. The tilting history for the Trøndelag area, for example, shows that the flexural rigidity 10^{23} Nm gives too steep shorelines, but 10^{24} Nm gives too low gradients (Fig. 5). Because the glacier thickness used in the modelling represent probable maximum thicknesses, it is reasonable to suggest that the flexural rigidity is less than 10^{24} Nm (50 km elastic thickness).

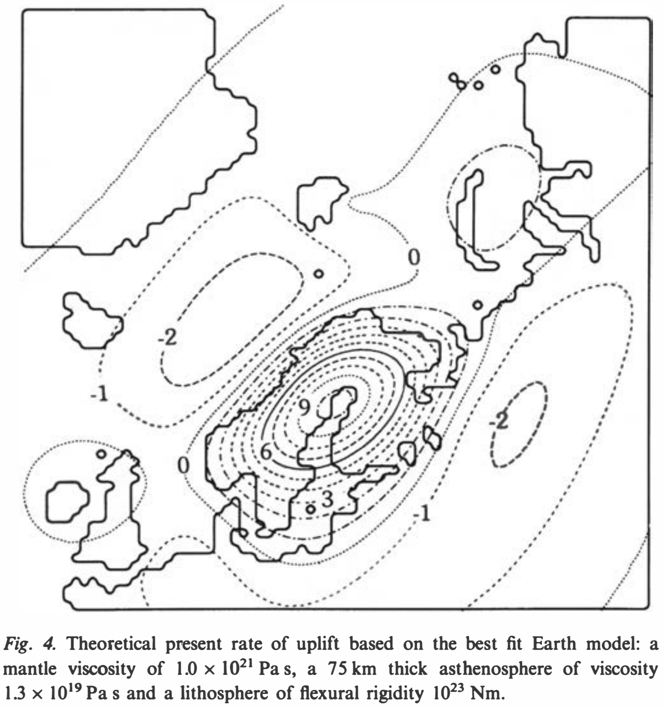


Fig. 4. Theoretical present rate of uplift based on the best fit Earth model: a mantle viscosity of 1.0×10^{21} Pa s, a 75 km thick asthenosphere of viscosity 1.3×10^{19} Pa s and a lithosphere of flexural rigidity 10^{23} Nm.

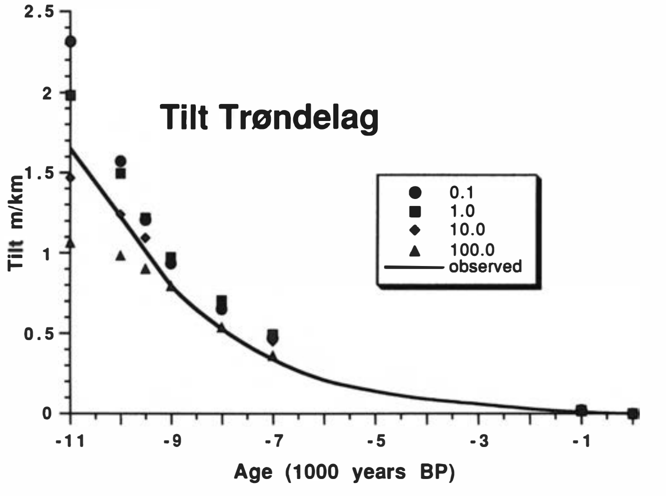


Fig. 5. Observed (from Kjemperud 1986) and theoretical shoreline tilting for Trøndelag. The theoretical shoreline tilting is calculated for a mantle viscosity of 1.0×10^{21} Pa s, a 75 km thick asthenosphere of viscosity 1.3×10^{19} Pa s and a flexural rigidity of 0.1, 1.0, 10.0 and 100.0×10^{23} Nm, respectively.

Forebulge

The theoretical uplift response for late- and post-glacial time based on this mantle viscosity profile and a lithosphere rigidity of 10^{23} Nm is shown in Fig. 6a–g. From 15,000 BP to the present the central area is uplifted by more than 650 m. The forebulge of that time was at the maximum 60 m above equilibrium position, located approximately 100 km from the maximum ice margin. It is often assumed that the forebulge will migrate inland during the deglaciation, but this is not the case for models incorporating a low-viscosity asthenosphere, as also pointed out by Cathles (1980). With such a model the zero uplift isoline in Fennoscandia is close to the maximum ice margin and will be relatively stationary with time, especially pronounced after 11,000 BP. This is also illustrated in Fig. 7a, which shows the uplift on a profile offshore from the Trøndelag area. The forebulge collapses smoothly while the forebulge maximum slowly moves further from the former ice margin.

The deglaciation and corresponding isostatic uplift cause a fall of the geoid from 15,000 to 9000 BP (when the ice was almost gone) of approximately 10 m over the profile, followed by a geoid rise up to the present level (Fig. 7b). The corresponding shoreline displacement, being the sum of the isostatic and eustatic change, is shown in Fig. 7c. The eustatic change is the sum of the glacial- and tectono-eustasy (approximated by the published eustatic curve of Shepard 1963) and the geoidal change (Fig. 7b). The minimum sea level at 15,000 BP is modelled to be 125 m below the present sea level and expected to be located 100 km from the former ice front. In a zone of distance 25 to 150 km from the former ice front the sea level is 115 m or more below present sea level. The modelled minimum sea level is somewhat lower than what is observed south of the modelled region, in the Møre area (Rokoengen et al. 1980), where the minimum

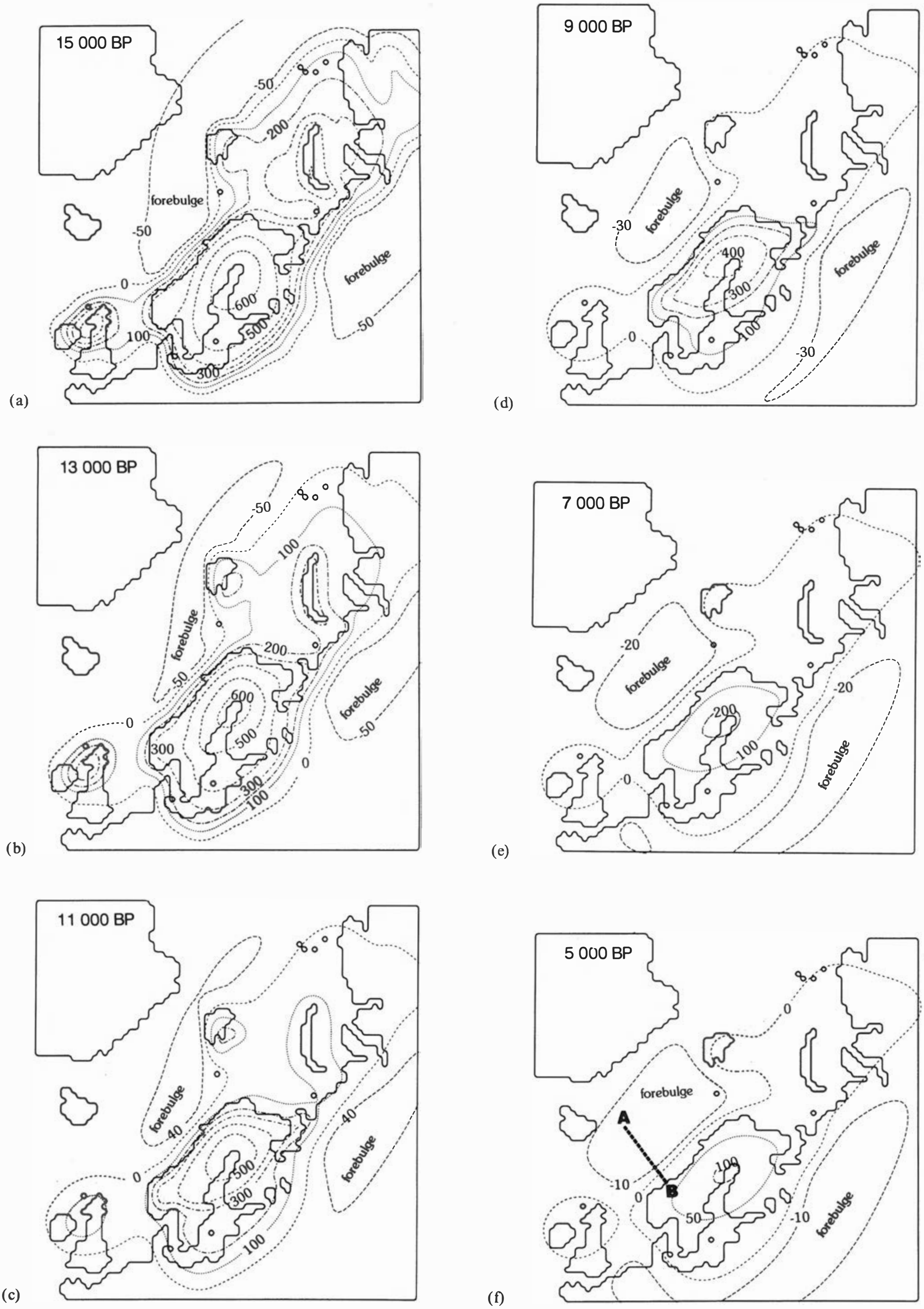
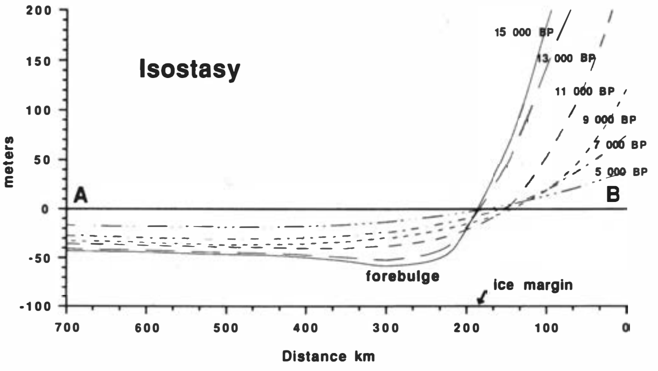
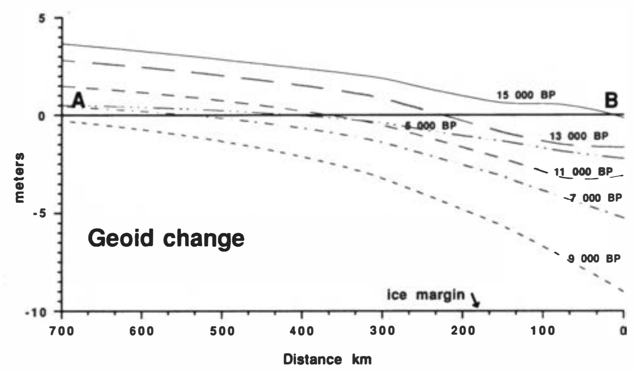


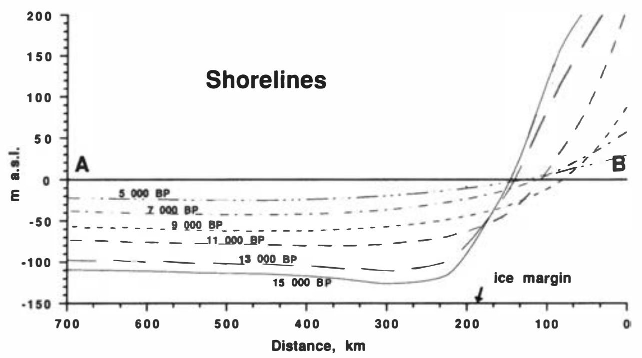
Fig. 6. Theoretical uplift response in metres for glacial and post-glacial time. (a) 15,000 BP; (b) 13,000 BP; (c) 11,000 BP; (d) 9000 BP; (e) 7000 BP; (f) 5000 BP.



(a)



(b)



(c)

Fig. 7. Isostatic (a), geoidal-eustatic (b) and shore-level displacement (c) along a profile offshore from the Trøndelag area (for location of the profile, see Fig. 6f).

sea level is reported to be at least 150 m below the present sea level. It is also interesting to note that the model predicts that the 9000 BP shoreline is lower than the younger shorelines in a zone close to the coast of Trøndelag (Tapes transgression). This is in accordance with the observations in the Frøya area (Kjemperud 1986).

Discussion

The minimum sea level at 15,000 BP is modelled to be 125 m below the present sea level, 100 km from the maximum ice margin. This is close to (but somewhat less than) the observed interpreted submerged beaches offshore Norway (Rokoengen et al. 1980), but probably too far from the coast. Decreasing the flexural rigidity will, however, give increased forebulge, and a minimum sea level at 15,000 BP, which is lower and closer to the coast. The palaeo-shoreline gradients for Trøndelag (cf. Fig. 5) indicate that the flexural rigidity is between 10^{23} Nm and 10^{24} Nm. The observed submerged beaches offshore Norway seem to indicate that the flexural rigidity is even lower, and may be due to decreasing flexural rigidity towards the west. However, there are several parameters in the modelling that can be questioned.

One of the uncertain parameters is the eustatic change. The modelled minimum sea level of 125 m is based on a eustatic (glacial- and tectono-eustasy) change of 70 m (a conservative extrapolation of Shepard's curve, 1963). However, by using the eustatic curve of Fairbanks (1989) the modelled minimum sea level would have been at least 40 m more.

One of the most uncertain parameters is the ice thicknesses. The ice model used here is probably a maximum model. If the ice thicknesses are decreased by 25%, the asthenosphere viscosity will have to be adjusted (to 1.6×10^{19} Pa s) for the uplift rate to match the observed maximum present rate of uplift of 9 mm/yrs in the central Baltic Sea. The modelled palao-shoreline gradients will be lower than the ones given in Fig. 5 using the same flexural rigidity. The resulting modelled minimum sea level at 15,000 BP would thus be somewhat higher than for the maximum ice model. This could be compensated for by reducing the flexural rigidity. It is thus reasonable to conclude that the amplitude and decay pattern of the forebulge will be very similar to what is reported here if the ice model differs by less than 25% from the real ice thicknesses.

Data from peripheral areas are important for better constraining the model parameters, particularly the flexural rigidity of the lithosphere. It is thus an important future research task to combine onshore sea-level data with offshore data, to be able to more precisely determine the size and nature of the forebulge offshore Norway. This could also give increased knowledge of the glacial history.

Conclusion

Comparison of theoretical models with the observed tilting of palaeo-shorelines and the pattern of present uplift indicates that the elastic lithosphere is less than 50 km thick, the mantle viscosity is 1.0×10^{21} Pa s and the asthenosphere has a viscosity 1.3×10^{19} Pa s. This model predicts (1) a maximum forebulge of 60 m at 15,000 BP that decays to 40 m at 11,000 BP, and (2) that the zero uplift isoline in Fennoscandia is relatively stationary over time and that the forebulge collapses smoothly without any migration. The minimum sea level at 15,000 BP is modelled to be 125 m below the present sea level and located 100 km from the former ice front.

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