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Evolution of the Central Tertiary Basin of Spitsbergen: towards a synthesis of sediment and plate tectonic history

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ABSTRACT

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The Central Tertiary Basin of Spitsbergen developed east of a strike-slip boundary between the Eurasian and Greenland plates in response to the initiation of rifting and later seafloor spreading in the Norwegian–Greenland Sea. Using a new plate model for the opening of the North Atlantic, we attempt to relate the sediment history of the Central Tertiary Basin to its plate tectonic framework and suggest a basin development in four phases. The Central Tertiary Basin was formed in the Early Paleocene during a right-lateral strike-slip phase, after the plate boundary between Greenland and Svalbard had jumped eastward to the Hornsund Fault Zone (phase 1). The onset of seafloor spreading between Greenland and Eurasia and a drastic counterclockwise change in spreading direction between Greenland and North America after chron 25 (Late Paleocene) induced compression-dominated transpression between Greenland and Svalbard until chron 24 (Lower Eocene) (phase 2). The first sedimentary evidence for tectonic uplift along the strike-slip zone is found in a wedge-shaped delta fan in the northwestern part of the basin. Strike-slip-dominated transpression prevailed from chron 24 to chron 21 (lower Middle Eocene), contemporaneous with successive basin narrowing and an eastward migration of the depocenter (phase 3). After chron 21 (lower Middle Eocene) the relative motion between Greenland and Svalbard first changed to strike-slip, terminating the transpressional phase, and then to transtension after chron 13 (Lower Oligocene) (phase 4).

Introduction

Tertiary rocks on Svalbard occur mostly on the main island Spitsbergen, covering a great part of the area south of Isfjorden. The "Central Basin" (Fig.1), discussed in this study, consists of a broad NNW–SSE-trending syncline, bounded in the west by the deformation belt of the "West Spitsbergen Orogeny" (Harland, 1965, 1969) and in the east by the Lomfjorden Fault Zone. The sedimentary succession of Carboniferous to Tertiary strata inside the basin rests on a Caledonian basement presently exposed on the west coast and in northern Spitsbergen (Birkenmajer, 1981). The Central Tertiary Basin of Spitsbergen is of particular interest because it was formed along the strike-slip boundary between two large plates, the Greenland and the Eurasian plates. As a result, the tectonic and sedimentary development of the basin can be expected to be intimately related to the rifting and seafloor spreading history between Greenland and Eurasia, particularly to the evolution of the triple junction between North America, Greenland and Eurasia. Many papers have been published in an attempt to relate the sedimentary development and subsidence history of the Central Tertiary Basin to the Tertiary tectonic regime between Svalbard and Greenland as predicted by the plate model from Talwani and Eldholm (1977). The most comprehensive syntheses, put forward by Kellogg (1975) and Steel et al. (1981, 1985), have led to a good general understanding of the basin history. One of the problems of reconstructing the basin history, though, consists in the correlation

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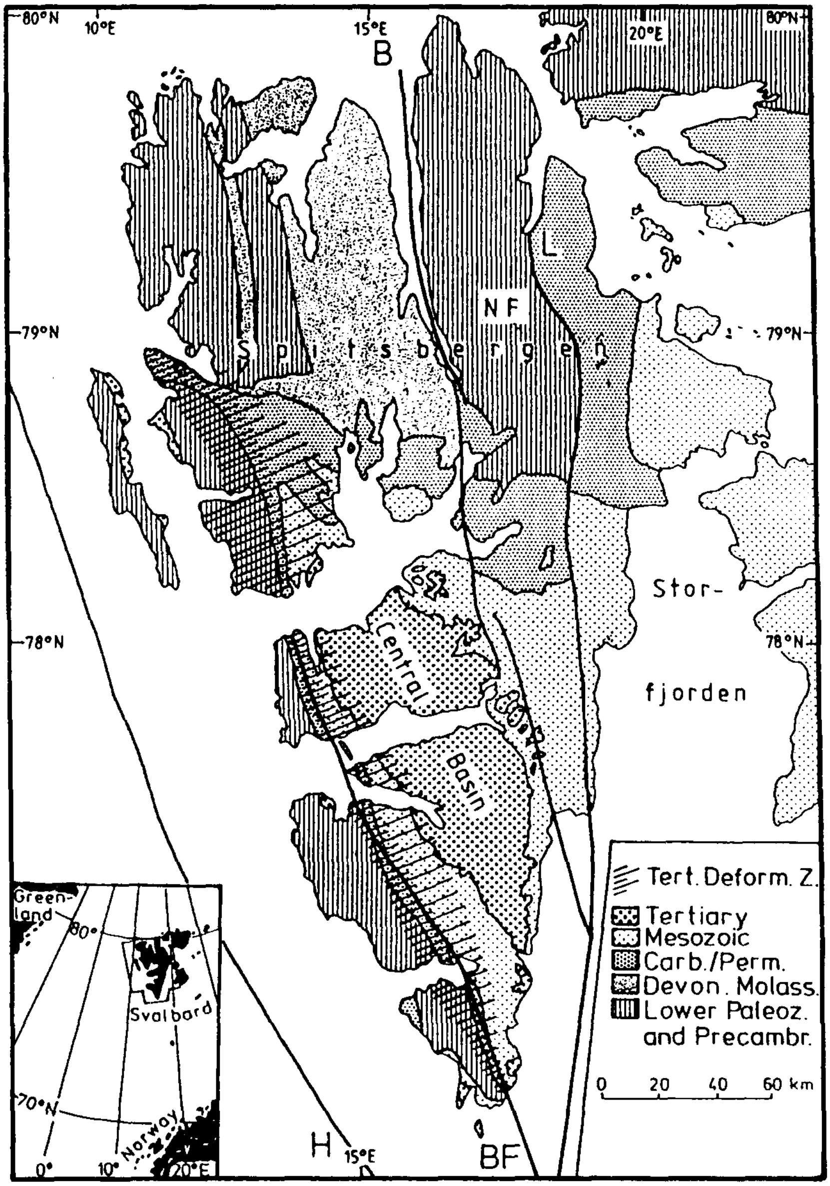


Fig. 1. General geological basement map of Spitsbergen. Structural features after Birkenmajer (1972) and Ohta (1982). *BF* = Western Boundary Fault, *H* = Hornsund Fault Zone, *L* = Lomfjorden Fault Zone, *NF* = Ny Friesland.

between plate tectonic events and the basin history. This is due to a general scarcity of datable fossils in the Tertiary succession, to an incomplete knowledge of the sediment distribution, and to problems in Talwani and Eldholm's (1977) plate model.

Objective

During four field seasons in the Central Tertiary Basin, several Kiel University Svalbard expeditions focussed on the depositional environment and distribution of Tertiary sediments on Svalbard

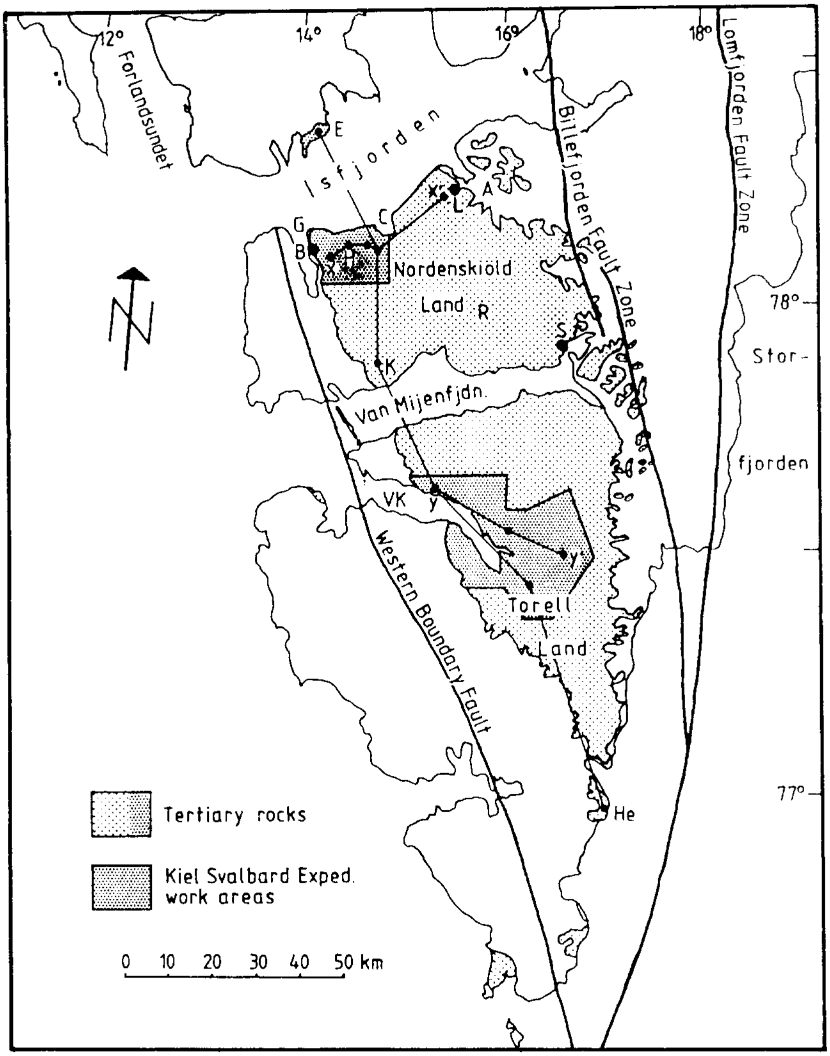


Fig.2. Map of southern Spitsbergen, displaying the area of Tertiary outcrops in the Central Tertiary Basin and the work areas of the Kiel Svalbard expeditions. Structural features after Ohta (1982). *A* = Adventdalen, *B* = Barentsburg, *C* = Colesbukta, *E* = Erdmanflya, *G* = Groenfjorden, *H* = Hollendardalen, *He* = Hedgehogfjellet, *K* = Kolfjellet, *L* = Longyearbyen, *R* = Reindalen, *S* = Svea, *VK* = Van Keulen-fjorden; *E–He*, *x–x'*, *y–y'* = profiles, see Fig. 4.

(Fig.2). The detailed mapping results will be published by Spielhagen and Müller (in prep.). A revised plate tectonic model for the opening of the North Atlantic has been published by Roest and Srivastava (1989a, b) and Srivastava and Roest (1989). This model incorporates recently acquired magnetic anomaly data north of the Charlie-Gibbs Fracture Zone and re-identifications of magnetic lineations 25 (59 Ma) and older in the Labrador Sea. All ages of magnetic anomalies referred to in this study are taken from the DNAG timescale (Kent and Gradstein, 1986). The main objective of this paper is to combine our mapping results with available data from the literature and to review the sedimentary evolution of the Central Tertiary Basin in light of the new North Atlantic plate model in order to better understand the impact of plate tectonic events on the basin's development.

Lithostratigraphy and facies interpretation of Tertiary rocks

The sedimentary succession of the Central Tertiary Basin of Spitsbergen has been divided into six lithologic series by Nathorst (1910). Generally following this scheme, Major (1964) and Livsic (1974) established two sets of formation names for the succession. Livsic (1974) subdivided the third series, the Sarkofagen Formations of Major (1964), into two different units, the Grumantbyen and Hollendardalen formations. Although Major's stratigraphic scheme (Fig.3) is currently used, some authors (Steel et al., 1981, 1985; Steel and Worsley, 1984) tried to integrate the Grumantbyen and Hollendardalen Formation into Major's Sarkofagen and Gilsonryggen Formation, regarding the former as sub-units of the latter formations and deviating from Livsic's (1974) nomenclature of the Hollendardalen Formation being equivalent to the upper part of Major's Sarkofagen Formation. In order to achieve a better integration of Livsic's "formation"-subunits into Major's formation system called the Van Mijenfjorden Group (Harland, 1969), we propose to reduce the rank of the Grumantbyen and Hollendardalen formations to member status. A north–south and two east–west sections through the Tertiary succession (Fig.4) illustrate the lateral thickness variation of lithologic units throughout the basin.

Firkanten Formation

Albian/Aptian strata belonging to the Carolinefjellet Formation, an alternation of shales and sandstones, underlie the base of the Tertiary on Spitsbergen. The lowermost Tertiary unit, the Firkanten Formation, can be divided into three members. In Nordenskiöld Land the basal Todalen Member begins with a conglomerate that thins from the western margin of the basin to the center and southwards (Vonderbank, 1970; Steel et al., 1981). The overlying fine gravel conglomerate represents the transition to a coal-bearing sequence of alternating shales, silt- and sandstones, the main part of the Todalen Member. The upper part of the formation consists of sandstone beds in the north (Endalen Member), which interfinger with siltstones and shales in the southern part of the basin (Kalthoffberget Member).

The three members of the Firkanten Formation reflect a predominantly transgressive trend from delta plain to prodelta/outer shelf facies, although the succession consists of several regressive segments (Steel et al., 1981). Initially two subbasins were separated by the Van Mijenfjorden Fault Block (Ytreland, 1980; Nøttvedt, 1985). The northern subbasin deposits exhibit a tide-dominated delta plain character for the lower part (Todalen Member) of the Firkanten Formation (Steel et al., 1981). In the more southerly parts of the Spitsbergen basin, sedimentation proceeded in a more distal environment under wave dominated conditions (Steel et al., 1981).

Basilika Formation

The overlying Basilika Formation consists of silty shales. Gastropods and bivalves with double-valve preservation occur in fossil lenses in the upper part of the formation in the Isfjorden area. Small pyrite nodules and scattered pebbles, as observed in the Firkanten Formation, are also found throughout the entire Basilika Formation.

The shales of the Basilika Formation have been interpreted as an "outer shelf mud complex" by Steel et al. (1985). More evidence for this interpretation is found in lenses of shells and foraminifera (also observed by Vonderbank, 1970), pointing to (seasonal?) storm events and redistribution by currents. Vonderbank (1970) interpreted the layer with double-valve preserved *Conchocele conradii* as a result of a sudden change from oxic to anoxic conditions. The scarcity of bioturbation and marine faunas, and the abundance of pyrite (up to 5%) are also indicative for an oxygen-depleted environment. Conditions may have resembled the present Baltic Sea where anoxic events in subbasins occur seasonally due to an estuarine circulation pattern (Grasshoff and Voipio, 1981).

Sarkofagen Formation

An upward increase in silt, sand and degree of bioturbation in the Basilika shales mark the transition to the silt- and sandstones of the Grumantbyen Member, the lower part of the Sarkofagen Formation. The Grumantbyen Member mainly consists of silt- or fine grained, matrix rich sandstones, which are intensely bioturbated. In the Adventdalen area (south of Isfjorden),

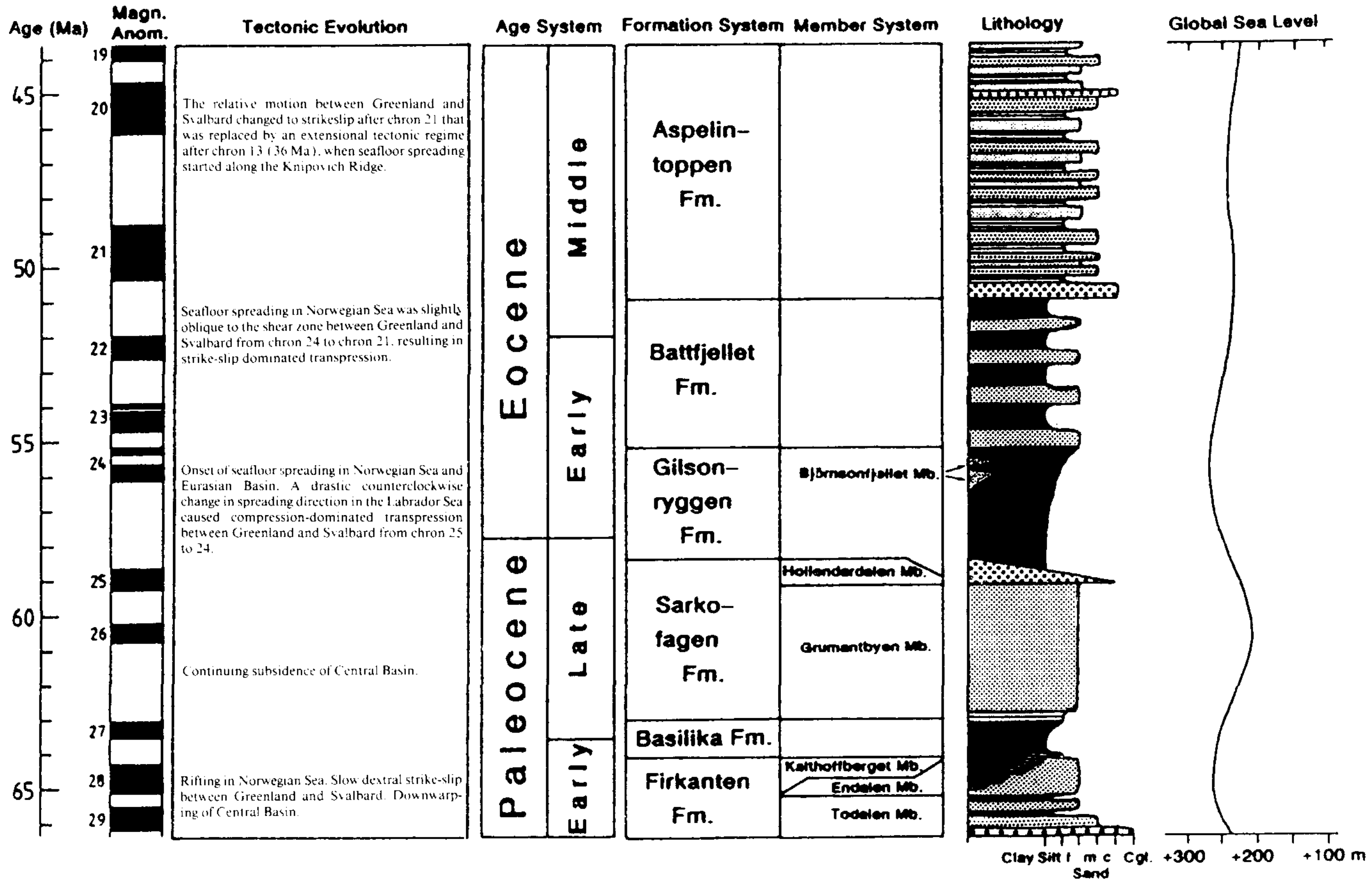
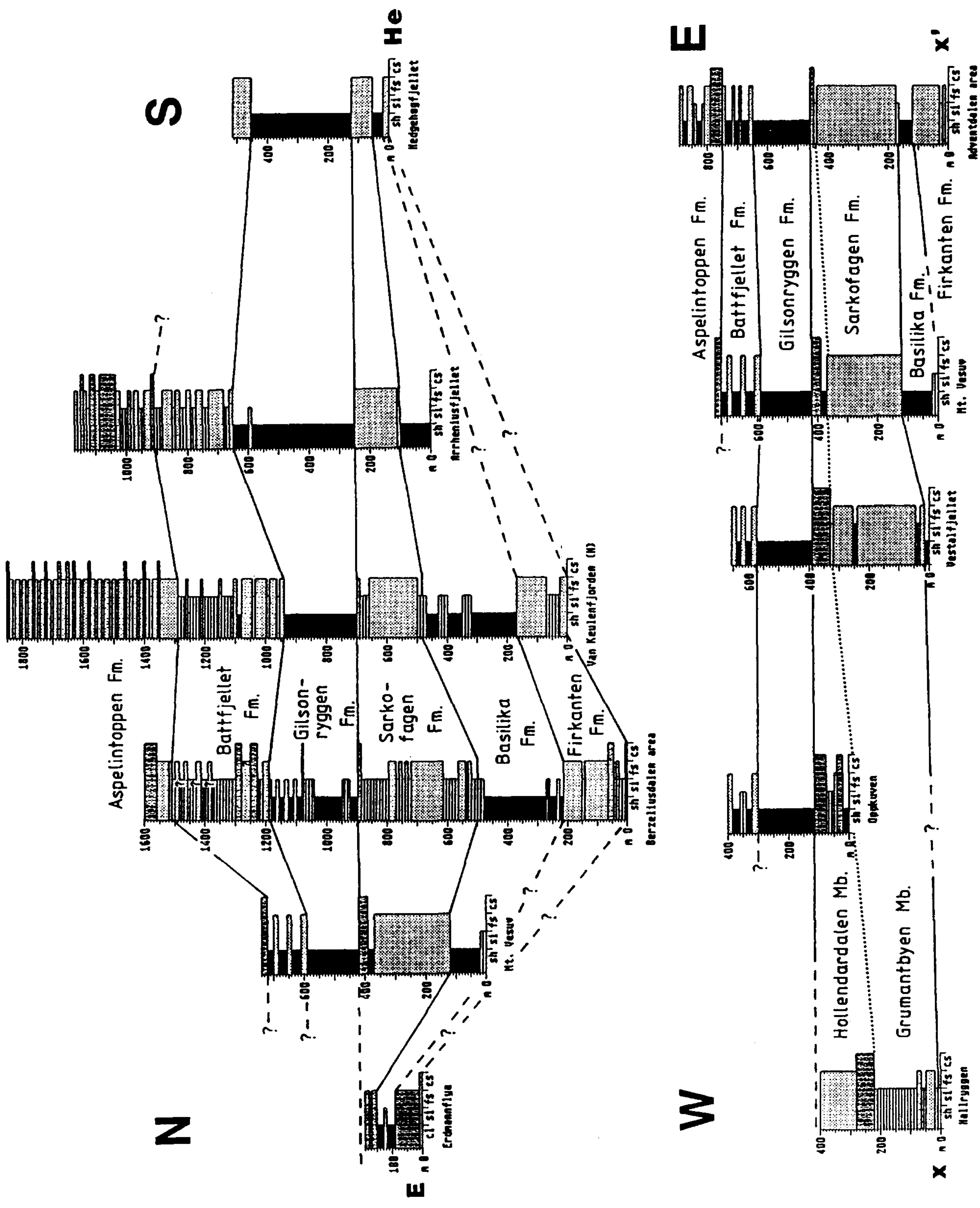


Fig.3. Standard profile of the sequence in the Tertiary Central Basin of Spitsbergen with correlation to magnetic anomalies (adapted from Kent and Gradstein, 1986), tectonic evolution in the NE Greenland–Svalbard area and global sea level curve (adapted from Haq et al., 1987). Age determinations are tentative, following Manum and Throndsen (1986) and own studies.


N
S
He
E
Aspelintoppen Fm.
Battfjellet Fm.
Gilson-ryggen Fm.
Sarko-fagen Fm.
Basilika Fm.
Firkanten Fm.
W
E
x
x'
Aspelintoppen Fm.
Battfjellet Fm.
Gilsonryggen Fm.
Sarkofagen Fm.
Basilika Fm.
Firkanten Fm.
Hollendardalen Mb.
Grumantbyen Mb.

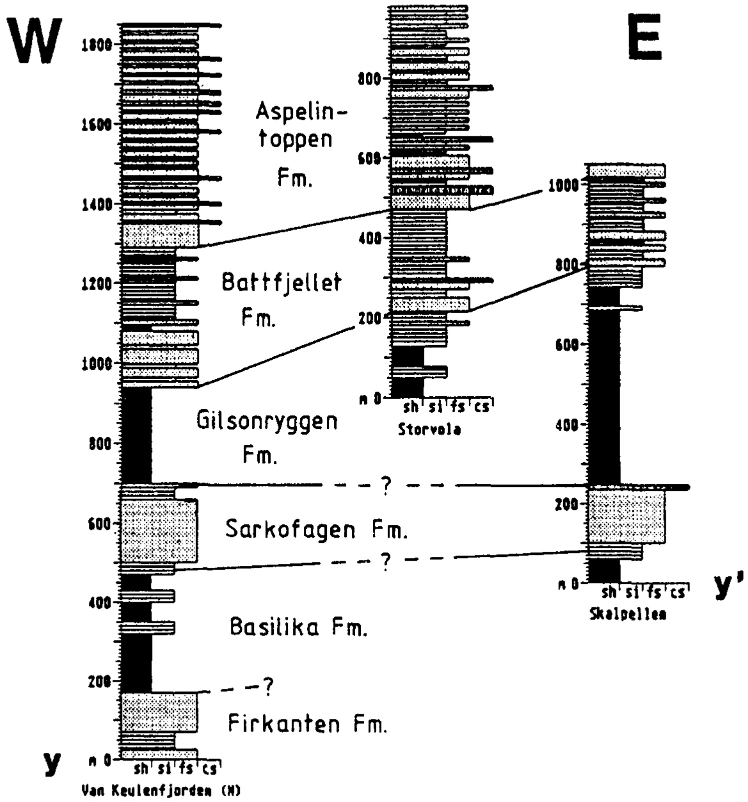


Fig.4. Profiles across the Central Tertiary Basin (see Fig.2. for locations). Adventdalen area profile after Major and Nagy (1972), Berzeliusdalen area profile after Croxton and Pickton (1976), Hedgehogfjellet profile after Birkenmajer and Narebski (1963). Profile *E–He* displays maximum thickness of most formations in the central part of the basin. Profile *x–x'* displays thickness homogeneity of Sarkofagen and Gilsonryggen formations in northern part of the basin and eastward wedging out of Hollendardalen Member. Profile *y–y'* displays eastward increasing thickness of Gilsonryggen shales in central part of the basin, corresponding to decreasing thickness of Battfjellet sandstones.

sandstones with intercalated conglomerates dominate (Major and Nagy, 1972; Dalland, 1979), indicating a near-shore facies. In Torell Land, medium-grained sandstones with interbedded conglomerates and enrichments of mudclasts were recorded. Sedimentary structures consist of infrequent ripple lamination and load casts.

The Grumantbyen Member most probably was deposited in a lower shoreface environment. Primary sedimentary structures in the lower shoreface environment are usually disturbed by storm events and bioturbation (Reinsson, 1984). The abundance of trace fossils and the formation of matrix-rich sandstones are typical for lower shoreface environments (Howard, 1972; Davies and Ethridge, 1975). The overall shallowing of the Spitsbergen basin at the end of Basilika time was coincident with an improvement in bottom water ventilation, as indicated by strong bioturbation in the transitional beds to the overlying sandy Sarkofagen Formation.

The upper part of the Sarkofagen Formation, the Hollendardalen Member, is restricted to northwestern Nordenskiöld Land. It consists of a variable sequence of shales overlain by alternating fine- to medium-grained sandstones, siltstones and subordinate shales. Our mapping results indicate that the basal shales, which are very similar to those of the Basilika Formation, are local intercalations.

The composition of the sandstones differs significantly from that of the Grumantbyen rocks (Fig.5). The feldspar and matrix content of the Hollendardalen sandstones is strongly reduced. Increased amounts of rock fragments and the

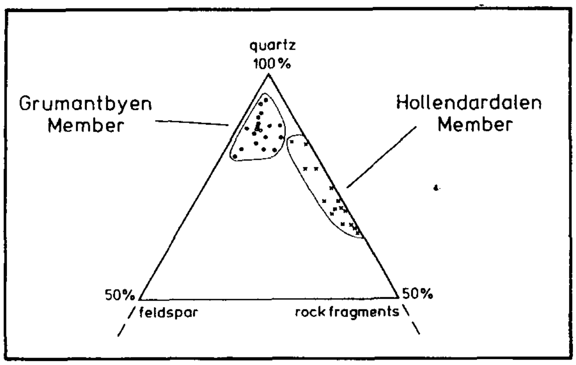


Fig.5. Composition of sandstones from the Grumantbyen and Hollendardalen Members of the Sarkofagen Formation in the Hollendardalen area. The increasing amount of rock fragments in sandstones of the Hollendardalen Member gives evidence for a nearby source area different from that of the Grumantbyen Member rocks.

mineral chloritoid, which is absent in the underlying beds, indicate a nearby source of eroded metamorphic and igneous rocks, probably in the area of the western fold belt of Spitsbergen.

Two different facies types are present in the upper part of the Hollendardalen Member. Facies A consists of interbedded and interfingering sandstones, siltstones, shales and coals. Frequent lateral and vertical changes within facies A are interpreted to be related to lagoonal environments, where coal development alternates with delta plain facies. Facies B, characterized by well-sorted fine- to medium- grained sandstones, sparse bioturbation and macrofauna, and a restricted assemblage of sedimentary structures, indicates high energy nearshore conditions. This facies represents reworked delta front sheet sands, which document the initial delta outbuilding phase from the western basin margin. East of Hollendardalen the sands partly prograded onto shales thereby giving evidence for the contemporaneous transgressive trend in the basin. The subenvironments of facies A and B of the Hollendardalen Member reflect a prograding, wave-dominated delta system with tidal influences in shallow, protected parts.

Gilsonryggen Formation

The contact to the overlying Gilsonryggen Formation is observed as a sharp lithological transition from silt/sandstones to shales throughout the entire basin. In Torell Land, a metersize slump mass of green sandstone was found in the lower part of the Gilsonryggen Formation. Scour marks on the lower surface of the slump block indicate WNW-directed gravity transport. Intercalations of sand and siltstone beds were recorded at several localities (cf. also Croxton and Pickton, 1976; Steel et al., 1981). The units, termed Bjørnsonfjellet Member by Steel et al. (1981), pinch out eastward and are mostly massive, with locally occurring graded bedding. Soft sediment deformations, such as loadcasts and large slump folds, are common and the bottoms of single units frequently show groove or flute casts.

The content of silt and fine-grained sand increases upwards towards the boundary of the overlying Battfjellet Formation. In western Nordenskiöld Land, the transition consists of a sandstone unit that pinches out eastward, where it is replaced by a set of sandstone clinoforms with successively higher stratigraphic positions (Kellogg, 1975; Helland-Hansen, 1990) each of them thinning out to the east. Hence the Battfjellet Formation in the Gröndalen area represents a time equivalent unit to parts of the Gilsonryggen Formation in the eastern areas.

The sharp transition from the Sarkofagen sandstones to the Gilsonryggen shales which do not contain coarse-grained material in the lower part indicate a rapid transgression. Increased content in organic matter (Manum and Throndsen, 1978) and microfossils (Dypvik and Nagy, 1979) suggests restricted circulation with low oxygen content in the bottom water for the Spitsbergen basin. The eastward pinching out of the sandstone beds of the Bjørnsonfjellet Member probably resulted from mass-gravity transport (Helland-Hansen, 1990) and/or turbidity currents (Steel et al., 1981) downslope. We interpret the scattered igneous and sedimentary pebbles and conglomerate blocks in the Gilsonryggen Formation as ice-rafted material (cf. Dalland, 1977).

Battfjellet Formation

The Battfjellet Formation displays a variety of facies types and successions throughout the Central Basin. In Nordenskiöld Land the lower unit of the overlying Battfjellet Formation consists of shales with subordinate thin, parallel or occasionally ripple-laminated silt and sandstone beds (Steel, 1977). The upper unit most frequently consists of interbedded fine sandstones and mudstones and grades upward into fine sandstones with alternating parallel- and ripple- lamination or trough-shaped lamina sets (Helland-Hansen, 1990).

In the Van Keulenfjorden area (central part of the basin) a succession of medium-grained sandstones alternating with planar laminated silt- and fine sandstones builds up the lower part of the Battfjellet Formation contrasting with the equivalent, mostly fine-grained facies in Nordenskiöld Land. The upper part of the formation is dominated by siltstones and shales, interbedded with coarsening-upward sequences of sandstones.

In Torell Land the lower part of the formation is dominated by interlayered shales, siltstones and fine-grained sandstones. It coarsens upward to low-angle, cross-bedded fine- and medium-grained sandstones. The sandstone bodies appear to pinch out to the northeast and the percentage of fine-grained material increases from central to northern Torell Land. Current directions measured from flute casts in Battfjellet sandstones in central Torell Land exhibit a northwesterly transport direction. Southeast of Torell Land at Hedgehogfjellet, the base of the Battfjellet Formation consists of shales with thin sandstone intercalations, grading upward into flaggy sandstones (Birkenmajer and Narebski, 1963).

The lower part of the formation was deposited mainly in an offshore/lower shoreface environment. In the basin center around Van Keulenfjorden the upper Battfjellet Formation is dominated by lower shoreface deposits. In Torell Land the upper Battfjellet Formation contains a diverse assemblage of local facies types such as upper shoreface environments, alternating with distributary channels and tidally-influenced sediments, indicating a generally shallower water depth than in the northeastern basin. Paleocurrents in this region mainly display directions towards the northeast, north and northwest.

The onset of the final regression in northern and southeastern parts of the basin is documented by a slow transition from offshore to lower shoreface conditions in the lower part of the Battfjellet Formation, followed by a sequence of upward coarsening foreset beds in the upper part. These clinoforms, successively reaching further eastward with increasing stratigraphic levels (Kellogg, 1976; Helland-Hansen, 1990), are evidence of a sequence of rapid coastline progradations.

Aspelintoppen Formation

The base of the overlying Aspelintoppen Formation usually consists of a fine- to coarse-grained sandstone unit, traceable throughout the entire basin, with alternating subparallel, planar cross- and ripple-lamination. The overlying unit is dominated by a flaser- or wavy-bedded succession of fine-grained sandstones, silt- and mudstones. Thin coal lenses and bands, coalified plant debris or wood fragments are common features in sand and siltstones, which occasionally contain well preserved tree leaves. This facies type is interbedded with different types of coarser-grained intercalations (for details see Spielhagen and Müller, in prep.).

The deposits of the Aspelintoppen Formation document the final regression in the Central Tertiary Basin. A thick succession of fluvial- and tide-dominated delta plain sediments is interlayered with some transgressive shoreface sequences. A flood tidal delta environment is indicated by thick units of subtidal and intertidal deposits little influenced by wave- and wind-generated processes. The Aspelintoppen Formation represents the youngest preserved Tertiary rocks in the Central Basin and is only occasionally overlain by Quaternary deposits.

Late Cretaceous to Neogene plate tectonic development between Greenland and Spitsbergen

We have reconstructed the positions of Svalbard relative to Greenland subsequent to the Late Cretaceous by using a new plate tectonic model for the opening of the North Atlantic (Roest and Srivastava, 1989a, b; Srivastava and Roest, 1989). This model improves the fit of magnetic anomalies for the early opening phases between Greenland, Eurasia, Rockall Plateau and the

Porcupine plate relative to North America by incorporating recently acquired magnetic anomaly data north of the Charlie-Gibbs Fracture Zone and re-identifications of magnetic lineations 25 (59 Ma) and older in the Labrador Sea. Figures 6a–f show the relative positions of Svalbard with respect to Greenland with Greenland held fixed for chrons 33 (80 Ma), 31 (69 Ma), 25 (59 Ma), 24 (56 Ma), 21

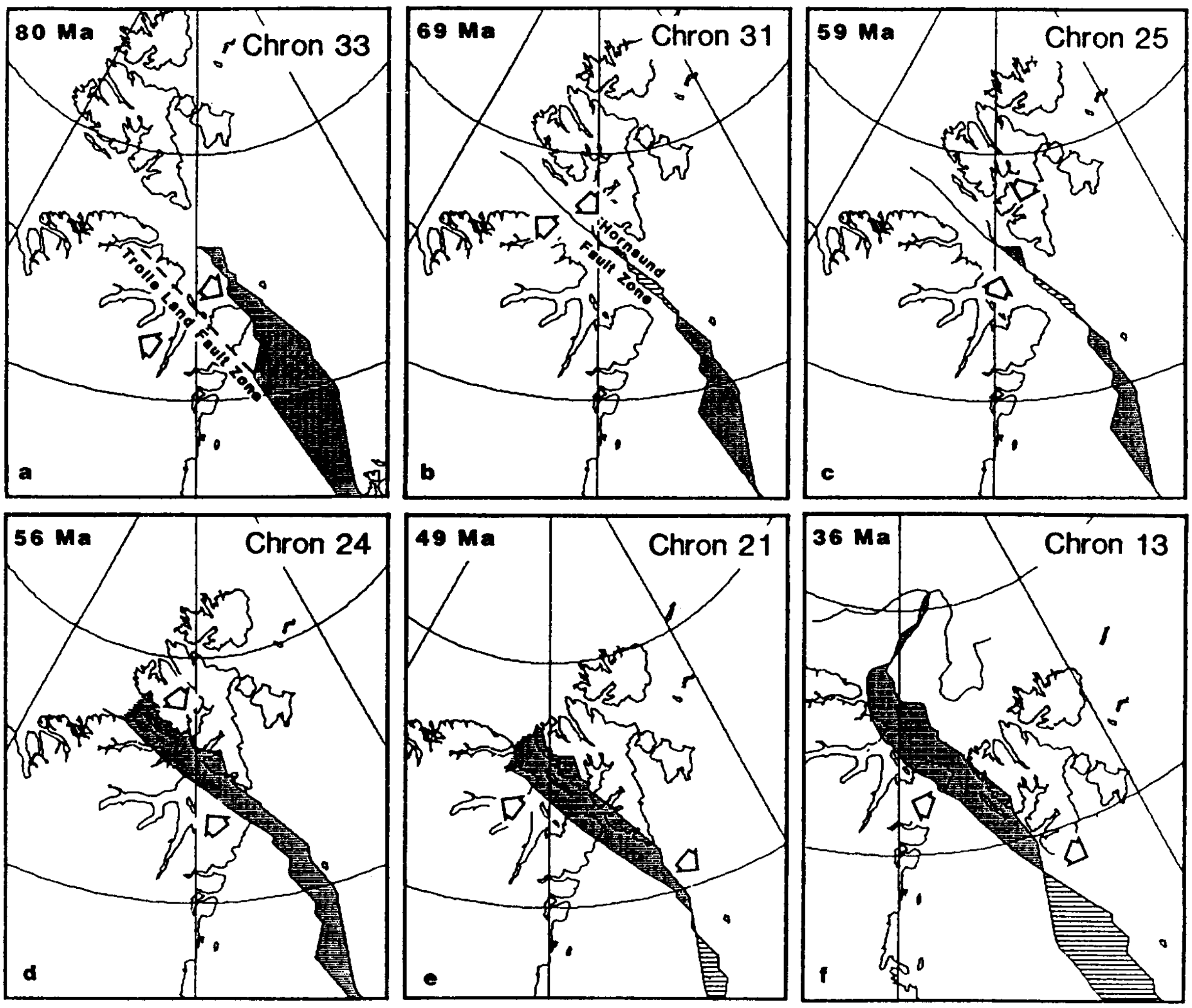


Fig.6. The motion of Svalbard relative to Greenland from the uppermost Cretaceous to Oligocene times is plotted with Greenland held fixed in the present-day coordinate system using finite reconstruction poles for Eurasia relative to Greenland from Srivastava and Roest (1989). Hatched areas mark gaps between continental margins, stippled areas mark overlaps. The motion between Greenland and Eurasia between chron 33 (80 Ma), chron 31 (69 Ma) and chron 25 (59 Ma) is calculated by using the reconstruction poles for Eurasia relative to North America and for Greenland relative to North America (based on magnetic lineations south of the Charlie-Gibbs Fracture Zone and in the southern Labrador Sea). For the time between chrons 33 and 25 the resulting differential motion between Svalbard and Greenland is characterized by right-lateral strike-slip. The plate boundary between Svalbard and Greenland in the Upper Cretaceous was most probably located in northeast Greenland at the Trolle Land Fault System as continuation of the Senja Fracture Zone (a, b). In the Lower Paleocene the plate boundary jumped eastward to the Hornsund Fault Zone. A drastic counterclockwise change in spreading direction in the Labrador Sea between chrons 25 (59 Ma) and 24 (56 Ma) caused transpression between Greenland and Svalbard, resulting in about 50–70 km shortening and 30 km of strike-slip motion (c, d). Strike-slip dominated transpression characterized the period from chron 24 to 21 (e), giving rise to 160 km of dextral strike-slip and 15–20 km of shortening. The relative motion between Greenland and Svalbard was dominated by strike-slip until chron 13 (36 Ma), subsequently followed by transtension, after seafloor-spreading in the Labrador Sea had ceased (f).

(49 Ma) and 13 (36 Ma). The new plate model implies a Late Cretaceous to Paleogene tectonic development between Svalbard and Greenland in four phases.

Phase 1: Strike-slip, chron 33 (80 Ma) to chron 25 (59 Ma)

Slow right-lateral strike-slip characterized the time-period between chron 33 and chron 25, similar to older models. In Srivastava and Roest's (1989) model, though, the revised fit position of Greenland relative to North America is about 100 km further south in comparison with previous models. As a result, the drastic overlaps of the positions of Svalbard relative to Greenland for reconstructions prior to chron 25 that are observed in the model of Talwani and Eldholm (1977) are no longer seen in the new reconstructions.

In the Late Cretaceous the Trolle Land Fault System in northeast Greenland is proposed to have been the plate boundary between Greenland and Svalbard as continuation of the Senja Fracture Zone (Eldholm et al., 1987). Geologic evidence indicates that the plate boundary between Greenland and Svalbard jumped eastward from the Trolle Land Fault System to the Hornsund Fault Zone in the Early Paleocene (Steel and Worsley, 1984; Steel et al., 1985). Only a few local occurrences of Lower Tertiary sediments in pull-apart basins have been mapped on northeast Greenland (Håkansson and Pedersen, 1982). On Svalbard, sediments from the Late Cretaceous are absent, a result of the uplifting, subaerial erosion and slight tilting of the Barents shelf (Birkenmajer 1981). The uplift might have been due to doming related to the onset of transtensional tectonics between Svalbard and Greenland (Steel and Worsley, 1984).

Kleinspehn et al. (1989) proposed a short sinistral strike-slip phase for the motion between Greenland and Svalbard in the early Danian (earliest Paleocene) based on paleostress analysis of the Todalen and Endalen Member of the Firkanten Formation. This short tectonic event that would have occurred between chrons 31 (69 Ma) and 25 (59 Ma), can not be detected by the plate model used here, because the relative plate motion is averaged over 10 m.y. for that time period.

Phase 2: Compression, chron 25 (59 Ma) to chron 24 (56 Ma)

The plate model by Srivastava and Roest (1989) implies a relatively short compressional event between Svalbard and Greenland from the latest Paleocene to early Eocene. This event was induced by a drastic counterclockwise change in spreading direction in the Labrador Sea between chrons 25 (59 Ma) and 24 (56 Ma), larger than inferred by previous models. During this time the average amount of shortening between Greenland and Svalbard is inferred to have been 50–70 km, along with a total strike-slip motion of about 30 km. These figures demonstrate that the transpressive regime at that time was dominantly compressive with a minor strike-slip component (Fig.6c, chron 25).

This model has important implications related to the observed structure of the in the core of the West-Spitsbergen Fold Belt, which are dominated by high-angle reverse faults, thrust faults and asymmetric faults, that merge eastward into decollement and thrust ramp strucutres (Nøttved et al., 1988). Several models have been suggested to explain the dominance of compressive structures in the Tertiary fold belt. Maher (1988) invoked a decoupling of dextral and convergent components during Tertiary transpression. Faleide et al. (1988) favor a model of separate deformation mechanisms between lower and upper crust. In their model a downward-flowing root is formed in the lower ductile part of the crust, separated from a thin-skinned fold and thrust belt in the upper crust by a sub-horizontal decollement surface. Based on the plate model as discussed above, we suggest that much of the observed shortening could be due to a transpressional phase dominated by compression between chrons 25 and 24 in the latest Paleocene/Early Eocene.

Phase 3: Transpression, chron 24 (56 Ma) to chron 21 (49 Ma)

The tectonic regime between Svalbard and Greenland changed to transpression dominated by strike-slip motion subsequent to chron 24. This phase resulted in approximately 160 km of dextral strike-slip motion from chron 24 (56 Ma) to chron 21 (49 Ma), accompanied by minor compression of about 15–20 km for this time interval. The magnitude of total Paleogene compression between Greenland and Svalbard estimated from the plate reconstructions used here is 65–90 km, of which 50–70 km occurred in the latest Paleocene/Early Eocene. Although these values are higher than former minimum estimates, they do not seem to be incompatible with the observed geological structures. Birkenmajer (1981) estimated a minimum of 10–15 km of total Eocene to Early Oligocene compression within the foldbelt. Nøttved et al. (1988) noted that adding the deformation in the Central Basin to Birkenmajer's (1981) estimate gives a total shortening of 15–20 km. Recent detailed mapping by Maher (1988) gives evidence that surface structures on Midterhuken (Bellsund) require at least 4 km and permit up to 8 km of stratal shortening. Maher (1988) pointed out that a simple extrapolation of this partial transect indicates a total shortening of tens of km, concurrent with the shortening implied by the plate model of Srivastava and Roest (1989).

Phase 4: Strike-slip/transtension after chron 21 (49 Ma)

After chron 21 (lower Middle Eocene) the relative motion between Greenland and Svalbard first changed to strike-slip, terminating the transpressional phase, and then to transtension after chron 13 (Lower Oligocene), when rifting and seafloor spreading started along the western margin of Spitsbergen.

In summary the tectonic development between Greenland and Svalbard presented here differs from previous models in two ways. Firstly the overlap between Svalbard and Greenland prior to chron 25 is minimal. This is in accordance with their narrow continental margins that do not indicate large amounts of stretching. Secondly, the transpression between chrons 25 and 21 can be divided into two distinct phases dominated by compression (chron 25–24) and strike-slip (chron 24–21), respectively, thereby giving an explanation

Fig.7. Paleogeographic maps of Spitsbergen in Early Paleocene to Middle Eocene time. *B* = Billefjorden Fault Zone, *BF* = Western Boundary Fault, *H* = Hornsund Fault Zone, *VMF* = Van Mijenfjorden Fault, *WFB* = Western Fault Block. Structural features after Ohta (1982).

(a) Early Paleocene: Deposition of lower/middle Firkanten Formation; transtensional tectonic regime. Northward directed transgression, establishment of two subbasins, divided by VMF. Deposition of coarse material (conglomerates at base) in northern part of northern subbasin, subsequent sand sedimentation in shoreface facies. Deposition of finer-grained material in southern subbasin. Source areas in the east, north, and northwest (WFB) of the subbasins.

(b) Middle Paleocene: Deposition of upper Firkanten Formation and Basilika Formation; strike-slip tectonic regime. Dominant shale deposition in offshore facies in the entire basin. Source areas in the east and northeast.

(c) Late Paleocene: Deposition of the Grumantbyen Member of the Sarkofagen Formation; strike-slip tectonic regime. Dominant sand deposition in shoreface facies in the entire basin. Source areas in the east and northeast.

(d) Latest Paleocene: Deposition of the Hollendardalen Member of the Sarkofagen Formation in the northwest and the lower Gilsonryggen Formation in the rest of the basin; tectonic regime changes to compression dominated transpression and causes uplift of WFB. Delta outbuilding and sand deposition in delta/offshore facies in the northwest of the basin; dominant shale deposition in the rest of the basin. Source areas for coarser-grained material in the northwest (WFB) and for shale in the east and northeast.

(e) Earliest Eocene: Deposition of the middle Gilsonryggen Formation; compression dominated transpressional tectonic regime. Dominant shale deposition in the entire basin, most likely due to rising global sea level with intercalations of turbidites/mass flows from the west. Source areas in the west (sand) and east and northeast (shale).

(f) Late Early Eocene: Deposition of the upper Gilsonryggen and lower Battfjellet Formations; transpressional tectonic regime. Continuing shale deposition in the east, uprising western deformation belt with eastward prograding sand deposition. Source areas in the east/northeast (shales) and west (sand).

(g) Early Middle Eocene: Deposition of the upper Battfjellet and lower Aspelintoppen Formations; last phase of transpressional tectonic regime. Eastward prograding deltas with sand deposition in delta facies, diminishing areas of open water sedimentation in the east. Source areas mainly in the west.

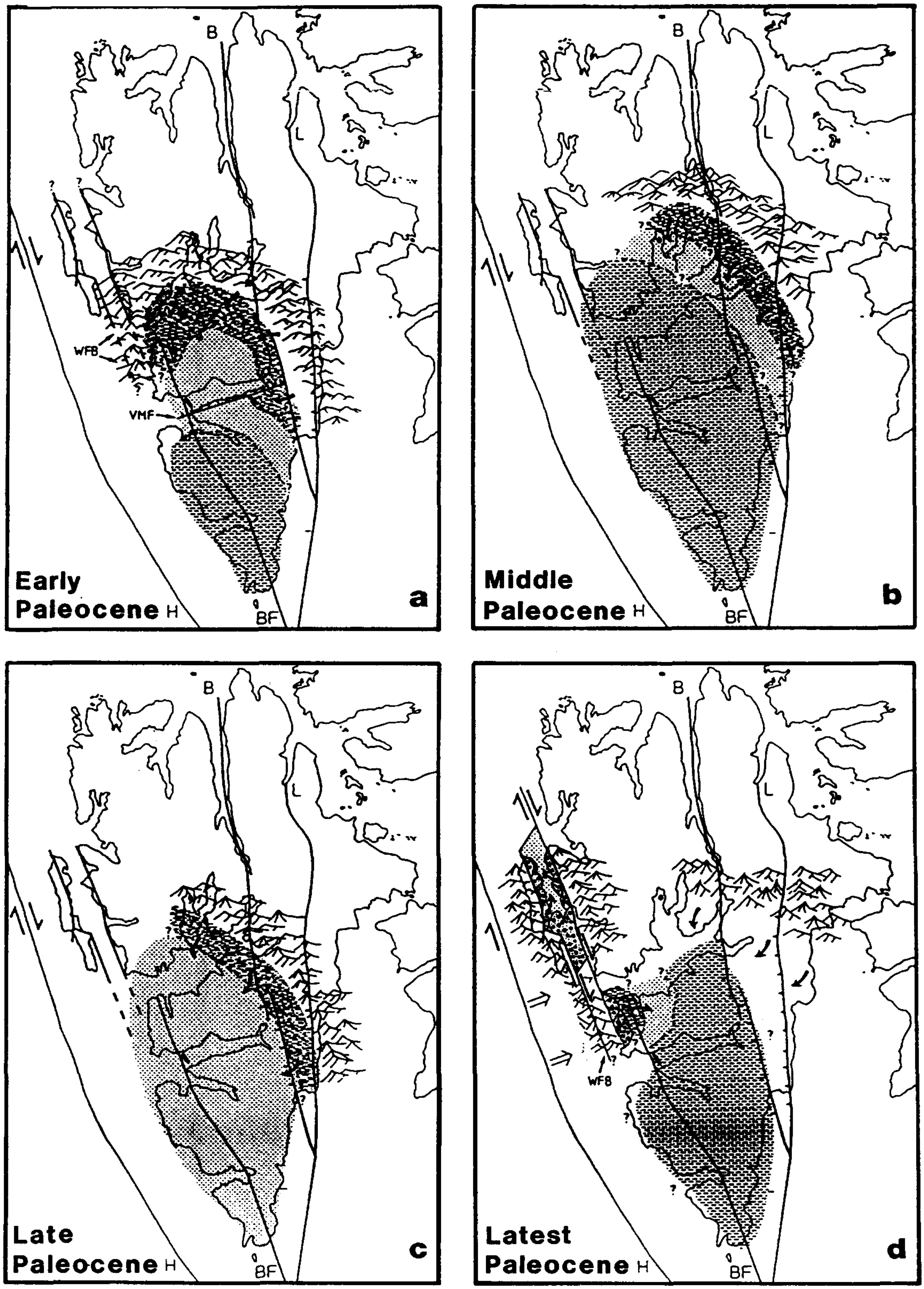

B
L
WFB
VMF
Early Paleocene
H
BF
a
Middle Paleocene
b
Late Paleocene
c
Latest Paleocene
WFB
d

Fig.7a–d.

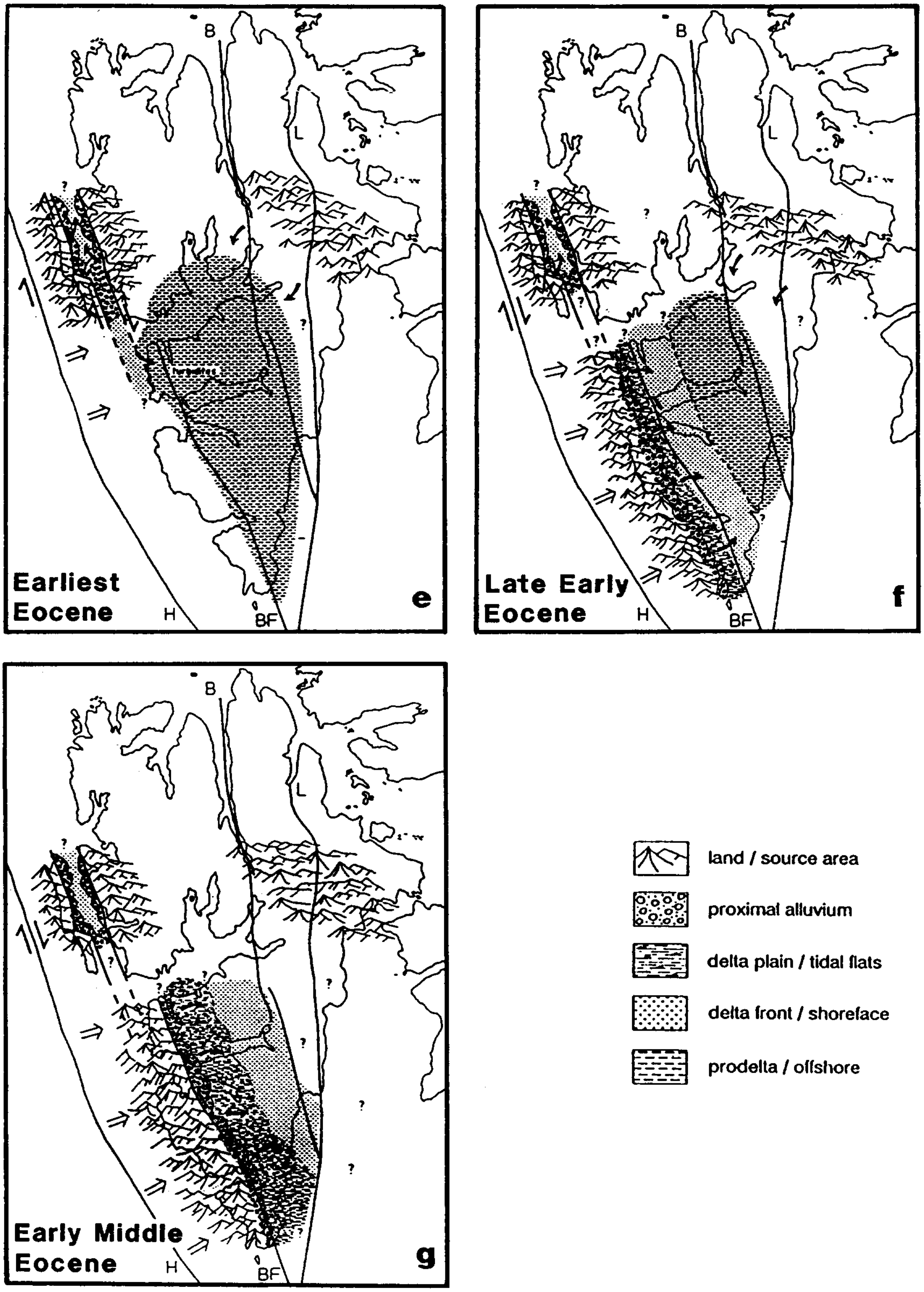


Fig.7 continued.

for the mostly compressional structures observed in the West Spitsbergen Fold Belt.

History of Tertiary Basin Development

In the following we attempt to outline the development of the Tertiary basin configuration and the sedimentary environment, illustrated by Figs.7a–g, in terms of tectonics, sediment input, and eustatic sea level changes (after Haq et al., 1987).

Paleocene

After a Late Cretaceous uplift, the Early Tertiary basin development started with reactivation of old north–south trending faults (Livsic, 1974). A transgression was caused by downwarping of the area west of the Billefjorden and Lomfjorden Fault Zones. These faults may have limited the basin to the east (Nøttvedt, 1985). The block movements are contemporaneous with a distinct increase in the spreading rate between Eurasia and North America (Srivastava and Roest, 1989) that may have resulted in ENE–WSW extension in the Svalbard area. The existence of a high-standing (horst-like?) western fault block (Fig.7a) northwest of the Van Mijenfjorden Fault Zone has been suggested by Nøttvedt (1985) for parts of Firkanten time (Earliest Paleocene). The transgression was directed northward and established two subbasins, separated by the Van Mijenfjorden Fault (Fig.7a; Ytreland, 1980; Nøttvedt, 1985). The main sediment source areas lay to the north and east of this basin (Ytreland, 1980). The eastward- and southward- thinning basal conglomerate in Western Nordenskiöld Land supports the concept of a western fault block as a supplementary source area for parts of Firkanten time. As there is no evidence for a westerly located sediment source area during deposition of the upper parts of the Firkanten Formation, the western fault block was probably flooded in Early Paleocene due to erosion plus sea level rise (cf. Fig.3) and/or subsidence caused by tectonic release. Although some indications for a (short?) regression in the uppermost Firkanten Formation have been reported (Kellogg, 1975; Steel, 1977), the deposition of the overlying Basilika shales can be seen in terms of the continued basin floor subsidence due to transtension along the Hornsund Fault Zone. The lack of nearshore deposits within the area of presently preserved outcrops (see also Major and Nagy, 1972) suggests a shoreline that was located east of the Billefjorden Fault Zone (Fig.7b).

The regression from the Basilika to the Sarkofagen Formation possibly corresponds to the worldwide sea level drop at the boundary Early/Late Paleocene (cf. Fig.3). During this regression water depth decreased to wave base and probably did not exceed 20–30 m in most parts of the basin (Fig.7c). The vertical lithologic homogeneity of the Grumantbyen Member indicates rather stable depositional conditions with a constant subsidence which compensated sediment accumulation. A minor increase in grain size from west to east and interfingering of medium-grained sandstones, conglomerates, and mudclasts at the eastern exposure margin indicate a shallow-relief shelf sea with an easterly shoreline located in the area of the Billefjorden Fault Zone.

The major sediment source area during deposition of the Grumantbyen Member probably lay to the east and northeast of the basin, where, in addition to the previously eroded Mesozoic and Paleozoic rocks (Kellogg, 1975), the nearshore deposits of the Basilika Formation were available for erosion. There is no evidence for a western source area.

It is important to note that until this point in the basin history (about the middle of Late Paleocene) all changes in basin geometry, water depth, and sedimentary facies can be explained by extensional basin subsidence and eustatic sea level changes. Until this time, except for the lowermost parts of the succession, there are no indications for a western coastline in the area of the recent Svalbard archipelago. The marked asymmetry of the Firkanten-Basilika-Grumantbyen succession with an increasing overall thickness to the west (Kellogg, 1975; Steel et al., 1981) gives evidence for greatest subsidence in this area. There is no indication for migration of the depocenter before deposition of the Gilsonryggen Formation, when a change in the subsidence pattern occurred (Manum and Throndsen, 1978).

Latest Paleocene–earliest Eocene

A reconstruction of the paleogeographic situation (Fig.7) in the latest Paleocene must reconcile two parallel developments. The Hollendardalen Member, a wedge-shaped deltaic sandstone unit (Fig.4), was deposited in Nordenskiöld Land. Decreasing thickness to the east indicates a western sediment source. In contrast, shales were deposited in southern parts of the basin at the same time (Steel et al., 1985).

The geometry and composition of the Hollendardalen Member are indicative of a westerly source area not far from the depositional site. The lack of similar deposits in the Van Keulenfjorden area and in northern Torell Land and a considerably reduced thickness of coarse sandstones north of Van Mijenfjorden (cf. Croxton and Pickton, 1976) suggest the existence of an erosional area west/northwest of Nordenskiöld Land, strikingly similar to the situation in earliest Paleocene (deposition of the lower Firkanten Formation, Fig.7a).

In the plate tectonic framework, the latest Paleocene upthrusting of the western fault block west of Nordenskiöld Land was contemporaneous with the onset of compression after chron 25 (59 Ma) due to a drastic change in seafloor spreading direction in the Labrador Sea. The correlation of these two events is in good accordance with Manum and Throndsen (1986), who dated the Lower Gilsonryggen Formation as upper Late Paleocene. Reflections of the incipient thrusting along the western margin of Svalbard can be seen in the meter-sized slump structures, found in the lowermost Gilsonryggen Formation in northern Torell Land, and in the increased amount of soft sediment deformation in the Hollendardalen Member.

Paleogeographic basin reconstructions for the latest Paleocene are complicated by the complexity of subenvironments and difficulty in reconstructing the paleoshoreline due to the lack of nearshore deposits in eastern and southern areas. For two reasons we suggest a wider basin extent than before. Firstly, the lower Gilsonryggen beds are free of coarse-grained intercalations and thus indicate a shoreline position outside the area of presently preserved Tertiary outcrops. Paleocene erosion must have flattened and lowered the relief which was successively flooded by the rising global sea level. Secondly, other evidence for a wide extent of the basin in northeastern Nordenskiöld Land was given by Birkenmajer et al. (1971), who compared findings of igneous and sedimentary pebbles and blocks in the Gilsonryggen Formation of Torell Land with possible source rock exposures in northeastern Svalbard. The area of provenance of this most likely ice-rafted material (Dalland, 1977) indicates location of the Gilsonryggen coastline in the Storfjorden and Ny Friesland regions. In most of the Central Basin areas south of Van Mijenfjorden, offshore conditions and shale deposition dominated from Latest Paleocene to Early Eocene times. The disappearance of the western fault block as a subaerial erosional area was most likely caused by the rising sea level at that time. Tectonic subsidence of the block was unlikely, since the transpressional tectonic regime persisted from the latest Paleocene to Middle Eocene. Thus the sedimentary development of the Central Basin in the latest Paleocene documents the onset of transpression along its western margin as well as a general trend of basin widening and deepening that we attribute to the contemporaneous global sea level rise (Haq et al., 1987).

Late Early Eocene–Middle Eocene

Renewed sediment discharge from an uplifted westerly source subsequent to deposition of the Hollendardalen Member is documented by eastward pinching out of massflow deposits and turbidites of the middle and upper Gilsonryggen Formation and especially the Bjørnsonfjellet Member. The final basin regression is illustrated in clinoforms prograding westward onto the Gilsonryggen Formation and in the eastward migration of the area of greatest subsidence (Steel et al., 1985). The clinoforms provide evidence of a sequence of rapid coastline progradations (Fig.7f, g) into the basin. This basin regression corresponds to the transpressional tectonic phase from chron 24 (56 Ma) to 21 (49 Ma). At this time the upthrusting of the western margin of Spitsbergen that had started at chron 25, had resulted in a subaerially-exposed area to the west of the basin that yielded the primary sediment supply for the basin. An eastward migration of the depocenter by about 20 km some time after deposition of the Grumantbyen Member is evidenced by lateral variation of the rank of coals in the Firkanten Formation (Manum and Throndsen, 1978) as well as by a comparison of the Late Paleocene margin and present day eastern margin of the basin (Steel et al., 1985). The main phase of basin narrowing and eastward migration of the depocenter corresponds to the upper Gilsonryggen/Battfjellet Formation, time-equivalent to a strike-slip-dominated tectonic regime (160 km strike-slip contrast only 15–20 km shortening from chron 24 to chron 21). We suggest that much of the eastward migration of the basin's depocenter may be explained by marine progradation of sediments from the Western Fold Belt to the east.

In Torell Land, northwestward-directed paleocurrents in Battfjellet sandstones, northeastward-pinching out sandstone bodies, and northward-decreasing grain sizes may indicate a basin geometry narrowing to the south. This contrasts with the Early Paleocene northward directed transgression that was possibly related to a shallow connection with the North Sea. The model of a possible southern basin closure is supported by the tectonic pattern as well (Fig.1). According to Ohta (1982), the southern termination of the Tertiary Basin was formed by the Billefjorden-/Lomfjorden Fault Zones that join east of Torell Land and form a junction with the Western Boundary Fault south of Spitsbergen. Hence we favor a Middle Eocene basin geometry closed to the south (Fig.7g), contrasting the generalized Eocene paleogeography of the Central Tertiary Basin as presented by Steel and Worsley (1984). Sediment input from the uprising deformation belt was roughly equivalent to subsidence rate throughout the subsequent deposition of the Aspelintoppen Formation, which consists of a thick package of deltaic and coastal plain sediments.

Summary of basin development

The development of the Spitsbergen Tertiary Basin in four phases can be summarized as follows:

Phase 1 (Paleocene)

Relative motion between Svalbard and Greenland was characterized by slow right-lateral strike-slip during this time period. Subsidence of the Tertiary basin floor started in the Early Paleocene and was asymmetric and greatest in the direction of the transform zone (see also Steel et al., 1981; 1985). The initial transgression was directed northward and supported by rising global sea level in earliest Paleocene (Fig.3). Rapid subsidence, exceeding the sedimentation rates, gave rise to deposition of the delta plain-shoreface–offshore sequence of the Firkanten and Basilika formations. The following regression and return to shoreface conditions over the entire basin (deposition of the Grumantbyen Member of the Sarkofagen Formation) was caused by decreasing subsidence and/or the falling sea level at the Early/Late Paleocene transition.

Phase 2: (Latest Paleocene–earliest Eocene)

Subsequent to chron 25 (latest Paleocene) the tectonic regime between Greenland and Svalbard changed drastically from strike-slip to compression-dominated transpression until chron 24 (earliest Eocene). This change in relative motion was related to a major plate reorganization in the North Atlantic, characterized by the onset of seafloor spreading between Greenland and Eurasia and by a severe counterclockwise change in spreading direction between Greenland and North America (Roest and Srivastava; 1989a). Beginning transpression at the western margin of Svalbard at that time is documented by the eastward prograding Hollendardalen Member in northwestern Nordenskiöld Land. Rising global sea level (Fig.3) may have favored the establishment of offshore conditions throughout the basin, resulting in deposition of the Gilsonryggen shales.

Phase 3: (Late Early Eocene–early Middle Eocene)

The time between chron 24 and chron 21 (Early to lower Middle Eocene) is characterized by strike-slip dominated transpression (Fig.7c–e), evidenced by intercalated mass flow deposits and turbidites in the Gilsonryggen shales and the overlying eastward prograding Battfjellet clinoforms. The sediments provided from the newly uplifted drainage area west of the basin caused successive basin narrowing and an eastward shift of the depocenter. The regression resulted in the final establishment of a delta plain environment. Within the western deformation belt, the Forlandsundet Graben (not discussed in this study) was generated, interpreted as a strike-slip basin formed in the transpressive tectonic phase (Steel et al., 1985).

Phase 4 (post early Middle Eocene)

A strike-slip phase from chron 21 and chron 13 (Fig.7e, f) terminated the previous transpression. After chron 13, rifting and seafloor spreading started along the western margin of Spitsbergen, giving rise to a transtensional tectonic regime. Upper Eocene Forlandsundet sediments (Manum and Throndsen 1986) give evidence for continued subsidence of the Forlandsundet Graben after termination of the transpressive strike-slip phase at chron 21.

It is unclear how long subsidence of the Central Tertiary Basin continued. The youngest, tentatively-dated sediments of the Tertiary succession are from the Middle Eocene upper Battfjellet Formation. The sedimentary pattern of the overlying Aspelintoppen Formation may have persisted up to the Late Eocene/Early Oligocene, since at least 1500 m of Tertiary sediments, now eroded, were deposited on top of the Aspelintoppen Formation (Manum and Throndsen, 1978).

Conclusions

The synthesis of sedimentary and plate tectonic data presented here generally supports earlier models describing the history of the Central Tertiary Basin of Spitsbergen. The use of a revised plate model for the North Atlantic, though, eliminates inconsistencies seen in older plate models such as unreasonably large overlaps between Svalbard and Greenland in Upper Cretaceous/Early Paleocene reconstructions, and yields a simple explanation for the compressive structures in the West Spitsbergen foldbelt. The successive tectonic phases of right-lateral strike-slip, compression and transpression for the plate boundary between Svalbard and Greenland as implied by the plate model correlate well with initial asymmetric basin subsidence, a wedge-shaped delta fan originating from an uplifted fault block along the western margin and continued basin narrowing and eastward progradation of a delta front, respectively.

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