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# Late Weichselian paleoceanography of the southeastern Norwegian Sea

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Downcore analyses of variations in planktonic and benthic foraminiferal distribution, sedimentology and stable isotopes in cores from the deep ocean basin and the continental slope of the southeastern Norwegian Sea show that the last glacial to postglacial transition occurred in two distinct steps. Prior to 13,000 BP the area was characterized by a year-round iceberg and pack-ice environment, low surface productivity and homogenous, low productivity deep water conditions. After ca. 13,000 BP the ocean became seasonally ice free with high productivity and low-scale vertical circulation. The surface remained cold until after 10,000 BP, in contrast to the warm North Atlantic during the period 13,000–11,000 BP. In this period benthic foraminifera indicate presence of stratified deep-waters dissimilar to the present. Two possible patterns of deep water exchange are presented. During the Younger Dryas chronozone the Norwegian Sea remained open. Constant, strong bottom currents were introduced on the upper continental slope, and large scale bottom water formation and overflows to the North Atlantic possibly started during the Y. Dryas. The final warming of the Norwegian Sea started around 10,000 BP as a rapid introduction of warm, temperate surface water in the eastern part, while the warming of central and western parts occurred in the early Holocene.

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A large number of publications dealing with the changing surface ocean circulation through the Quaternary have been published during the last decade, especially through the CLIMAP project. Information on the deep circulation is, however, much more sparse than for the surface circulation. One explanation for this might be that there does not exist precise knowledge of the ecological preferences of deep-sea benthic foraminifera, thus prohibiting the use of transfer functions for paleoecological interpretations. Since some of the models which have been proposed for explaining the cause of Pleistocene glaciations involve theories for changing bottom water circulation and formation (Weyl 1968, Newell 1974, Lamb & Woodruffe 1979), it is important to obtain more information on the deep-sea environmental changes.

Streeter et al. (1982) showed that benthic and planktonic records from the Norwegian Sea display a covariance, but the exact relationship between surface changes and deep-water circulation is not yet clearly understood. Further progress seems to depend on more detailed studies of cores with high stratigraphic resolution within intervals showing large climatic fluctuation. Our strategy has been to study in detail benthic and planktonic environmental signals during the deglaciation period. This paper presents models on the surface and deep-water circulation based on investigations of cores from the continental slope off Norway. We are able to detect events at different depths and to correlate a high resolution late-glacial deep-sea record with the climatic evolution of the shelf areas outside Norway.

## Bathymetry and present hydrography

The Norway Basin (max depth 3900 m) is situated in the southeastern part of the Norwegian Sea (Fig. 1), and covers the area between Iceland and Western Norway. A deep, irregular sill

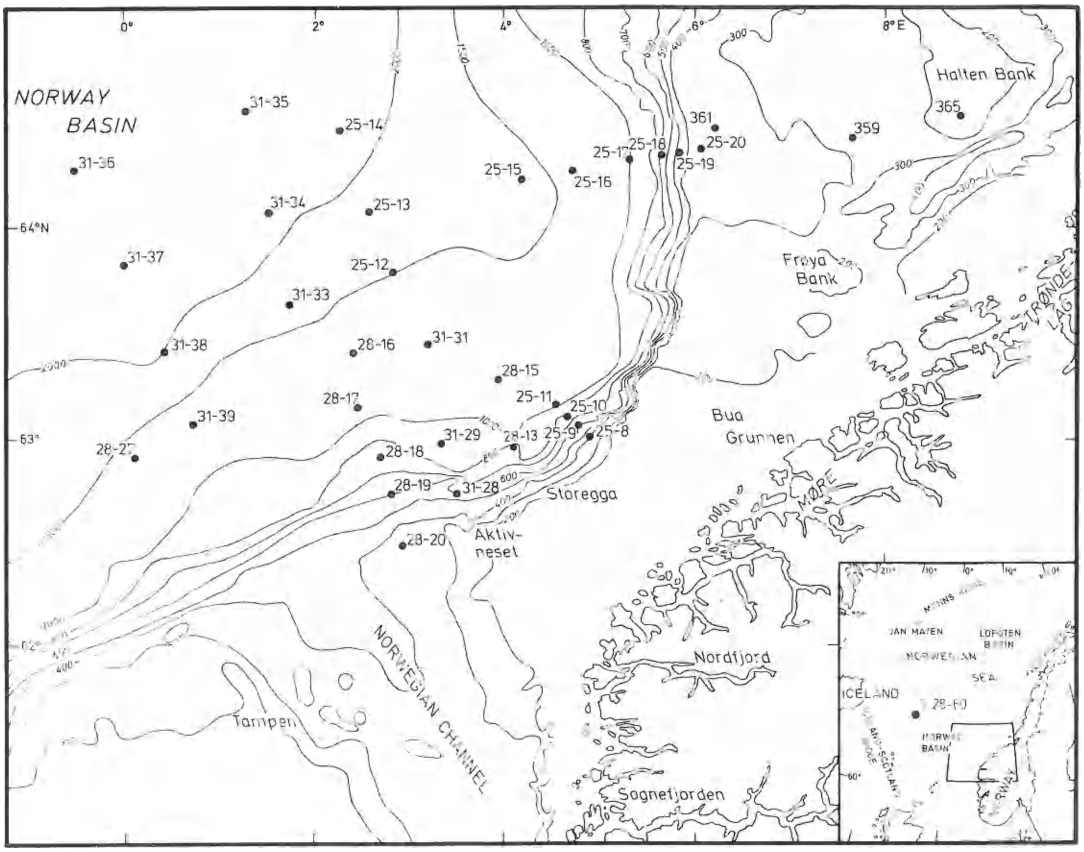

*Fig. 1.* Map showing core locations in the southeastern Norwegian Sea. Depth contours in metres.

along the Jan Mayen Fracture Zone separates this basin from the Lofoten Basin in the north. The southern rim of the basin is formed by the Iceland-Scotland Ridge with a maximum depth of 800 m in the Faeroe/Shetland Channel. Parts of the area west of Storegga (63°–64°N) have very steep and rugged topography (Fig. 1). Acoustic profiling indicates that the area has been exposed to at least one large submarine slide (Holtedahl 1971, Bugge et al. 1978, Hald 1980). Cores from depths down to more than 2000 m contain preconsolidated material (Løvlie & Holtedahl 1980). With the exception of core 25–9, we did not use cores from this area for stratigraphical studies.

The southeastern Norwegian Sea is today the main entrance of warm, saline waters from the North Atlantic Current which forms the northwards flowing Norwegian Current (NC) along the Norwegian Continental Margin. The NC water covers the shelf and upper continental slope (down to 600–700 m) along the Norwegian coast (Fig. 8). Nearshore, the Coastal Current forms a low salinity layer above the NC water. The deep water of the Norwegian Sea (NSDW) is formed as saline Atlantic water is cooled and sinks during winter and autumn in the northern Norwegian and Greenland Seas. NSDW being recently formed, is well oxygenated and homogenous. The water below 800 m is homohaline, while the water below 1200 m is both homohaline and homothermal (Mosby 1959). The present deep-water formation in the Norwegian-Greenland Seas leads to substantial overflows of dense, cold water southwards across the Greenland-Scotland ridge into the North Atlantic where it forms a major part of the North Atlantic Deep Water (NADW). Estimates based on tritium and $^{14}$C-profiles indicate that NSDW is renewed within 100 yrs, in contrast to 30 yrs

*Table 1.* Core selection and core locations

| Core | Latitude | Longitude | Depth (m) | G = Gravity core<br>P = Piston core |
|---|---|---|---|---|
| 359 | 64°23'N | 09°11'E | 270 | G |
| 361 | 64°23,5'N | 06°38'E | 365 | G |
| 365 | 64°28'N | 07°48'E | 190 | G |
| 25–8 | 62°59'N | 04°58'E | 300 | G + P |
| 25–9 | 63°03'N | 04°47'E | 600 | G + P |
| 25–10 | 63°05'N | 04°41'E | 790 | G + P |
| 25–11 | 63°09'N | 04°31'E | 1000 | G + P |
| 25–12 | 63°49'N | 02°50'E | 1500 | G + P |
| 25–13 | 64°04.5'N | 02°19'E | 2000 | G + P |
| 25–14 | 64°26'N | 02°19'E | 2400 | G + P |
| 25–15 | 64°13,5'N | 04°14'E | 1690 | G + P |
| 25–16 | 64°15,7'N | 04°43'E | 1400 | G + P |
| 25–17 | 64°17,5'N | 05°23'E | 1000 | G + P |
| 25–18 | 64°20'N | 05°40'E | 650 | G + P |
| 25–19 | 64°21'N | 05°50'E | 450 | G + P |
| 25–20 | 64°21'N | 06°01'E | 365 | G + P |
| 28–13 | 62°56'N | 04°06.5'E | 800 | G + P |
| 28–15 | 63°17'N | 03°58,3'E | 1200 | G + P |
| 28–16 | 63°25'N | 02°28'E | 1200 | G |
| 28–17 | 63°09'N | 02°32'E | 1000 | G |
| 28–18 | 62°56'N | 02°44'E | 770 | G |
| 28–19 | 62°44'N | 02°50,5'E | 570 | G |
| 28–20 | 62°29,5'N | 02°58'E | 375 | G |
| 28–27 | 62°48'N | 00°06'E | 1190 | G |
| 31–28 | 62°44,9'N | 03°21,5'E | 510 | G |
| 31–29 | 62°58,0'N | 03°17,0'E | 800 | G |
| 31–31 | 63°37'N | 03°10'E | 1146 | P |
| 31–33 | 63°38,2'N | 01°46'E | 1580 | P |
| 31–34 | 64°02'N | 01°32'E | 2200 | P |
| 31–35 | 64°32'N | 01°14'E | 2750 | P |
| 31–36 | 64°15'N | 00°31'E | 2620 | P |
| 31–37 | 63°46,7N | 00°00'E | 2285 | P |
| 31–38 | 63°25'N | 00°23'E | 1880 | P |
| 31–39 | 63°02'N | 00°44,6'E | 1309 | P |
| V28–60 | 64°05'N | 04°02'W | 3244 | P |

for that of the Greenland Sea, and that the overflow water is younger than NSDW and probably originates from water lying above the pycnocline at 1200 m (Peterson & Rooth 1976).

## Material and methods

The core selection consists of 24 piston and 26 gravity cores (Fig. 1, Table 1). All cores have been studied for grain size distribution, and most cores for foraminiferal content. $CaCO_3$-determinations have been carried out on most cores. Piston cores were collected with a system that did not include 'trigger weight corer' (except V28–60). The record in core 31–35 and stratigraphic studies of trigger weight cores from the Lamont-Doherty core collection indicate that some of the top sediments are lost during coring in most piston cores, and that the recorded thicknesses of Holocene sediments generally are too small. However, the biostratigraphical investigations strongly indicate that Holocene faunas are present in the top samples of the cores.

Most cores were x-ray photographed before opening in order to detect possible 'flow-in' and disturbed sequences. Sections where these features were suspected were not analyzed. X-ray photographs were also used as a preliminary stratigraphic tool in the selection of cores and sampling intervals. Grain-size analyses were performed using the sieving/pipette technique described in Mangerud (1977). $CaCO_3$ content refers to bulk sample. The results were obtained using a gas-volumetric method. Foraminiferal samples were prepared according to the method given by Feyling-Hanssen (1958). Samples rich in terrigenous sand were separated with $CCl_4$, density 1.6 g/cm$^3$. Foraminiferal counts were performed on the 150–1000 micron fractions in cores 28–27 and V28–60. In all other cores the 125–1000 micron fraction was used. Cores 31–33, 31–36 and V28–60 were chosen for stable isotope analysis. Left-coiling specimens of the planktonic foraminifer *Globoquadrina pachyderma* were analyzed. The chronostratigraphic terminology and division follow the suggestions of Mangerud et al. (1974) with alterations proposed by Mangerud & Berglund (1978).

## Lithostratigraphy

### *Stratigraphic units*

Lithostratigraphic investigations of the cores yield 4 distinguishable units (Fig. 2) which, according to their regional distribution (Hedberg 1976), are defined as formal stratigraphic units. The names of the units are picked from geographical names in or close to the study area (Fig. 1).

*Møre Diamicton:* The diamicton is found in the lower part of the cores at all depths in the investigated area, and as no cores penetrate it, only a minimum thickness of 5 m can be given (Fig. 3). It is homogenous, very poorly sorted and contains all grain sizes from clay to gravel (Fig. 2). Average grain-size distribution is.: 55 % clay, 29 % silt, 15 % sand, 1 % gravel. The fossil con-

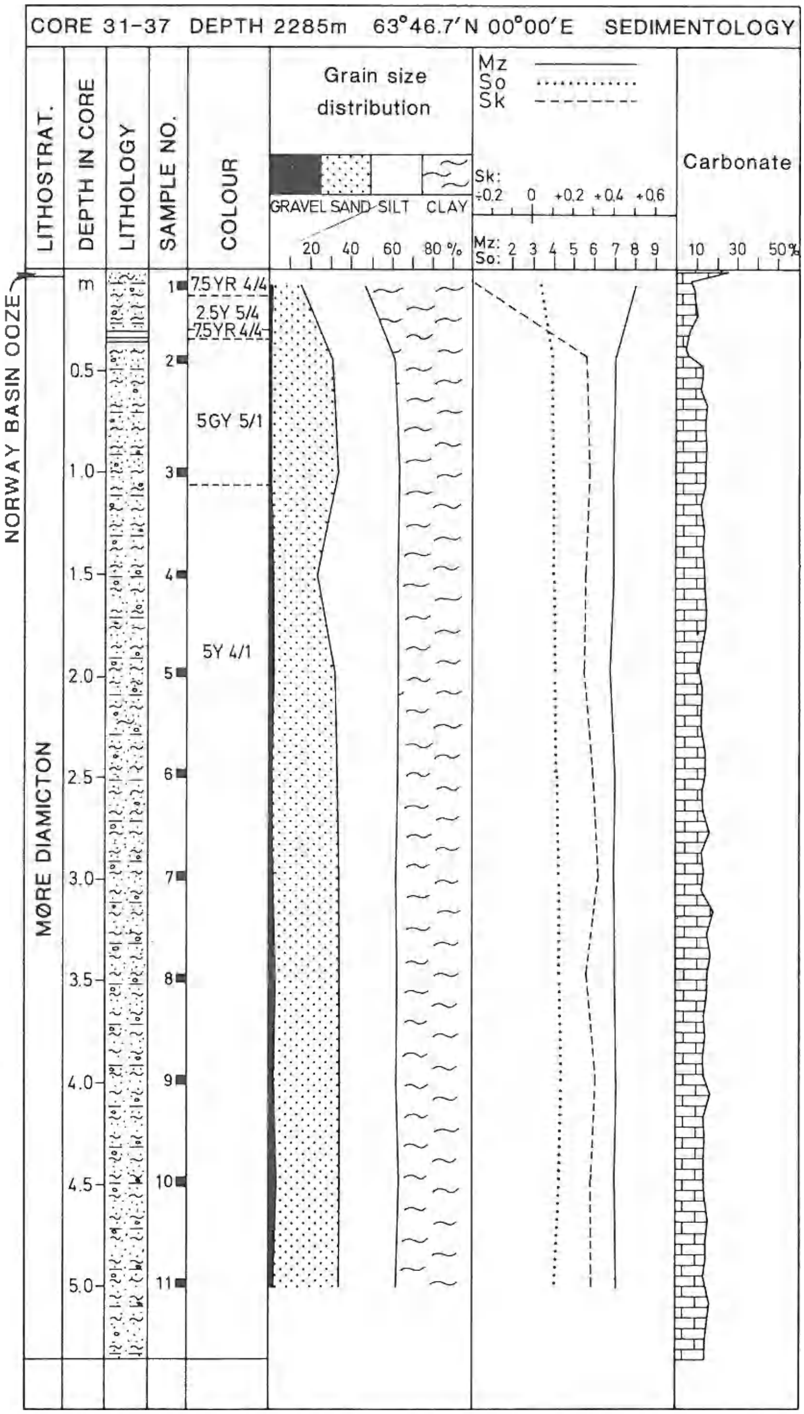

*Fig. 2.* Downcore sedimentological properties and lithostratigraphical subdivision of cores 31-37 (2a), 31-33 (2b), 28-17G (2c), and 25-09P (2d). Mean grain size (Mz), Sorting (So) and Skewness (Sk) as defined in Folk & Ward (1957). Colour code from Munsell Soil Color Chart (1975).

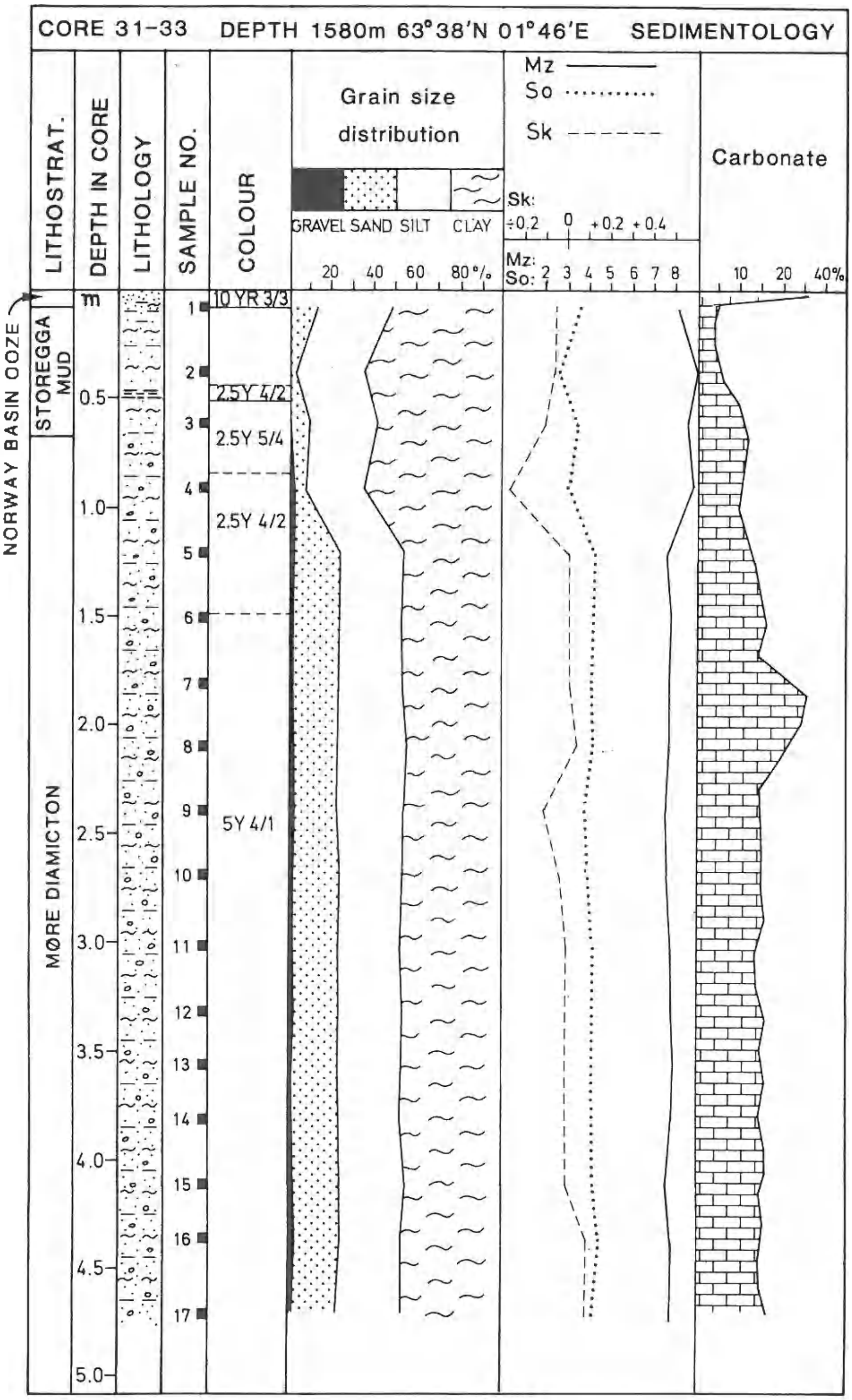
CORE 31-33 DEPTH 1580m 63°38'N 01°46'E SEDIMENTOLOGY
LITHOSTRAT.
DEPTH IN CORE
LITHOLOGY
SAMPLE NO.
COLOUR
Grain size distribution
GRAVEL SAND SILT CLAY
20 40 60 80 %
Mz
So
Sk
Sk: ÷0.2 0 +0.2 +0.4
Mz: So: 2 3 4 5 6 7 8
Carbonate
10 20 40%
NORWAY BASIN OOZE
STOREGGA MUD
MØRE DIAMICTON
m
0.5
1.0
1.5
2.0
2.5
3.0
3.5
4.0
4.5
5.0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
10 YR 3/3
2.5Y 4/2
2.5Y 5/4
2.5Y 4/2
5Y 4/1

*Fig. 2b.*

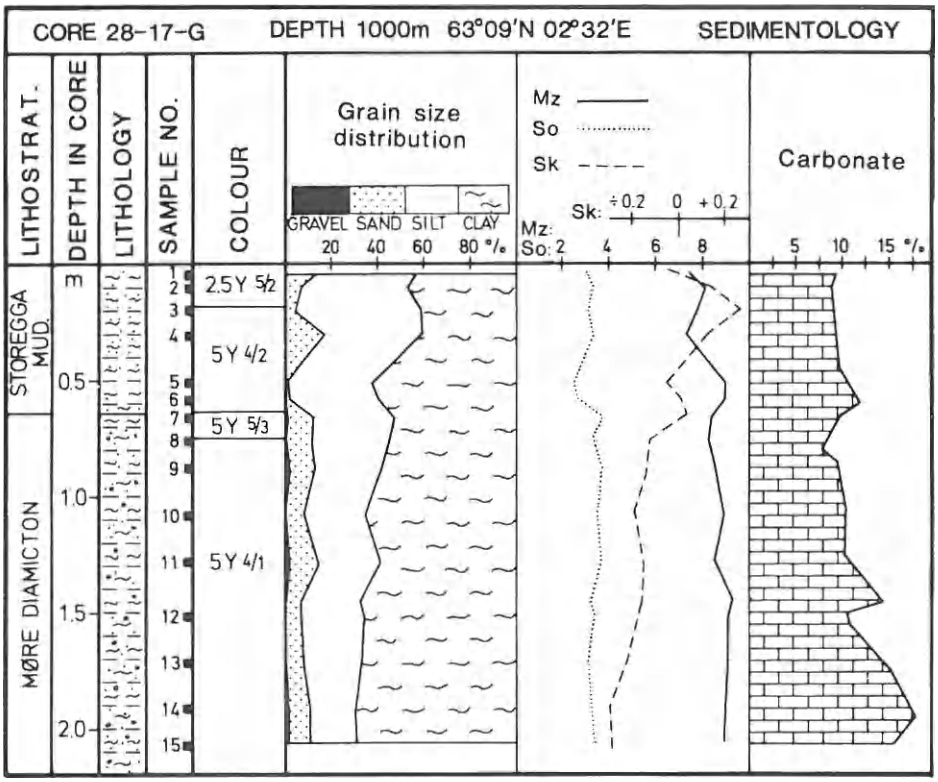

*Fig. 2c.*

tent is very low, usually less than 10 foraminifera per gram sediment, except in the upper sections where the fossil content increases. Carbonate content averages between 10 % and 15 %, but samples containing up to 30 % are found in some cores. Sand lenses and more sandy or clayey parts are found within Møre Diamicton in some cores from the upper slope (Fig. 2d). Core 31–33 (Fig. 2b) is chosen as stratotype for the unit.

The very even distribution together with the poorly sorted character strongly indicate a glacial origin for this sediment. For cores on the continental slope and in the deep basin, the only plausible genesis is that caused by ice-rafting and sedimentation by suspended matter. Some sand lenses may occasionally occur, but outside the submarine slide area no structures indicating turbidity current deposition have been recognized. Investigations of regional trends give no clear indications with respect to depth or geographical position except a small tendency towards higher clay content at about 1000 m depth. The regional stratigraphic distribution of Møre Diamicton is shown in Fig. 3. At depths less than 600–800 m Møre Diamicton is overlain by Aktivneset Sand (Fig. 2d), between 600–800 and 1600 m by Storegga Mud (Fig. 2b & 2c), and at depths greater than 1600 m it is succeeded by Norway Basin Ooze (Fig. 2a).

*Storegga Mud.* – The unit is found in cores between 800 and 1600 m depth. It is overlain by Norway Basin Ooze between 1200 and 1600 m (Fig. 2b), whereas the unit forms the surface sediment within the depth interval 600–800 m to 1200 m. The unit is characterized by high pelitic content (average 92 %), low sand content (average 8 %), and contains no gravel (Fig. 2). The sediment is poorly sorted and has relatively low carbonate content (average 10 %) which in some cores increases upwards. Fossil content is low to moderate and tends to increase upwards. The thickness of the unit is variable; the thickest found is 4 m (Fig. 3) at about 1000 m depth. Core 28–18G is chosen as a stratotype for the unit (Fig. 2c).

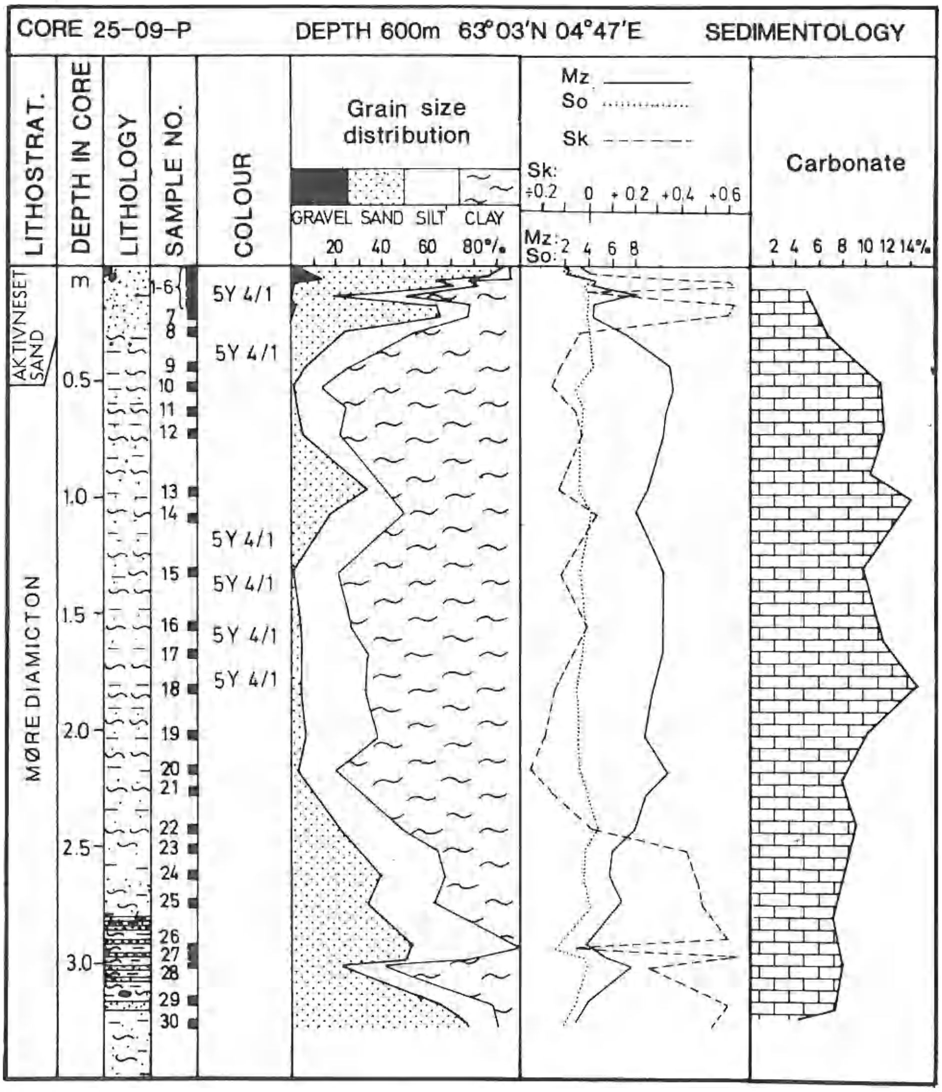

*Fig. 2d.*

Storegga Mud shows a distinct trend of higher clay content at increasing depths, indicating rapid sedimentation in a relatively quiet hydraulic regime. These observations indicate that the sediments mainly accumulated by redeposition of material brought into suspension by the action of strong currents on the upper slope. The fact that Storegga Mud is overlain by Norway Basin Ooze at depths between 1200 and 1600 m implies that the unit previously was deposited at greater depths than today. This shallowing of sedimentation province may have been caused by decreased supply of material exposed to current erosion on the upper slope through the formation of a lag deposit on top of glacigenic diamictons, and also by the postglacial sea level rise.

*Norway Basin Ooze.* – This unit forms the surface sediments at depths greater than 1200 m. The thickness found is usually 5–10 cm, but thickness up to 30–40 cm has been recorded (Fig. 3). The sediments of the unit are fine grained, with a coarse fraction composed mainly of foraminiferal shells. $CaCO_3$ content (>30 %)

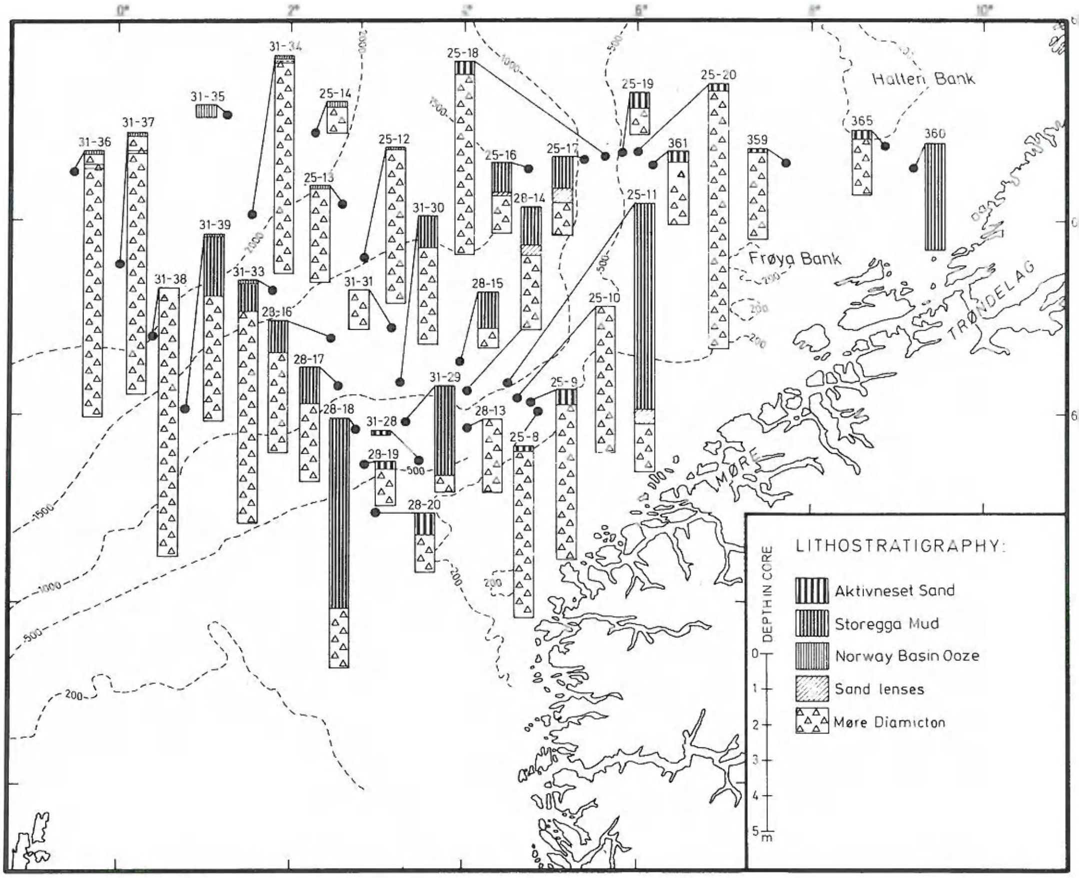

*Fig. 3.* Regional lithostratigraphy in study area. See text for definitions of lithostratigraphic units.

and foraminiferal content (>1.500 tests/g dry sediment) are high (Fig. 2). Core 31–33 is chosen as a stratotype for the unit (Fig. 2b). Kellogg (1975, 1976) demonstrated that there is a clear relation between high carbonate content of the surface sediments of the Norwegian Sea and the presence of warm, saline Atlantic surface water. The deposition of Norway Basin Ooze is thus closely related to the postglacial warming and increased productivity of the Norwegian Sea.

*Aktivneset Sand.* – This sand unit forms a clear lag-deposit on top of Møre Diamicton on the upper slope and on the shelf. The sediments are well to moderately sorted, have an average sand content of 70 % and contain in many cases considerable amounts of gravel. Core 25–09P is chosen as stratotype. (Fig. 2d). Its close relation to the presence of the Norwegian Current along the bottom indicates that it is formed by winnowing of glacial sediments by current action, a conclusion also reached by Holtedahl (1981) and Holtedahl & Bjerkli (1982).

## *Carbonate stratigraphy*

Generally there is a high carbonate content within Norway Basin Ooze and low content within most of Storegga Mud and Møre Diamicton (Fig. 2). In the latter unit the content is variable and values as high as 30 % are measured in sediments that contain less than 10 foraminifera per gram sediment. Most parts of Møre Diamicton contain chalk fragments of sand and gravel size, and Cretaceous coccoliths and foraminifera of the genera *Heterohelix* are common in the unit. We thus conclude that large quantities of Mesozoic limestone, glacially eroded from the continental shelf, have been included in Møre Diamicton. Elverhøi (1979) reached the same conclusion in a study of glacial diamictons off West Norway. 'Peak to peak' correlation of carbonate curves has been used to correlate stratigraphic levels in Norwegian Sea cores (Kellogg 1973, 1975, 1976, 1977, 1980). Based on biostratigraphical comparison with the results of Kellogg et al. (1978) and Streeter et al. (1982), it is evident that none of the cores in the present study penetrate isotope stage 4. The high carbonate samples found in Møre Diamicton are thus from somewhere in stages 2, 3 or 4. This result and also a carbonate minimum caused by early diagenetic processes (Jansen 1981) bear limitations on the use of bulk carbonate maxima and minima as a tool for correlation. Especially in areas close to the shelf one should be careful to test carbonate fluctuations against foraminiferal content to establish whether these are true productivity events or are caused by other processes.

## Benthic foraminiferal assemblage zones

### *Systematics*

Sejrup et al. (1981) studied the recent distribution of benthic foraminifera in the investigated area. We refer to this publication for taxonomic references. Belanger & Streeter (1980) and Streeter et al. (1982) practise a different nomenclature for two important species: *Cassidulina teretis* and *Oridorsalis tener*. Based on the similar recent distribution and on the descriptions we believe that these must be the same species as we have called *Cassidulina laevigata* and *Eponides umbonatus*. However, dissimilarities exist between our *C. laevigata* and the *C. laevigata* of some Holocene coastal deposits. *C. laevigata* from the investigated area often have shorter chambers not always extending across the umbilical area, in contrast to those described from the Oslofjord area which tend to have more elongated chambers (Feyling-Hanssen 1964). We thus feel that more taxonomic studies of the species should be undertaken until a definite nomenclature can be established, and that both *C. laevigata* and *C. teretis* are possible alternatives. *Cibicides wuellerstorfi* displays some variant forms. The typical large, flat type with bended sutures is the most frequent, but a smaller, hyaline variant is also present, especially in the lower part of the zone characterized by *C. wuellerstorfi*.

### *Zone division*

The benthic foraminiferal assemblages have been divided into 7 zones. These are defined by the dominant and/or characteristic species.

### *Zone A:* Elphidium excavatum *zone*

*Elphidium excavatum* dominates the faunas within Møre Diamicton, the lower parts of Storegga Mud and also in a few samples higher up in this unit. In most cases zone A does not contain taxa known to dominate the present fauna of the core location. Characteristic *Elphidium* fauna (Figs. 4–7) is: 50–80 % *Elphidium excavatum,* significant abundances of *Cassidulina reniforme*, *Bulimina marginata, Islandiella spp.*, different *Elphidiids* and species known from recent shallow water faunas of the continental shelf. The zone is poor in specimens, often less than 10 per gram sediment. The number increases towards the top where the *Elphidium* dominance is weakened and species known from recent faunas in the area (*Cassidulina laevigata, Eponides umbonatus,* and *Cibicides wuellerstorfi*) begin to constitute significant parts of the fauna. Very few planktonic foraminifera are found in association with zone A. They are mainly left-coiling *G.pachyderma.*

Whether the *Elpidium* fauna represents autochthonous or allochthonous faunas will govern the interpretation of the glacial bottom environment in the area. The common view is that *Elpidiids* are shallow water indicators and that whenever they are found in deep-sea sediments this is caused by resedimentation (e.g. Phleger et al. 1953, Streeter et al. 1982). This assumption is based on the genera's common occurrence in shallow water environments, but this is not a straightforward conclusion. Matoba (1976) reported observations of stained (i.e.living) *Elphidiids* from depths up to 2000 m off Japan. Culver & Buzas (1980) note that *E. excavatum* is not associated with *Ammonia beccarii* in recent deep-sea samples off NE America, whereas the two species are commonly associated in samples from the shelf, which could be an argument against *E. excavatum* interpreted as an allochthonous species. One of the authors of the present paper has also argued the *Elphidium excavatum* is an autochthonous element in the glacial deepsea sediments of the Norwegian Sea (Skarbø

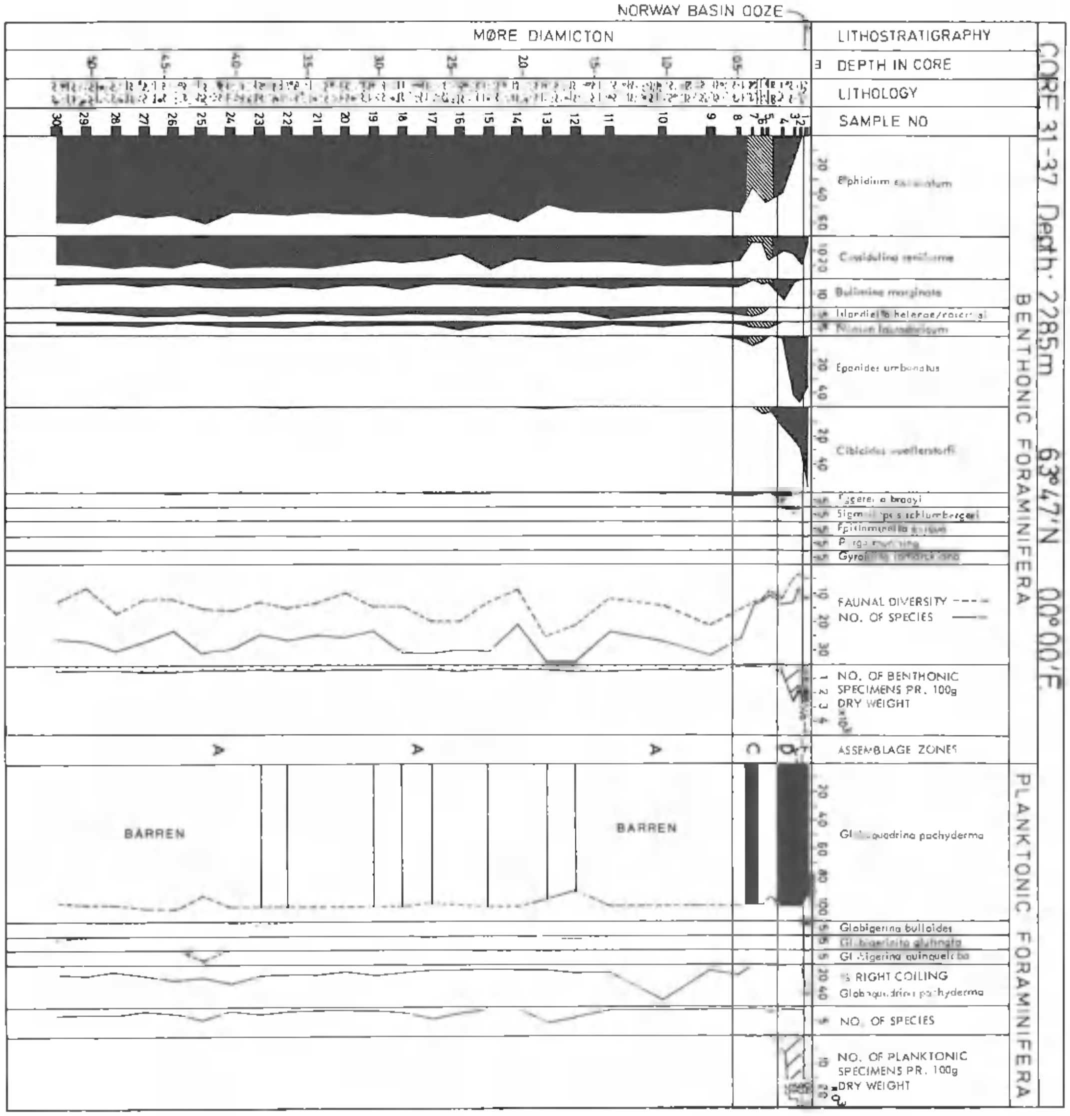

*Fig. 4.* Downcore foraminiferal variation and subdivision of benthic assemblage zones in core 31–37. Faunal diversity as defined by Sanders (1960). Hatched section indicates interval with less than 100 foraminifera per sample.

1980). However, the argument of Culver & Buzas may not be relevant if the resedimentation along NE America mainly included pre-Holocene sediments that do not contain *A. beccarii*, such as those reported by Vilks & Rashid (1976) and Vilks & Mudie (1978) from the shelf off Labrador and Nova Scotia.

We have no indications that *E. excavatum* lives in the southeastern Norwegian Sea today. The species occurs very sparsely and is only found in recent samples in areas with present erosion and redeposition (Sejrup et al. 1981). Neither have we found any modern analogue to the zone A assemblage. The zone displays greater diversity than overlying zones and definitely contains displaced elements such as thermophilic shelf species (*Trifarina angulosa* and *Bulimina marginata*) and pre-Quaternary species. Consequently it is difficult to distinguish possible autochthonous elements from the rest of the assemblage. Gla-

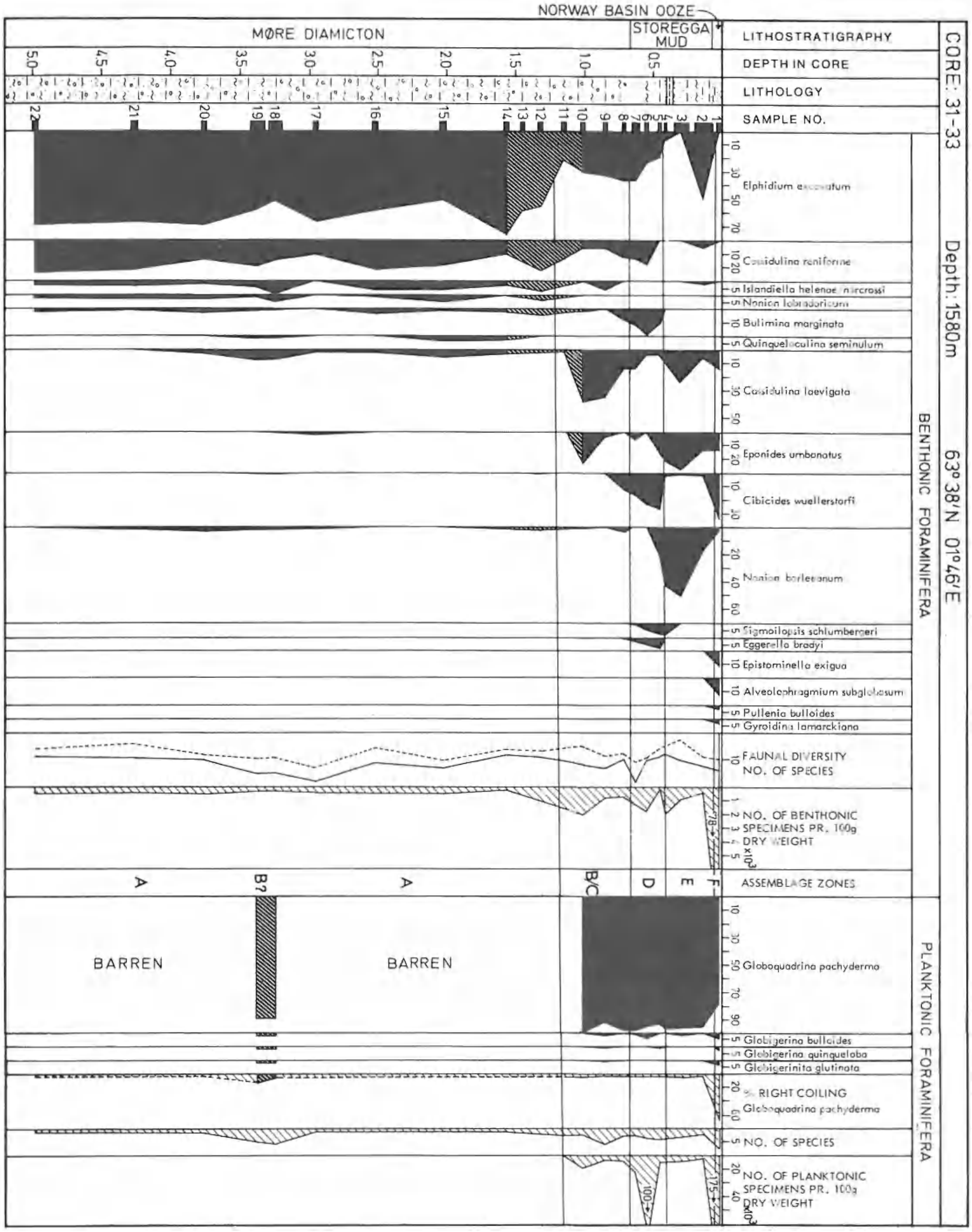

*Fig. 5.* Downcore foraminiferal variation and subdivision of benthic assemblage zones in core 31–33.

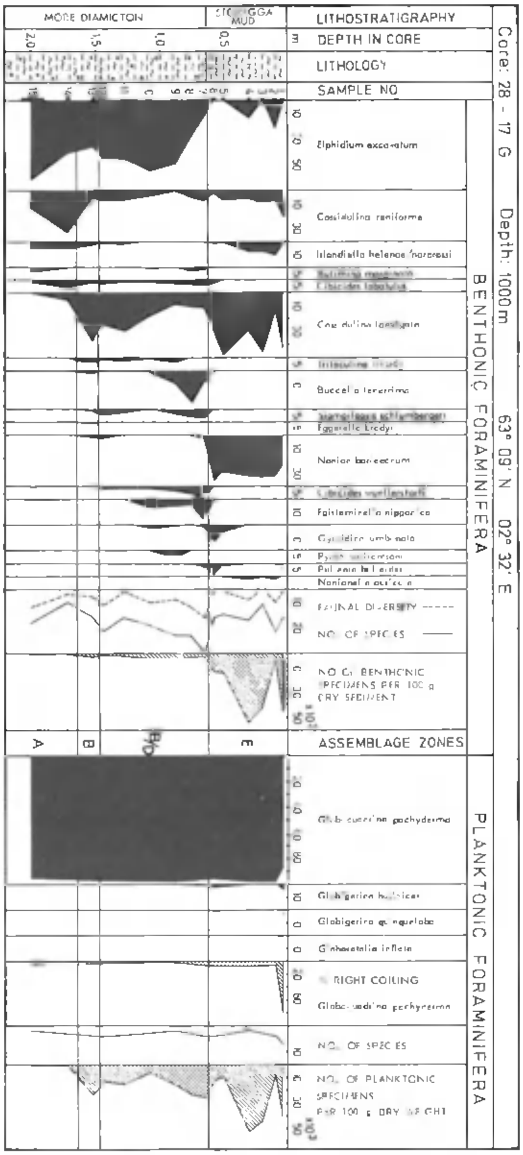

*Fig. 6.* Downcore foraminiferal variation and subdivision of benthic assemblage zones in core 28–17G.

ciomarine sediments from shelf areas bordering the investigated area contain *Elphidium* assemblages very similar to our findings (Morvik 1979, Norvik 1980, Godvik 1981). The shelf thus serves as a potential source area for retransported *Elphidium* assemblages in the deeper Norwegian Sea. The very even distribution over the whole Norway Basin of zone A faunas (Holtedahl 1959, Streeter et al. 1982) and the distinct "thinning out" tendency towards greater depth are probably best explained by ice rafting of shelf faunas.

The sediments of zone A indicate heavy ice rafting, high sedimentation rates and only weak bottom currents. Sedimentation rates were definitely higher than the average Weichselian sedimentation rate of 3 cm/kyr estimated by Kellogg et al. (1978) in cores from the central Norwegian Sea. We have estimated glacial sedimentation rates in core 31–33 to be larger than 10 cm/kyr. High input of ice rafted detritus documents a surface ocean favourable for iceberg drifting; thus the surface cannot have been tightly frozen through the year, as such conditions impede iceberg movement. Likewise the high sedimentation rate points against the existence of a continous ice-shelf covering the entire Norwegian Sea as proposed by Hughes et al. (1979) and Denton & Hughes (1981). Low sedimentation rates rather than high should be expected under large ice shelves (Orheim & Elverhøi 1981). Zone A sediments are almost barren of all planktonic organisms, indicating suppression of surface productivity by a low saline upper water layer produced by glacial meltwater. Total dominance of sinistral *G.pachyderma* indicates surface temperatures below 5–6°C, and no pronounced exchange with warmer, subpolar surface water from the Atlantic.

Bottom environments of zone A are difficult to interpret due to lack of reliable benthic autochthonous fauna. The absence of common deep-sea foraminifera may have been caused by reduced supply of oxygen due to impeded bottom water formation. There are, however, no indications that the Norwegian Sea became azoic, a conclusion also reached by Streeter et al. (1982). Another possibility is that the low surface productivity combined with strong dilution of nutrients by rapid sedimentation of terrigenous matter could have reduced nutrient levels at the bottom, below that required for benthic foraminiferal reproduction. A third possible factor is the temperature. Glacial bottom-water temperatures could definitely not have been much below the present -1°C. On the contrary, cessation of deep-water formation could have led to significantly higher temperatures, as suggested by Duplessy et al. (1975) and Duplessy et al. (1980), based on isotope ratios of deep-sea benthic foraminifera. Increased temperature should not be expected to restrain bottom life.

In conclusion we will argue that zone A indicates a polar environment with extensive ice-rafting, stratified upper water column, low sur-

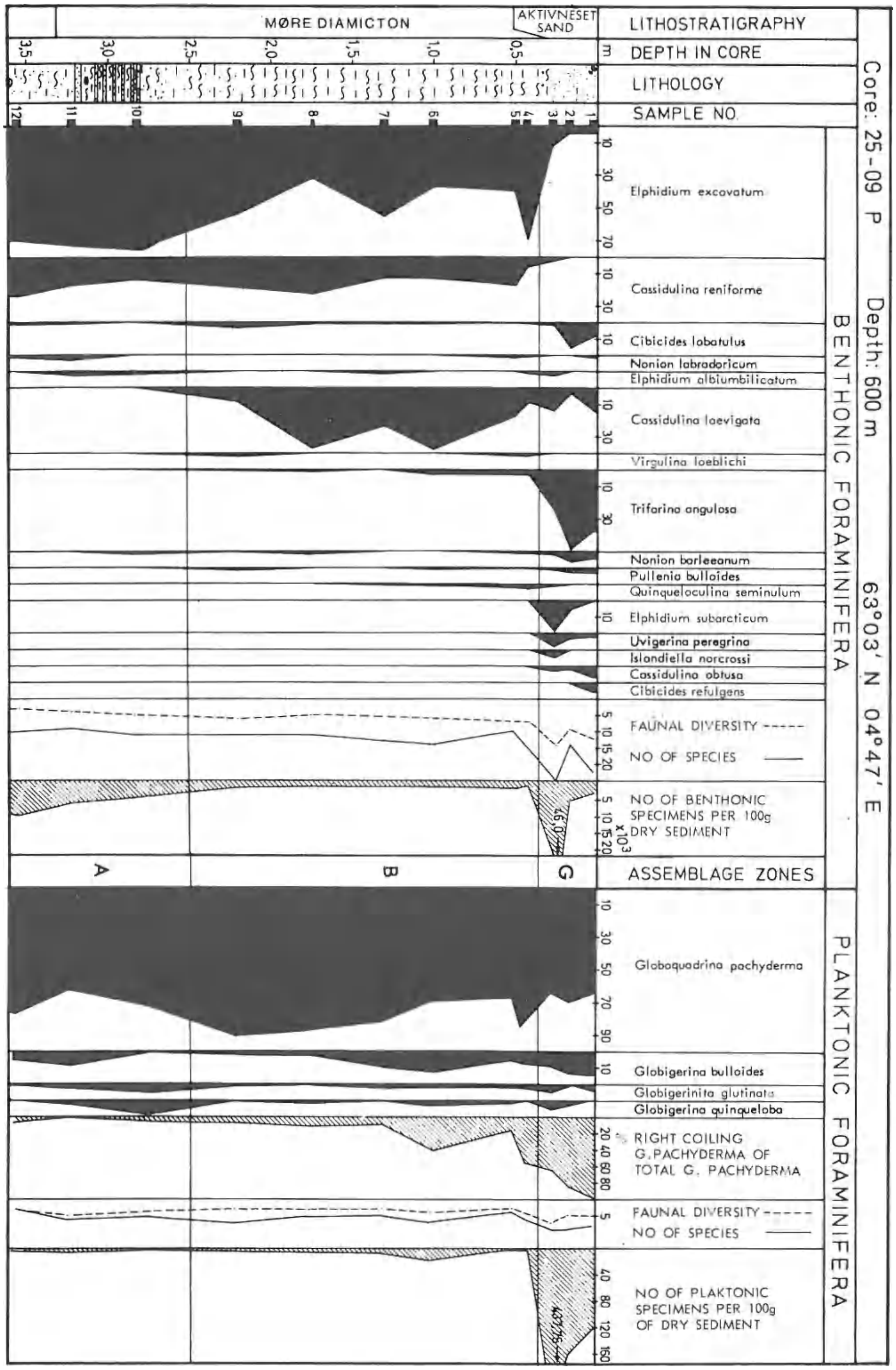

*Fig. 7.* Downcore foraminiferal variation and subdivision of benthic assemblage zones in core 25–09P.

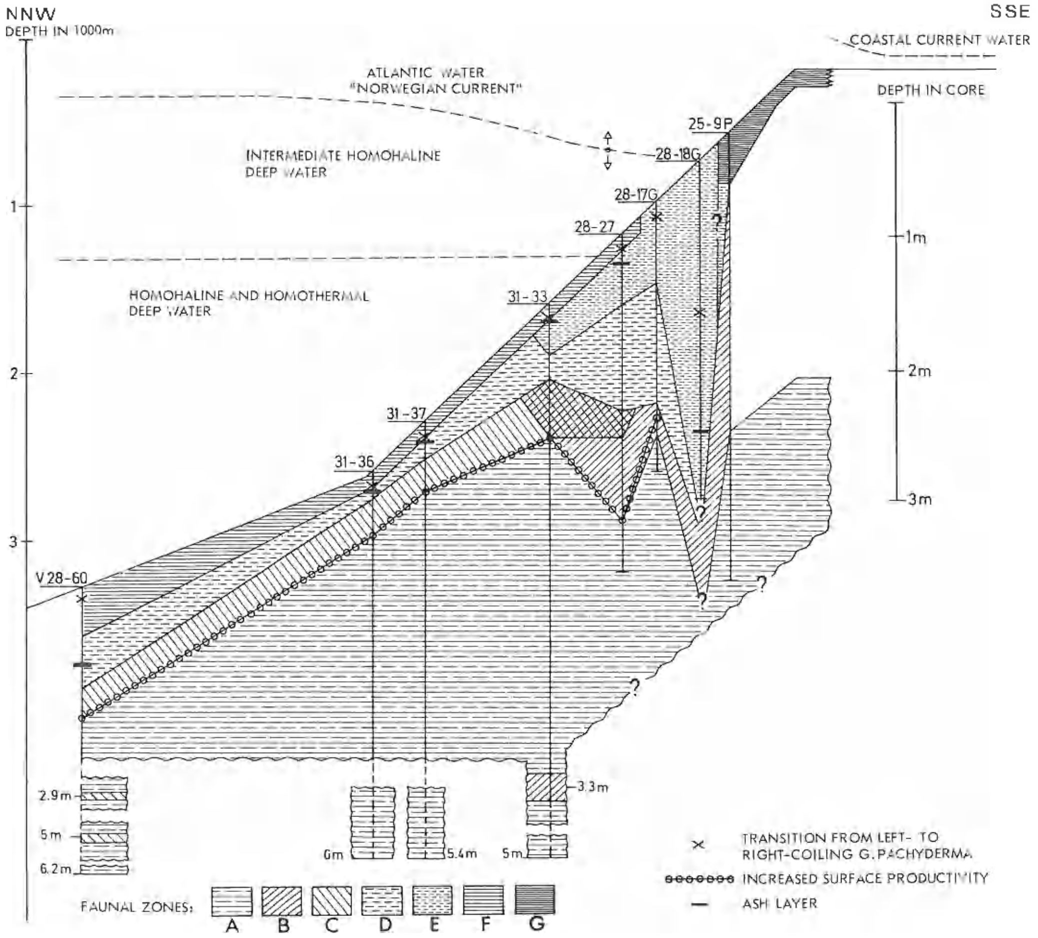

*Fig. 8.* Idealized transect across study area showing vertical and horizontal distribution of benthic faunal zones, planktonic foraminiferal events and ash-layer. The present hydrographic boundaries are indicated.

face productivity and only limited water exchange with the Atlantic. The bottom water was poorly renewed, low in nutrients and possibly also oxygen. Only twice in the zone is there evidence of more favourable conditions (indicated on Figs. 8, 10 & 11). These were small changes, but indicate increased surface productivity and some improvement of bottom conditions.

## *Zone B:* Cassidulina laevigata *zone*

From depths around 1600 m and to the upper slope zone A is succeeded by a fauna marked by a strong increase in *C.laevigata* content. *C.laevigata* and *E.excavatum* dominate, but the two species seem to occur in opposite phase to each other. If *E.excavatum* is excluded, *C.laevigata* shows a more homogenous appearance. Mesozoic foraminifera are also found in zone B, but appear more sporadically than in zone A. The content of planktonic foraminifera increases drastically with the transition from zone A to B, and is in the order of 10 to 100 times higher in the latter zone. The planktonic fauna is totally dominated by sinistral *G.pachyderma* (>95 %) (Figs. 4–7 and 10–12). Zone B is mainly found within the upper parts of Møre Diamicton, but may also be present in lower parts of Storegga Mud. Also in some shelf areas the transition from *Elphidium* assemblages is characterized by immigration of *C.laevigata:* outside Trøndelag (Løfaldli & Rokoengen 1980) and in the Norwegian Channel (Godvik 1981). The appearance of *C.laevigata* is thus a pronounced faunal signal with regional significance.

A fining upwards in the sediments of zone B indicates reduced ice-rafting frequency. The increase in planktonic foraminiferal content is a clear indication of increased surface productivity and thus reduced influence of glacial meltwater. Although ice-rafting persists through the zone, the increased surface productivity indicates at least seasonally ice-free waters and possible surface water exchange with the subpolar gyre system of the North Atlantic. Surface temperatures were low, deduced from the total dominance of sinistral *G.pachyderma,* and influence of warm Atlantic surface water must have been negligible. The planktonic increase is less marked in cores from above 1000 m, which indicates that glacial meltwater still dominated the eastern parts. The immigration of *C.laevigata* marks an environmental event that affected depths less than 1600 m on the continental slope and also parts of the neighbouring shelf areas. This species is found to dominate the present benthic foraminiferal fauna in the transitional water masses between the warm Norwegian Current water and the cold, homogenous deep-water on the continental slope off Norway. Its present deepest occurrence also fits well with the distribution of zone B (Belanger & Streeter 1980, Sejrup et al. 1981). The immigration of the species may thus imply the onset of a depth stratified water column in the deeper parts of the Norwegian Sea, renewal of bottom water on the slope, increased nutrient availability at the bottom and it marks the start of the transition from glacial to modern conditions along the Norwegian Continental Margin. In the Beaufort Sea, *C.laevigata* dominates the recent fauna related to presence of the Atlantic Undercurrent Water (T: 0–2°C) (Vilks et al. 1979). Zone B may thus also be interpreted as an indication of an undercurrent circulation similar to the present pattern in the Arctic Ocean.

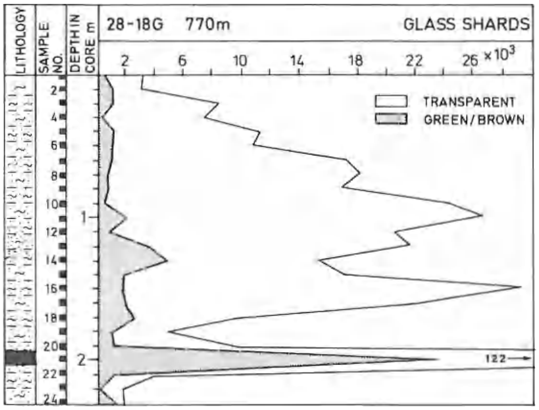

*Fig. 9.* Core 28–18G: Number of volcanic glass shards per gram dry sediment and position of ash-layer (dark area in lithology column).

### *Zone C:* Eponides umbonatus *zone*

This assemblage is found in cores deeper than 1200 m. The zone denotes the first shift to a fauna with significant elements of the present deep-water fauna. *E.excavatum* may still dominate, but to a smaller extent than in zone A, and the dominance decreases within zone C. *E.umbonatus* contributes significantly to the fauna with the highest content of the species found in the deepest cores. At depths between 1200 and 1600 m, zone B and C overlap and cover mostly the same lithostratigraphic sections (Figs. 8 & 11). Core 28-27 (1200 m) is an exception. At this depth *E.umbonatus* seems to immigrate later than *C.laevigata* (Fig. 9). Zone C is found in the upper part of Møre Diamicton but the same fauna is also present in two deeper intervals in core V28–60 (Fig. 8). The content of Mesozoic foraminifera is pronounced, but lower than in zone A. Concurrent with the transition from zone A to C, the content of planktonic foraminifera (sinistral *G.pachyderma* >95 %) increases by several orders of magnitude, comparable to the increase at the zone A/B transition. The immigration of *E.umbonatus* thus coincides with increased surface productivity and the same surface environment that was deduced for zone B also applies to zone C. *E.umbonatus* is the most common deep-sea benthic foraminifera in the present Norwegian/Greenland Sea and Arctic Ocean and does not seem to be connected with any specific water mass or environmental parameter (Lagoe 1977, Lohmann 1978, Belanger & Streeter 1980, Sejrup et al. 1981). The species is common up to 1200 m in recent sediments, which is the same shallow boundary as the boundary of zone C. In our material *E.umbonatus* shows positive correlation with planktonic foraminiferal abundance. However, it responds earlier to the late-glacial changes than does *C.wuellerstorfi.* This could imply that *E.umbonatus* is more adaptable to conditions with significant, but reduced surface productivity, and that *C.wueller-*

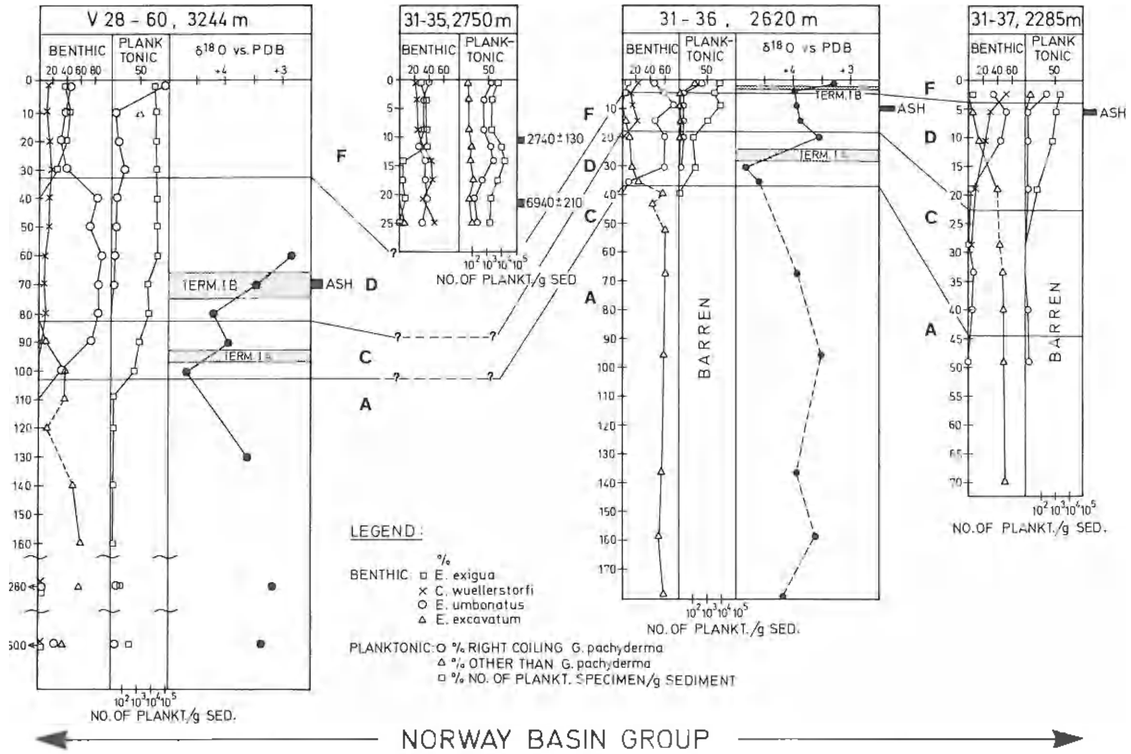

*Fig. 10.* Selected cores from the deep Norway Basin showing main foraminiferal signals, correlation of benthic assemblage zones, oxygen isotopic ratios of left-coiling *G.pachyderma*, ash layer position and $^{14}$C-dates. Vertical scale in cm.

*storfi* requires surface conditions with longer open, productive seasons that allow more nutrient transfer to the bottom and possibly also requires conditions with more rapid bottom water renewal.

In conclusion we will note that both zone B and C were formed under definitely glacial conditions with extensive ice-rafting, but different from zone A in the existence of seasonally open water and high surface productivity. The bottom environment experienced a drastic environmental change that affected both deep waters (zone C) and shallower areas (zone B). This event was probably caused by the introduction of water masses more similar to the present with respect to nutrients and oxygen content. This period also produced a benthic faunal boundary on the slope, reflecting a possible transition between deep waters and slope waters.

### *Zone D:* Cibicides wuellerstorfi *zone*

This assemblage is found in cores deeper than 1000 m. It is defined by high content of *C.wuellerstorfi* and also contains the arenaceous and porcellaneous species *Sigmoilopsis schlumbergeri* and *Eggerella bradyi*, which are very rare in surface sediments from the area (Sejrup et al. 1981). At depths less than 1300 m *Epistominella nipponica* is a significant member of the fauna. *E.excavatum* normally shows a drastic decrease and is often not present in the zone. The content of *E.umbonatus* and *C.laevigata* also generally decreases in zone D. Mesozoic foraminifera are rare. The content of planktonic foraminifera is high and the planktonic fauna is strongly dominated by left-coiling *G.pachyderma* (Figs. 4–7 and 10–12). Zone D covers the uppermost part of Møre Diamicton and/or the lowermost part of Storegga Mud. The sediments reflect decreasing ice-rafting intensities and the onset of down-

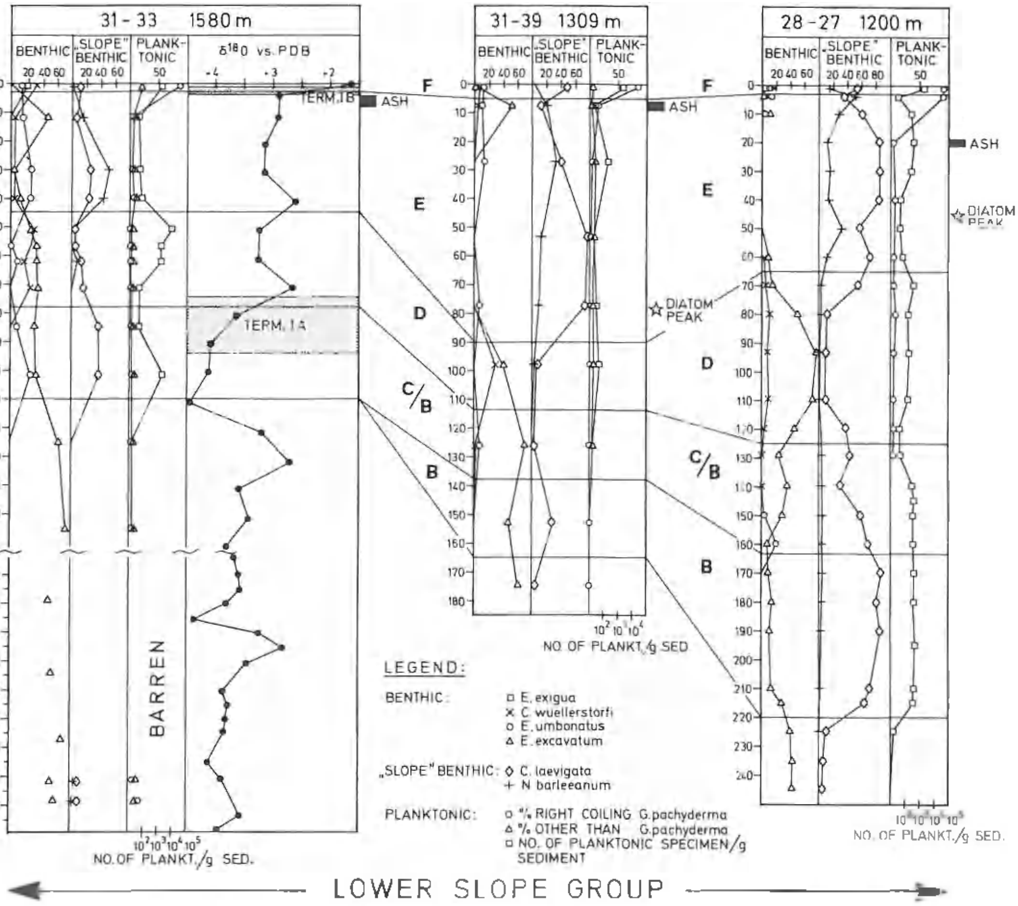

*Fig. 11.* Selected cores from the lower continental slope off Middle Norway, showing main foraminiferal signals, correlation of benthic assemblage zones, diatom peaks, oxygen isotopic ratios of left-coiling *G.pachyderma* and ash-layer. Vertical scale in cm.

slope mud transport produced by current erosion on the upper slope. The planktonic foraminiferal assemblage was strictly polar throughout the formation of zone D. However, surface productivity reaches a maximum, and the coincidence of *C.wuellerstorfi* with this productivity peak and thus with high nutrient supply to the bottom may indicate that feeding conditions are important regulators for the distribution of this species. Prolonged ice-free seasons reflect less meltwater influx, which may have enabled some bottom water formation. The deep-sea fauna is, however, not identical to that living under the present deep-water formation and overflow conditions. Several of the characteristic Holocene species are not present, and *E.bradyi* and *S.schlumbergeri* are recorded. These species, being rare in the present Norwegian Sea, are very common in the deep North Atlantic (Phleger et al. 1953, Caralp et al. 1970, Pujos-Lamy 1972). The deep water of the North Atlantic is about 3° warmer than the NSDW, and the presence of the two species in zone D may indicate a possible warmer watermass which may have originated from the North Atlantic. The boundary between *C.wuellerstorfi* (zone D) and *C.laevigata* (zone B) was situated at about 1000 m (Fig. 9), somewhat shallower than the present faunal boundary at 1200 m (Sejrup et al. 1981), suggesting that the transition between slope waters and deep waters was shallower during the formation of zone D than at present.

### *Zone E:* Nonion barleeanum *zone*

The assemblage is dominated by *N.barleeanum* and *C.laevigata* and is found in cores between 800 and 1600 m. Faunal diversity is lower than in the other zones. At depths greater than 1300 m, the zone also includes *E.umbonatus*. Zone E succeeds zone B and D depending on depth (Fig. 8). Between 1200 and 1600 m the zone is overlain by zone F, while it constitutes the top assemblage at depths between 800 and 1200 m. This distribution coincides with the present and past distribution of Storegga Mud, and zone E seems to be closely related to the deposition of this sediment. Mesozoic foraminifera occur very seldom in zone E. The content of planktonic foraminifera associated with the zone is lower than in zone D (see core 31–33, Figs. 5 & 11). In cores from 800 to 1200 m there are tendencies towards increasing amounts near the top of the zone. At the same depth interval there is a change from dominance of left-coiling to dominance of right-coiling *G.pachyderma* within zone E. In cores situated deeper only planktonic assemblages dominated by left-coiling *G.pachyderma* are associated with zone E. High content of *G.pachyderma* (right) is associated with the species: (5 % to 30 % abundance) *Globigerina bulloides, G.quinqueloba, Globigerinita glutinata, Globigerina falconensis* and *Globorotalia inflata*.

Zone E is strongly related to its sedimentary habitat, and follows the retreat of Storegga Mud deposition upslope. All of the zone below 1200 m and the lower part between 1200 and 600–800 m were formed while polar surface conditions still prevailed. We record a reduced planktonic content, but ascribe this to dilution by increased terrigenous deposition. At depths shallower than 1200 m the zone continued to form after the surface warmed (change to right-coiling *G.pachyderma*) and the zone here constitutes the surface sediment assemblage. Both *N.barleeanum* and *C.laevigata* seem to adapt well to conditions with rapid pelitic sedimentation, and to a smaller extent this also applies to *E.umbonatus*. On the other hand *C.wuellerstorfi* disappeared during zone E formation, and did not reimmigrate until the rapid mud deposition had ceased. Zone E is thus interpreted as a faunal response to the introduction of a constant current regime on the upper slope with the capacity of eroding glaciomarine sediments and causing down-slope fine-grained transport. The onset of this regime was before the surface water warmed to the present level.

### *Zone F:* Epistominella exigua *zone*

The zone is characterized by *E.exigua* and forms the top assemblage in cores deeper than 1200 m. Other diagnostic species are: *Pullenia bulloides* (depths between 1200 and 1600 m), *Gyroidina lamarckiana, Triloculina frigida, Alveolophragmium subglobosum* and *Pyrgo murrhina*. The zone contains significant amounts of *C.laevigata* (depths less than 1600 m), *E.umbonatus* and *C.wuellerstorfi*. The *E.exigua* content is variable and increases with increasing depth. There is also an increase in content in the upper part of the zone (core 31–35, Fig. 10). In the deepest core (V28–60) *E.exigua* arrives later than the other characteristic species of the assemblage. In the other cores the immigration of the different species seems to be simultaneous. We have not found Mesozoic foraminifera in this zone. The content of planktonic foraminifera is very high in samples from the zone. The planktonic assemblage is dominated by right-coiling *G.pachyderma* and the additional species mentioned under zone E. On the continental slope and in the Eastern Norway Basin, the shift from left- to right-coiling *G.pachyderma* occurs simultaneously with the benthic transition from zone D or E to F. In core V28–60, however, the shift occurs *within* zone F.

Zone F is identical to the Holocene deep-sea assemblage in the Norwegian Sea and is only found within Norway Basin Ooze. Planktonic foraminifera indicate warm, productive surface waters during the formation of the zone on the lower slope and in the eastern deep basin. In more western parts of the Norwegian Sea the formation of zone F started while polar surface conditions still prevailed. The immigration of *E.exigua* occurred in a step-wise manner with an increase in content within the zone (at about 3000 BP, Fig. 10). This may indicate that *E.exigua* requires very stable bottom conditions and that the bottom environment did not stabilize before well into the Holocene. In the Atlantic there seems to be some relation between this species and presence of the oxygen-rich NADW (Lohmann 1978, Schnitker 1979, Streeter & Shackleton 1979), but results from the Indian Ocean do not relate it to specific oxygen-rich water masses (Corliss 1979). It is tempting to interpret occurence of *E.exigua* as indicative of bottom-water formation and overflow conditions, but this is not without problems. Both Kellogg (1980) and Steeter et al. (1982) regard the circu-

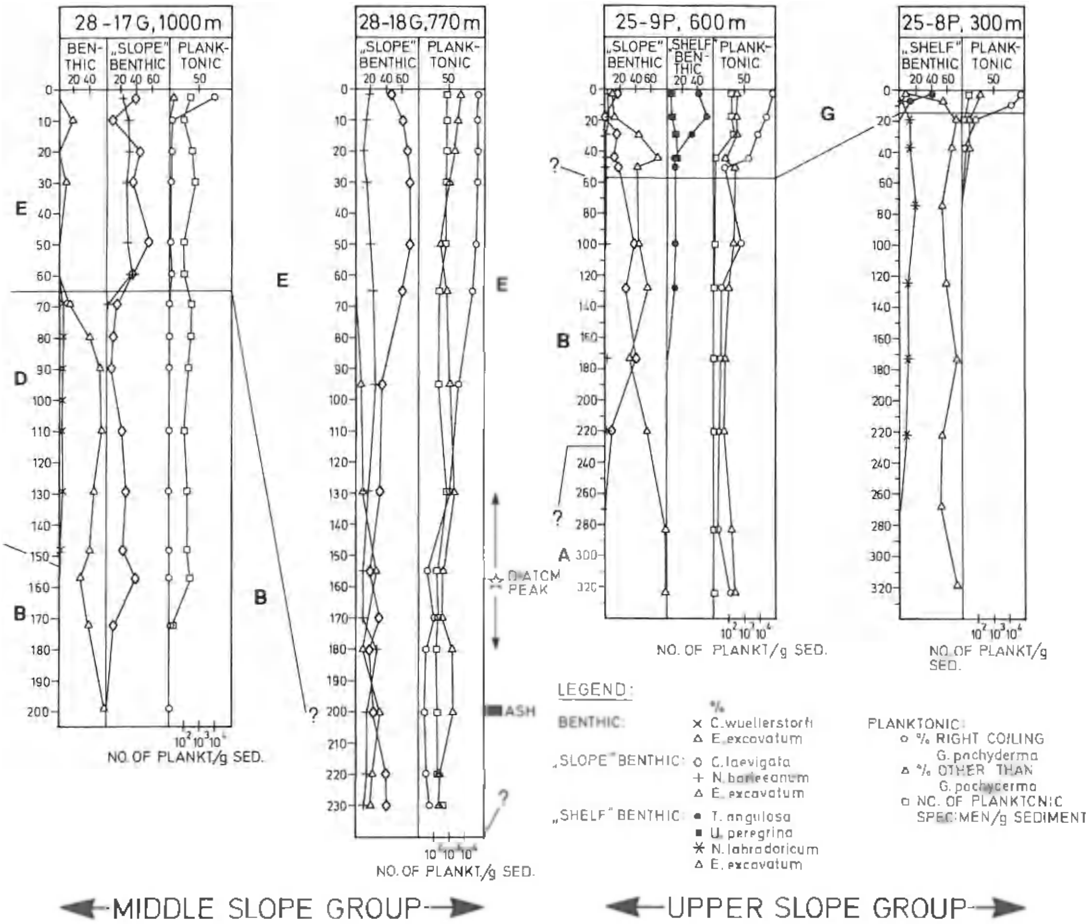

*Fig. 12.* Selected cores from the middle and upper continental slope off Middle Norway, showing main foraminiferal signals, correlation of benthic assemblage zones, diatom peak, and ash-layer. Vertical scale in cm.

lation pattern of stage 5e and Holocene as similar. However, *E. exigua* was not present in significant amounts during stage 5e, although Streeter et al. (1982) describe the benthic faunas of stage 5e and the Holocene as similar. The bottom water of the Arctic Ocean and the Norwegian Sea is identical with respect to temperature and salinity (Coachman & Aagaard 1974), yet *E. exigua* is not found in Arctic Ocean surface samples (Green 1960, Lagoe 1977). This should, together with the variable occurrence of the species with depth, imply that *E. exigua* responds to a complex set of environmental conditions rather than just to presence of newly formed bottom water. Zone F is indicative of Holocene conditions with warm, productive surface water in the east, but the exact environmental patterns that govern the formation of the assemblage are not well understood.

### *Zone G:* Trifarina angulosa *zone*

This assemblage constitutes the top fauna at depths less than 800 m. It is characterized by *T. angulosa* and contains in addition the common species *Cibicides lobatulus, Cassidulina obtusa, Uvigerina peregrina*, and *C. laevigata*. The assemblage is only found in Aktivneset Sand. Mesozoic foraminifera are normally absent. The planktonic foraminiferal content associated with zone G is very high. The planktonic assemblage is dominated by right-coiling *G. pachyderma* and the transition from left- to right-coiling *G. pachyderma* occurs simultaneously with the transition

*Table 2.* Selected $^{14}$C-dates

| | $^{14}$C age BP | M = Molluscs<br>F = Foraminifera<br>O = Ooze | Location | Benthic foram. fauna | Reference |
|---|---|---|---|---|---|
| 2a: Dates from study area | 2740 ± 130 (T–3693) | O | Core 31–35 | *E.exigua* | E |
| | 6940 ± 210 (T–3694) | O | Core 31–35 | *E.exigua* | E |
| | 2270 ± 170 (T–2383) | F | Core 25–11 | *N.barleeanum* | B |
| | 5010 ± 260 (T–2382) | F | Core 25–11 | *N.barleeanum* | B |
| | 10040 ±730 (T–2381) | F | Core 25–11 | *N.barleeanum* | B |
| 2b: Dates from shelf areas | 9450 ±340 (T–2488) | M | Off Troms N-Norway | *T.angulosa* | A |
| | 9810 ± 180 (T–2859) | M | Off Nordfjord W-Norway | *T.angulosa* | C |
| | 9920 ±110 (T–3236) | M | Off Trøndelag Middle Norway | *T.angulosa* | D |
| | 11060 ±210 (T–2860) | M | Off Nordfjord | *N.labradoricum* | C |
| | 11230 ±100 (T–2929) | M | Off Trøndelag | *C.lobatulus*<br>*C.laevigata* | D |
| | 11770 ± 110 (T–3237) | M | Off Trøndelag | *C.lobatulus*<br>*C.laevigata* | D |
| | 12390 ± (T–2928) | M | Off Trøndelag | *C.lobatulus*<br>*C.laevigata* | D |
| | 12570 ±260 | M | Off Nordfjord | *N.labradoricum* | C |
| | 13350 ± 340 (T–2708) | M | Off Nordfjord | *E.excavatum*<br>*C.reniforme* | C |

**References:*
A: Vorren et al. (1978)
B: Bjørklund et al. (1979)
C: Holtedahl & Bjerkli (1981)
D: Løfaldli & Rokoengen (1980)
E: This paper

from underlying benthic zones to zone G (Figs. 7, 8 & 12). The distribution of zone G coincides with the areal extent of the Norwegian Current along the bottom. Planktonic foraminifera indicate relatively warm, postglacial conditions and the benthic assemblage, being boreal-lusitanean, also indicates relatively warm conditions. We thus reach the same conclusion as Sejrup et al. (1980) that this benthic assemblage denotes the latest warming of the shallow-water areas outside Western Norway produced by the warm Norwegian Current.

## Time control

The chronostratigraphic framework is based on several parameters. Available $^{14}$C-dates are compared with biozones and isotope stragigraphy. An ash layer serves as a time marker. A summary of the proposed ages for the faunal zones is presented in Table 4.

*$^{14}$C-dates.* – The strong terrigenous dilution gives in most cases insufficient carbonate for dating bulk samples. Contamination by old carbonate in Møre Diamicton also invalidates possible $^{14}$C-dates in this unit. The most useful $^{14}$C-dates are those performed on molluscs. Unfortunately, significant amounts of dateable molluscan carbonate are only found in cores from the shelf area (Table 2). Only 5 $^{14}$C-dates are available from the continental slope and the deep basin. Three dates were published by Bjørklund et al. (1979) from core 25–11 (Location on Fig. 1), all within zone E.

The oldest dated sample (10,040 ± 740 BP) is taken close to the shift from left- to right-coiling *G.pachyderma*. Due to the large standard deviation and a possible hiatus in this core, the date only roughly places the planktonic shift to about 10,000 BP. Two dates were obtained from Norway Basin Ooze (2740 ± 130 BP & 6940 ± 210 BP). Calculated from sedimentation rate extrapolation these dates appear to be at least 1000 years too young, possibly due to bioturbation processes. If we extrapolate to the bottom of the core, and include a possible error of 1000 years, we date this to minimum 9300 BP. As the core does not penetrate the polar/subpolar planktonic transition, this should be considered as being the *minimum* age for this event in core 31–35 (Fig. 10). The coarse fraction of a sample from zone E in Storegga Mud was dated to 14,900 ± 270 (Fjæran 1980), which is considered too old compared with other means of dating (ash-layer and isotope stratigraphy).

Faunal correlation to shelf cores which contain dateable molluscan shell provides a means for dating the faunal zones in the investigated area. The lower boundary of *T.angulosa* fauna (corresponding with zone G) has been dated to 9920 BP and 9810 BP outside Middle Norway and 9450 BP outside Troms, N-Norway (Table 2B). Dated molluscs in *Elphidium* fauna outside West Norway gave an age of 13,350 ± 340 BP (Table 2b). In nearshore and bank cores *Elphidium* assemblages are succeeded by *N.labradoricum* dominated assemblages, while *C.laevigata* and *Cibicides lobatulus* dominate the assemblages that succeed *Elphidium excavatum* in deeper parts of the shelf. Dates from *N.labradoricum* assemblages have given ages between 11,060 BP and 12,570 BP, and dates from sediments containing *C.laevigata* and *C.lobatulus* give ages between 11,230 BP and 12,390 BP (Table 2b). These ages indicate that the *Elphidium* fauna was replaced outside Western and Middle Norway about 13,000 BP.

*Ash-layer.* – A very distinct horizon of volcanic ash is observed in many cores (Figs. 8, 10, 11, 12). The ash-layer is usually a few cm thick and is little disturbed by bioturbation. Glass shards are very conspicuous in the layer and consist of two different types: One which constitutes about 75 % of the total is platy, thin and transparent; and another less frequent bubbly, green-brown variant (Fig. 9). Counts in core 31–33 gave 1450 shards per gram sediment in the fraction >125 μm. Similar ash-layers have been reported from the Norwegian Channel (Godvik 1981), the outer part of the Sognefjord (Seland 1981, Stensland 1982) and in a large number of lacustrine and raised marine sediments from Møre, West Norway (Mangerud et al. in press). The latter authors name their ash Vedde Ash Bed, and have dated it to 10,600 BP. This age corresponds well to the age indicated from shelf cores, as the layer here is situated slightly beneath the faunal boundary dated to 9900 BP. Ash from core 31–33 has a geochemical composition identical to the Vedde Ash, and origins from Iceland (Mangerud et al. in press). Mangerud et al. also demonstrate that the North Atlantic Ash Zone 1 dated to ca. 9800 BP (Duplessy et al. 1981) contains shards from two eruptions, and that one is similar to the Vedde Ash.

The planktonic transition from polar to subpolar assemblages always occurs above the ash-layer (Figs. 8, 10, 11, 12). On the slope and in the eastern basin there seems to be a short duration between these events, whereas there appears to have been a significant time lag between ash deposition and planktonic change at the more westerly location of core V28–60 (Fig. 10). The ash-layer also precedes zone F, but not much. Together with the extrapolated age in core 31–35, the ash-layer pin-points the concurrent transition to benthic zone F and a subpolar planktonic assemblage on the lower slope to about 10,000 BP. Zone E is found both below and above the ash, and the formation of this zone and hence also of, Storegga Mud started before middle Younger Dryas. On the deeper slope, the deposition of Storegga Mud had ceased by the time of ash deposition. Both zone B and C occurred before the ash deposition, while zone D started to form below the ash, and the ash-layer is often found near the top of this zone. Also ice rafting over the area had ceased at the time of ash deposition.

*Isotope stratigraphy.* – Oxygen-isotope values in δ-notation related to PDB for left-coiling *G.pachyderma* are measured in three cores (Table 3 and Figs. 10 & 11). The transition from glacial (heavy) values to Holocene (light) values occurs in a step-wise manner: First a decrease in $\delta^{18}O$ (Termination Ia, varying between 0.8 and 1.8 per mil.). This is followed by a smaller increase in $\delta^{18}O$. In cores 31–33 and 31–36 the isotopic values then remain quite constant before the final decrease to Holocene values (Termination Ib). However, the relatively constant plateau is about 1‰ heavier in core 31–36 than in 31–33. This may indicate more deglacial meltwater influence affecting the upper water column at the position of core 31–33 which is situated closer to the coast. A similar two step transition is recorded and $^{14}C$-dated in core CH 73139C, collected west of Ireland. This record indicates that Termination Ia started at about 16,000 BP and finished at

*Table 3.* Oxygen-isotope ratios from *G. Pachyderma* (left-coiling)

| Core | Depth (cm) | $\delta^{18}$O vs. PDB |
|---|---|---|
| 31–33 | 0–2 | + 1.64 |
| | 3–5 | + 2.89 |
| | 10–13 | + 2.90 |
| | 20–23 | + 3.13 |
| | 30–33 | + 3.14 |
| | 40–43 | + 2.61 |
| | 50–53 | + 3.23 |
| | 60–63 | + 3.26 |
| | 70–73 | + 2.66 |
| | 80–83 | + 3.64 |
| | 90–93 | + 4.09 |
| | 100–104 | + 4.08 |
| | 110–114 | + 4.44 |
| | 120–124 | + 3.18 |
| | 130–134 | + 2.62 |
| | 140–144 | + 3.56 |
| | 150–154 | + 3.42 |
| | 160–164 | + 3.80 |
| | 170–174 | + 3.60 |
| | 180–184 | + 3.59 |
| | 190–194 | + 3.55 |
| | 200–204 | + 3.80 |
| | 210–214 | + 4.33 |
| | 220–224 | + 3.24 |
| | 230–234 | + 2.82 |
| | 240–244 | + 3.68 |
| | 260–264 | + 3.87 |
| | 270–274 | + 3.77 |
| | 280–284 | + 3.80 |
| | 290–294 | + 3.83 |
| | 310–314 | + 4.09 |
| | 320–324 | + 3.87 |
| | 350–360 | + 3.26 |
| | 360–364 | + 3.94 |

| Core | Depth (cm) | $\delta^{18}$O vs. PDB |
|---|---|---|
| 31–36 | 0–2 | + 3.32 |
| | 3–4 | + 4.00 |
| | 8–10 | + 3.97 |
| | 13–15 | + 3.90 |
| | 17–19 | + 3.54 |
| | 29–31 | + 4.88 |
| | 34–35 | + 4.62 |
| | 65–67 | + 3.99 |
| | 95–96 | + 3.51 |
| | 135–138 | + 3.94 |
| | 155–159 | + 3.79 |
| | 175–179 | + 4.16 |

| Core | Depth (cm) | $\delta^{18}$O PDB |
|---|---|---|
| V 28–60 | 60 | + 2.84 |
| | 70 | + 3.43 |
| | 80 | + 4.21 |
| | 90 | + 3.92 |
| | 100 | + 4.69 |
| | 130 + 140 | + 3.59 |
| | 280 | + 3.15 |
| | 500 | + 3.39 |

*Table 4.* Estimated age for benthic zones and planktonic faunal events

| | Age | Method* |
|---|---|---|
| Zone A: | >~ 13.000 BP | B, D |
| Zone B: | ~ 13–~ 11,000 BP | B, C, D |
| Zone C: | ~ 13–~ 12,000 BP | B, D |
| Zone D: | ~ 12–10,000 BP | C, D, E |
| Zone E: | ~ 11,000 BP – Present | A, C, D |
| Zone F: | 10,000 BP – Present | A, B, C, D |
| Zone G: | 10,000 BP – Present | |
| Planktonic productivity increase: | | |
| | ~ 13,000 BP | B, D |
| Change from left-<br>*G.pachyderma:* | to right-coiling<br>10,000 BP (eastern area) | A, C, D |

*Method used for age determination:
A: $^{14}$C-dated.
B: Faunal correlation to $^{14}$C-dated cores outside study area.
C: Ash layer.
D: Oxygen-isotope curve.
E: Interpolation.

13,300 ± 700 BP, and that Termination Ib started ca. 10,000 BP and ended about 6300 BP (Duplessy et al. 1981). This core is situated close to the entrance to the Norwegian sea, and the dating of this isotopic record should thus give an approximate age for the similar features in the southern Norwegian Sea.

Termination Ia roughly coincides with the commencement of benthic zones B and C (Figs. 10 & 11). The change from zone A to B and C, and the corresponding planktonic productivity increase can thus be timed to around 13,000 BP or shortly before, and seems to have occurred simultaneously on both the slope and in the deep basin. Termination Ib is situated slightly above the ash-layer, indicating that the age proposed by Duplessy et al. (1981) for Termination Ib agrees well with the age of 10,600 BP obtained for the ash-layer. Zone D was established before both Termination Ib and the ash-layer, suggesting that the zone may have been initiated at about Allerød time. However, some time-transgressive movement involved with the formation of the zone cannot be outruled. In core 31–33 where both zone E and Storegga Mud is present, we note that the plateau of isotopic values between Termination Ia and Ib is extended. This supports our interpretation that the deposition of Storegga Mud took place rapidly (Fig. 11), probably at the beginning of Y. Dryas Chronozone. The isotope results also indicate that the transition to dominance of right-coiling *G.pachyderma* occurred shortly after 10,000 BP on the slope and in the eastern deep basin, and significantly later in core V28–60 in the central deep-basin.

The isotopic record of core 31–33 indicates that the core does not penetrate the isotopic stage 4/5 boundary (ca. 73,000 BP). Two peaks of moderately light values can be distinguished (130 cm and 240 cm levels), but the foraminifera in these samples do not correspond with the benthic and planktonic assemblages recorded from stage 5 by Kellogg et al. (1978) and Streeter et al. (1982). The values recorded in the deeper part of our cores are on the average lighter than those obtained from stages 2, 3 and 4 by Kellogg et al. (1978). Due to the very few foraminifera available for analysis in many samples from the deeper parts of the cores, a variable number of juvenile specimens had to be included. As juvenile planktonic foraminifera normally crystallize their carbonate nearer the surface than adult specimens (Bé & Tolderlund 1971), this could lead to more meltwater influenced isotopic composition. Also retransportation of foraminifera from shelf areas could contribute with low $\delta^{18}O$ values to the samples.

## Diatom peaks

Samples containing abnormally high numbers of diatoms have been recorded in cores from 1300 m to 770 m depth. The diatom peaks are very pronounced and usually appear in short depth intervals (2–40 cm). The broadest zone was observed in core 28–18G (Fig. 12), which is believed to have experienced very high sedimentation rates. Diatom abundance is as high as 3800/g sediment, while diatom abundances in other sections of the cores are usually lower than 100/g sediment. The diatom flora is identical in all cores. *Thalassiosira gravida* (70–80 % of total) and *Thalassiotrix longissima* are the most prominent members of the flora. The diatom peaks are found in Storegga Mud, benthic zone E and both below (28–27 & 31–39) and above (28–18G) the ash-layer (Figs. 11 and 12). This may indicate a time transgressive movement of high diatom production, or it may indicate more than one event of high primary productivity over the investigated area.

A similar peak was described by Bjørklund et al. (1979) from core 25–11 (1000 m) (Location on Fig. 1). Biostratigraphical investigations indicate coherency with the peak in core 28–18G. Also diatom peaks slightly above the ash-layer in Norwegian Channel cores (Godvik 1981) indicate presence of high-productive waters off western Norway about 10,600 BP (age of ash-layer). The diatom flora indicates cold-water environments similar to those of the present Labrador Sea, close to the polar front (Schrader, pers.comm., Bjørklund et al. 1979). It is thus likely that the diatom peak denotes the boundary between polar surface waters of the Norwegian Sea, and strong intrusion of subpolar Atlantic surface water.

## Correlation

### *North Atlantic record*

During the period 20,000 to 13,000 BP the oceanic polar front, as defined by polar/subpolar planktonic foraminifera, was located far south in the North Atlantic. At about 13,000 BP the polar front abruptly retreated northwestwards, and warmer, temperate surface waters entered the northeast North Atlantic. Between 13,000 and 11,000 BP the polar front was situated southwest of Iceland and the Northeast Atlantic surface waters became nearly as warm as today (Ruddiman & McIntyre 1981, Duplessy et al. 1981). This configuration could have allowed relatively warm surface waters to enter the southeastern Norwegian Sea through the Iceland/Scotland passages. Such early deglacial warming of the southeastern Norwegian Sea has been proposed by Ruddiman & McIntyre, based on the conclusions of Mangerud (1977) that littoral and sublittoral faunas from the Bølling and Allerød Chronozones in western Norway indicate presence of warm Atlantic water along the Norwegian coast. However, by using the same criteria as those used for the North Atlantic when defining the polar front (left- to right-coiling *G.pachyderma*), we can clearly demonstrate that the oceanic polar front was *not* situated in the Norwegian Sea until after 10,000 BP, which indicates a lag of about 3000 yrs between the initial warming of the NE-Atlantic and the Norwegian Sea.

The warm North Atlantic surface prevailed until 11,000 BP when a substantial readvance of the polar front is recorded by Duplessy et al. (1981) and Ruddiman & McIntyre (1981). Not until after 10,000 BP did warm surface waters invade the Northeast Atlantic again. This severe and rapid cooling corresponds to the Y. Dryas cooling known from NW European continental records. In the Norwegian Sea we are not able to define this cooling by planktonic foraminifera, as polar assemblages dominated both the Y. Dryas and previous periods.

Benthic foraminiferal faunas both in the Norwegian Sea and the North Atlantic show marked changes in response to the last glacial cycle (Streeter & Shackleton 1979, Streeter et al. 1982). Both the present study and Schnitker (1979) demonstrate extensive and rapid faunal changes in the late glacial deep-sea environment of the two ocean basins. It seems evident that the North Atlantic did not receive deep water from the Norwegian Sea during the period 24,000 to 12,500 BP (Schnitker 1979). Large parts of the North Atlantic sea floor were inhabited by a fauna indicating presence of Antarctic Deep Water (AADW), and some bottom water could have been formed within the North Atlantic itself (Schnitker 1979, Duplessy et al. 1980). Between 12,500 and 9000 BP the western North Atlantic started to receive larger quantities of newly formed deep water as recorded by a possible initiation of the Western Boundary Undercurrent off North America. However, the modern extent and strength of this current was not reached until after 9000 BP (Schnitker 1979, Balsam 1981). With the problems of interpreting benthic deep-sea records and the problem of dating deep-sea sediments in mind, we note a considerable consistency between the Norwegian sea and the North Atlantic record: Only after warm surface waters of the North Atlantic Drift had invaded the Norwegian Sea (10,000 BP) is there evidence of large scale overflows in the North Atlantic (Western Boundary Undercurrent). The transitional period of the Western Boundary Undercurrent between 12,500 and 9000 BP agrees with the transitional period between 13,000 and 10,000 BP in the Norwegian Sea and may reflect bottom water formation and overflow on a smaller scale, or deep-water formation in the North Atlantic itself. At the present we are not able to distinguish between these two deep-water sources, and one must seriously consider the possibility of deep- and intermediate waters entering the Norwegian Sea *from* the North Atlantic during this period, i.e. a circulation opposite to the present. The onset of extensive overflows from the Norwegian Sea is also reflected by increased bottom transport and sedimentation on the Reykjanes Ridge (Ruddiman & Bowles 1976). They date this event to 11,000 BP by extrapolation from $^{14}$C-dates. We feel that this is only an approximate age, the true age possibly being younger due to incorporation of old carbonate from retransported glacigenic material (Durazzi et al. 1980).

### *Coastal and shelf areas*

The margin of the Scandinavian Ice Sheet was possibly situated on the shelf break outside western Norway at about 13,000 BP (Rokoengen 1980). After this time the ice margin retreated rapidly towards the coast, coincidently with the seasonal opening of Norwegian Sea surface waters. The outer coastline of the Møre area was deglaciated not later than 12,400 BP (Mangerud et al. 1981). Once the ice margin had retreated to the coast, the density difference between glacial meltwater and marine waters could have produced a northward flowing coastal current of low-salinity waters which entrained icebergs in a coastal northern drift route and diminished icerafting over the deep Norway Basin. The glacial retreat was followed by a succession of benthic foraminiferal faunas on the shelf and in coastal areas; near-glacial *E.excavatum* and *C.reniforme* faunas were followed by *N.labradoricum* and *C.laevigata* (Morvik 1978, Norvik 1980, Godvik 1981, Seland 1981), reflecting increased influence of marine waters in coastal areas during the period 13,000–11,000 BP. Benthic foraminiferal assemblages do not indicate environments similar to the present boreal/lusitanean conditions; they consist of relatively diverse faunas with arctic and cosmopolitan elements. This agrees with the oceanic record.

Based on littoral and sublittoral mollusc-faunas from raised marine sediments at Ågotnes and other localities outside Bergen, Mangerud (1977) concluded that the oceanic polar front was situated north in the Norwegian Sea and that warm, Atlantic water was present outside Western Norway during the Bølling/Allerød chronozones. We have earlier demonstrated that the conventional way of defining the oceanic polar front (i.e. polar/subpolar planktonic foraminifera) indicates a polar front position south of the Norwegian Sea and surface temperatures of the Norwegian Sea not in excess of 6°C during this period. It is difficult to relate coastal littoral faunas directly to open ocean temperatures and circulation due to possible influence of more local factors, i.e.

sea-ice influence and feeding conditions. Yet, the Ågotnes section contains molluscs that indicate less sea-ice than found on Svalbard today (Feyling-Hanssen 1955) and possibly also higher temperatures. The benthic foraminiferal assemblages of the section are quite similar to those found on the nearshore sea-bottom from the same period (Godvik 1981, Seland 1981) and indicate cold, but not high-arctic conditions. This taken together should imply that the coastal areas were exposed to cold, normal marine waters at least during summer seasons as indicated by reduced sea-ice influence. The most plausible cause for this is an increased flux of water masses from the Atlantic into the Norwegian Sea. The coastal record is thus in agreement with the cold but seasonally high-productive surface waters prevailing in the southern Norwegian Sea during the Bølling/Allerød chronozones. Judging from both mollusc and foraminiferal data we suggest that the coastal environment may have been quite similar to that found north of Bjørnøya (74°N) today. The coastal record from northern Norway also agrees with our oceanic record. Vorren et al. (1983) concludes that the ice margin was situated on the continental shelf outside Troms in Late Weichselian time, and a second, less extensive ice advance occurred shortly before 13,000 BP. The period 13,000 to 10,000 BP was characterized by icebergs, more open and possibly more temperate waters and stronger bottom currents than the previous period. The final warming of the coastal areas of Troms did not occur until after 10,000 BP. This age also seems to apply to the Glomfjord area in the more southern part of North-Norway (Rasmussen 1981).

The late glacial record of the Atlantic coasts of the British Isles is dissimilar to that of West-Norway. Benthic foraminifera, molluscs and fossil *Coleoptera* show a clear response to the warm surface of the adjacent North Atlantic during the period 13,000 to 11,000 BP (Coope 1975, Coope & Joachim 1980, Haynes et al. 1977, Peacock et al. 1978). Peacock (1983) shows evidence for a considerable temperature gradient across the shelf west and northwest of Scotland, which is in agreement with a winter polar front position in the vicinity of the Scotland/Iceland ridge 13,000 to 11,000 BP. This is supported by the large temperature gradient between the still cold surface of the Norwegian Sea and the surface waters west of Ireland, which at least for parts of the 13,000 to 11,000 BP period reached temperatures close to the present (Duplessy et al. 1981, Ruddiman & McIntyre 1981).

### *The Arctic Ocean record*

Most deep-sea cores from the Arctic Ocean contain a foraminifera-rich top sediment with a glacigenic fossil-poor zone beneath. This lower zone is rich in *Elphidium* while the upper part is rich in both benthic deep-sea species and planktonic foraminifera (Herman 1974). The Arctic Ocean stratigraphy is in this respect similar to that found in the present study. However, based on paleomagnetic studies, most investigators of Arctic Ocean cores conclude that sedimentation rates have been extremely slow (1–2 mm/kyr) and that the cores often penetrate the whole Quaternary (Hunkins et al. 1971, Herman 1974, Clark et al. 1980, Herman & Hopkins 1980). Magnetic measurements have also been performed on many of the cores from this paper and they show a very confusing picture both in inclination and declination (Løvlie, pers.comm.). One should expect some similarity between Arctic Ocean and Norwegian Sea cores due to the open interchange of water between the two basins. Preliminarily we will thus suggest that the Arctic Ocean cores may cover shorter timespans than hitherto proposed.

## Conclusions: a circulation model for the Late Weichselian Norwegian Sea

Based on the previous discussions we propose a model for the circulation patterns of the Norwegian Sea during the Late Weichselian / Holocene transition (Figs. 13–16).

### *Phase I: Prior to 13,000 BP*

During this period a permanent pack-ice and iceberg environment was present over the Norwegian Sea. A stable, stratified water column inhibited bottom water formation. The deep waters seem to have been very homogenous. Deep water exchange with the North Atlantic was probably negligible, leaving the Norwegian Sea with a slowly renewed deep water body with low oxygen and nutrient content. Temperatures were probably above 0°C in the deep water. As there are no physical barriers between the Norwegian/Greenland Sea and the deep basins of the Arctic Ocean, this type of deep water was probably also

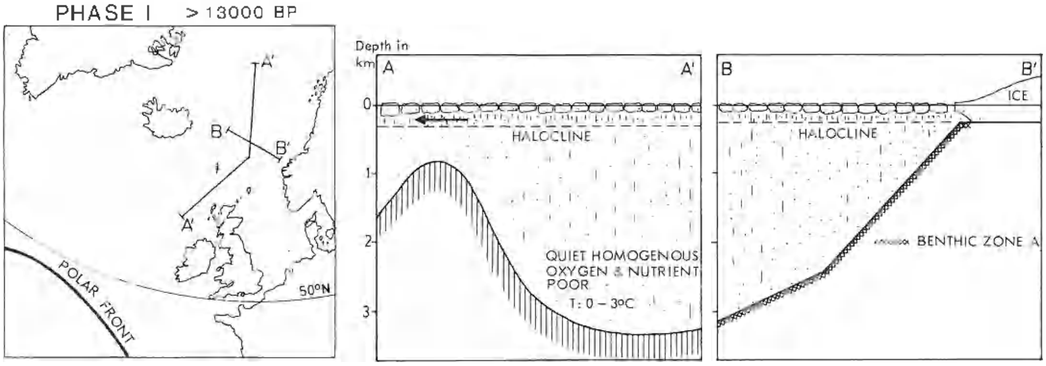

*Fig. 13.* Proposed circulation pattern, bottom water properties and depositional environment in the SE-Norwegian Sea prior to 13,000 BP. Polar front position from Ruddiman & McIntyre (1981).

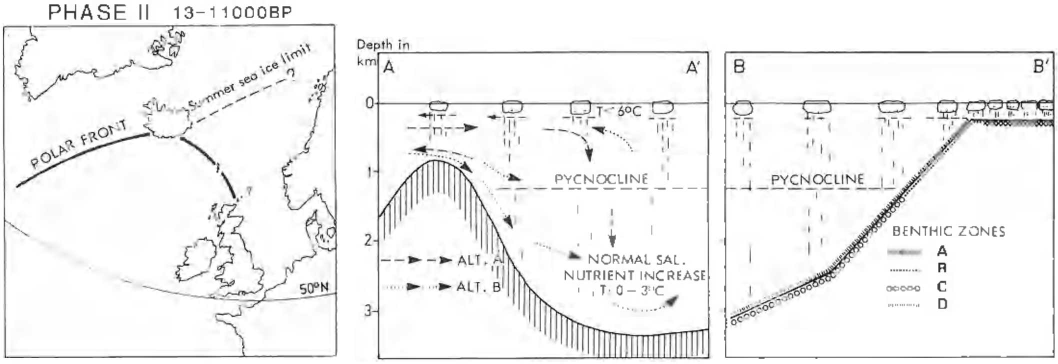

*Fig. 14.* Proposed Polar Front position, summer sea ice limit, circulation pattern with two different alternatives (A och B), bottom and surface water properties and depositional environment in the SE-Norwegian Sea 13,000 to 11,000 BP. Polar Front position SW of Iceland from Ruddiman & McIntyre (1981).

present in the Arctic Ocean. Considerable amounts of meltwater from the ice sheets caused a net outflow of low-saline surface waters as indicated by the current arrow above the halocline on Fig. 13.

## *Phase II: 13,000–11,000 BP*

At about 13,000 BP the Norwegian Sea experienced drastic environmental changes at all depths. The ice-sheet margin retreated towards the coast, the oceanic polar front moved northwestward in the North Atlantic. The sea ice cover over the Norwegian Sea diminished, giving open summer seasons which gave rise to high surface productivity. The decrease in ice cover was probably connected with influx of normal marine Atlantic surface waters which may also have caused warmer, ice-free summers along the west coast of Norway. The connection between deep-sea and surface circulation is, however, difficult to model. We have therefore on Fig. 14 indicated two possible alternatives that may have existed during the period.

*Alternative A* postulates influx of Atlantic surface or near-surface water. The inflowing water must have been significantly colder than the present Norwegian Current, as indicated by planktonic foraminifera. To what degree this circulation regime could have produced bottom water is dependent on the salinity of the upper layers and the possibility of cooling while exposed to the atmosphere during cold seasons. Surface waters were definitely quite cold, so the limiting factor must have been salinity. We would expect that

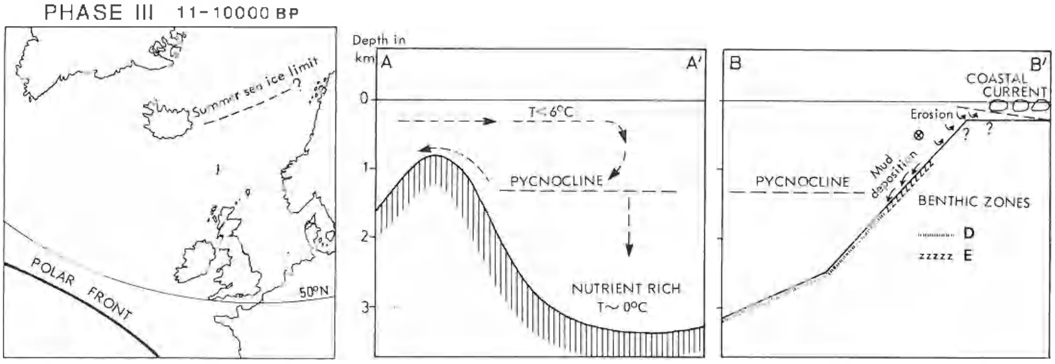

*Fig. 15.* Proposed Polar Front position and summer sea ice limit, circulation pattern, bottom and surface water properties and depositional environment in the SE-Norwegian Sea 11,000 to 10,000 BP. Polar front position in the N-Atlantic from Ruddiman & McIntyre (1981).

meltwater from the ice-sheet, melting sea ice and icebergs did reduce salinity and thus possibilities for bottom water formation. During such periods water masses originating from the North Atlantic could have circulated as an undercurrent in the Norwegian Sea in a manner similar to the present Arctic Ocean. Yet the marked environmental change which also affected the deep basin of the Norwegian Sea strongly indicates the presence of vertical water exchange either as in Alternative A or Alternative B. Possibilities for bottom water formation probably increased during this period due to decreased iceberg production and the onset of a coastal current along the Norwegian Coast which may have reduced meltwater influence in the central Norwegian Sea. Bottom temperatures were possibly still above 0°C due to less intense bottom water formation than at present.

*Alternative B* implies the introduction of dense water entering over the sills from the North Atlantic and into the deep Norwegian Sea. This could have pressed overlying water masses upwards, destroying the surface halocline causing a surface outflow into the North Atlantic. Such a 'reversed' circulation could have been the initial prosess that led to the dimunition of the pack-ice cover over the Norwegian Sea. Still, as there are two possible patterns for this initial deglaciation period, the nexus of cause and effect between surface and bottom conditions of the Norwegian Sea is still an open question.

In this connection we will also underline the similarity between the record from this deglacial period and the planktonic and benthic record from isotope stage 5d–5a (Kellogg et al. 1978, Streeter et al. 1982), which indicate similar environments. The seasonal open surface far to the north during the period from 13,000 BP would also imply a possibility for moisture to feed glaciers, e.g. on Svalbard and is in this respect in accordance with theories for a late glacier advance in these northern areas (Boulton 1979).

### *Phase III: 11,000–10,000 BP*

The Younger Dryas cooling did not leave as great impact on the Norwegian Sea record as in the North Atlantic. The cold surface conditions of the previous period in the Norwegian Sea also prevailed during the Y. Dryas. The Norwegian Sea remained ice-free at least parts of the year. A more intense current regime was introduced on the upper slope (Fig. 15), indicating increased surface water exchange with the North Atlantic. This caused a larger advection of saline waters into the Norwegian Sea and enlarged possibilities for bottom water formation and lowered bottom temperatures. The onset of large scale overflows to the deep North Atlantic thus most likely started during this period.

### *Phase IV: 10,000 BP to Present*

The warm water of the Norwegian Current was introduced rapidly at about 10,000 BP in the eastern part of the Norwegian Sea. Coincidental with this event, the present deep water fauna started to appear, possibly responding to Holocene nutrient levels and bottom water forming conditions. The northward retreat of the oceanic

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