A structural outline of Forlandsundet Graben, Prins Karls Forland, Svalbard

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Structural investigations have shown that the basin fill of the Forlandsundet Graben, Svalbard, has suffered stronger deformation than previously assumed. From plate-tectonic reconstructions it is known that the basin was initiated in a regional, dextral transpressional environment. During its early history, the basin may have been wider than its present outline, and syn-depositional deformation of the basin sediments was common. Later, the basin fill was affected by regional folding and local thrusting, probably in relation to continued strike-slip movements. The most intense deformation is localized in certain areas which may have acted as restraining bends of the graben margin. This event was followed by mild inversion of the basin margins. The inversion is now traced as compressional, margin-vertgent structures. Several systems of extensional fractures are developed. These clearly post-date the compressional event.

Forlandsundet Graben represents one of three exposed areas of Tertiary sediments within the West Spitsbergen fold-and-thrust belt. The other two are found at Renardodden and at Øylandet (Fig. 1). The structural position, development and timing of the Forlandsundet Graben differ from those of the Tertiary Kongsfjorden Basin area to the northeast (Midbøe 1985) and the Central Basin to the southeast (Fig. 1). Eiken & Austegard (1987), using reflection seismic data, pointed to the existence of possibly similar offshore basins along more or less the entire west coast of Spitsbergen.

The Tertiary sediments on the east shore of Forlandsundet were first recognized and described by Holtedahl (1913), whilst those on Prins Karls Forland were discussed by Craig (1916) and Tyrell (1924). Atkinson (1962) documented the tectonic influence on the sedimentological evolution of the Forlandsundet Graben, and defined its setting as ‘taphrogeosynclinal’, thereby indicating an extensional origin for the basin. Birkenmajer (1972, 1981), Harland & Horsfield (1974) and Kellogg (1975) argued in favour of Oligocene post-orogenic collapse as the driving mechanism, whereas later workers, building on Harland (1969), have suggested that the Forlandsundet Graben developed in Eocene to Oligocene times, first in a transpressive setting, and later in transtensive and extensional tectonic regimes (Steel et al. 1981, 1985). In concert with the shear movements, compressional components of stress have influenced the deformation of the basin sediments (Lepvrier & Geyssant 1985; Lepvrier 1988, 1990).

Forlandsundet Graben is filled by an approximately 5000 m thick sequence of sediments (the Forland-sundet Group of Harland 1969; see Fig. 2). The basin is bordered by marginal alluvial fan sequences (Selvågen, Sarsbukta and Sarstangen Formations), which grade laterally into nearshore and shallow marine units (Sesshøgda, Reinhardpynten, Krokodillen and Mæhalslaguna Formations). To the north and in the central part of the basin the sub-marine fan-dominated Aberdeenflya Formation is found (Livsčič 1967; Rye Larsen 1982; Steel et al. 1985).

The rocks within the graben are reported to be mildly deformed (Rye Larsen 1982). NW–SE-trending folds of ‘apparent Tertiary age’, interpreted to be related to dextral wrench movements, were mentioned but not described in any detail by Kellogg (1975, p. 483).

The age of the rocks of the Forlandsundet Group has been estimated as Eocene to early Oligocene (Manum 1962; Livsčič 1974), but this is still disputed. The Sarsbukta Formation, which constitutes the youngest sequence (Fig. 2), has been dated to Oligocene, based on foraminiferæ assemblages (Feyling-Hansen & Ulleberg 1984), whereas dinoflagellae indicate a mid to late Eocene age for this unit (Manum & Thordsen 1986). The latter interpretation is supported by the recent study of palynomorphs (Nøttvedt et al. pers. comm). Schlüter & Hinz (1978) and Steel et al. (1985) have shown that post early Oligocene deposits are lacking in the onshore basins, and units of this age are found in offshore basins east of the Hornsund fault zone only.
The more fine-grained sediments of the Forlandsundet Graben are well exposed in cliffs up to 10 m high along the eastern coastline of Prins Karls Forland, whereas exposures in the conglomerates are abundant along the foot of the mountains bordering the strandflats on both shores of Forlandsundet. In the northern part of the area described here (Aberdeenflya and Fuhrmeisterstranda, Fig. 3), a broad strandflat separates the coastal cliffs from the mountain chain which forms the spine of the island. From a distance, exposures may seem very sparse on the strandflat, but closer inspection often reveals small exposures, or, where layers of sandstones are interbedded in the shales, even several hundred metres of continuous outcrops. Even though the rocks may be affected by congelifraction, the fragments are in situ, and, as seen from aerial photographs (e.g. Fig. 12), may still be used to identify structural elements.

The present description derives mainly from field data gathered during Norsk Hydro expeditions in 1989 and 1990. These data were supplemented by field check of some of the master faults in 1991. The work concentrated on Prins Karls Forland, particularly on Aberdeenflya, Fuhrmeisterstranda, Buchananryggen and between Krokodillen and Selvågen. The eastern graben margin was only visited on one occasion (Dahltoppen east of Sarstangen). The present work therefore emphasizes the western margin of the graben. For the present study two seismic lines have been released by Barents Oil.

Key localities and the position of the seismic lines are given in Fig. 3.

Structural description of the Forlandsundet Graben

The Forlandsundet Graben is approximately 30 km wide and 80 km in length. Principally it has a half-graben geometry in east-west sections, with changing polarities along strike (Fig. 4). As illustrated in Fig. 1 it is one element in a system of Tertiary graben structures situated along the western coast of Spitsbergen.

Structural architecture of the northern part of Forlandsundet Graben

The present study indicates that the Forlandsundet Graben has a more complex architecture than previously assumed. The exposed part of the basin may be subdivided into four graben units (in the context of Rosendahl et al. 1986 and Rosendahl 1987), of which the northernmost NW–SE-striking unit (graben unit A, Fig. 3) is delineated to the west by fault segment W1. The eastern margin of this graben unit is difficult to locate precisely, but may be defined by a submerged NW–SE trending fault (fault segment E1, proposed by Steel et al., 1985, their Fig. 11) situated north of the well-exposed fault segment E2. Unfortunately, no seismic or other data are available to the present authors to confirm the existence of fault segment E1.

The next graben unit to the south (graben unit B, Fig. 3) trends slightly west of north and is bordered by fault segments W2 and W3 to the west and fault segments E2 and E3 to the east. Of particular interest is fault segment E3, reported by Wójcik (1981), which seems to splay out from the easterly master fault system towards the graben axis. This indicates that a further structural subdivision of the graben into subunits may be justified.

Farther to the south, graben unit C is delineated by fault segments W4 and E4 to the west and east, respectively.
Fig. 4 illustrates the interpretations of two reflection seismic profiles across the Forlandsundet Graben between Selvågen and Sarsbukta (Fig. 4a) and between Brucebukta and St. Jonsfjorden (Fig. 4b) (for location of profiles, see Fig. 3). Line NP-FO-85/08 crosses graben unit C in a SW–NE direction. It shows that the strata have a general westerly tilt. A large extensional fault, with a flexure above it, occurs near the centre of the basin. It is natural to assume that the transition zone between the two graben units is to be found near the kink in the master fault trends, i.e. near Peter Winterbukta and Sarsbukta. Graben geometries documented to the south of Prins Karls Forland (Eiken & Austegard 1987) indicate that similar polarity shifts may also occur in graben units further south along the west coast of Spitsbergen.

The original total width of the Forlandsundet Graben may have been greater than presently defined by the master faults. For example at Geddesfjellet, conglomerates of the Tertiary Selvågen Formation appear with primary depositional contact to the folded and metamorphosed Hecla Hoek rocks. The depositional surface is tilted strongly and paralleled by post-depositional faults which belong to fault segment W4 of the present master fault system (see also Kleinspehn & Teyssier, this volume). This suggests that the present outline of the graben is at least partly defined by faults which have been active after the deposition of the Selvågen Formation. Thus, the true master faults (if they existed) which delineated the Forlandsundet Graben at the time of deposition of the Selvågen Formation should be sought west of the present graben margin, although not very far, as indicated by the immaturity of the conglomerates. It is noted that similar relations are reported from Renardodden (Dallmann 1989; Norsk Hydro pers. comm. 1991).

Both graben margins are associated with a pronounced normal drag, but this is best developed in the west.

**The master fault systems**

Regionally, the present extension of the Forlandsundet Graben is defined by NNW–SSE-striking faults separating the Tertiary clastic sediments from basement rocks ('Hecla Hoek'). The master faults dip steeply (Atkinson 1962; Livšč 1974; Lepvrier 1990) and are composed of systems of en échelon structures (Wójcik 1981; Morris 1989). In some cases termination of the fault segments coincides with NW–SE-trending semi-regional faults (Hjelle & Lauritzen 1982, see Fig. 3).

The exposed part of the Forlandsundet Graben is structurally dominated by the pronounced normal drag along the graben margins (Fig. 4). For the western part of the graben, this is reflected in the plots of bedding planes, as shown in Fig. 8. The average maximum dip near the master faults is in the order of 40–50°, but may locally approach vertical orientation, as described by Wójcik (1981) at Kafføya near Gråfjellet. In contrast, the graben floor is rather flat in the more central part of the basin (Fig. 4).

As seen from the following descriptions, strong deformation, partly brecciation, is common in the footwalls of the master faults along the western as well as the eastern margins of the Forlandsundet Graben. It is emphasized that the bulk of this deformation is believed to be pre-Tertiary, and probably Caledonian, as demonstrated in Fuglehukfjellet by Manby (1986). This indicates that the present margins of the Forlandsundet Graben are old, rejuvenated zones of weakness.
Fig. 3. Structural key map with localities. In addition to present field data, data from Atkinson (1962), Wójcik (1981), Hjelle & Lauritzen (1982), Manby (1986) and Morris (1989) have been used in the compilation. Fault segments and graben units A, B, C and D (see text), and location of lines NP-FO-85/08 and NP-FO-85/11 (see Fig. 4) are indicated. Arrows near the seismic profiles indicate dip direction of sediments in the graben units. Letters indicate tectonic subareas. Stereoplots of data from the subareas are given in Fig. 8.
This point has been stressed also in previous literature (Lepvrier 1990).

**Western margin.** – Along the western graben margin five separate fault segments may be identified. All master fault segments are commonly paralleled by small faults with abundant horizontal or oblique slickenside lineations.

**Fault segment W1** defines the border of the northernmost part of the presently exposed graben margin. It strikes NW–SE to N–S, and runs along the base of the mountains west of Aberdeenflya southwards to the southwestern shore of Richardlaguna (Fig. 3). At the base of Fuglehukfjellet in the north, the Hecla Hoek rocks close to the fault consist of heavily brecciated reddishly weathering quartzites of the Fuglehukfjellet Formation of the ‘Grampian Group’ as defined by Harland et al. (1979), cut by white quartz veins (Fig. 5). The brecciation is believed to represent a syn-metamorphic Caledonian deformational event (Manby 1986), and supports the view that the master faults have had a history also prior to the Tertiary deformation. In this locality, exposures in the brecciated rocks and undeformed Tertiary sediments on Aberdeenflya, which here consist of sandstones with occasional conglomeratic horizons and mudstones, suggest that the contact between the Hecla Hoek rocks and the Tertiary sequence is situated a few hundred metres east of the foot of the mountain. The Tertiary beds in these localities typically dip 40–50°ENE.
Fault segment W2 trends NNW–SSE and is identified in the area southwest of Richardlaguna southwards to Buchananryggen (Fig. 3). South of Richardlaguna the fault disappears beneath Murraybreen. It has been suggested to reappear at the shore of Grimaldibukta (Atkinson 1962, his Fig. 1), but has also been indicated to run off the coast in this area by Hjelle & Lauritzen (1982). In the southerly part of Buchananryggen, the fault seems to split into two (?) major branches, thus defining an isolated sliver of Tertiary conglomerates bordered on both sides by Hecla Hoek rocks. This implies that the master fault delineating the Hecla Hoek phyllonites against the continuous Tertiary sequence is found rather far to the east on Buchananryggen.

Strongly deformed Hecla Hoek rocks are seen in river creeks east of Laurantzfonfjellet. At Laurantzfonfjellet itself, phyllites, probably of the Kaggen Formation of the Scotia Group (Harland et al. 1979), are exposed along the nearly N–S-striking fault. The contact between the Tertiary sediments and the Hecla Hoek rocks is complex; east of Laurantzfonfjellet an abrupt, faulted contact is seen, and no drag effects in the Tertiary sequence are observed (Fig. 6a). In a couple of small exposures on Buchananryggen, however, conglomerates are found to lay unconformably on the top of the basement, demonstrating that also here there is a primary depositional contact which was later affected by faulting.

The Tertiary strata on Trocaderostranda immediately east of Buchananryggen are affected by open to close NW to N, and W-plunging folds (Figs. 7 and 8).

Fault segment W3 trends southwards from Buchananryggen below Søre Buchananisen to Krokodillen (Fig. 3). Three major N–S striking faults, of which the two western are extensional and the eastern probably compressional,
Fig. 8. Stereoplots of poles of bedding planes, fold axes and poles of fractures, Tertiary, Prins Karls Forland. (1) All data, Prins Karls Forland, (2) all data northern Prins Karls Forland; Aberdeenfløya and Fuhrmeisterstranda, (3) all data southern part of mapped area, (a) Aberdeenfløya, (b) Fuhrmeisterstranda, (c) Buchananryggen/Trocaderostranda, (d) Seshægda/Geddesfjellet/Peter Winterbukta/Krokodillen, (e) Seshægda, (f) Geddesfjellet, (g) Peter Winterbukta, (h) Krokodillen. Numbers and letters refer to locations given in Fig. 3. Lower hemisphere, equal area net.
cross the ridge of Krokodillen. Also Atkinson (1962) and Rye Larsen (1982) indicated that the master fault splays into several major branches in this area, with major strands cutting into both the hanging wall and the footwall.

The westernmost fault in this system defines the contact between the Hecla Hoek rocks and Selvågen Formation. The dips of the Tertiary strata are generally 40° to 70° to the SE or E (Fig. 8). In the present interpretation, the western fault on Krokodillen has been linked with the eastern fault on Buchananryggen to constitute Segment W3, because both these faults separate the Tertiary rocks from the basement. However, no conclusive evidence for this correlation exists, and other alternatives for the correlation may be just as likely.

**Fault segment W4**, which is situated between Geddesfjellet and Peterbukta, defines a NW–SE-striking strand of the western master fault system. On Geddesfjellet, where the fault-contact traditionally has been drawn, the conglomerates of the Selvågen Formation rest unconformably on Hecla Hoek chloritic schists (Fig. 6b), indicating that, during deposition, the synsedimentary master fault was located to the west of the present western limit of the Tertiary graben sediments. The tilted contact locally strikes NW–SE in this locality. On a larger scale the Tertiary strata are seen to dip 45–50° to the east, i.e. towards the present graben axis.

The contact at Sesshøgda farther to the southeast is more obscure, and a combination of NW–SE- and N–S-trending faults has been suggested (Rye Larsen 1982). The present authors were unable to identify the N–S-trending fault in the field. On the other hand, two separate branches of the NW–SE-trending fault were observed. Farther to the southeast Segment W4 crosses Ferrierryggen and probably goes offshore at Peterbukta (Fig. 3).

**Fault segment W5**, Atkinson (1962) proposed the existence of a N–S-trending fault in Scotiadalen. Discussion has been going on in the literature on the interpretation of this structure (Hjelle et al. 1979); Morris (1989) interpreted it as a right lateral strike-slip Palaeocene–Eocene fault composed of a large number of left-stepping Riedel shears. The Scotiadalen fault links up with fault Segment W4 south of Selvågen, and is accordingly included as Segment W5 in Fig. 3. It should, however, be noted that there is stratigraphic continuity in the Hecla Hoek rocks in Scotiadalen, and that no large displacement across this fault is possible (G. M. Manby pers. comm. 1991).

**Eastern margin.**— Also the eastern margin of the Forlandsundet Graben is segmented. The vertical separation on the eastern margin is estimated to be in excess of 700 m (Harland & Horsfield 1974). East on Sarstangen (Fig. 3), NNW–SSE to NW–SE-trending fault branches are separated by ENE–WSW to NE–SW possibly younger faults (Birkenmajer 1972; Wójcik 1981).

**Fault segment E1** is not observed onshore, but is inferred to run NW–SE along the coastline north of Engelsbk Bukta (Steel et al. 1985). The present authors did not have data to confirm this interpretation.

**Fault segment E2** trends NNW–SSE and is suggested to run from Kapp Graarud and southwards at the base of Alexanderfjellet and Kuppelryggen to Cissybreen (Wójcik 1981).

**Fault segment E3**. According to Wójcik (1981), the fault zone makes a pronounced westerly step towards Sarstangen, and trends WNW–ESE towards the northern margin of Aavatsmarkbreen. Wójcik (1981, p. 33) also reports Quaternary fault activity along this segment. During a visit at this locality in 1991, the present authors were unable to confirm the Quaternary faulting reported by Wójcik (1981). These faults are therefore not included in Fig. 3.

At Gråfjellet and Dahltoppen the master fault cuts well into the Hecla Hoek rocks, which are characterized by pre-Tertiary (?Caledonian) folding and brecciation (Fig. 6c). Slickenside striae on fault surfaces in the Tertiary conglomerates indicate both strike-slip (dextral) and oblique movements. Faults parallel to the graben margin are developed also in the Tertiary conglomerates, which typically reveal normal drag towards the fault zone. Here, the westerly dips of the beds change from approximately 20° to (locally) 70° within a few hundred metres.

**Fault segment E4** trends NNW–SSE from the southern margin of Aavatmarkbreen towards Farmusundet. Even this system is offset by a number of transverse faults according to Wójcik (1981).

**Compressional structures in the Tertiary graben sediments**

Even though the regional dip in the Tertiary sediments along the western margin of the Forlandsundet Graben is relatively consistent, folding is not uncommon. This deformational style is best exposed in the shore sections. The folds are seen on two scales. Large, gentle, upright folds with amplitudes up to a few tens of metres and wavelengths of several hundred metres are seen in shore sections and on the strandflats, e.g. at Furhmeisterstranda and Reinhardtpynten. In addition to these large-scale gentle structures more intense folding characterized by amplitudes in the order of 1 to 5 m and wavelengths of up to a few tens of metres are seen. These are frequently associated with reverse faulting or thrusting.

Folds and reverse faults or thrusts seem to be developed in two settings:

1. A complicated deformational pattern is seen in areas where there are major steps in the graben margin. This is particularly evident at Buchananryggen where the master fault, according to our interpretation, splits into several major fault branches, and an easterly step in the fault system is indicated. It is noted, however, that the step may be associated with two generations of fractures where (older) N–S-trending faults are cross-cut by
Fig. 9. (a) Schematic cross-section with reverse south-vergent faults south of Rottenburgpynten. (b) Schematic cross-section with reverse north-vergent faults at Fuhrmeisterstranda. Continuous lines are observed, stippled lines are inferred. Scales are approximate.

(younger) ENE–WSW trending faults. Also, the Tertiary sediments at Trocaderostranda, which is situated immediately east of this step (Fig. 3), are deformed by steep reverse faults and associated folds. As shown in Fig. 8, the fold axes plunge towards the N and NW.

At Krokodillen, which consists of shales interbedded with sandstones of the Krokodillen Formation (Livsic 1967; Rye Larsen 1982), strong deformation is recorded. Several approximately N–S-trending faults cross the ridge of Krokodillen. One of these seems to be a thrust fault with a dip of 35–40° to the SE and with northwesterly vergence. As the fault more or less follows the contact between a sandstone bench and the shales within the Krokodillen Formation, the amount of movement is difficult to determine.

(2) NW–SE-trending folds with SE-plunging axes associated with reverse faults or small thrust faults are commonly exposed along the eastern coastline of Prins Karls Forland (Fig. 8). Examples of the deformational style from Fuhrmeisterstranda and the area south of Rottenburgpynten are shown in Figs. 9a and b, 10 and 11. Folds and thrusts appear in zones separated by undeformed areas, suggesting that the deformed zones represent fault branches ramping up-section from a deeper floor fault, and that the steep reverse faults define separate horses within a duplex. It should be noted that northerly as well as southerly vergence of these structures are observed, and that direction of transport may vary considerably (Fig. 3). However, a tendency exists for vergence towards the graben margin to dominate, although E–W to NE–SW-trending structures with northerly as well as southerly vergence are seen (Fig. 9).
Fractures

The compressional structures are crossed by several generations of steep fractures characterized by extensional, oblique and also shear movements. Fracture sets parallel to the graben margins are common, but this simple pattern is complicated by conjugate sets of WNW–ESE and ENE–WSW-trending fractures.

Shear fractures and oblique fractures are frequent. One system of shear fractures has been mapped close to the master faults, and runs sub-parallel to them. Horizontal and oblique slicken-side striae are common on the fracture surfaces, which are frequently crossed by more or less E–W-trending fractures, also with sub-horizontal or oblique slickensides.

On the strandflats at Aberdeenflya and Fuhrmeisterstranda, the Tertiary strata are cut by faults with apparent strike-slip components. These faults seem to constitute conjugate NW–SE and NE–SW-trending sets, which are easily mapped from aerial photographs (Fig. 12) and on the ground (Fig. 13). Typical apparent horizontal separations are in the order of 5–30 m. As the fault planes are seen in horizontal sections only, it has not been possible to decide whether or not these structures are true strike-slip or oblique faults, or if they are extensional non-vertical faults developed in a tilted rock sequence. However, in exposures along the shore, indications of strike-slip movements along minor fractures with similar trends are abundant.

Fractures with sub-parallel trends (WNW–ESE and NE–SW to ENE–WSW) are also seen northeast of Geddesfjellet. Clear indications of shear are seen where these fractures cross the conglomerates of the Selvågen Formation. Unfortunately, the present data are insufficient to make any firm interpretation of these fracture systems, but it is noted that the fractures have a surprisingly fresh appearance to have suffered strong glacial erosion, and that the relief across the faults is frequently preserved (Fig. 14).

Extensional structures are seen in the coastal exposures, and can, in some cases, be followed as topographic depressions a few tens of metres inland. The extensional faults are usually steep (dips of 65–75° are common), are frequently seen to strike N30°–75°E, and clearly post-date the compressional structures (Fig. 9b). They have a vertical separation from 0.5 to 2 m. Relative to the systems of minor fractures, the extensional faults mapped are few, and they have little impact in the stereoplots presented in Fig. 8.

Soft-sediment and syn-depositional deformation

Soft-sediment and syn-depositional deformation are seen in several places in the finer-grained parts of the Tertiary sequences. These structures include flow in mudstones as well as slumping and growth faulting. The general deformational style and the lack of mineralization on fault surfaces make it possible to separate the soft-sediment or syn-depositional deformation effects from the tectonic structures.

Structural synthesis

Dog-leg geometry, as seen in the Forlandsundet Graben, is common in most graben systems (e.g. Johnson 1930; Freund & Merzer 1976; Barry 1984; Harding 1984; Mulugeta 1985) and should not, as such, be taken as definite proof of a strike-slip origin of the structure. Gibbs (1984), Bosworth et al. (1986), Rosendahl et al. (1986) and Rosendahl (1987) have demonstrated that, depending on geometries, pattern of over-lap and vergence of the master faults, dog-leg geometries may be associated with either shifting or constant polarity of graben units. The graben units are connected by transfer faults (Gibbs 1984) or accommodation zones (Rosendahl et al. 1986; Rosendahl 1987).

As shown in Fig. 4, shift in polarity occurs between graben units C and D, and a general subdivision of the basin into graben units may be defended. According to this model, accommodation zones should be looked for in the areas around Murreyppynen, Geddesfjellet and south of Peterbukta on the western margin, and near Engelskbuhta, possibly Sarstangen, and around Farmssundet along the eastern margin of the graben. At the present stage, these structures have not yet been analysed in detail, but the most obvious candidate for an accommodation zone is the area between Murreyppynen and Sarstangen (Fig. 3).

However, until a detailed investigation of reflection seismic data from within the graben has been carried out, it is not possible to decide whether the general structure is built in an overlapping or non-overlapping, opposing half-graben configuration, in the context of Rosendahl et al. (1986) and Rosendahl (1987).

Investigations on Prins Karls Forland demonstrate that folding has occurred, and that NW–SE striking
axial traces are dominating, whereas the vergences of reverse faults and thrusts vary more (Fig. 3). It is suggested that the thrusts and reverse faults are subdivided into two genetic groups in the investigated area. *Type 1-structures* strike at a large angle to the graben margin and are characterized by both northerly and southerly vergence. *Type 2-structures* are associated with transport towards the western graben margin.

It may be reasonable to associate the generation of these structures with a strike-slip event and moderate basin inversion, respectively. The reverse faults and thrusts are frequently seen to occur in groups, indicating that they are parts of duplexes associated with ramps from deeper sub-horizontal floor faults.

Most fold axes plunge towards the southeast or east, indicating a possible rotation in connection with the normal drag developed along the present graben margin. This suggests that the drag was developed after the compressional event. If this assumption is correct, and an average dip of 35° associated with the later drag is assumed, removal of this effect in a stereoplot shows that the original orientation of most of the fold axes was nearly horizontal NW–SE to E–W.

As shown in the previous section, a set of extensional fractures post-dates the compressional structures, indicating a final relaxation of the compressional component in the stress system.

**Discussion**

Palaeontological constraints on the timing of the tectonic stages affecting the Tertiary West Spitsbergen are scanty (e.g. Atkinson 1962; Steel et al. 1985) and, hence,
interpretations rely heavily on plate tectonic reconstructions of the Norwegian-Greenland Sea. These models (Talwani & Eldholm 1972, 1977; Myhre et al. 1982; Eldholm et al. 1984, 1987; Myhre & Eldholm 1988) suggest that transpression dominated in the northern part of the Hornsund fault zone at anomaly 23-time (54 ma; earliest Eocene), whereas further south the Senja Fracture Zone acted as a leaky (transtensional) transform at the same time. The transtensional regime spread northwards and reached the southern tip of Spitsbergen at about anomaly 21-time (49.5 ma; mid-Eocene) when the tectonic activity as associated with a compressional component peaked. At anomaly 13-time (36 ma; early Oligocene) a shift in the relative plate movements brought an end to the shear movements (Eldholm et al. 1987; Nøttvedt et al. 1988a; Müller & Spielhagen 1990).

Although some of the interpretations upon which these ideas are based may at present be under revision (Andresen et al. 1988; Maher & Craddock 1988; Nøttvedt et al. 1988b; Maher et al. 1989; Bergh & Andresen 1990; Haremo & Andresen in press), it has been generally accepted that dextral shear associated with the break-up of the Laurasia as the Greenland and Eurasian plates were sliding past each other in the Tertiary was also responsible for the deformation of western Spitsbergen and development of the West Spitsbergen fold-and-thrust belt (e.g. Harland 1965, 1969; Lowell 1972; Harland et al. 1974; Kellogg 1975). Within this plate-tectonic framework, and bearing the available datings of the sedimentary fill of the basin in mind (see introductory paragraphs for a short summary), it seems obvious that the Tertiary Forlandsundet Graben was initiated during a phase of prevailing regional dextral transpression.

To explain its subsidence, Steel et al. (1985) proposed a dextral transpressional situation in late Palaeocene to Eocene times, prior to formation of the Forlandsundet basin. Deformation peaked in mid-Eocene times with formation of the West Spitsbergen fold-and-thrust belt (see also Steel & Worsley 1984). In this model, the transpressional system switched back to tension in late Oligocene via a period of late Eocene–mid-Oligocene transtension during which the Forlandsundet Graben was formed, probably in a releasing bend position.

The large apparent stratigraphic thickness compared with the vertical thickness of the basin fill was taken as a suggestion of strike-slip origin. This was supported by studies of conglomerate clasts in the Selvågen Formation that may have been transported at least 3 km along strike of the graben margin from their source area (Rye Larsen 1982). However, recent studies show that stratigraphical thicknesses, as measured from reflection seismic data, correspond roughly to maximum stratal
thick (Nøttvedt et al. this volume). Also, as stated by Steel et al. (1985), since the sedimentological evidence for a transcurrent component is sparse, the regional tectonic setting of the basin may be the strongest argument for a strike-slip origin.

Based on fracture studies Lepvrier & Geyssant (1985) and Lepvrier (1988, 1990) suggested dextral transpression to have dominated the Forlandsundet area in late Cretaceous times, switching locally to sinistral in early Palaeocene. Further, by clockwise rotation of the maximum horizontal stress axis from approximately N20°N to N70–80°N, compression took over. A relaxation of the horizontal stress then took place, and the compressional regime was followed by extension from early Eocene time on. For the dating of the Tertiary events Lepvier (1990) partly relied on correlations to a parallel study of the Central Basin (Kleinspehn et al. 1989), where sediments as old as Palaeocene are preserved. It is noted here that the sequence of events described by Lepvier (1990) is in accordance with our structural observations. However, since the available biostratigraphical data suggest that the sediments of the Forlandsundet Graben were deposited during the Eocene, we will try to use this time-framework in the following.

It should also be mentioned that vitrinite reflectance data (Manum & Thronsdøen 1986; Kleinspehn & Teyssier, this volume, Norsk Hydro pers. comm. 1991) suggest that differential subsidence (or differential uplift) has characterized the graben development. Thus, the rocks of the western margin may have been buried several thousand metres deeper than the sediments of the eastern margin. This is supported by the two seismic lines (Fig. 4) which demonstrate that the sequence exposed along the eastern shores of Prins Karls Forland is situated stratigraphically higher than those in the Selvågen area.

The present investigation has suggested that the development of the Forlandsundet Graben may be subdivided into four stages:

1. During the initial stage deposition took place in a basin somewhat broader than that now delineating the Forlandsundet (Fig. 15a). This is suggested by the rotated, primary sedimentary contacts between the Hecla Hook rocks and the Tertiary conglomerates, which were later to be cut by the younger master faults of the present Forlandsundet Graben. It is not known whether or not an active fault system existed outside the present graben area at this time. We have no evidence from within the Forlandsundet Graben to indicate the tectonic setting of the basin at this time, but from regional evidence a dextral strike-slip regime is proposed.

According to Steel et al. (1985) the initiation of the sedimentation in Forlandsundet Graben may have taken place in (early?) Eocene time. This is in accordance with the present investigation, which has shown that the conglomerates of the Selvågen Formation occur in a position stratigraphically lower than that of the Sarstangen Formation (Fig. 4, see also Nøttvedt et al. this volume). The Selvågen Formation has been dated to be ‘not younger than Eocene’ (Manum 1962; Livsč 1965, 1974; Feyling-Hansen & Ulleberg 1984). It is also noteworthy that Kleinspehn et al. (1989), based upon paleostress investigations in the Central Basin, suggested a phase of extension (or transtension) in early Eocene time.

2. Following the initial stage, the early phase of graben formation involved rapid subsidence and the establishment of (?) discontinuous fault scarps in a generally right-stepping configuration of graben units (Fig. 15b).

Even though it is acknowledged that dextral shear was prevailing on a regional scale, it seems clear that an extensional component contributed to the basin development at this stage. This may be supported by the general architecture of the present graben. Steel et al. (1985) preferred a model which involved extension in the lee of a wrench fault curvature around a resistant land mass to explain this stage in the formation of the Forlandsundet Graben. However, when the extension of similar graben structures along the coast of Spitsbergen is considered (Eiken & Austegard 1987), a transtensional event of more regional significance may seem more likely. Also, in a system of right-stepping faults exposed to dextral shear, a series of separate pull-apart basins should be expected (Rodgers 1980; Aydin & Nur 1982, 1985). So far, isolated basins associated with the early phase of basin formation have not been observed (Eiken & Austegard 1987).

Dextral shear would be consistent with subsequent development of the observed NW–SE to NE–SW striking folds and faults within the graben (type 1-structures, see ‘Structural synthesis’) as well as the complications and strain concentration which is seen in the vicinity of steps in the master fault systems. On the other hand, many of these structures are characterized by northerly vergence. It may be speculated as to whether a local phase of sinistral shear (as proposed by Lepvrier (1990) and tentatively dated to Palaeocene by him) has contributed to this complex pattern. Tilt of the graben floor may be associated with this event, or alternatively with event (4) (see below).

It may be concluded that, even though the basin formation took place in a dextral transcurrent environment, the total amount of strike-slip in the Forlandsundet Graben was probably not very large, possible a few kilometres.

3. It is proposed that the stress changed to transpressional or compressional at the stage when the type 2-structures were developed, possibly in late Eocene–early Oligocene times (Fig. 15c). This led to narrowing of the basin, and a mild phase of inversion, which may have been noticeable only in the vicinity of the master faults along the graben margin.

Again, it is noted that a compressional event with a N70°–80°E oriented maximum horizontal axis has been reported by Lepvrier (1990).

4. The extensional structures that are seen to crosscut compressional structures may represent the transition
Fig. 15. Schematic development of the Forlandsundet Graben. Shadowed area illustrates anticipated area of deposition. Note that the basin area extended outside the present graben as defined by its present master faults at the stage of initiation. See text for further explanation.
to transtension and later extension to prevail during the remaining part of the Oligocene (Fig. 15d). The flexuring of the basin margins is attributed to this event. Klein­sphehn et al. (1989) and Lepvrier (1990) reported a N–S directed extensional phase to have been active in the Oligocene.

Conclusions

The present structural investigation of the Forlandsundet Graben has revealed no unequivocal evidence for the mechanism of initiation of the graben formation. However, local dextral transtension in a regional dextral transpressional regime is considered the most likely envi­ronment. There are also indications that subsidence started in (?)-early Eocene times, resulting in a basin that may have been somewhat wider than the present Forlandsundet Graben.

During continued deformation the northern part of the basin was subdivided into at least four graben units, probably with opposite polarities in an overall right-step­ping graben configuration. As can be seen from sedimentology ev­idence (Rye Larsen 1982; Steel et al. 1985) and the present structural analysis, indications are that the horizontal component of the movements were subordi­nate. A reasonable estimate may be in the order of a few kilometres. The graben sediments were then de­formed by folding, reverse faulting and thrusting in a continuing dextral strike-slip environment, possibly with a phase of sinistral shear.

A late Eocene (?)-early Oligocene episode of compres­sion led to mild inversion of the basin margins before accelerated subsidence resulted in the development of large-scale drag structures along the graben margins. During this deformation the older border faults acted as hinge-lines. This deformation, which may correspond to the transition from dextral transpression to transtension on a regional scale, was accompanied, or followed, by extensional faulting within the graben itself.

It is acknowledged that the resolution of the tectonic events within the Forlandsundet Graben is no better than that of the available biostratigraphical data. It is also realized that dateable fossils are rare, and that detailed dating of the sediments is difficult. Nevertheless, it is suggested that future structural geological studies in the area should be accompanied by a dating programme to obtain a good correlation to the plate tectonic events that governed the development of the Forlandsundet Graben.

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Structural outline of the Forlandsundet Graben

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