On the magnitude of the Late Tertiary and Quaternary erosion and its significance for the uplift of Scandinavia and the Barents Sea

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Geological evidence related to the Late Tertiary uplift and erosion of the Barents Sea and Scandinavia includes marine seismic data, well data and field observations on land. The amount of erosion has been studied using geochemical and sedimentological data from wells and topographical data and data on the block fields onshore. The amount of erosion has been compared to the amount of deposition in the huge Pliocene-Pleistocene wedges situated offshore Norway and in the Barents Sea.

The data suggest that there is a close relation between the uplift and erosion of Scandinavia and the Barents Sea, and that the uplift is related to two main events: (1) mountain building in Central South Norway and along the trend Lofoten-Bjørnøya-Svalbard due to Palaeocene to Oligocene tectonic phases; and (2) glaciation in the Late Pliocene and Pleistocene, which was related to plateau uplift of large areas.

Based on lithospheric flexural rheology parameters from studies of the post-glacial uplift, modelling shows that the compensating isostatic uplift is close to 70% of the magnitude of erosion where crystalline basement is eroded. The magnitude of the observed uplift in Scandinavia seems, however, to be greater than the erosion in some areas, and therefore mechanisms other than pure isostatic compensation must be sought to explain this. One part of the uplift is thought to date back to Paleogene tectonism, but observations suggest that there is an uplift in excess of these two components. We have investigated the possibility that migration of phase boundaries in the lithosphere may be the mechanism of this additional uplift.

The response of a phase transition in the lithosphere to pressure relief caused by erosion leads to thickening of the upper lighter phase and thus greater uplift than expected from isostasy alone. Migration of the phase boundary is a time-dependent process. Theoretical models show that for load cycles of the order of 1 Ma, the surface deflections associated with the movement of the phase boundary can be large, up to 50% of the isostasy. Therefore, modelling of the phase migration combined with isostasy in response to erosion might lead to an uplift larger than the amount of eroded rocks. Thus, theoretically the phase migration model is of interest, although it is not supported by independent geological evidence.

Introduction

The late phase of erosion in the Barents Sea significantly affected the petroleum accumulated in traps (e.g., Nyland et al., this volume). Also, in large parts of the platform areas, Triassic and Jurassic source rocks situated in the oil window have been uplifted and cooled. Therefore, much effort has been put into the task of quantifying the uplift and erosion in the Barents Sea.

This study was initiated as a consequence of new dating results of the Late Cenozoic sedimentary wedge penetrated by two wells on the Senja Ridge (Fig. 1). Our results indicate the wedge to be of the Late Pliocene and Pleistocene age (Eidvin and Riis, 1989). The size of the wedge corresponds to an erosion of 500–1000 m in the Southern Barents Sea (Vorren et al., 1988; Vorren et al., 1990; Nøttvedt et al., 1988).

A similar wedge is situated on the shelf offshore Mid-Norway (Fig. 2), although it contains a smaller volume of sediments. This wedge has been penetrated by several wells on Haltenbanken, and its age has been determined to be Late Pliocene and Pleistocene. Thus, this age correlates with the wedge outside Bjørnøyrenna.

Results from ODP wells on the Vøring Plateau (Jansen et al., 1988) indicate that the first major glaciations took place at approximately 2.5 Ma. This age correlates well with the age dating of the wedges. Examination of the coarse fraction of the cutting material has yielded large amounts of angular fragments of crystalline basement in the wedges (Eidvin and Riis, 1989).

In both areas, the Miocene and Early Pliocene deposits appear to be thin, on average in the order of a few hundred metres at most. Thus, geological evidence indicates that in the Late Pliocene and

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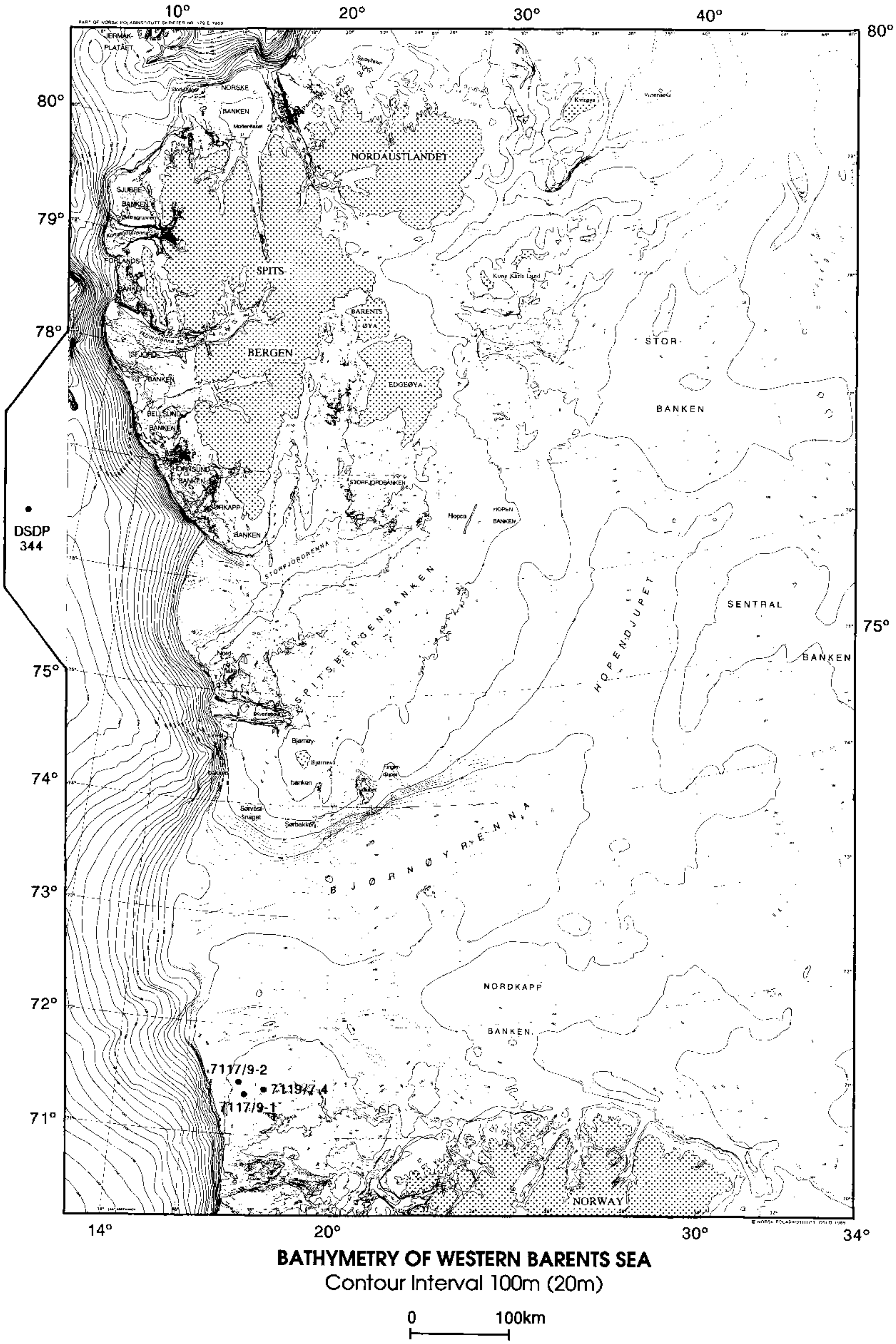

Fig. 1. Location map of Barents Sea, showing key wells and bathymetry. Map compiled by Y. Kristoffersen, B. Beskow, M. Sand and Y. Ohta, Norsk Polarinstitutt (1989).

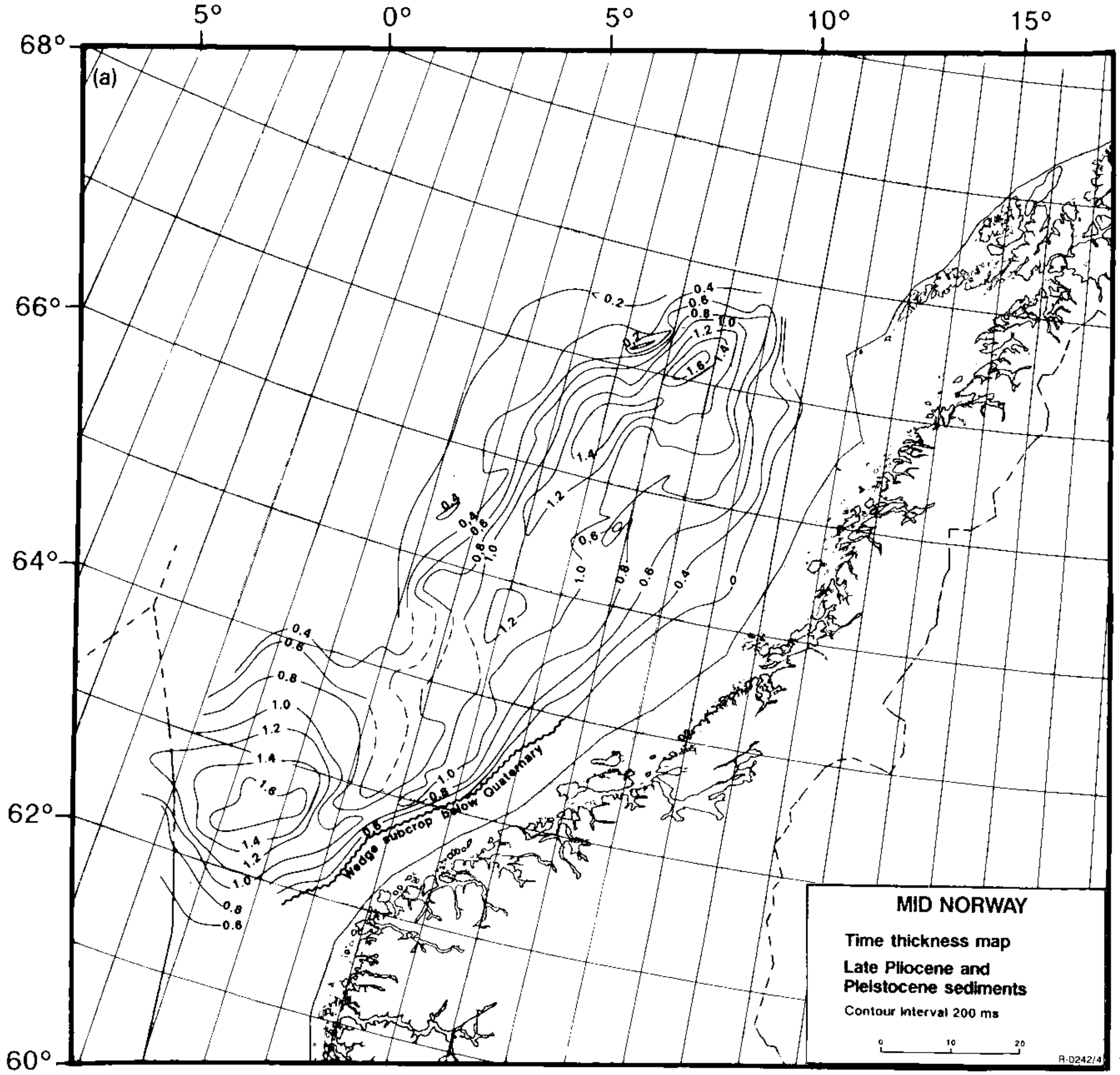

Fig. 2. (a) Time thickness map of Late Pliocene-Pleistocene wedges, Mid-Norway shelf.

Pleistocene period a dramatic increase in deposition rates along the shelf was related to glaciation of the Barents Sea and of Scandinavia. The late phase of erosion in the Barents Sea must, therefore, be seen in a larger context. The observation that large amounts of erosion were related to glaciations led us to investigate the thermal and isostatic effects of the rapid loading and unloading resulting from the erosion process. The study is based on simplified maps showing estimated amounts of erosion and deposition in Scandinavia and in the Barents Sea in glacial times (the last 2.5 million years).

The aim of this study is to determine whether the observed uplift pattern can be modelled using simple assumptions about the effect of erosion and glaciation, and whether the interpreted uplift related to erosion and glaciation can be separated from uplift due to older tectonic phases related to the opening of the North Atlantic.

Late Pliocene-Pleistocene geology

Erosional and depositional pattern

Occurrence and dating of sedimentary wedges

Along the Barents Sea margin, large sedimentary wedges occur at the mouth of bathymetric troughs (Fig. 1). The wedge fed by Bjørnøyrenna is more than 4 km thick (Fig. 3), and the wedge off Storfjordrenna reaches the same thickness, although its areal extent

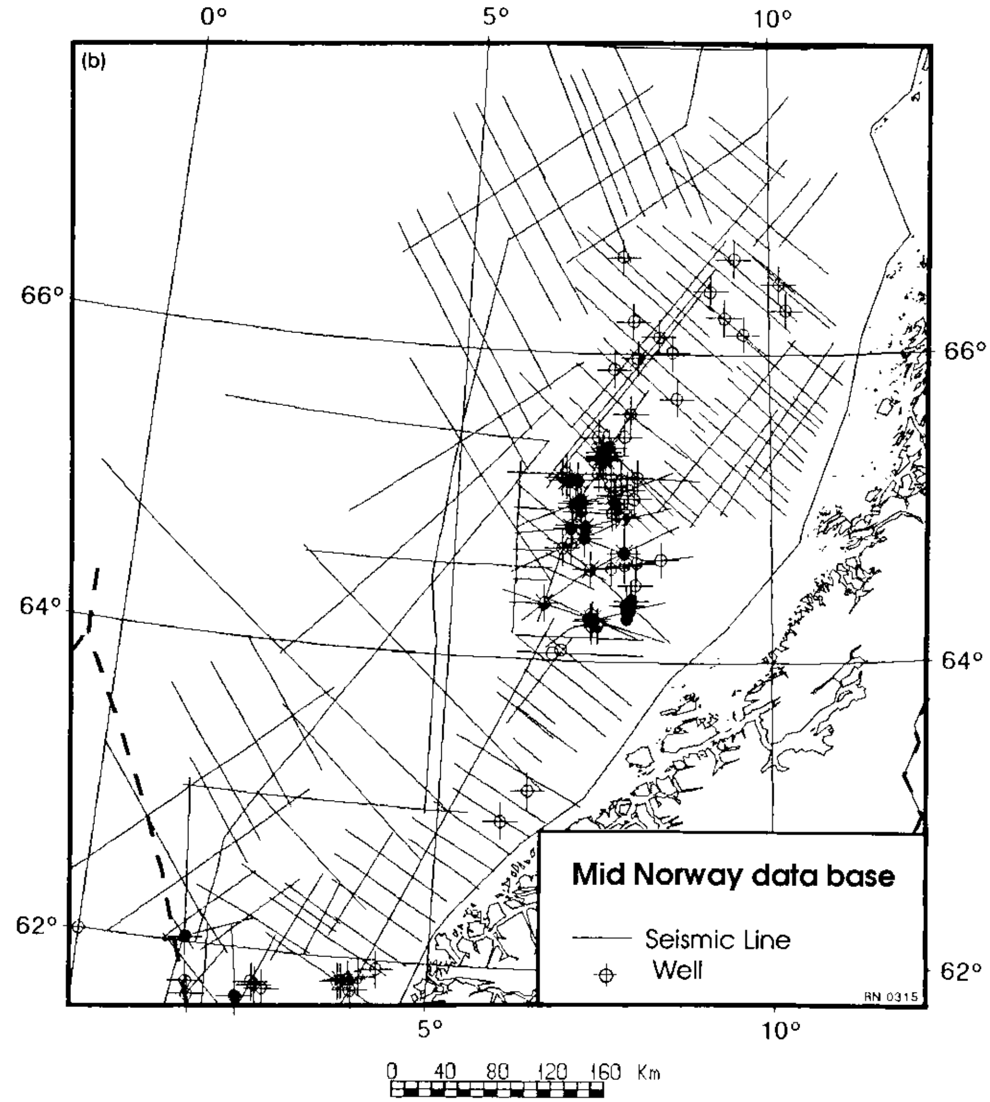

Fig. 2 (continued). Maps (b) and (c) show location of wells and seismic lines and bathymetry.

is smaller. These two wedges are associated with large positive Bouguer gravity anomalies. Fjeldskaar and Riis (1988) relate parts of these anomalies to a lack of isostatic compensation due to rapid recent deposition.

Along the northern Barents Sea margin, a similar pattern appears. Sparse seismic data collected by NPD and by Baturin (1988) and regional bathymetry (Perry and Fleming, 1985) indicate that sedimentary wedges occur in a similar pattern, although probably with smaller sediment volumes.

The wedge off Bjørnøyrenna (Fig. 1) was sampled by the two wells, 7117/9-1 and 2, which penetrate approx. 800 m Upper Pliocene and Pleistocene (Eidvin and Riis, 1989). The dating of this sequence has been much debated due to resedimentation (e.g., Spencer et al., 1984; Nøttvedt et al., 1988). However, the micropalaeontology, Sr isotope analysis and study of the basement fragments all strongly indicate that the depositional age of the sequence is Late Pliocene and Pleistocene.

The wedge off Storfjordrenna was penetrated by one DSDP well in a distal position close to the Knipovich Ridge (Fig. 1). The section was dated as Pliocene (Talwani et al., 1976). The uppermost, mainly Quaternary part of the wedge can be correlated seismically to the wedge off Bjørnøyrenna.

In the Mid-Norwegian shelf, there are two main wedges which are also linked to major bathymetric troughs — the wedge at the mouth of Norskerenna and the "Mid-Norway wedge" at the mouth of troughs extending from Vestfjorden and Ranafjorden (Figs. 2 and 4). These two wedges have a maximum thickness in the order of 2000 m. Both have been penetrated by a number of wells (Fig. 2), and they may be assigned a Late Pliocene–Pleistocene age (Rokoengen and Rønningsland, 1983; Dalland et al., 1988) (Naust Formation).

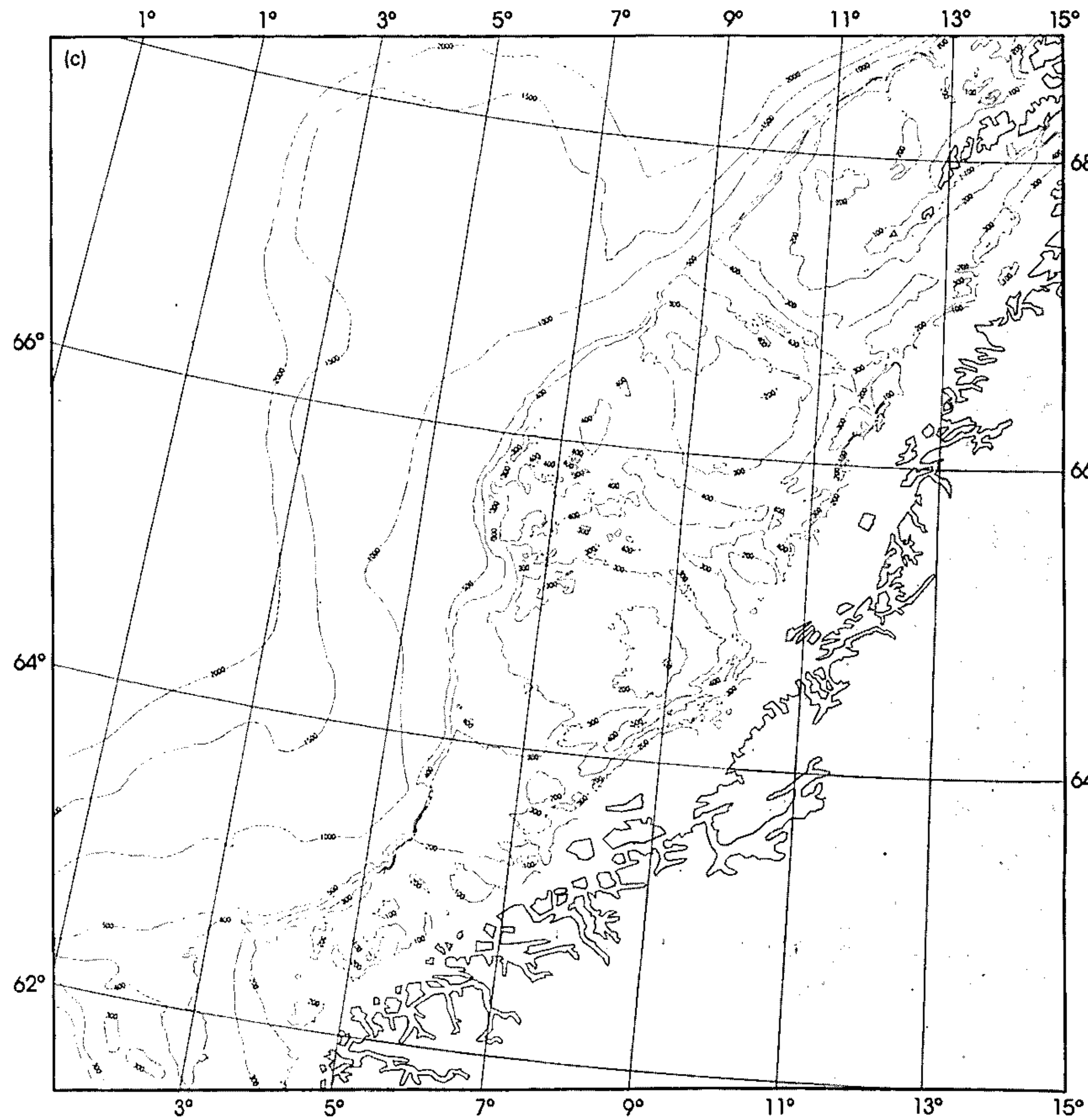

Fig. 2 (continued).

Estimates of erosion

In the Barents Sea and in Scandinavia, the amount of erosion has been estimated by different methods. For the Southern Barents Sea and for Mid-Norway, it has been assumed that all erosion products which were transported to the west have been deposited in the respective offshore wedges. This mass balance gives an independent control on the erosion map, although the drainage divides cannot be established with certainty. We have concentrated on applying the method of mass balance on Mid-Norway-North Sweden, because the uncertainty of the size of the drainage area is not too great. For the southern Barents Sea and Finnmark, the drainage area is less well known, and for South Norway-South Sweden, the area of deposition is not easily mappable. Furthermore, there is no major fault activity along the coast of Nordland post Triassic times. This means that the geometry of the uplift should be simple. However, Lofoten was an active structural element in the Late Cretaceous and Paleogene.

(1) Barents Sea. The magnitude of erosion in the Barents Sea was estimated by geochemical data (vitrinite reflectance and T_{max}) from the deep wells (Fig. 5). In the western parts, the map is based on the occurrence of a seismic reflection which cuts across the seismic stratification in the Late

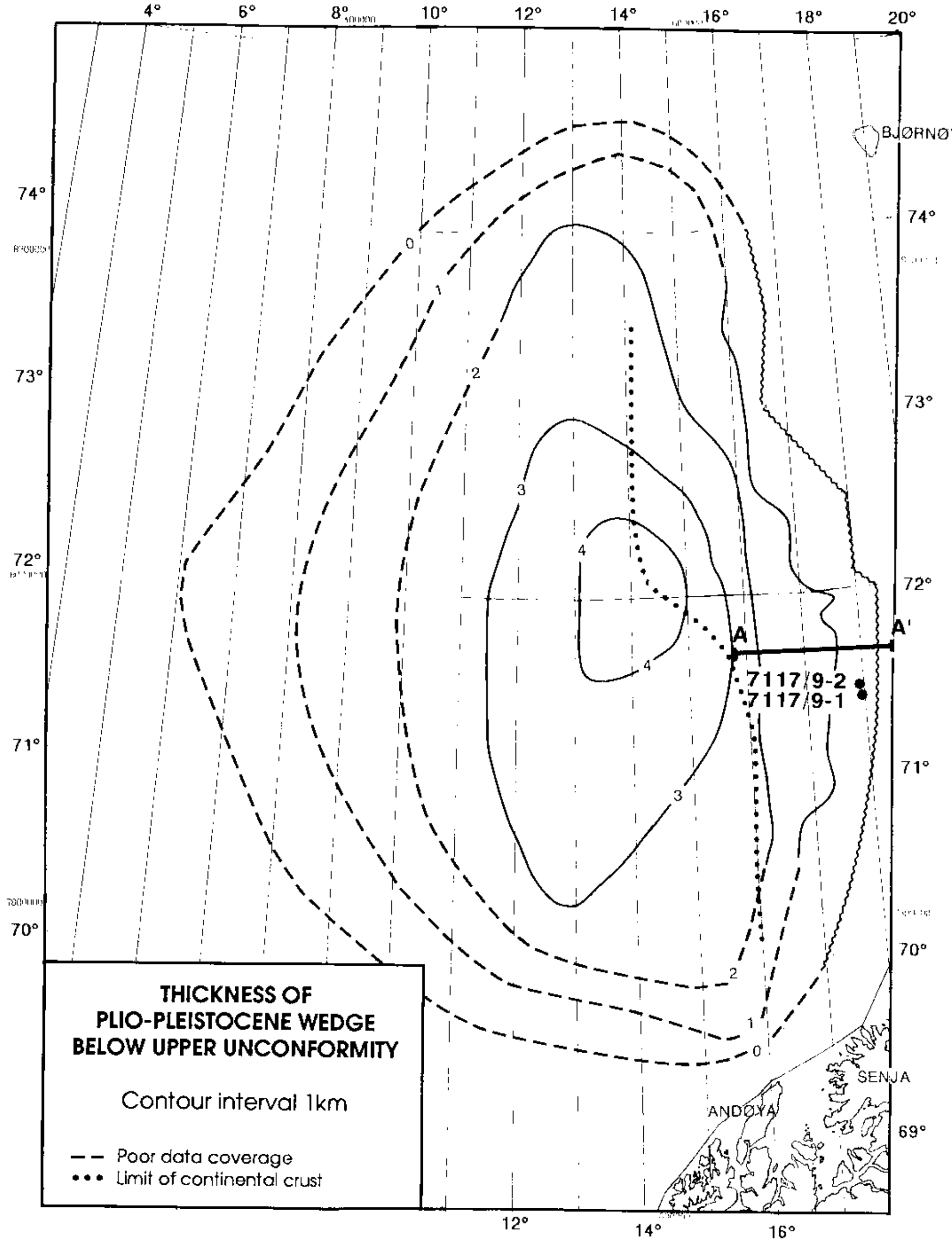

Fig. 3. Thickness map of Late Pliocene and Pleistocene sediments below the Upper Regional Unconformity, southwestern Barents Sea.

Paleocene and Early Eocene layers. This reflection shows a regional dip to the west (Fig. 6), and is correlated to the transition from Opal A to microcrystalline quartz (opal CT) observed in well 7117/9-1 (Ramberg Moe et al., 1988). This transition takes place at approx. 50°C, and the dip of the reflector indicating the diagenetic change may be explained by tilting and erosion after its formation. The deepest level observed for the reflector was assumed to be the level of no erosion, and the erosion in the rest of the area was estimated by the difference in elevation, assuming no lateral change in temperature gradient.

In the eastern and northern parts of the Barents Sea, there are very few data points (Antonsen et al., 1991). Seismic stacking velocities (Norwegian Petroleum Directorate, unpublished work) and geological data from Svalbard (Manum and Throndsen, 1978) suggest a general decrease in erosion from west to east in this area.

It is well known that Spitsbergen was uplifted and eroded as a consequence of Eocene-Oligocene tectonism (e.g., Steel and Worsley, 1984). Also, the Bjørnøya-Stappen High area was uplifted along the Knølegga Fault in different tectonic phases in the

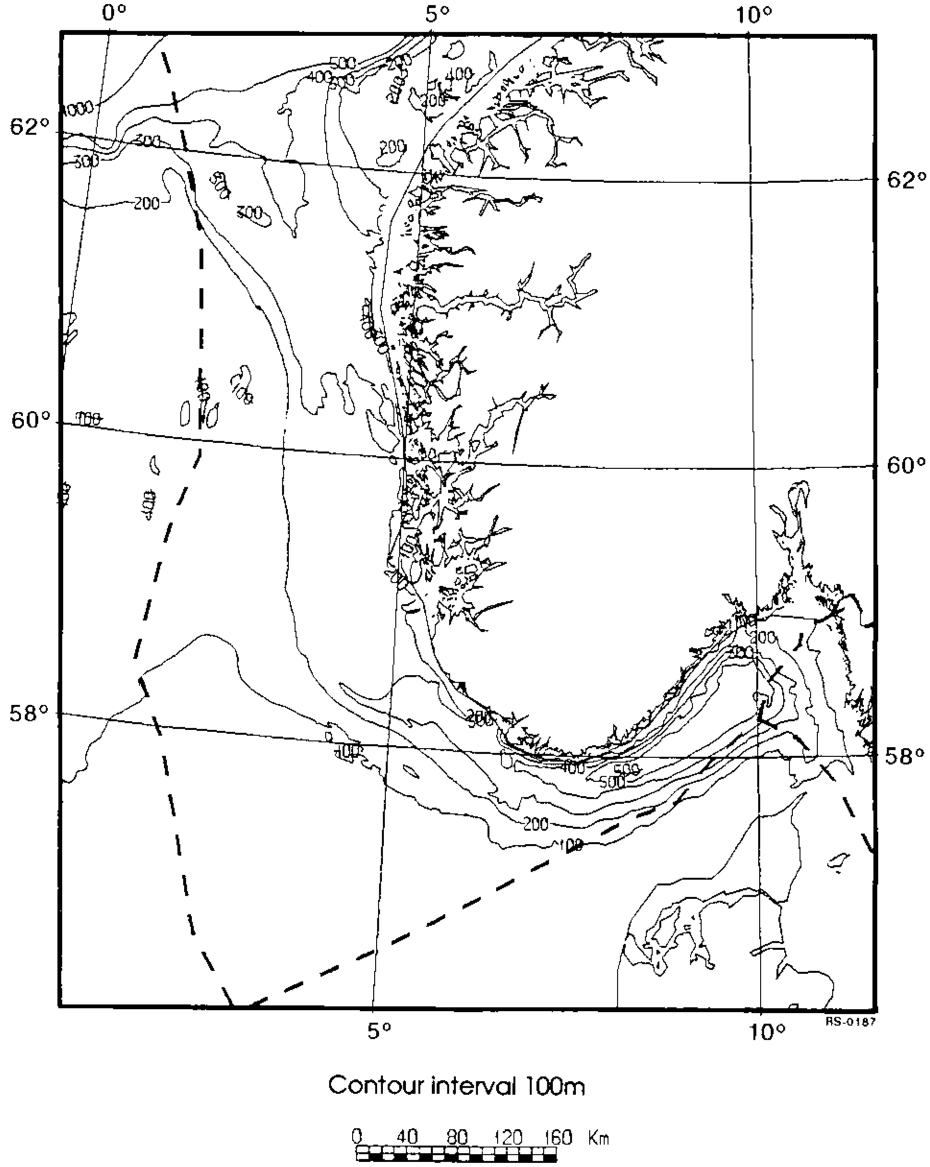

Fig. 4. Bathymetry of the North Sea. Based on data from Statens Kartverk.

Eocene and Oligocene (Fig. 7). This happened also to the Senja Ridge and Lofoten islands farther south (Brekke and Riis, 1987). As a consequence, the large amount of erosion observed in the western and northern parts of the Barents Sea can be related partly to tectonic uplift in the Paleogene and partly to the glacial effects we are concerned with in this study.

Farther east and south in the Barents Sea, the Early Tertiary tectonic effects can be expected to be more subdued. Thus, an estimate of 800–1000 m of Late Pliocene-Pleistocene erosion based on the sediment volume in the wedge off Bjørnøyrenna seems reasonable for the central parts of the Southern Barents Sea.

It should be pointed out that these values are surprisingly high compared to studies of present day erosion rates at Svalbard, estimated at 0.2–0.5 mm/year (Løvø et al., in preparation), and at 0.5–1 mm/year for active glaciers (Elverhøi et al., 1980, 1983). 1000 m of erosion is equivalent to 0.4 mm/year, suggesting that the Barents Sea was not submerged for much of the glaciation period.

(2) Mid Norway. The calculated erosion is based on the assumption that a pre-glacial surface can be mapped onshore in Scandinavia. If the present height and topography of the pre-glacial surface can be established, the amount of erosion can be calculated by subtracting the present mean heights from the height of the pre-glacial level.

Block fields are observed on many of the high mountains in South Norway. Nesje et al. (1988) suggest that the block fields were formed by weathering processes prior to glaciation. It is assumed that the lowest limit of the block fields approximates a pre-

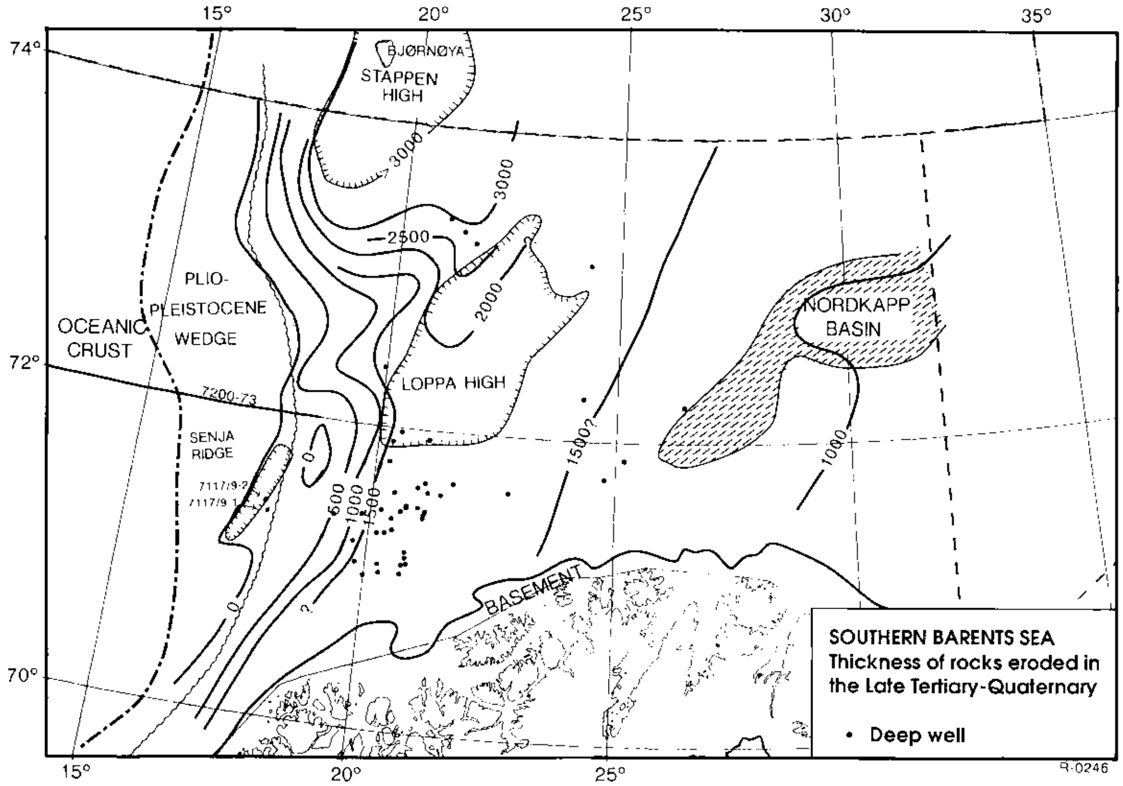

Fig. 5. Map of southern Barents Sea showing interpreted amount of Tertiary erosion (in metres). This interpretation is based on vitrinite and pyrolysis data from deep wells as well as on Figs. 6 and 12, profiles EE' and FF'.

glacial surface. In Møre, these pre-glacial surfaces may be traced almost to sea level.

Fig. 8 shows a simplified map of mean heights contoured from an 8 × 8 km grid prepared by the Geographical Survey of Norway and the Geographical Survey of Sweden (1963). The summit level map (Fig. 9) is based on smoothed heights of the block fields where they exist, combined with summit levels in areas where there are few data on block fields or where block fields do not occur. This map may be regarded as an approximation of a pre-glacial surface, but as it does not take into account topography in this surface, it could be regarded as a pre-glacial surface of maximum probable elevation.

A. Nesje (University of Bergen, unpublished work) has compiled the paleic surface in Norway (Fig. 10). The interpretation is based on geomorphological field work in Southern and Mid-Norway, and mainly on map studies in Northern Norway. This map may be interpreted as an approximation of the present day level of the pre-glacial topography, and it must be regarded as a better approximation than the summit level map.

Onshore erosion (Fig. 11) was calculated by subtracting the mean height map (Fig. 8) from the summit level map (Fig. 9), and therefore it must be regarded as a maximum case of eroded rock in glacial times. In the offshore areas, erosion has been calculated by extrapolating the eroded part of the sedimentary sequence to the shore. The pre-glacial pinch out line of sedimentary rocks on the crystalline basement was inferred by using the profiles AA′ to FF′ (Fig. 8), and is shown in Fig. 11. The pinch out line was drawn so that the block field regions were not given a pre-glacial sediment cover.

A comparison was made between the deposited volume of sediments in the wedge off Mid-Norway and the eroded volume in Mid-Norway suggested in Fig. 11, using tentative drainage divides as shown on Fig. 9. The result is given in Table 1. The eroded thicknesses suggested in Fig. 11 are in the order of 1.5–2 times too large to fit into the wedge. This sup-

TABLE 1

Erosion and accumulation Mid-Norway wedge

	Area (km^2)	Volume (km^3)
Glacial wedge (Fig. 2)	139000	100000
Eroded sedimentary rocks (Fig. 11)	88600	64900
Eroded basement (Fig. 11)	108000	89300
Corrected basement (2.7/2.2)	108000	109000

Total erosion: 173900 km^3
Total deposition: 100000 km^3

The assumptions are seismic velocity of 2200 m/s and density of basement/density of sediment = 2.7/2.2.

ports the idea that the paleic surface map (Fig. 10) may be a better approximation of the pre-glacial surface than the summit level. If this is the case, the topography before glaciation may have been quite rough even in Nordland, with local relief in the order of 500–700 m. In this case, a "summit level" surface must be interpreted as older than Neogene. Alternative ways of explaining the volume difference would be to change the drainage divides and to reduce the amount of eroded sedimentary rocks.

Our present knowledge of Tertiary tectonic events and thickness of the Oligocene to Early Pliocene section on the shelf, suggests fairly low deposition rates and, thus, a fairly low relief at the onset of glaciation. It seems that more detailed studies are needed to resolve this problem.

In conclusion, Fig. 11 is a maximum case for the amount of material eroded in the glacial period, and the volume could be reduced by 30–40%. Even so, the amount of erosion in the time of glaciation is higher than was expected.

Structural pattern

Six profiles were constructed across Scandinavia in different structural positions (Fig. 12). Profiles AA′ and DD′ describe the situation with a fairly large uplift onshore, no obvious tectonic activity along the margin and, for profile DD′, a thick wedge of Plio-Pleistocene sediments deposited on the shelf. Note that for those profiles the summit level can be projected into the sedimentary sequence, suggesting a pinch out of sediments in a position presently onshore before erosion and uplift took place.

Profile BB′ shows a large uplift onshore, tectonism offshore, and a small Plio-Pleistocene wedge. Onshore, the paleic surface can be traced almost to sea level, fitting the position of the base of the wedge offshore. This supports a pre-glacial, but Tertiary age of the paleic surface.

Profile CC′ traverses the anomalously low Trondheim area, where Jurassic sediments are known in a fault-bounded basin in Beitstadfjorden (Bøe, 1989). Here the summit level projects into the basement-sediment boundary offshore. Accordingly, the sedimentary cover in this profile has been extended farther onshore, to give the onshore dome a shape similar to the other sections. The sediments removed were included in the calculation of erosion (Fig. 11).

Profile EE″ and profile FF′ have been constructed by adding a sedimentary cover in the Barents Sea according to Fig. 5. The profiles suggest that the summit level inferred from Troms and Finnmark correlates fairly well with the erosion interpreted in the Barents Sea, given that there was not a marked topographic break between these areas prior to the onset of major erosion.

The summit level map (Fig. 9) and the cross-sections AA′ to FF′, combined with a regional overview (Fig. 13), show that there are some common structural features related to the uplifted Scandinavia and Barents Sea.

– The sequence of sedimentary rocks surrounds the coast from Skagerrak to the White Sea with a fairly uniform dip away from the mainland. The boundary between sediments and crystalline basement is at a fairly regular distance from the coast line and commonly coincides with a bathymetric low.

– The base of Late Pliocene (glacial) sediments combined with the surface of the summit level together form an asymmetric dome structure with the steep side facing offshore. The structuring of the base of the glacial sediments shows that the dome formation is recent. The dome has two maxima — in North Sweden and in South Norway — and in map view it turns around the coast (Figs. 9 and 13). The boundaries of the dome do not seem to be faulted to a large extent, although they coincide with the boundary faults along some segments of the coast.

– The maximum amount of erosion seems to occur in a zone along the coast. The extremal value is commonly located close to the boundary of sedimentary cover and crystalline basement.

The pattern of a structural dome in Scandinavia may be regarded as part of a geomorphological/structural pattern of concentric highs and lows, as pointed out by White (1972). The following observations may be of importance.

– The exposed crystalline shield is geographically linked to the areas of glaciation.

– A central shallow depression is found in Bottenvika (Fig. 13), (even when allowing for a residual glacial rebound of approx. 100 m, as proposed, e.g., by Balling, 1980).

– Marginal highs occur concentrically, including Scandinavia, Kola and the hills at the Finland-Soviet boundary. The outcrop of the Cretaceous in Denmark and Scania indicates that even in the south there has been a large amount of erosion and uplift.

– In the periphery outside the marginal highs we find shallow marginal morphological depressions. Offshore, they include the bathymetric lows along the coast of Norway and Kola, the White Sea and the Gulf of Finland. Onshore, they include the great lakes of Vänern, Vättern, Ladoga and Onega (Fig. 13). This zone of marginal lows is connected to the boundary between crystalline basement and sedimentary cover.

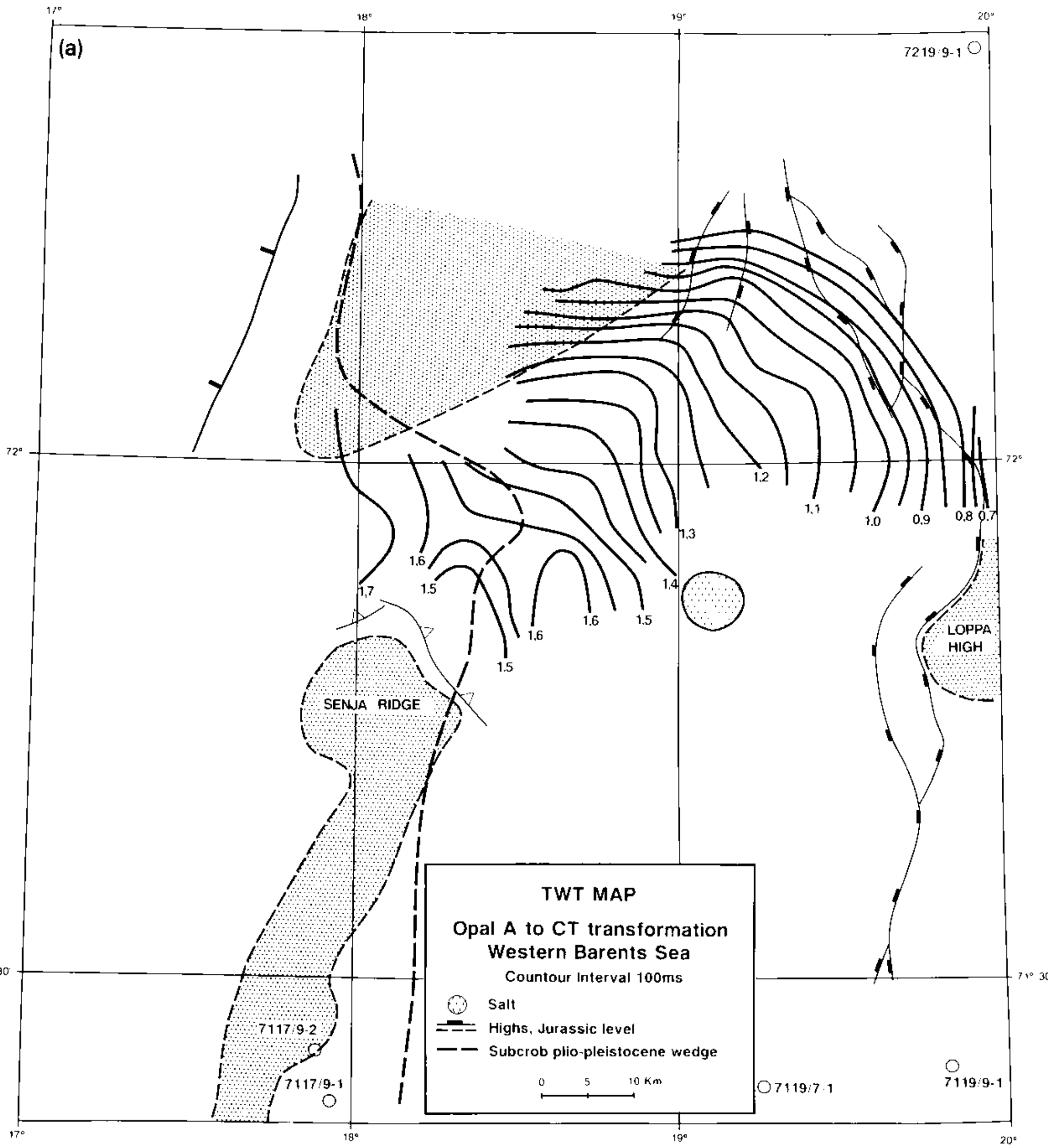

Fig. 6. (a) Map of two-way travel time to seismic anomaly representing the Opal A to CT transition in the Barents Sea.

– The sedimentary cover surrounding the glaciated area in general dips away from the exposed core of basement rocks.

White (1972) noted such a concentric geomorphic pattern in Fennoscandia and compared it with a similar pattern in North America. He connected this pattern to glaciation, but his explanation involved unrealistically high values of erosion (Sugden, 1976).

It may briefly be commented that a similar pattern of a central depression, marginal highs and huge sedimentary wedges occurs even in connection with the Greenland ice sheet (Weidick, 1976). The Barents Sea ice sheets were developed on thick deposits of sedimentary rocks and the present topography differs from the other areas by its low elevation 100–400 m below sea level. However, in this context, Svalbard, Bjørnøya, Franz Josef Land and Novaya Zemlya may be regarded as marginal highs.

It is concluded that the development of the concentric pattern of highs and lows is related to glaciation both in time and space, and hence we suggest that its development should be explained in terms of effects related to glaciation itself.

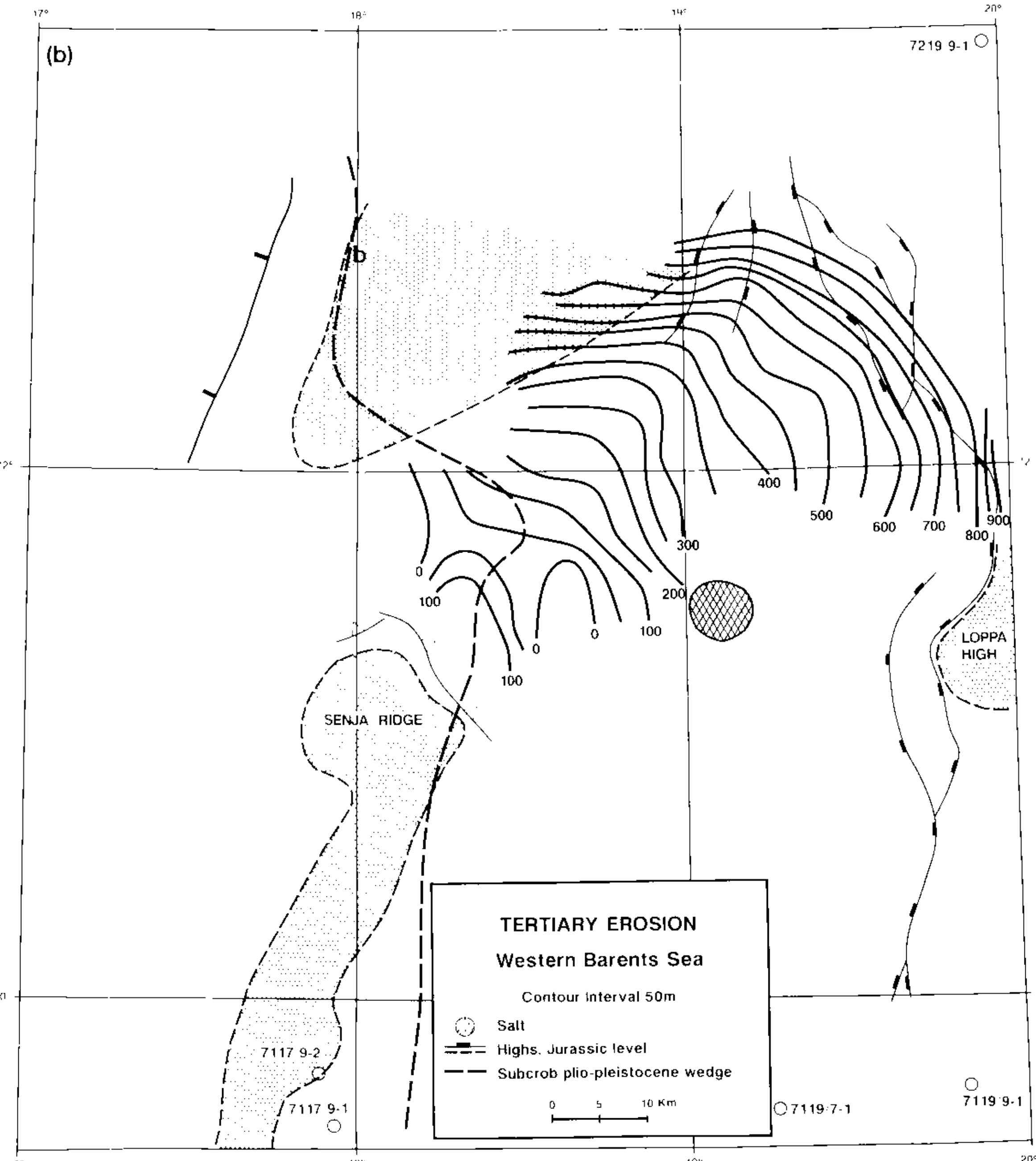

Fig 6. (continued). (b) A conversion of the map to show the eroded thickness of the sediment.

Modelling of crustal response

The removal of load in eroded areas and additional loading in the areas of deposition causes isostatic response of the Earth's lithosphere. The isostatic response is the most obvious mechanism for the uplift/subsidence in the study area. The magnitude of the observed uplift in some areas, however, seems to be greater than the erosion. The isostatic process alone cannot explain an uplift of this magnitude. An additional process is needed. Here we have looked at one possible candidate which may contribute to the uplift. This is the migration of phase boundaries in the lithosphere in response to pressure relief caused by erosion.

In addition, rifting in the North Atlantic causing uplift in Scandinavia (Sales, 1989) is regarded as a part of the story. The modelling presented in this paper is an attempt to calculate the magnitude of the effects expected from glacial erosion, so that these fundamentally different tectonic events can be quantified and separated.

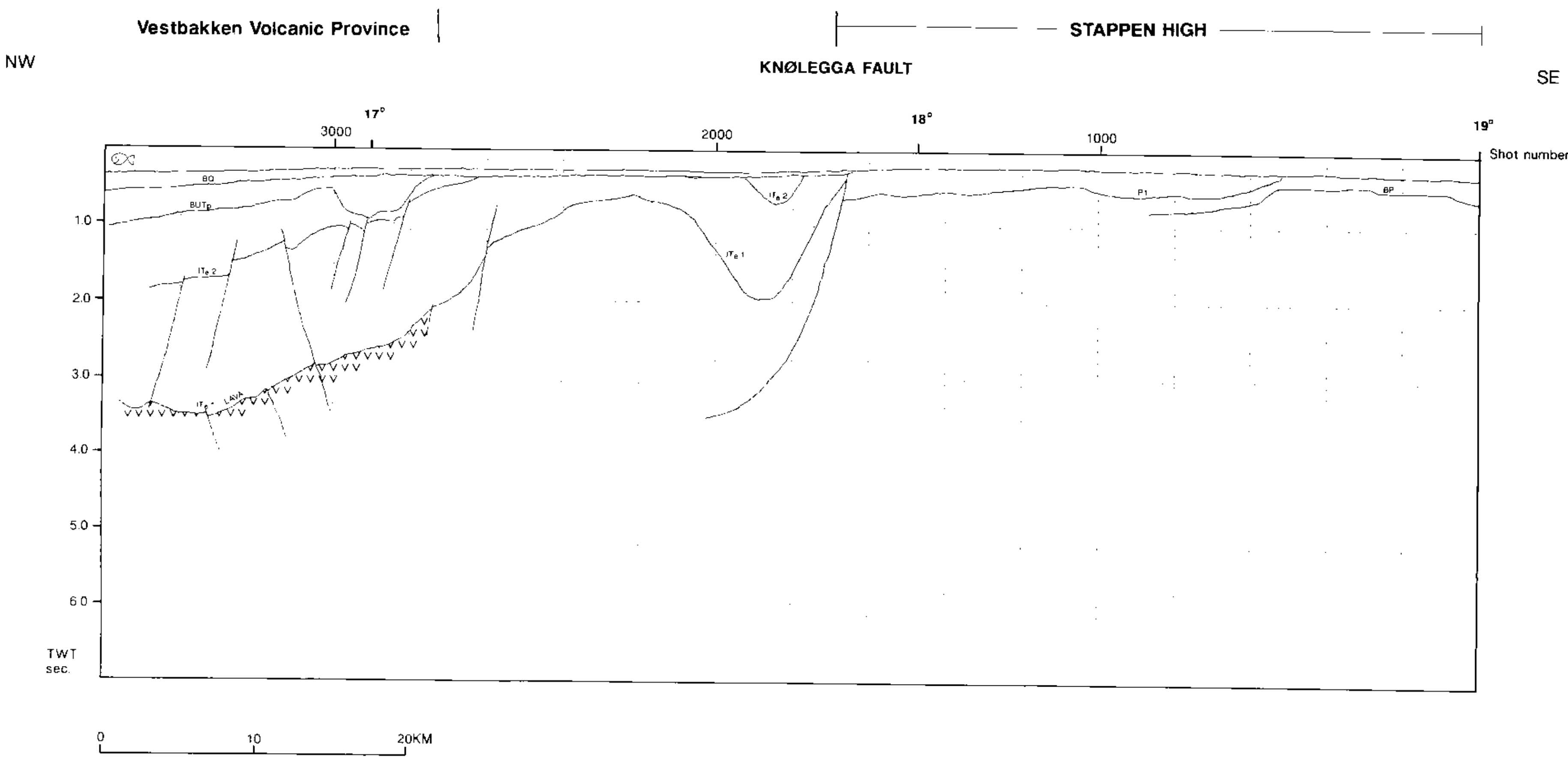

Fig. 7. Cross-section across the Knølegga Fault southwest of Bjørnøya indicating motion in the Eocene and post-Eocene. Line BV-12-86. *BQ* = base of Quaternary; *BUTp* = base of Upper Pliocene; ITe_1 and ITe_2 = intra Eocene; P_1 = Permian; *BP* = base of Permian. From Gabrielsen et al. (1990).

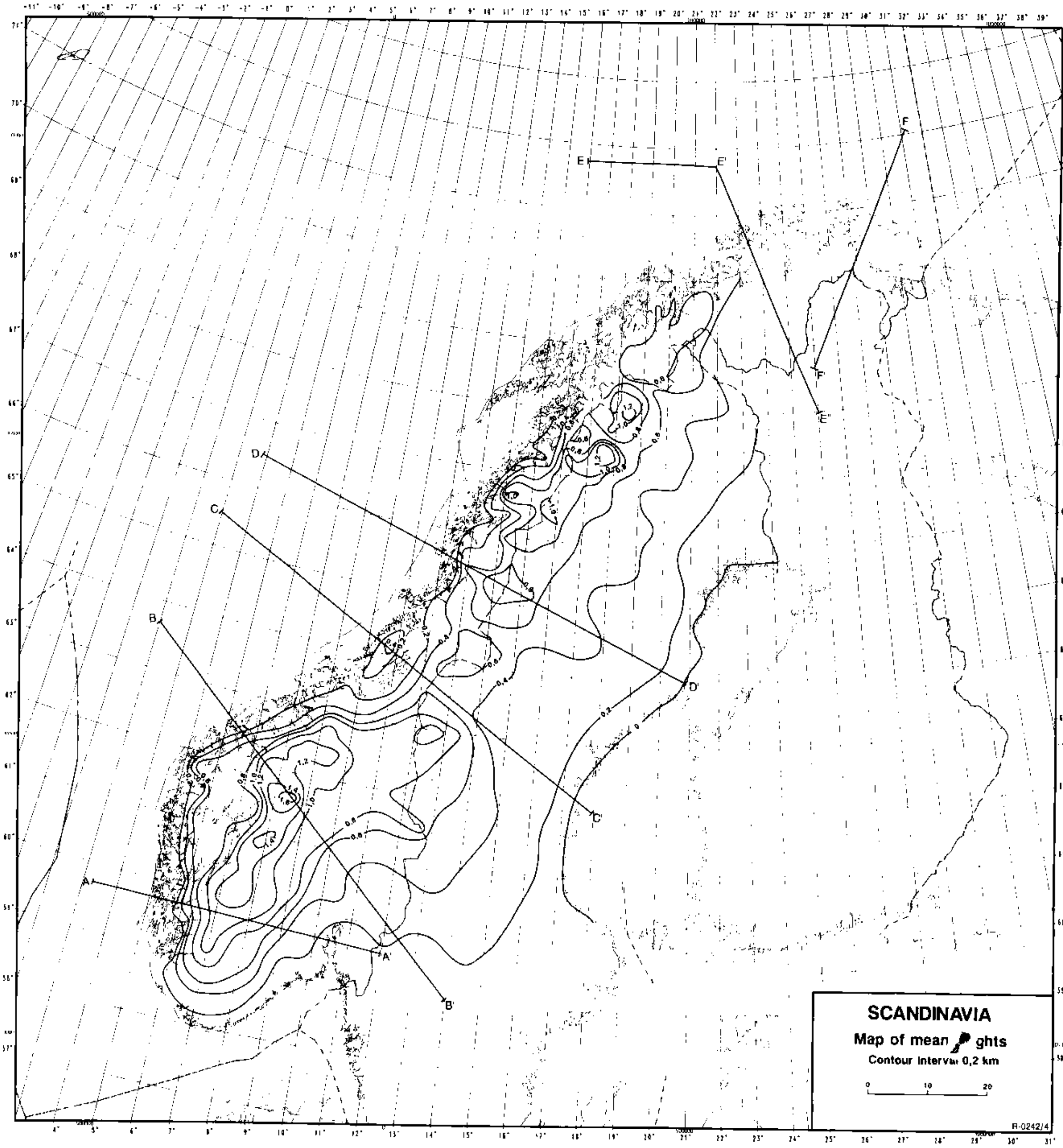

Fig. 8. Mean height map of Scandinavia, based on contouring of 8 x 8 km grid values.

Isostasy

The isostatic process is simulated by a non-spherical three-dimensional model, in which the lithosphere is a uniformly thick elastic layer overlying a viscous asthenosphere. With this model, we are able to treat the isostatic problem analytically using the Fourier transform technique. The mathematical model is described in Appendix A.

A (positive or negative) load applied to the Earth's surface will be partly balanced by the elastic lithosphere and partly by the buoyant forces of the asthenosphere. Loads of short wavelengths will be balanced by the lithosphere, so that the lithosphere acts as a low-pass filter. The characteristics of this filter depend on the flexural rigidity. In this study we used a value for the flexural rigidity of $D = 10^{23}$ Nm. A study based on data on the present rate of uplift and postglacial sea level changes has revealed a value of the flexural rigidity of Scandinavian mainland of less than 10^{24} Nm (Fjeldskaar and Cathles, 1991). The results of the isostatic response reported here

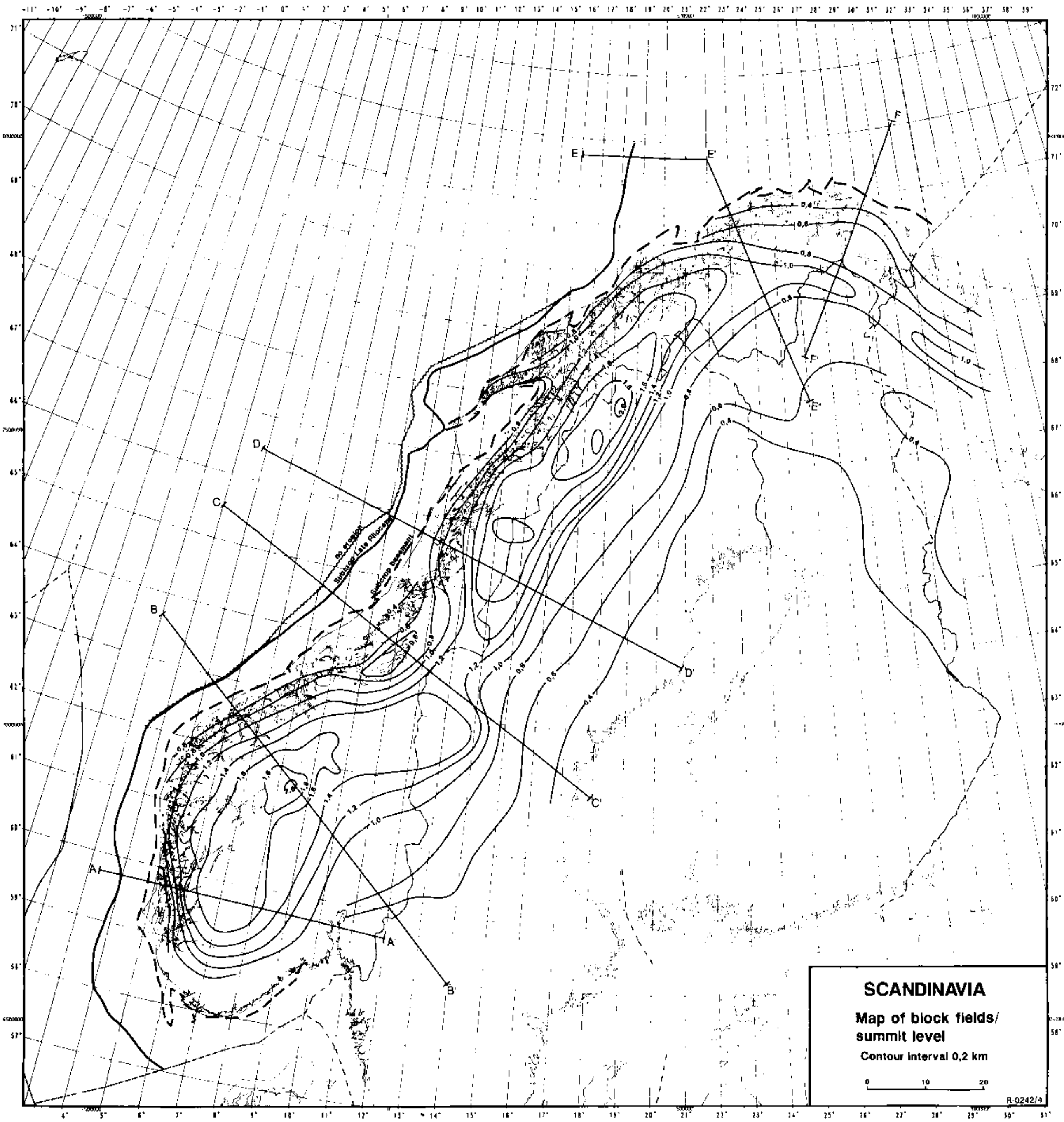

Fig. 9. Summit level map of Scandinavia. Constructed from height of block fields (Nesje et al., 1988) and by drawing isolines through the highest summits.

(Fig. 14) can, therefore, be considered as maximum values.

The relaxation time for the isostatic flexural compensation estimated from the postglacial uplift is only a few thousand years (Fjeldskaar and Cathles, 1991). Thus it is reasonable to assume that the flexural equilibrium was achieved during the erosional process. As the time period considered here is less than 5 million years, the viscous properties of the lithosphere (as described by Fjeldskaar and Pallesen, 1989) are not taken into account.

The resulting isostatic adjustments of the Earth's surface due to Pleistocene/Pliocene erosion/deposition are shown in Fig. 14a. The modelling was based on the erosion model presented in Fig. 13 and the parameter values are given in Table 2.

Calculation based on the amounts of erosion inferred show that the isostatic uplift of Late Pliocene/Pleistocene age amounts to 900–1400 m in the western Barents Sea, and the subsidence of the sediment

Fig. 10. Map of altitude of paleic surface in Norway. Compiled by Atle Nesje (University of Bergen, unpublished work). 500, 1000 and 1500 m contours are reinforced.

wedge is 700–1800 m. Onshore, the most severe erosion took place in the Nordland area, which caused an uplift of 200–800 m, depending on the lithosphere rigidity of the area.

Phase boundary migration

It has been proposed (e.g., Kennedy, 1959) that phase transitions in the Earth's upper layers provide

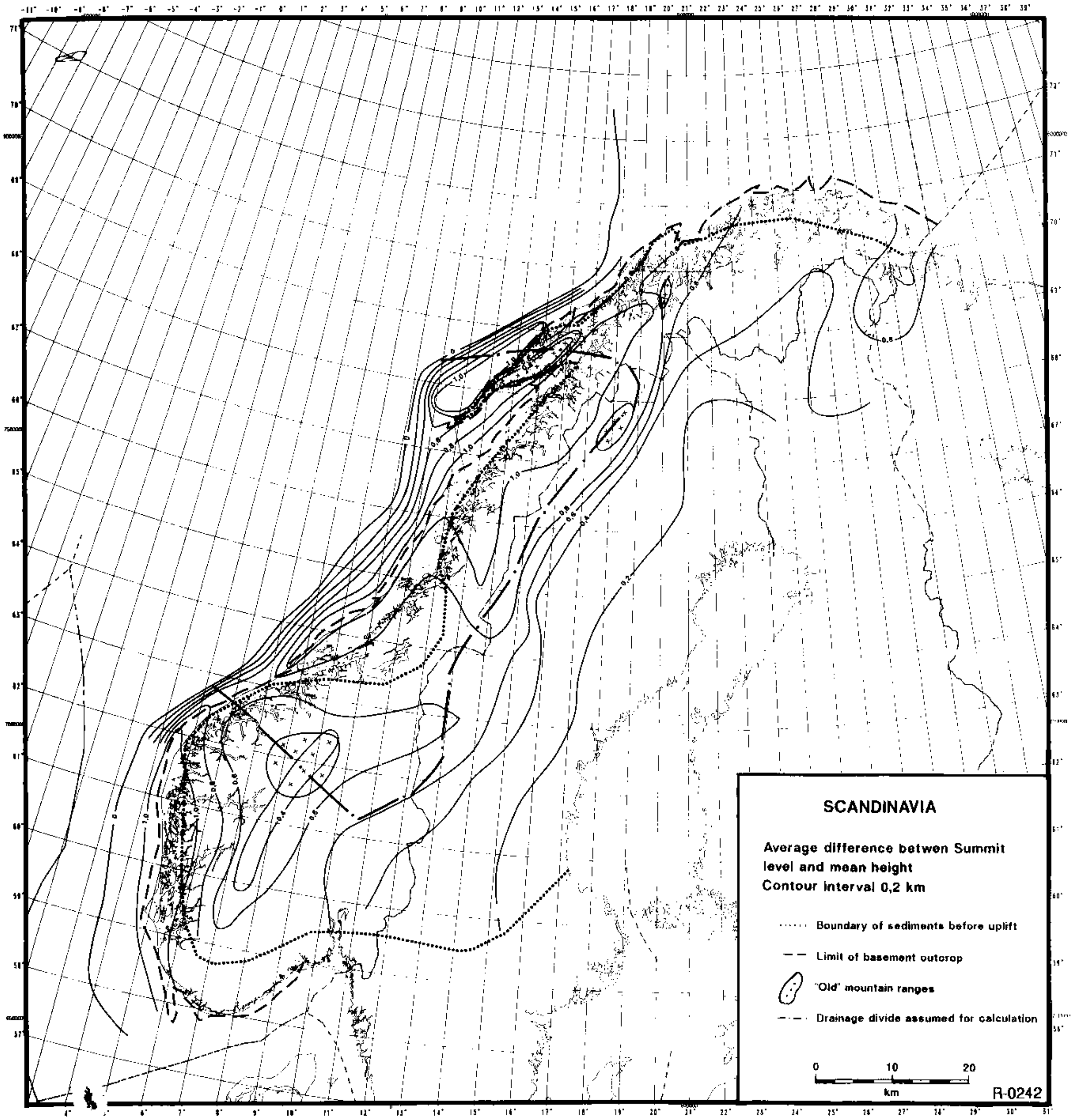

Fig. 11. Smoothed map showing difference between summit levels and mean height.

TABLE 2

Parameter values

Young's modulus	(E)	10.3×10^{10} Nm^{-2}
Poisson's ratio	(υ)	0.25
Lame's parameter	(μ, λ)	3.34×10^{10} Nm
Asthenosphere density	(ρ)	3.3×10^3 kg/m^3
Inverse slope of Clausius Clapeyron curve	(γ)	75 K/kbar
Geothermal gradient	(β)	12 K/km
Density of upper phase	(ρ_1)	2.8×10^3 kg/m^3
Density of lower phase	(ρ_2)	3.2×10^3 kg/m^3
Bulk modulus	(K)	5.56×10^{10} Nm

Partly based on O'Connell and Wasserburg (1967).

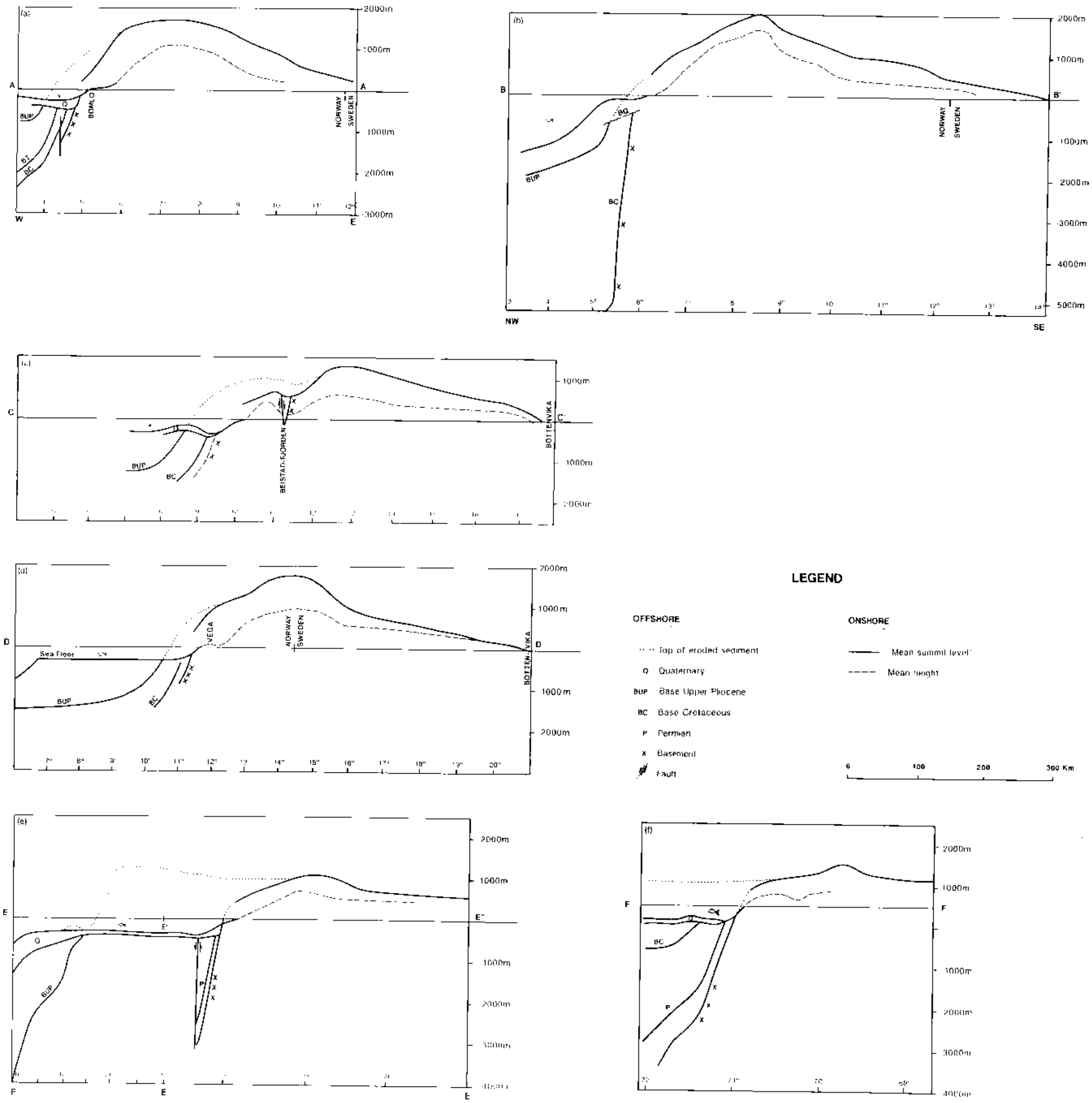

Fig. 12. Structural cross-sections showing summit level (solid line), mean height (dashed line), basement outcrop and Pliocene-Pleistocene deposits. For location of profiles, see Fig. 8.

a mechanism of uplift and subsidence of the Earth's surface. This is based on the assumption that phase transitions at depth respond to pressure changes at the surface. Subsidence at the surface occurs when an increase in pressure (by sedimentation) causes the upward migration of the phase boundary. Uplift would follow a decrease in pressure (by erosion) causing downward motion of the phase boundary. The mechanism has been studied by several authors (e.g., O'Connell and Wasserburg, 1967, 1972; Gjevik, 1972; O'Connell, 1976; Mareschal and Gangi, 1977; Mareschal and Lee, 1983). In these works, analytical and numerical approximations of the movement of the Earth's surface by phase boundary migration have been established.

The phase transition considered here is the transition from gabbro to eclogite at the base of the crust at a depth of 30–50 km. Migration of this phase boundary has been suggested as an explanation for the formation of sedimentary basins (e.g., Mareschal and Lee, 1983).

We have adopted an equilibrium approximation, since it is easily shown, using the analytical approximations of O'Connell (1976), that the phase bound-

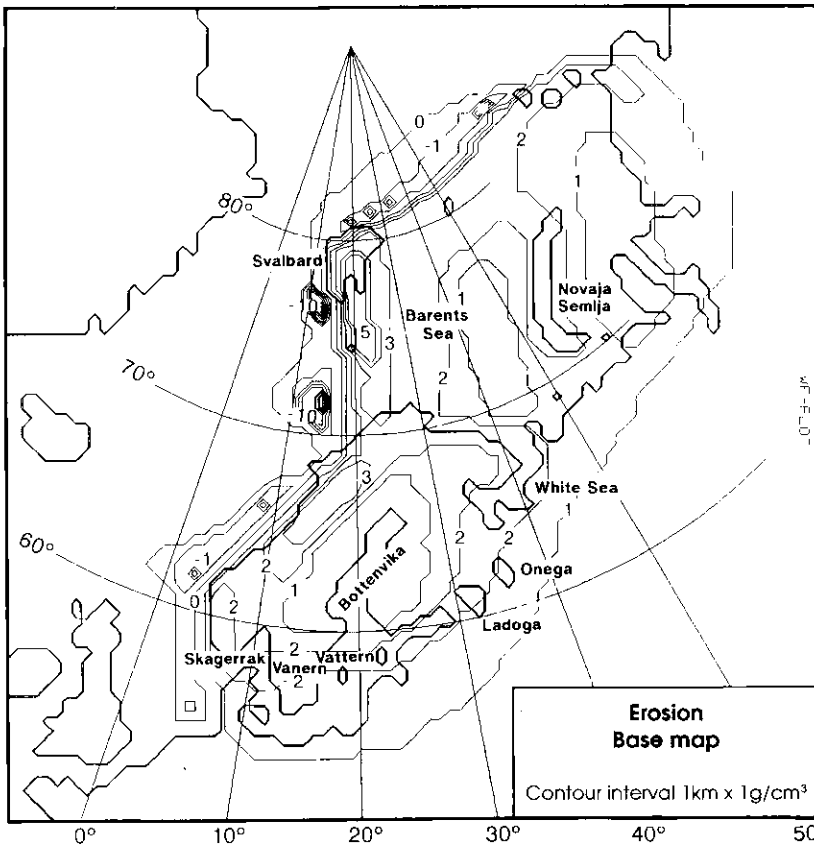

Fig. 13. Contoured grid showing values of erosion in the area covered by the Scandinavian and Barents Sea ice sheets in the glacial period. Erosion is given in terms of thickness (km) multiplied by density. The map is based on computer contouring of grid values. Contour interval: 1 km x 10 kg/m^3. Areas of sedimentation are given with negative values.

ary migration will be 90% compensated in 3 million years, using the parameter values given in Table 2 and a phase transition temperature of 1000 K. The results shown in Fig. 14b are based on an assumption of 50% phase transformation in the area. In spite of this, we consider the resulting surface adjustments to be maximum values. The mathematical model is formulated in Appendix B.

We are aware that field studies of eclogites indicate that granulites may be metastable for a long time, and eclogitization may take place only along shear zones where fluid has moved (Jamtveit et al., 1990).

The modelling of surface adjustments caused by possible changes in phase boundaries shows that this effect may be significant, of the order of 500 m of uplift in the western Barents Sea, and similar amounts of subsidence in the depositional wedge. Approximately 300 m of the uplift in the Nordland area may be explained by phase changes in the crust/mantle boundary.

The resulting adjustments of the Earth's surface due to movements of the phase transition are shown in Fig. 14b. The calculation was based on the erosion/deposition model in Fig. 13, and the parameter values given in Table 2.

Discussion of modelling results

The calculation of crustal response was based on a grid representing the thickness of material eroded in the glacial period multiplied by its density (Fig. 13), as well as the amount of sediment accumulations in the offshore wedges (negative numbers).

For the modelling, it was necessary to input information for the total area of glaciation and erosion and, therefore, very tentative values have been given in the eastern areas on which we have no information. However, errors in the eastern parts will not seriously affect the results in Norway and the Barents Sea.

The contoured values in the western Barents Sea do not coincide with data in Fig. 5, as we have tentatively tried to separate the last (glacial) phase from the Paleogene phase of uplift and erosion.

The three-dimensional modelling results indicate that the uplift of the Scandinavian mountains should be comparable with the amount of erosion. The areas of sediment accumulation subside substantially, allowing space for continued rapid deposition.

In the Barents Sea and other areas where sediments are removed, the net effect of the erosion

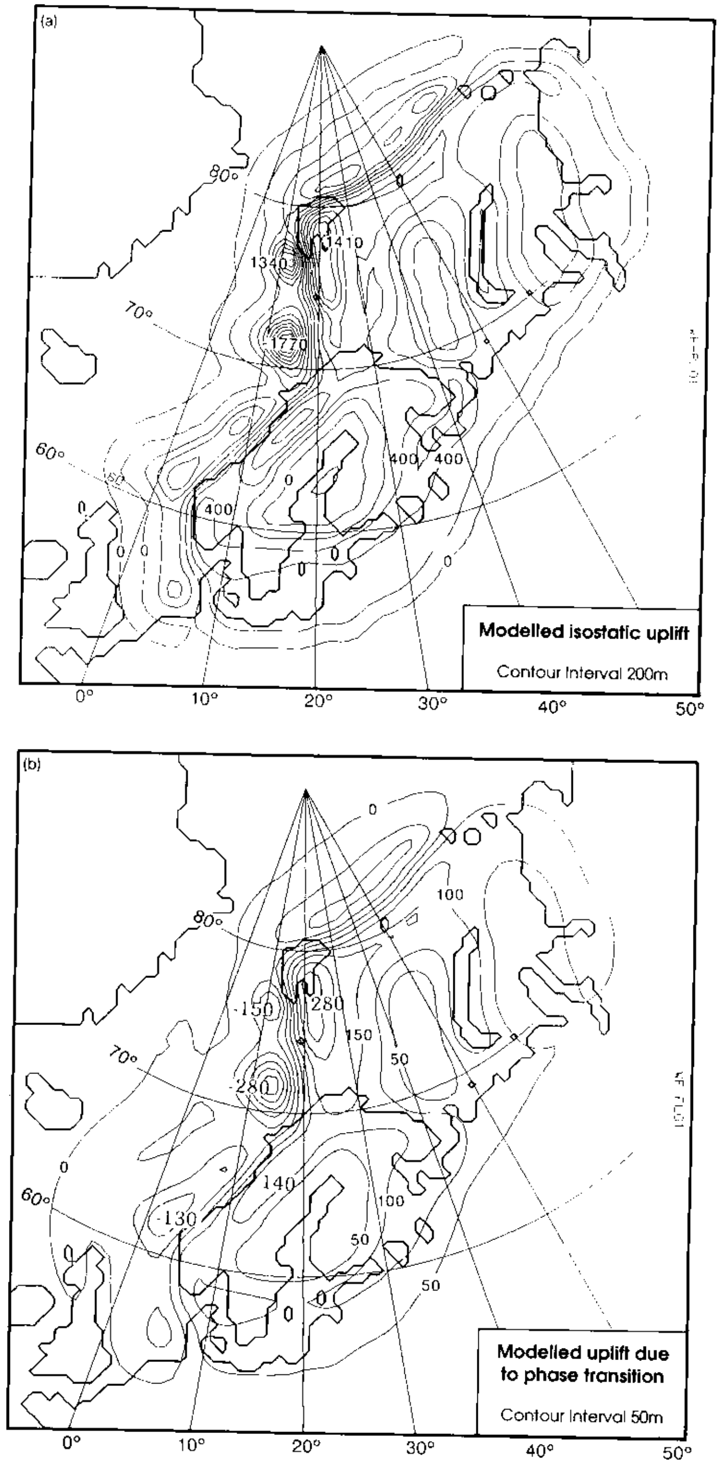

Fig. 14. Three-dimensional modelling of crustal response in the Scandinavia-Barents Sea areas. (a) Uplift and subsidence due to isostatic adjustment, using a flexural rigidity of 1×10^{23} Nm. The contour interval is 200 m. (b) Uplift and subsidence as a result of phase boundary migration, using a crustal thickness of 35 km and a density contrast of 0.4×10^3 kg/m^3. The contour interval is 50 m.

process is lowering of the surface. In the southern Barents Sea, the model predicts that removing 1200–1400 m of sediments (assuming a density of 2.2), results in an uplift of 600–800 m (compare Figs. 13 and 14). Hence, a pre-glacial elevation of 300–400 m above sea level is needed to explain the present elevation. In the western and northern Barents Sea, the elevation must have been higher. Similar results are obtained from the southern part of Sweden and Denmark, where mainly sediments were removed.

The three-dimensional modelling gives results which simulate quite well the observed structural pattern with the marginal highs and peripheral lows. The marginal highs are formed where dense base-

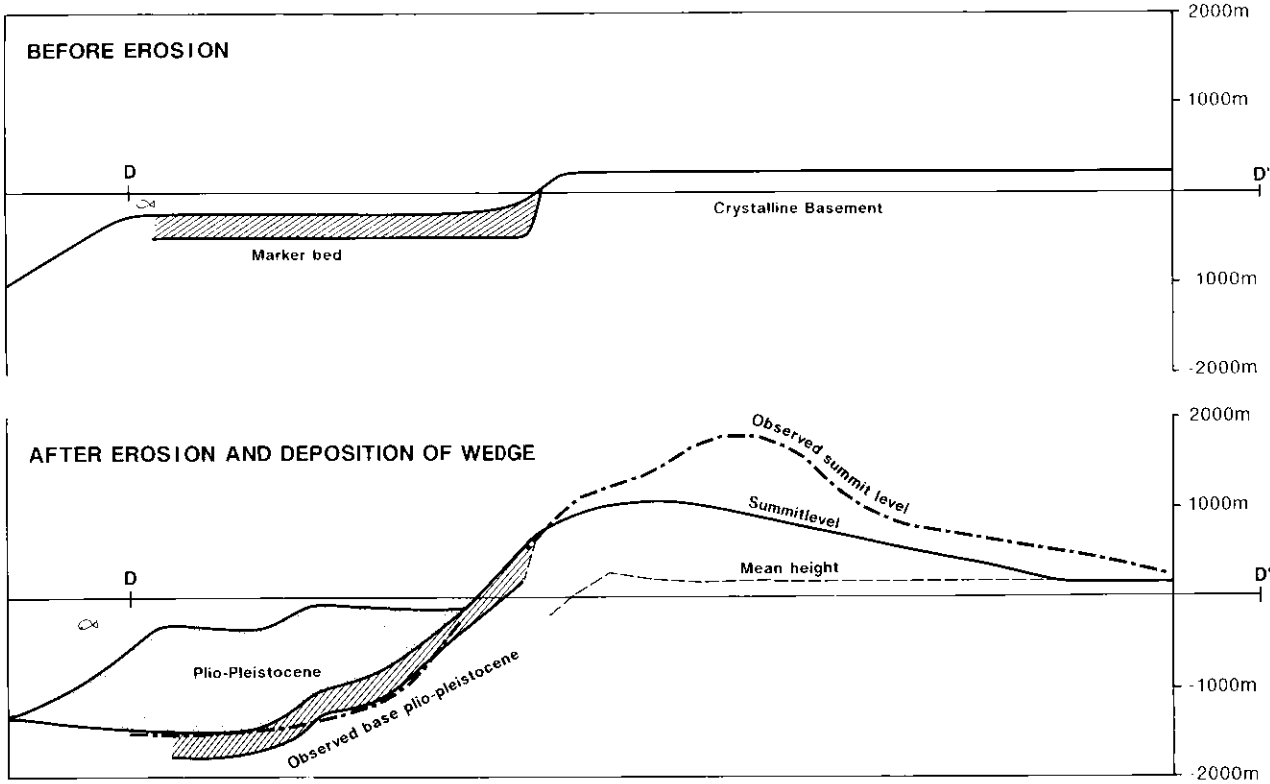

Fig. 15. Two-dimensional modelling of crustal response along profile *DD'*. The shape of a simplified pre-glacial topography is shown after erosion and deposition. The modelled results are compared with the observed profile *DD'*. For location, see Fig. 8.

ment rocks are eroded, and they are slightly displaced towards the central parts of the ice sheet from the area of maximum erosion. The peripheral lows seem to form where the erosion is at a maximum, while the subsidence/uplift is small due to the hinge line position at the boundary between crystalline basement and sedimentary cover. Where sediments only are eroded, the model predicts surface lowering, as is observed. In the central depression of Bottenvika, subsidence is predicted, and observed.

The low elevation of the Trondheim area can be explained by the assumption that it was covered by sedimentary rocks considerably lighter than the basement before onset of erosion.

However, the total theoretical adjustment is not in accordance with observation everywhere. Adjustment due to isostasy and phase change was calculated on profile DD′ crossing Vega, Nordland, extending from the shelf edge to Bottenvika (Fig. 12). This profile is regarded as a cross-section in an area which was not affected strongly by Paleogene tectonic events, and the uplifted dome as well as the depositional wedge are well developed.

The results of calculation (Fig. 15) show that the isostatic and phase change effects may comprise almost the amount of erosion in the onshore area. The calculated uplift is considerably smaller than observed, leaving a difference in the order of 500 m. Offshore, the subsidence was calculated by adding the isostatic and phase change effects to the effect of compaction of the underlying sediments. For the calculation of compaction, we used a general porosity trend based on well data from Haltenbanken. Figure 15 demonstrates that the subsidence modelled is slightly smaller than that observed. However, the fit is better than for the onshore part.

There are many possible explanations for the discrepancy observed. One way out of the problem could be to fit the onshore part by assuming that the pre-erosional elevation was much higher than 250 m. This explanation is supported by the paleic surface map, Fig. 10. However, the low sedimentation rates on the shelf from the Oligocene through Early Pliocene suggest that the relief was not high at this time. This points in the opposite direction.

It is thought that some anomalies can be explained by assuming a pre-erosional high relief. The best example is South Norway. Figure 10 indicates that the paleic surface is elevated to more than 1500 m between Sognefjorden and the high peaks of Jotunheimen. These high peaks represent a topography where the summit level is elevated several hundreds of metres above the paleic surface. In the North Sea and Møre Basin, thick Paleogene sedimentary wedges are observed with a progradational direction roughly from the east. Thus, there is independent evidence that the uplift of the highest mountains in South Norway took place largely prior to the Pliocene.

Hence, separation of the Paleogene uplift from the Plio-Pleistocene is problematic. Our calculations indicate that an important part of the concentric morphology can be explained by rapid erosion and deposition of the material in the glacial periods. However, if the calculated uplift is removed from the present topography, we will still be left with a concentric morphology, although more subdued. We consider it unlikely that plate tectonic effects related to the opening of the Atlantic in the Paleogene should create such a topography. Also, the deformation of the base of the Upper Pliocene and even Quaternary layers (Fugelli and Riis, in press) into the regional dome structures is such that it cannot be explained totally using the present model.

Conclusions

– The uplift of Scandinavia and the Barents Sea is related to Paleogene thermal effects due to the North Atlantic rifting and opening, and to isostatic adjustments due to Pliocene/Pleistocene glaciation and erosion.

– The models presented do not explain satisfactorily the total magnitude of the observed uplift of the Scandinavian mountains, although they simulate well the observed structural pattern.

– Observation and modelling suggest that the erosion effects are very important in the study of recent structures in glaciated areas worldwide.

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Appendix A

Isostatic model

The lithosphere is modelled as a thin elastic plate overlying an incompressible fluid substratum, expressed in the Fourier domain with the initial topographic load $F(k)$ and the resulting flexural deformation $h_0(k)$ written as a function of wavenumber k. The equilibrium isostatic compensation is achieved by subsidence (Cathles, 1975):

$$h_0(k) = \frac{F(k)}{\rho g \alpha(k)}$$

where g is gravity, ρ is density of the mantle, and $\alpha(k)$ is a parameter characterizing the flexure response of the lithosphere. The lithosphere flexure depends on the elastic strength of the lithosphere and $\alpha(k)$ is given by the equation (Cathles, 1975):

$$\alpha(k) = \frac{\frac{2\mu k}{\rho g}\,[S^2 - (kH)^2] + (CS + kH)]}{S + kHC}$$

where S is $\sinh kH$, C is $\cosh kH$, and μ is Lamé's parameter.

The mechanical thickness of the lithosphere, H, is given by the equation:

$$H = \left[\frac{12\,(1-\nu^2)\,D}{E}\right]^{1/3}$$

where D is flexural rigidity, E is Young's modulus, and ν is Poisson's ratio.

Appendix B

Theoretical formulation of phase boundary migration

Let us first, for the purpose of illustration, consider a one-dimensional case assuming a phase boundary at depth in the mantle, separating two phases of the same component. Suppose the temperature at which the phase change takes place (T_c) is related to the pressure by the following equation (Clausius-Clapeyron equation):

$$T_c = T_0 + \gamma P_c = T_0 + \gamma g \rho_1 z$$

T_0 is the transition temperature at the Earth's surface, ρ_1 is the density of the upper phase, z is the depth of the phase boundary and γ is the inverse slope of the Clausius-Clapeyron curve.

Consider a uniform erosion of a half-space which undergoes a phase change at depth z_p. The decrease in hydrostatic pressure due to the erosion is ΔP. If the geothermal gradient is initially zero, the phase boundary will migrate downward until the pressure decrease is cancelled. The equilibrium migration of the phase boundary is:

$$\Delta z_p = \frac{\Delta P}{\rho g}$$

This migration will cause an uplift of the Earth's surface of:

$$\Delta z = \Delta z_p \cdot \frac{\Delta\rho}{\Delta\rho + \rho}$$

where $\Delta\rho$ is the density difference between the upper and lower phases.

In this study we use an analytical approximation for the equilibrium position of a phase boundary under the horizontally varying surface loads given by Mareschal and Gangi (1977). Under the assumptions that the Earth behaves as an elastic solid above the phase boundary, that the equilibrium position does not depend on deviatoric stress and that the two phases have the same thermal properties, they found that the vertical displacement of the phase boundary due to the applied load $P(k,a)$ is in the Fourier domain:

$$S(k) = Y(k) \cdot P(k,a)$$

where:

$$Y(k) = \frac{\left[\dfrac{-\gamma}{(\gamma g \rho_1 - \beta)}\right] \exp(-|ka|)}{1 + \left[\dfrac{\mu}{(\lambda + \mu)}\right]\left(\dfrac{\alpha K}{a}\right)\left[\dfrac{ka[\sinh(2ka) - 2ka]}{D(ka)}\right]}$$

and:

$$D(y) = \left[\frac{\mu}{\lambda + \mu}\right]^2 + \left[\frac{\lambda + 3\mu}{\lambda + \mu}\right] \cosh^2 y + y^2$$

$$\frac{\alpha K}{a} = \frac{\rho_2 - \rho_1}{\rho_1} \frac{\left(\dfrac{\gamma K}{a}\right)}{\gamma g \rho_1 - \beta}$$

γ is the inverse slope of Clausius-Clapeyron curve, β is the geothermal gradient, ρ_1 is the density of the upper phase, ρ_2 is the density of the lower phase, a is the depth of the phase change, K is the bulk modulus, and λ and μ are Lamé's parameters.

References

Antonsen, P., Elverhøi, A., Dypvik, H. and Solheim, A., 1991. Shallow bedrock geology in the Olga Basin area, northwestern Barents Sea. Am. Assoc. Pet. Geol., Bull., 75(7): 1178–1194.

Balling, N., 1980. The land uplift in Fennoscandia. In: N.A. Mörner (Editor), The Earth Rheology, Isostasy and Eustasy. J. Wiley and Sons, New York, N.Y., pp. 297–321.

Baturin, G.N., 1988. Dokl. Akad. Nauk SSSR, 299, No. 2.

Brekke, H. and Riis, F., 1987. Tectonics and basin evolution of the Norwegian shelf between 62°N and 72°N. Nor. Geol. Tidsskr., 67: 295–321.

Bøe, R., 1989. The structural development of the Mesozoic rocks in Beitstadfjorden, Edøyfjorden and Frohavet, Central Norway. Terra Abstracts (Eur. Union Geosci.), 1: 22–23.

Cathles, L.M., 1975. The Viscosity of the Earth's Mantle. Princeton University Press, Princeton, N.J, 386 pp.

Dalland, A., Worsley D. and Ofstad, K., 1988. A lithostratigraphic scheme for the Mesozoic and Cenozoic succession offshore mid- and northern Norway. Norw. Pet. Directorate, NPD Bull., 4, 65 pp.

Eidvin, T. and Riis, F., 1989. Nye dateringer av de tre vestligste borehullene i Barentshavet. Resultater og konsekvenser for den tertiære hevingen. NPD Contrib., 27, 43 pp.

Elverhøi, A., Antonsen, P., Flood, S.B., Solheim, A. and Vullstad, A.A., 1988. The physical environment Western Barents Sea 1:1500000. Shallow bedrock geology. Nor. Polarinst. Skr. 179 D.

Elverhøi, A., Liestøl, O. and Nagy, J., 1980. Glacial erosion, sedimentation and microfauna in the inner part of Kongsfjorden, Spitsbergen. Nor. Polarinst. Skr., 172: 33–61.

Elverhøi, A., Lønne, Ø. and Seland, R., 1983. Glaciomarine sedimentation in a modern fjord environment, Spitsbergen. Polar Res., 1: 127–149.

Fjeldskaar, W. and Cathles, L.M., 1991. Rheology of mantle and lithosphere inferred from postglacial uplift in Fennoscandia. In: R. Sabadini et al. (Editors), Glacial Isostasy, Sea Level and Mantle Rheology. Kluwer Academic Publishers, Dordrecht, pp. 1–19.

Fjeldskaar, W. and Riis, F., 1988. Modellering av tertiærbasseng på den vestlige marginen av Barentshavet. In: K. Binzer (Editor), Abstracts, 18. Nordi. Geol. Vintermøde, København, 1988. Danmarks Geologiske Undersøgelse.

Fjeldskaar, W. and Pallesen, S., 1989. The application of a viscoelastic lithosphere model to isostatic subsidence in backstripping. In: J. Collinson (Editor), Correlation in Hydrocarbon Exploration. Graham and Trotman, London, pp. 53–59.

Fugelli, E.M.G. and Riis, F. in press. Neotectonism in the Jæren area, southwest Norway. In: L.N. Jensen and F. Riis (Editors), Post-Cretaceous Uplift and Sedimentation Along the Fennoscandian Shield. Nor. Geol. Tidsskr.

Gabrielsen, R.H., Færseth, R.B., Jensen, L.N., Kalheim, J.E. and Riis, F., 1990. Structural elements of the Norwegian continental shelf. Part I: The Barents Sea Region. NPD Bull., 6, 33 pp.

Geographical Survey Office of Sweden, 1963. Mean height map of Sweden. Rikets Allmänna Kartverk, Meddelande, Ser. A, No. 30.

Gjessing, J., 1967. Norway's paleic surface. Nor. Geol. Tidsskr., 21: 69–132.

Gjevik, B., 1972. Surface adjustment due to a subcrustal phase transition. Phys. Earth Planet. Inter., 5: 403–408.

Jamtveit, B., Bucher, K. and Austrheim, H., 1990. Fluid controlled eclogitization of granulites in deep-crustal shear zones, Bergen Arcs, Western Norway. Contrib. Mineral. Petrol., 104: 184–193.

Jansen, E., Bleil, U., Henrich, R., Kringstad, L. and Slettemark, B., 1988. Paleoenvironmental changes in the Norwegian Sea and the Northeast Atlantic during the last 2.8 Ma: DSDP/ODP Sites 610, 642, 643 and 644. Paleoceanography, 3: 563–581.

Kennedy, G.C., 1959. The origin of continents, mountain ranges and ocean basins. Am. Sci., 47: 491–504.

Løvø V., Elverhøi, A., Antonsen, H.P., Solheim, A., Butenko, G., Gregersen, O. and Liestøl, O., 1990. Submarine permafrost and gas hydrates in the Northern Barents Sea. Nor. Polarinst., Rapportser., 56, 171 pp.

Manum, S. and Throndsen, T., 1978. Rank of coal and dispersed organic matter and its geological bearing in the Spitsbergen Tertiary. Nor. Polarinst., Årbok 1977, pp. 159–177.

Mareschal, J.C. and Gangi, A.F., 1977. Equilibrium position of a phase boundary under horizontally varying surface loads. Geophys. J.R. Astron. Soc., 49: 757–772.

Mareschal, J.C. and Lee, C.K., 1983. Initiation of subsidence in a sedimentary basin underlain by a phase change. Geophys. J.R. Astron. Soc., 74: 689–712.

Nesje, A., Dahl, S.O., Anda, E. and Rye, N., 1988. Block fields in southern Norway: significance for the Late Weichselian ice sheet. Nor. Geol. Tidsskr., 68: 149–169.

Nyland, B., Jensen, L.N., Skagen, J.I., Skarpnes, O. and Vorren, T., 1992. Tertiary uplift and erosion in the Barents Sea; magnitude, timing and consequences. In: R.M. Larsen, H. Brekke, B.T. Larsen and E. Talleraas (Editors), Structural and Tectonic Modelling and its Application to Petroleum Geology. Norwegian Petroleum Society (NPF), Special Publication 1. Elsevier, Amsterdam, pp. 153–162 (this volume).

Nøttvedt, A., Berglund, T., Rasmussen, E. and Steel, R., 1988. Some aspects of Tertiary tectonics and sediments along the western Barents Shelf. In: A.C. Morton and L.M. Parson, (Editors), Early Tertiary Volcanism and the Opening of the NE Atlantic. Geol. Soc. London, Spec. Publ., 39: 421–425.

O'Connell, R.J., 1976. The effects of mantle phase changes on postglacial rebound. J. Geophys. Res., 81: 971–974.

O'Connell, R.J. and Wasserburg, G.J., 1967. Dynamics of the motion of a phase change boundary to changes in pressure. Rev. Geophys. Space Phys., 3: 329–410.

O'Connell, R.J. and Wasserburg, G.J., 1972. Dynamics of submergence and uplift of a sedimentary basin underlain by a phase change boundary. Rev. Geophys. Space Phys., 10: 335–368.

Perry, R.K. and Fleming, H.S., 1985. Bathymetry of the Arctic Ocean. Map sheet, scale 1:4704073. Naval Research Laboratories, Washington D.C.

Ramberg Moe, H., Roaldset, E. and Gjelsvik N., 1988. Siliceous sediments and seismic response — a correlation tool in exploration on Norwegian shelf. Norwegian Petroleum Society, Correlation in Hydrocarbon Exploration, 3–5 October 1988 (abstract).

Rokoengen, K. and Rønningsland, T.M., 1983. Shallow bedrock geology and Quaternary thickness in the Norwegian sector of the North Sea between 60°30′N and 62°N. Nor. Geol. Tidsskr., 63: 83–102.

Sales, J., 1989. Late Tertiary uplift — West Barents shelf. Character, mechanisms and timing. Paper presented at the Norwegian Petroleum Society Workshop, 18–20 October 1989, Stavanger.

Spencer, A.M., Home, P.C. and Berglund, L.T., 1984. Tertiary structural development of the western Barents Shelf: Troms to Svalbard. In: A.M. Spencer et al. (Editors), Petroleum Geology of the North Atlantic Margin. Graham and Trotman, London, pp. 99–109.

Steel, R.J. and Worsley, D., 1984. Svalbard's post-Caledonian strata — an atlas of sedimentational and palaeogeographical evolution. In: A.M. Spencer (Editor), Petroleum Geology of the North European Margin. Graham and Trotman, London, pp. 98–107.

Strøm, K.M., 1948. The geomorphology of Norway. Geogr. J., 62: 19–27.

Sugden, D.E., 1976. A case against deep erosion of shields by ice sheets. Geology, 4: 580–582.

Talwani, M., Udintsev, G. et al., 1976. Initial reports of the Deep Sea Drilling Project, Vol. 38. U.S. Government Printing Office, Washington, D.C., 1256 pp.

Vorren, T.O., Lebesbye, E., Henriksen, E., Knutsen, S.M. and Richardsen, G., 1988. Cenozoisk erosjon og sedimentasjon i det sørlige Barentshav. Abstr., XVIII Nord. Geol. Vintermøte, København, 1988, 424 pp.

Vorren, T.O., Lebesbye, E., Andereassen, K. and Larsen, K.-B., 1989. Glacigenic sediments on a passive continental margin as exemplified by the Barents Sea. Mar. Geol., 85: 251–272.

Vorren, T.O., Richardsen, G., Knutsen, S.-M. and Henriksen, E., 1990. The western Barents Sea during the Cenozoic. In: U. Bleil and J. Thiede (Editors), Geologic History of the Polar Oceans: Arctic versus Antarctic. NATO ASI Series. Kluwer Academic Publishers, Dordrecht, pp. 95–118.

Weidick, A., 1976. Glaciation and the Quaternary of Greenland. In: A. Escher and W. Stuart Watt (Editors), Geology of Greenland. Geological Survey of Greenland, Copenhagen, pp. 430–458.

White, W.A., 1972. Deep erosion by continental ice sheets. Geol. Soc. Am. Bull., 83: 1037–1056.

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