Structural development along the Billefjorden Fault Zone in the area between Kjellströmdalen and Adventdalen/Sassendalen, central Spitsbergen

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The Billefjorden Fault Zone represents a major lineament on Spitsbergen with a history of tectonic activity going back into the Devonian and possibly earlier. Recent structural, sedimentological and stratigraphical investigations indicate that most of the stratigraphic thickness variations within the Mesozoic strata along the Billefjorden Fault Zone south of Isfjorden are due to Tertiary compressional tectonics related to the transpressive Eocene West-Spitsbergen Orogeny. No convincing evidence of distinct Mesozoic extensional events, as suggested by previous workers, has been recognized. Tertiary compressional tectonics are characterized by a combined thin-skinned/thick-skinned structural style. Décollement zones are recognized in the Triassic Sassendalen Group (Lower Décollement Zone) and in the Jurassic/Cretaceous Janusfjellet Subgroup (Upper Décollement Zone). East-vergent folding and reverse faulting associated with these décollement zones have resulted in the development of compressional structures, of which the major are the Skolten and Tronfjellet Anticlines and the Adventelva Duplex. Movements on one or more high angle east-dipping reverse faults in the pre-Mesozoic basement have resulted in the development of the Juvdalskampen Monocline, and are responsible for out-of-sequence thrusting and thinning of the Mesozoic sequence across the Billefjorden Fault Zone. Preliminary shortening calculations indicate an eastward displacement of minimum 3–4 km, possibly as much as 10 km for the Lower Cretaceous and younger rocks across the Billefjorden Fault Zone.

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The Billefjorden Fault Zone (Fig. 1), one of the most prominent lineaments of Spitsbergen, displays a complex tectonic history going back into the Precambrian-Early Palaeozoic (Orvin 1940; Harland et al. 1974).

The early Caledonian tectonic events (Ny-Friesland and Svalbardian Orogenies) probably include both compressional and strike slip movements. However, the relative importance and magnitude of these tectonic events have been debated (McWhae 1953; Harland 1965; Harland et al. 1974; Harland et al. 1984; Lamar et al. 1986; Reed et al. 1987). Renewed movement along the Billefjorden Fault Zone during Early/Middle Carboniferous time, as evidenced by erosion and graben formation along the fault zone (Orvin 1940; Cutbill & Challinor 1965; Gjelberg & Steel 1981), resulted from the establishment of a regional extensional setting on Spitsbergen during Early Carboniferous time (Birkenmajer 1975; Steel & Worsley 1984).

From Permian to Palaeocene time only minor extensional movements appear to have taken place across the Billefjorden Fault Zone, sup-

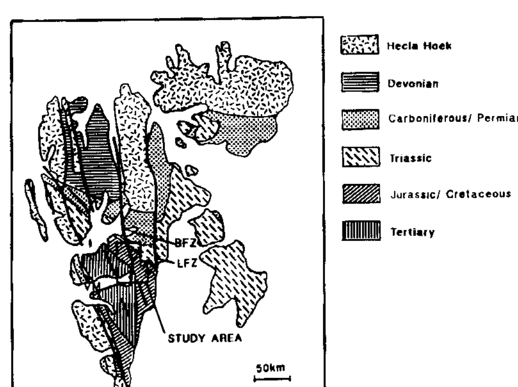


Fig. 1. Simplified geological map of Svalbard. BFZ: Billefjorden Fault Zone, LFZ: Lomfjorden Fault Zone, I: Isfjorden, VM: Van Mijenfjorden.

porting the idea that rather stable platform conditions prevailed, as indicated by the sedimentary sequences present (Buchan et al. 1965; Parker 1967; Harland et al. 1974; Mørk et al. 1982; Steel & Worsley 1984; Dypvik et al. in press). A marked thickness reduction of the Jurassic and Cretaceous sequences along the Billefjorden Fault Zone south of Isfjorden has been interpreted as the result of normal (down to the west) movements along the Billefjorden Fault Zone during an early Cretaceous tectonic event (Parker 1966, 1967; Harland et al. 1974).

Tertiary tectonic movements along the Billefjorden Fault Zone have been known for more than a hundred years (De Geer 1866), and have been confirmed by several papers based on fieldwork in the area between Isfjorden and Van Mijenfjorden (De Geer 1919; Hagerman 1925; Orvin 1940; Livshits 1965; Parker 1966). However, neither the geometry nor the kinematics of the Tertiary structures along the Billefjorden Fault Zone have been thoroughly described. How these structures relate to the narrow belt of intense deformation along western Spitsbergen is also unclear.

This paper focuses on the post-Palaeozoic tectonic history of the Billefjorden Fault Zone, and is based on the results of recent detailed, structural and stratigraphical mapping along the Billefjorden Fault Zone in the area between Adventdalen/Sassendalen and Kjellströmdalen (Fig. 1).

The main topics concerning the structural development of the study area have been: (1) To describe in detail the geometric and kinematic development of the Tertiary structures along the Billefjorden Fault Zone, and (2) to determine whether the observed thickness variations within the Mesozoic sequence are related to syn-sedimentary extensional tectonics or are a result of a later compressional event.

Area description and stratigraphy

The mapped area is characterized by high, steep mountains (up to 1,100 metres high) separated by broad valleys, of which Adventdalen, Reindalen and Kjellströmdalen are the most prominent. These mountains expose excellent east–west sections across the Billefjorden Fault Zone (Figs. 2 and 3). Bedrock exposures are fairly good, except in the valley floors, which are filled with glacial till, and in the glaciated areas, which cover about 10% of the study area (Fig. 2).

The rocks exposed in the mapped area have a gentle regional dip towards SW and progressively older rocks are exposed from south to north (Fig. 2). Mesozoic and Tertiary rocks cover most of the area. A thin zone of Permian strata, in Sassendalen, to the north-east, represents the only pre-Mesozoic rocks exposed (Fig. 2).

The sedimentary succession exposed in the study area consists generally of terrigenous clastics deposited in varying offshore marine, marginal marine to fluviodeltaic environments of Triassic to Tertiary age (Mørk et al. 1982; Steel & Worsley 1984; Dypvik et al. in press).

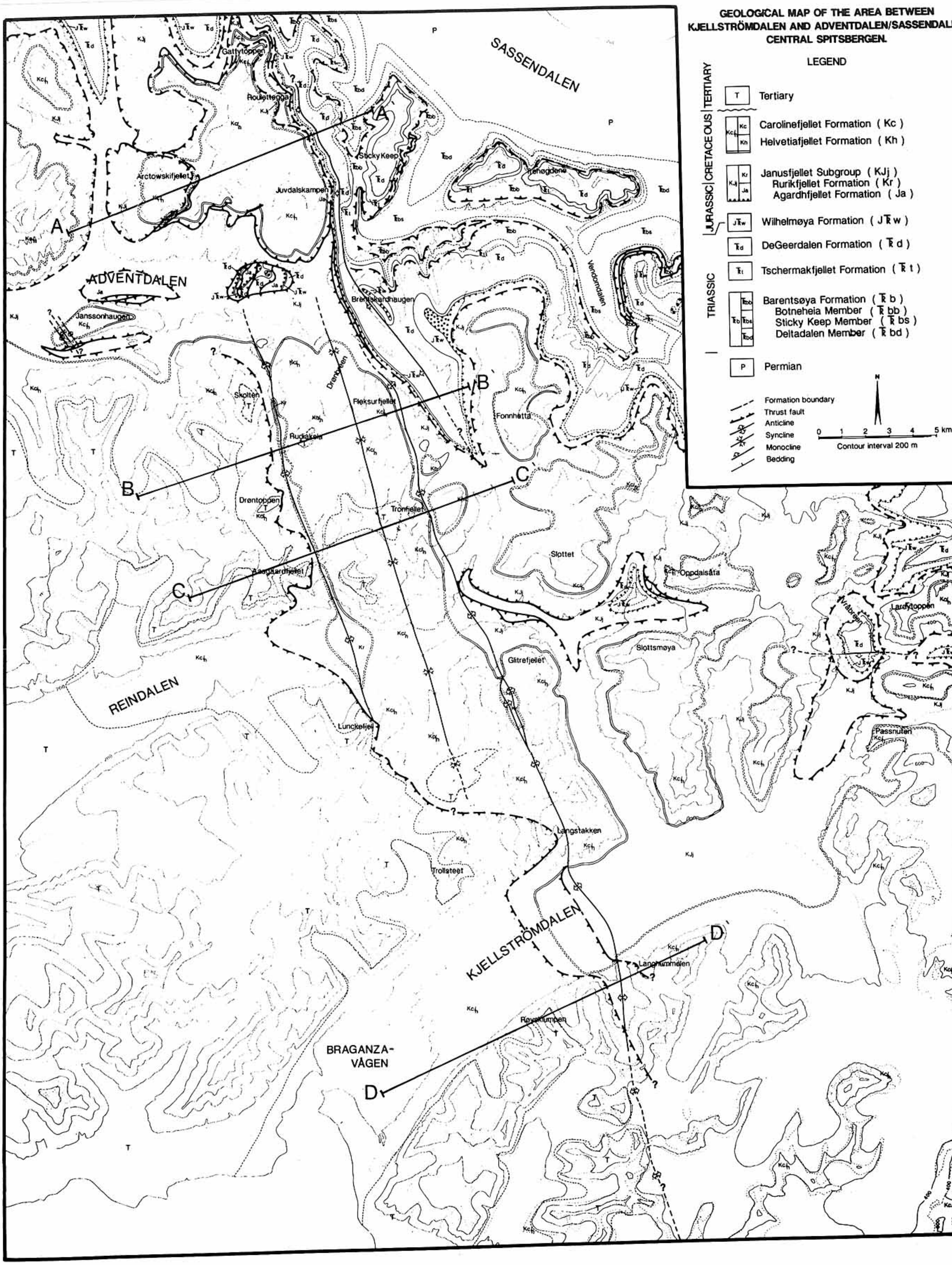
Sassendalen Group (Lower to Middle Triassic)

The Sassendalen Group is exposed in the north-eastern part of the mapped area. The lower part of the group is generally poorly exposed or covered by glacial till and rock scree in the lower hillsides southwest of Sassendalen (Fig. 2). The upper part of the group is represented by the dark, easily recognized, cliff-forming Botneheia Member (Fig. 2).

The Sassendalen Group comprises the Barentsøya Formation (Fig. 4; Mørk et al. 1982), consisting mainly of black shales and siltstones of Lower to Middle Triassic age (Buchan et al. 1965; Mørk et al. 1982). The thickness of the formation increases from about 300 m west of the Billefjorden Fault Zone to about 400 m east of the lineament. This thickness variation has been explained by periodic (down to the east) subsidence along the Billefjorden Fault Zone through Early and Middle Triassic time (Harland et al. 1974; Mørk et al. 1982).

The formation is divided into three members (Fig. 4; Mørk et al. 1982). The lower Deltadalen Member consists of greenish grey silty shales, with thin rippled and planar laminated sandstones grading up into shales with thin siltstone or very fine sandstone beds. Limestone concretions are common. The overlying Sticky Keep Member consists of dark shales with thin beds of calcareous siltstone, which weather yellowish. Calcareous nodules ranging from 10 cm to 1 m in diameter are common, and occur generally at distinct horizons.

Fig. 2. Geological map along the Billefjorden Fault Zone in the area between Kjellströmdalen and Adventdalen/Sassendalen.


GEOLOGICAL MAP OF THE AREA BETWEEN KJELLSTRÖMDALEN AND ADVENTDALEN/SASSENDALEN, CENTRAL SPITSBERGEN.
LEGEND
TERTIARY
T Tertiary
CRETACEOUS
Kc Carolinefjellet Formation (Kc)
Kh Helvetiafjellet Formation (Kh)
JURASSIC
Kr Janusfjellet Subgroup (KJj)
Rurikfjellet Formation (Kr)
Ja Agardhfjellet Formation (Ja)
Wilhelmøya Formation (JTRw)
TRIASSIC
DeGeerdalen Formation (TRd)
Tschermakfjellet Formation (TRt)
Barentsøya Formation (TRb)
Botneheia Member (TRbb)
Sticky Keep Member (TRbs)
Deltadalen Member (TRbd)
P Permian
Formation boundary
Thrust fault
Anticline
Syncline
Monocline
Bedding
N
0 1 2 3 4 5 km
Contour interval 200 m
SASSENDALEN
ADVENTDALEN
REINDALEN
KJELLSTRÖMDALEN
BRAGANZA-VÅGEN
Gattytoppen
Roulettegga
Arctowskifjellet
Juvdalskampen
Sticky Keep
Tenøgdene
Brentskarhaugen
Janssonhaugen
Skolten
Fleksurfjellet
Rudakela
Fonnhetta
Drøntoppen
Tronfjellet
Slottet
Oppdalsåta
Aasgaardfjellet
Slottsmøya
Glitrefjellet
Lunckefjell
Passnuten
Langstakken
Trollsteet
Langhummelen
Røyskampen
Vendomdalen
A
B
C
D

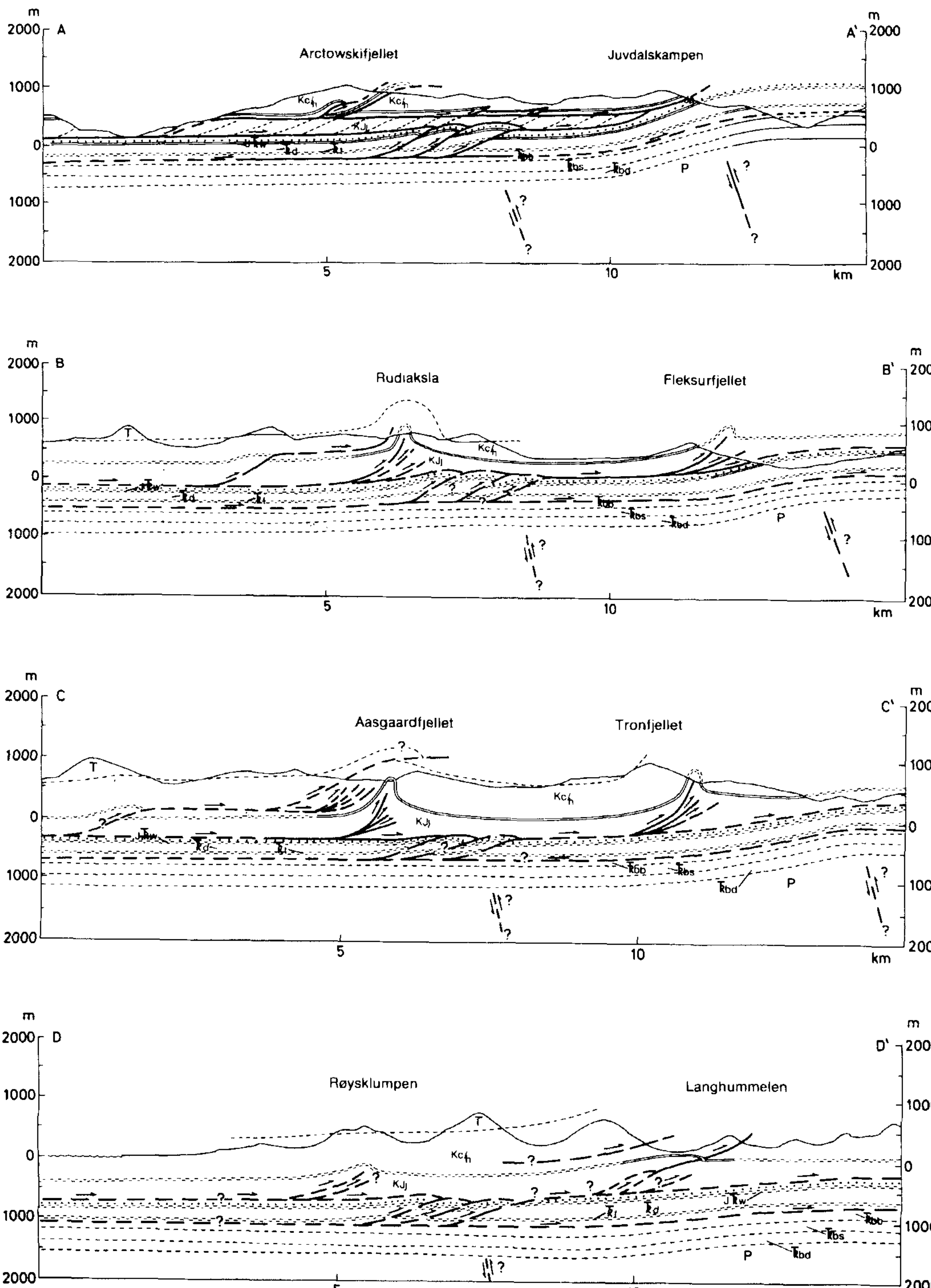


Fig. 3. Geometrical cross-sections across the Billefjorden Fault Zone. A) north of Adventdalen (Profile A on Fig. 2), B) south of Adventdalen (Profile B on Fig. 2), C) north of Reindalen (Profile C on Fig. 2), D) south of Kjellströmdalen (Profile D on Fig. 2). B: Brentskardhaugen Bed, F: Festningen Sandstone, P: Permian. For letter code see Fig. 2.

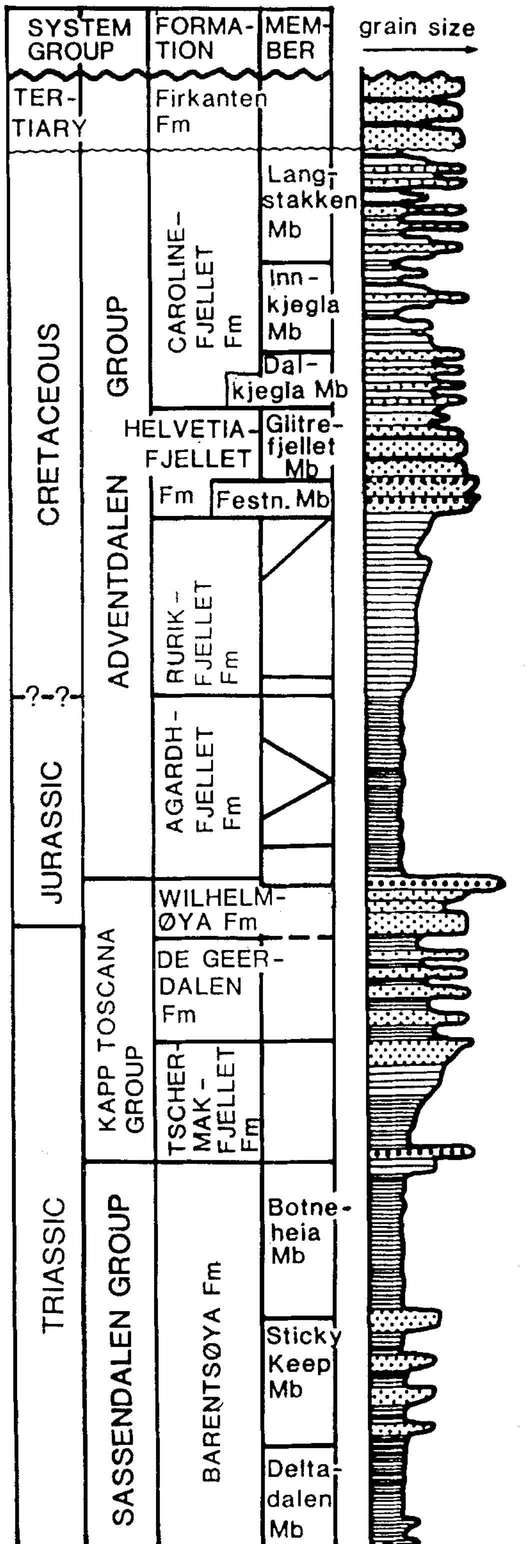


Fig. 4. Stratigraphy in the study area.

Siltstone beds are more common towards the top of the unit. The Botneheia Member consists of black shales with phosphate nodules. The shales are commonly papery and calcareous in the upper part of the unit. The middle part of the member consists of dark shales with a few thin calcareous siltstones. The upper part is highly calcareous and cliff forming. The top of the cliff is marked by a siltstone bed rich in phosphate nodules and trace fossils. This unit is overlain by grey shales with red weathering siderite nodules belonging to the Tschermakfjellet Formation.

Kapp Toscana Group (Middle Triassic to Lower Jurassic)

The Kapp Toscana Group is mainly exposed to the northeast (Fig. 2). It is easily recognized between the underlying, cliff-forming, uppermost part of the Sassendalen Group and the basal, phosphatic conglomerates of the overlying Adventdalen Group.

The Kapp Toscana Group has a total thickness of 300–350 m in the mapped area, and was divided into three formations by Mørk et al. (1982) (Fig. 4). The Tschermakfjellet Formation (Carnian to Norian) is generally not well exposed in the study area. It consists of 100–125 m of dark grey shales with siltstone interbeds in the upper part. Red-weathering sideritic concretions are common in the lower portion (Mørk et al. 1982). The formation contains a marine fauna of ammonites and bivalves, indicating an offshore marine to prodeltaic and deltafront environment (Mørk et al. 1982).

The De Geerdalen Formation, of upper Triassic (Norian to Rhaetian) age, consists of a 170–200 m thick sequence of dark, predominantly greenish shales and light brown sandstones characterized by both coarsening upwards and fining upwards sequences. Although previous investigators (Buchan et al. 1965; Mørk et al. 1982) have indicated a dominantly fluvial environment, our work suggests a brackish environment for most of the formation in the mapped area.

The Wilhelmøya Formation (Rhaetian to lower Jurassic) is made up of 15–45 m of medium grained, light grey, commonly strongly bioturbated sandstones interbedded with dark grey claystones. The formation was deposited in a shallow marine setting (Eikeland & Dypvik unpublished).

Adventdalen Group (Middle Jurassic to Lower Cretaceous)

The Adventdalen Group is exposed over much of the mapped area (Fig. 2). It consists of four formations (Fig. 4), with a total thickness of about 800 m. The easily recognized base of the group is placed at the base of the characteristic phosphatic Brentskardhaugen Conglomerate, which represents a hiatus comprising parts of the Lower Jurassic and much of the Middle Jurassic (Bäckström & Nagy 1985). The top of the group is marked by an angular unconformity towards the overlying Tertiary sequence.

The lower part of the Adventdalen Group, the Janusfjellet Subgroup (Dypvik et al. in press), consists of a 300–400 m thick sequence, dominated by black and grey shales deposited in a shelf/shallow marine environment. The subgroup is divided into the Agardhfjellet and Rurikfjellet Formations (Fig. 4).

The Agardhfjellet Formation (Bathonian to Volgian) has a thickness of about 150 m, but tectonically thickened and thinned sections characterize the sequence in the study area. The formation is divided into four members: the Oppdalen, Lardyfjellet, Oppdalsåta and Slottsmøya Members (Fig. 4). The Oppdalen Member consists of phosphatic conglomerates (Brentskardhaugen Bed), glauconitic sandstones (Marhøgda Bed) and upward fining sandy siltstones (Drønbreen Bed). The overlying Lardyfjellet Member is dominated by black paper shales with scattered carbonate concretions, while the overlying Oppdalsåta Member is made up of several upwards coarsening sands and sandy shale units. The Slottsmøya Member consists of black paper shales at some locations, with faint developments of coarsening upwards sequences in its upper part.

The Rurikfjellet Formation (Volgian to Hauterivian) has a thickness of about 150 m and is divided into the Wimanfjellet and Ullaberget Members. The Wimanfjellet Member consists in its lower part of the light grey to yellow soft clays of the Myklegardfjellet Bed (Fig. 4). The rest of the member is made up of silty grey to dark grey shales characterized by 'cannon-ball' carbonate concretionary horizons. The first appearing sand bed marks the transition to the overlying Ullaberget Member, which contains silty/sandy shales and several sandstone beds which are commonly crossbedded and bioturbated.

The fluviodeltaic Helvetiafjellet Formation (Barremian), overlying the Janusfjellet Subgroup, is about 100 m thick in the study area, and is characterized by light grey quartzitic sandstones. The formation consists of a lower Festningen Member and an upper Glitrefjellet Member (Parker 1967). The 5–15 m thick Festningen Member (Festningen Sandstone) consists of light grey, fine- to coarse-grained crossbedded sandstone, usually conglomeratic in its lower part. The Festningen Sandstone is easily distinguished as a characteristic cliff-forming sandstone above the shales of the Janusfjellet Subgroup. The Glitrefjellet Member consists of about 100 m of medium- to coarse-grained sandstones interbedded with shales and siltstones. The shales are commonly carbonaceous and the sandstones and siltstones contain carbonized plant detritus at several horizons.

The overlying shallow marine sediments of the 250–300 m thick Carolinefjellet Formation (Aptian-Albian) consist of alternating shales, siltstones and sandstones divided into the Dalskjegla, Innkjegla and Langstakken Members (Parker 1967; Fig. 4). The lower contact with the Helvetiafjellet Formation is not easily recognized. The boundary towards the overlying Tertiary siliciclastics is marked by a slight angular unconformity formed during the pre-Tertiary uplift of the Spitsbergen area.

The Tertiary

Tertiary rocks are preserved in the southwestern part of the study area (Fig. 2) in a succession consisting of alternating sandstone and shales of marine, estuarine and terrestrial facies (Major & Nagy 1972). The sequence is considered to be of Palaeocene-Eocene age (Ravn 1922; Major & Nagy 1972; Manum & Throndsen 1978, 1986; Steel et al. 1985). Due to the erosional level, only the lower part (up to 500 m) of the sequence is exposed in the study area. The Tertiary sequence preserved in the Central Basin of Spitsbergen contains as much as 2.3 km of clastic deposits. Manum & Throndsen (1978) claimed, based on vitrinite reflectance studies, that an additional overburden of 1.7 km has been eroded since the Eocene.

Fold and fault geometries along the Billefjorden Fault Zone

The most characteristic structures developed along the Billefjorden Fault Zone in the area studied are two N-S trending anticlines separated by a broad open syncline (Fig. 2). The anticlines are easily recognized in the field and their development have earlier been described by several workers (De Geer 1919; Hagerman 1925; Orvin 1940; Livshits 1965; Parker 1966; Harland et al. 1974). In the following text the two anticlines will be referred to as the Skolten (in the west) and Tronfjellet (in the east) Anticlines, respectively. The syncline is named the Drønbreen Syncline (Fig. 2).

Skolten Anticline

The Skolten Anticline is particularly well developed between Adventdalen and Reindalen (Fig. 2). In this area it is developed as a tight, slightly east-vergent anticline with an amplitude of 600–700 m (Figs. 2, 3b, c and 5). At Skolten, Drøntoppen and Rudiaksla the hinge zone is eroded (Fig. 2). However, the structure is thought to be closed at the Festningen Sandstone level. This is supported by observations to the south, at Aasgaardfjellet (Fig. 2), where the fold closure is exposed at an elevation of about 600 m (Festningen Sandstone level).

Structural data sampled along the Skolten Anticline between Adventdalen and Reindalen are presented in Fig. 6a, and demonstrate that the Skolten Anticline in this area has a trend of about N160°E and a plunge of less than 5° south-southeast.

The Skolten Anticline changes geometry across Adventdalen (Figs. 2 and 3a, b). North of the valley, at Arctowskifjellet, two distinct reverse-faults cutting upsection through the Helvetiafjellet and Carolinefjellet Formations accomplish the shortening seen as folds southward across the valley. The Festningen Sandstone ramps upsection approximately 300 m across the eastern fault (Figs. 3a and 7). Minimum shortening estimates across these faults are 1.5 km. The reverse faults are interpreted as splaying off a major sole thrust in the incompetent shales of the Janusfjellet Subgroup (Figs. 2 and 3a). Structural data suggesting

Fig. 5. The Skolten Anticline at Rudiaksla, viewed from the south (Drøntoppen). The Festningen Sandstone is developed as a tight, slightly east-vergent anticline. Compare with cross-section (Fig. 3b) for structural interpretation.

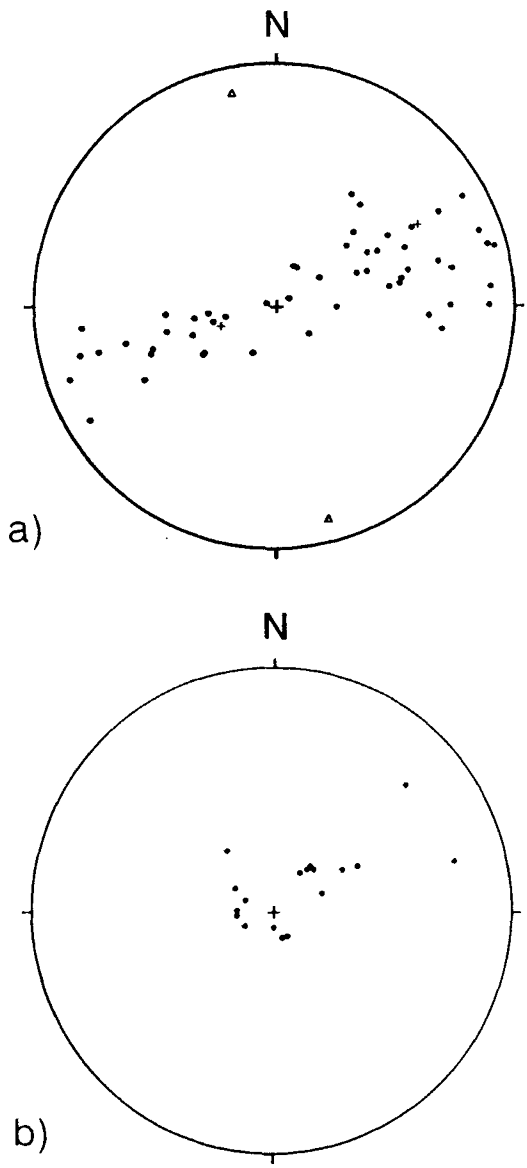


Fig. 6. Structural data collected along the Skolten Anticline. a) Adventdalen-Reindalen area, b) Arctowskifjellet area. Circles: poles to bedding, triangles: fold axis, crosses: poles to reverse faults. Schmidt equal area projection.

a north-northwest strike of these faults are presented in Fig. 6b.

At Aasgaardfjellet, north of Reindalen, numerous tight folds and reverse faults are developed west of the Skolten Anticline within the Helvetiafjellet and Carolinefjellet Formations (Fig. 8). Fold-amplitudes are 70–80 m while fold axes cluster around N160°E. Axial surfaces dip steeply to the west. Reverse faults located above the Festningen Sandstone have also been recognized at Drøntoppen and Rudiaksla (Fig. 2). However, folding and thrusting above this stratigraphical level is generally difficult to recognize and trace over long distances.

The geometry of the Skolten Anticline southward across Reindalen is somewhat uncertain. At Lunckefjell, just south of the valley, the fold closure at the Festningen Sandstone level is exposed at an elevation of about 400 m, suggesting a simple continuation of the anticline across Reindalen (Fig. 2). However, as exposure is poor across the valley and because steeply dipping and folded sandstones, probably belonging to the Glitrefjellet Member, have been found at one locality along the trend of the anticline in the valley floor, a more complex relation may exist. At Lunckefjell the Festningen Sandstone occurs as an upright, fairly tight symmetric anticline. Measured bedding readings are too few for foldaxe-construction, but an axial trend close to N160°E is inferred (Fig. 2). In this area the fold geometry in the less competent units of the Helvetiafjellet and Carolinefjellet Formations is more open than in the underlying Festningen Sandstone. The internal deformation at this stratigraphical level, however, appears to be a combination of small scale folding and reverse faulting. A major reverse fault is recognized on the western flank of the anticline, and is interpreted as a continuation of the reverse fault recognized at the same stratigraphical level at Aasgaardfjellet to the north (Fig. 2).

The Skolten Anticline can be traced southward to Kjellströmdalen as an open anticline in the Carolinefjellet Formation and overlying Tertiary formations. The anticline continues southward across Kjellströmdalen with an even smaller amplitude/wavelength ratio than further north. We consider the Skolten Anticline to exist as a subsurface anticlinal structure at the Festningen Sandstone level south of Kjellströmdalen (Fig. 3d).

Tronfjellet Anticline

The Tronfjellet Anticline occurs as an upright, asymmetric, slightly east vergent anticline involving Middle Jurassic and younger rocks. It is located approximately 5 km east of the Skolten Anticline (Figs. 2 and 3c, d), and is particularly

Fig. 7. The eastern reverse-fault at Arctowskifjellet viewed from the south (Adventdalen). The reverse fault ramps up-section through the Helvetiafjellet and Carolinefjellet Formations with a throw of at least 300 m. Shortening of the Festiningen Sandstone is estimated to at least 1 km across this fault.

Fig. 8. Photography showing the structural relationship between the Tertiary and Cretaceous strata along the western limb of the Skolten Anticline at Aasgaardfjellet. A thrust separates weakly deformed Tertiary rocks to the left from isoclinally folded rocks of the Carolinefjellet/Helvetiafjellet Formation. Viewed from the south (Reindalen).

Fig. 9. The Tronfjellet Anticline at Tronfjellet (viewed from the south, Reindalen), showing the Festningen Sandstone as an upright, slightly east vergent anticline. Note thickening of the Janusfjellet Subgroup in the core of the anticline. The anticline recognized in the Cretaceous and younger strata dies out downsection, leaving the Agardhfjellet Formation almost flatlying. Compare with cross-section (Fig. 3C) for structural interpretation.

Fig. 10. The Tronfjellet Anticline at Glitrefjellet (viewed towards the south). The Festningen Sandstone and the uppermost part of the Rurikfjellet Formation are folded into an overturned anticline. Note the continuous but locally inverted position of the Festningen Sandstone.

well-developed north and south of Reindalen. In this area the anticline, as outlined by the Festningen Sandstone, has an amplitude of 400–500 m and a wavelength of about 600 m. To the north erosion has removed most of the structure except its western flank (Fig. 2). The fold-hinge is eroded at Tronfjellet (Fig. 9), whereas south of Reindalen, at Glitrefjellet, the fold-closure is well exposed, illustrating that shortening of the Festningen Sandstone across the anticline at least south of Reindalen is by folding only (Fig. 10). Structural data collected along the Tronfjellet Anticline north and south of Reindalen are presented in Fig. 11a. An axial trend close to N160°E and a plunge of less than 5° toward south-southeast are indicated.

A simple continuation of the Tronfjellet Anticline is indicated across Reindalen (Fig. 2), as no evidence supporting the existence of a structural discontinuity across the valley has been recognized along the Tronfjellet Anticline. However, along strike variations in fold geometry do occur as exemplified at Glitrefjellet, where the single anticline defined by the Festningen Sandstone on the northern slopes of the mountain, southward, changes into two minor anticlines separated by a syncline (Figs. 2 and 12).

Janusfjellet Subgroup shales 'fill' the core of the Tronfjellet Anticline through the development of numerous minor folds and reverse faults thickening the subgroup. Most folds are overturned towards the east. Most reverse faults dip west, but locally east-dipping reverse faults are also found. Folds and thrusts become less frequent downsection and deformation apparently dies out towards the underlying Kapp Toscana Group, indicating the existence of a structural discontinuity in the lower part of the Janusfjellet Subgroup.

The Tronfjellet Anticline is traced southward to Kjellströmdalen (Fig. 2). Here the tight anticline recognized to the north has changed to an open, slightly asymmetric anticline with an amplitude of less than 200 m and a wavelength of about 1 km, involving Cretaceous and Tertiary strata. Bedding readings along the anticlinal flanks north and south of Kjellströmdalen are presented in Fig. 11b.

The anticlinal asymmetry is related to a west-dipping reverse fault rooted in the Janusfjellet Subgroup, ramping upsection through the Festningen Sandstone (Figs. 2 and 3d). North of the valley, at Langstakken, the Festningen

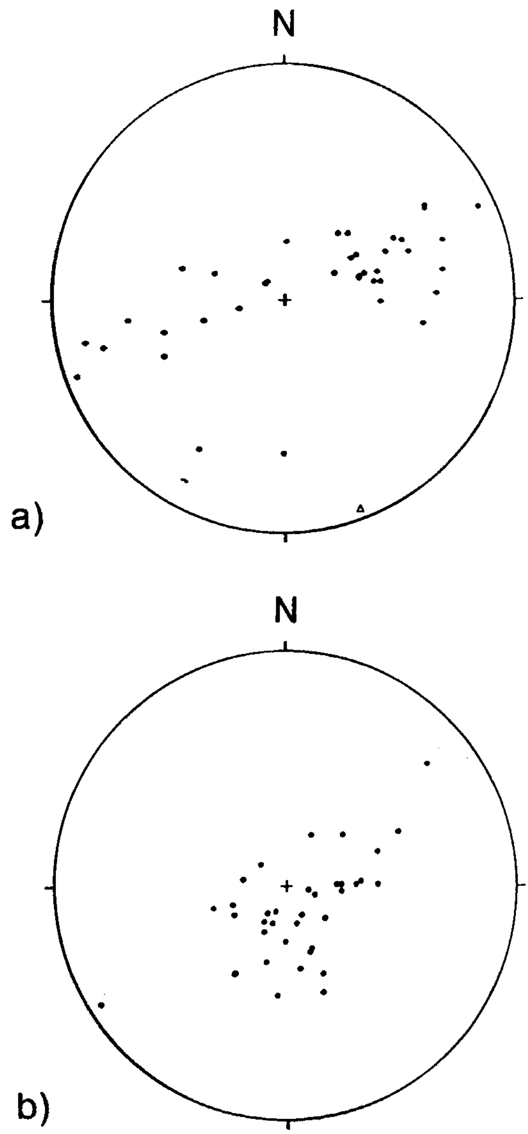


Fig. 11. Structural data collected along the Tronfjellet Anticline. a) Reindalen, b) Kjellströmdalen. Circles: poles to bedding, triangles: fold axis. Schmidt equal area projection.

Sandstone is displaced 400–500 m eastward along this fault (Fig. 13). Above the Festningen Sandstone the fault flattens and displacement gradually 'dies out' through small scale folding and imbrication in the core of the anticline.

A similar reverse fault is recognized in the core of the Tronfjellet Anticline at Langhummelen, south of Kjellströmdalen. However, displace-

Fig. 12. The Tronfjellet Anticline at the southern slopes of Glitrefjellet (viewed towards the north). The Festningen Sandstone occurs as an upright M-folded structure.

ment along the fault is minor and a distinctive, highly asymmetric kink-like structure is superimposed on the regional fold (Fig. 14). The length of the short and steeply dipping eastern limb diminishes upsection, suggesting that displacement 'dies out' upsection.

Due to the erosional level and because the valley floor is covered by glacial till, exposures of Janusfjellet Subgroup rocks are poor across Kjellströmdalen. However, local occurrence of steeply dipping Janusfjellet Subgroup shales in the middle of the valley suggests that the tight, often disharmonic folding observed in the Janusfjellet Subgroup to the north also continues southward.

Another reverse fault, occurring above the Festningen Sandstone and west of the Tronfjellet Anticline, has been observed both north and south of Kjellströmdalen (Figs. 2 and 3d). Displacement on this fault is uncertain due to poor exposure. We believe that this fault represents a continuation of the fault recognized at this stratigraphical and structural level west of the Skolten Anticline to the north.

Drønbreen Syncline

The Drønbreen Syncline is a broad, almost symmetric syncline bordering the Skolten- and Tronfjellet Anticlines (Figs. 2 and 3). The syncline has an amplitude of about 500 m and a wavelength of about 5 km. Its geometry is clearly controlled by the development of the two adjoining anticlines.

Juvdalskampen Monocline

A west-facing strongly asymmetric monocline (Juvdalskampen Monocline) located slightly east of the Tronfjellet Anticline is another major structural element recognized along the Billefjorden Fault Zone in the Adventdalen-Reindalen area (Fig. 2). Along the short western limb bedding dips as much as 45° to the west (Fig. 15). Beds on the eastern sub-horizontal limb are elevated about 400 m relative to the beds west of the monocline. Bedding readings collected along the flanks of the monocline are presented in Fig. 15.

The Juvdalskampen Monocline is genetically unrelated to the Skolten and Tronfjellet Anti-

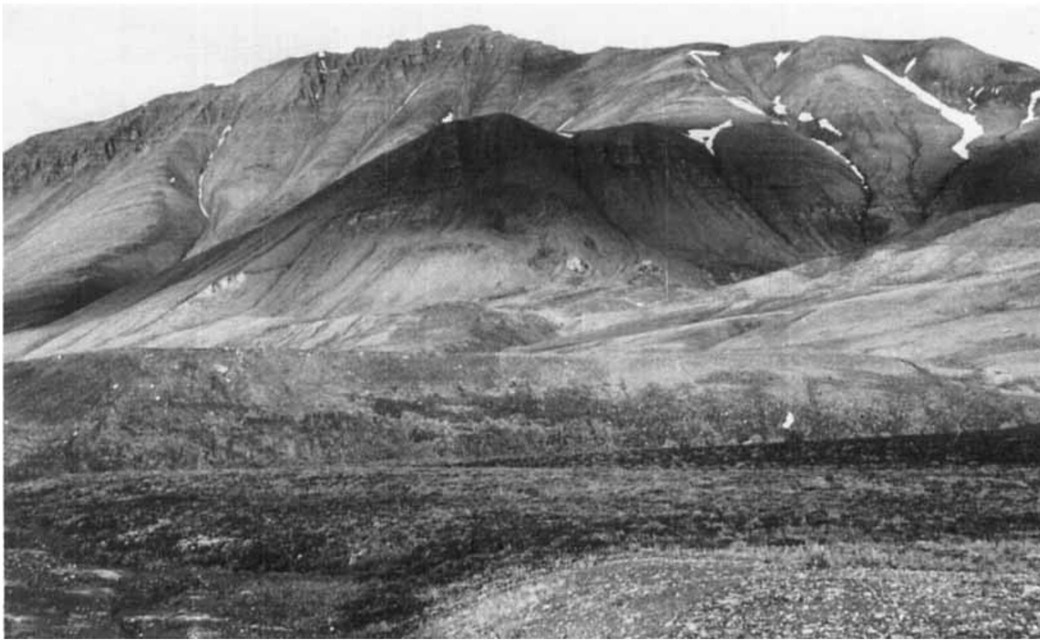


Fig. 13. Langstakken viewed from the south (Kjellströmdalen). The Tronfjellet Anticline is seen as an open anticline developed in the Helvetiafjellet and younger formations. In the central lower part of the picture a reverse fault, ramping through the Festiningen Sandstone, is developed.

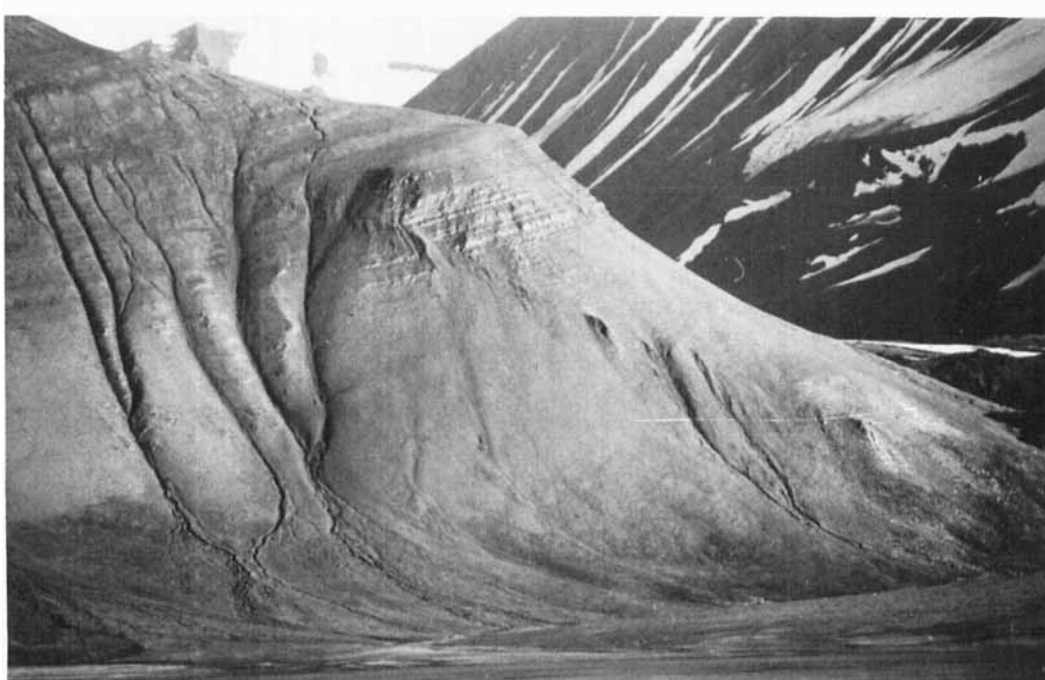


Fig. 14. A reverse fault on the eastern limb of the Tronfjellet Anticline at Langhummelen south of Kjellströmdalen. Displacement across the fault appears to die out upsection and the fault is developed as a kink-fold above the Festningen Sandstone level.

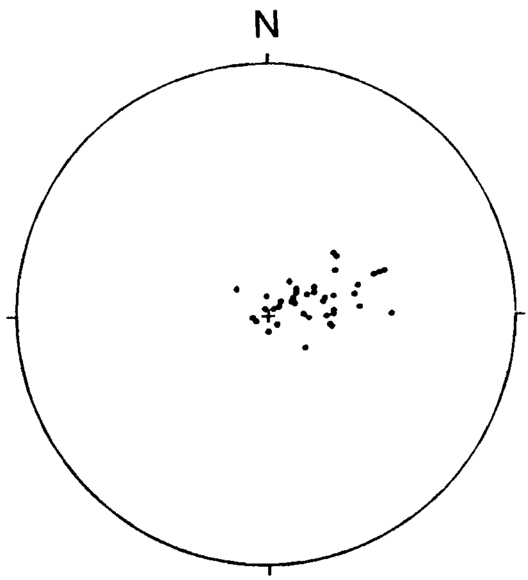


Fig. 15. Bedding readings (poles to bedding) collected along the Juvdalskampen Monocline in the Adventdalen-Reindalen area. Schmidt equal area projection.

clines and it involves the entire Mesozoic sequence. At Juvdalskampen (Figs. 2 and 3a) the Juvdalskampen Monocline is associated with a thinned Janusfjellet Subgroup, emphasizing that the monocline played an important role during the tectonic evolution along the Billefjorden Fault Zone.

The Juvdalskampen Monocline has a north-northwest trend, and intersects the Tronfjellet Anticline at an acute angle (Fig. 2). The monocline is less developed southward, but a link to a similar, but east–west trending monocline in the inner Kjellströmdalen area may exist (Fig. 2).

Upper Décollement Zone

The Janusfjellet Subgroup, consisting mainly of soft, black shales, is characterized by great thickness variations in the study area. Tectonic thickening of the subgroup is especially well documented through folding and reverse faulting in the core of the Skolten and Tronfjellet Anticlines (Figs. 16 and 17). Outside the anticlines and stratigraphically downsection toward the underlying Kapp Toscana Group deformation is less pronounced as bedding is generally rather flat. However, careful investigations have documented that the lower part of the Janusfjellet Subgroup (Agardhfjellet Formation) is locally intensely deformed with numerous minor low-angle and bedding parallel reverse faults (Fig. 18). Locally well developed imbricate fans and duplex structures are developed (Fig. 19). Bedding readings, fold axis and reverse faults associated with the Janusfjellet Subgroup are presented in Fig. 20.

Tectonic transport of the hanging-wall is generally towards the east, although back-thrusts, with tectonic transport towards the west, are observed locally (Fig. 19). Small scale out-of-sequence thrusts, locally cutting downsection in transport direction, are also a characteristic feature throughout the Agardhfjellet Formation.

A marked tectonic thinning of the Janusfjellet Subgroup has been recognized at Juvdalskampen north of Adventdalen (Figs. 2 and 3a; Parker 1966; Major & Nagy 1972). Structural observations demonstrate that this thinning is the result of a regional out-of-sequence thrust cutting downsection as approaching the Juvdalskampen Monocline (Figs. 3 and 21, see discussion).

The intense deformation, although localized to certain narrow zones in some areas, suggests that the Janusfjellet Subgroup acted as a major décollement zone during deformation (Upper Décollement Zone).

Adventdalen Duplex

Relatively little is known about the structural style and deformation below the Upper Décollement Zone due to the erosional level in the study area. However, in a river section along Adventelva, in the inner part of Adventdalen, deformed and tilted Kapp Toscana Group rocks crop out (Fig. 2). We suggest these imbricated Kapp Toscana rocks represent the upper part of a duplex structure, the Adventelva Duplex (Figs. 1 and 3). Structural observations associated with the Adventelva Duplex are presented in Fig. 22. The roof thrust associated with this duplex is located in the lower part of the Agardhfjellet Formation. As the floor thrust is not exposed, it has not been possible to determine with certainty at which stratigraphical level the Adventelva Duplex is rooted. However, based on structural observations in Eskerdalen and in Sassendalen (Fig. 2) we believe that the floor thrust is located within the upper shaly part of the Barentsøya Formation.

Fig. 16. Imbricate reverse faults developed in the core of the Tronfjellet Anticline at Tronfjellet (upper part of Agardhfjellet Formation).

Fig. 17. Photography showing an overturned tight fold along the eastern limb in the core of the Tronfjellet Anticline at Glitrefjellet. The folded strata belong to the upper part of the Agardhfjellet Formation.

Fig. 18. Nearly bedding parallel fault (flat) in the Agardhfjellet Formation NE of Drønbreen. A reverse fault which ramps up-section is seen on the left.

Fig. 19. Example of leading edge imbricate faults associated with a back-thrust (tectonic transport towards the west) developed in the Agardhfjellet Formation in the Advent River NW of Janssonhaugen.

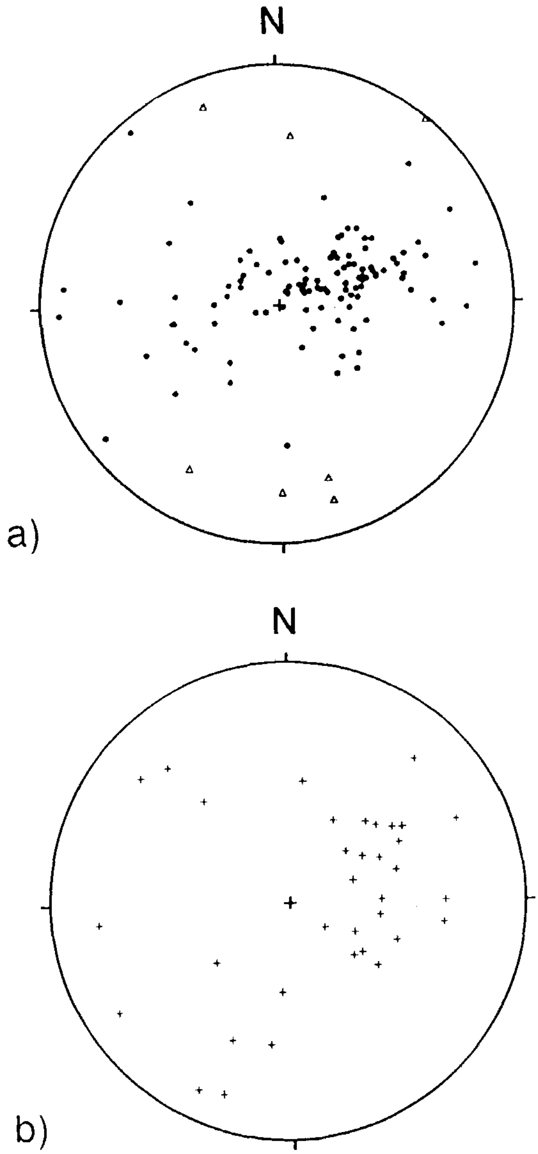


Fig. 20. Structural data collected from the Upper Décollement Zone. a) Bedding readings (poles to bedding, circles) and fold axis (triangles), b) reverse faults (crosses). Schmidt equal area projection.

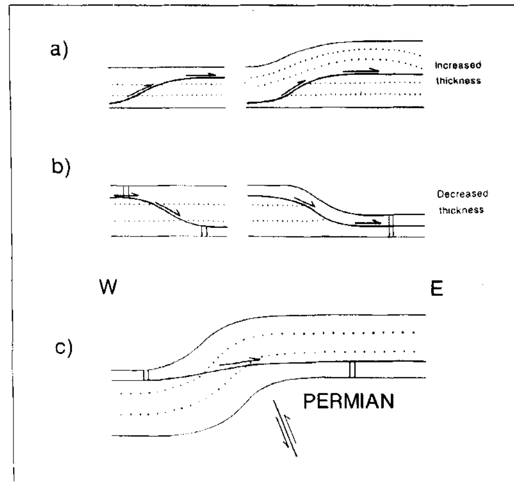


Fig. 21. a) Reverse fault cutting upsection. b) Reverse fault cutting downsection (out-of-sequence thrust). c) Model which explains the reduced thickness of the Janusfjellet Subgroup in the Juvdalskampen area by an out-of-sequence thrust.

Lower Décollement Zone

The floor thrust of the Adventelva Duplex is interpreted to be linked with a regional décollement zone located in the upper part of the Barentsøya Formation. Evidences for such a Lower Décollement Zone are well documented through the development of numerous small-scale, west-dipping reverse faults, generating imbricate fans and duplexes within the Botneheia Member in Sassendalen (Fig. 2). In an approximately 3 km long east–west traverse in Vend-

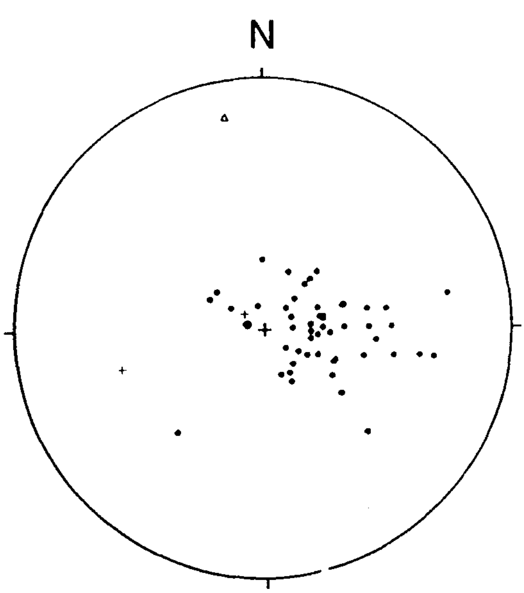


Fig. 22. Structural data collected along the Adventelva Duplex. Circles: poles to bedding, triangles: fold axis, crosses: poles to reverse faults. Schmidt equal area projection.

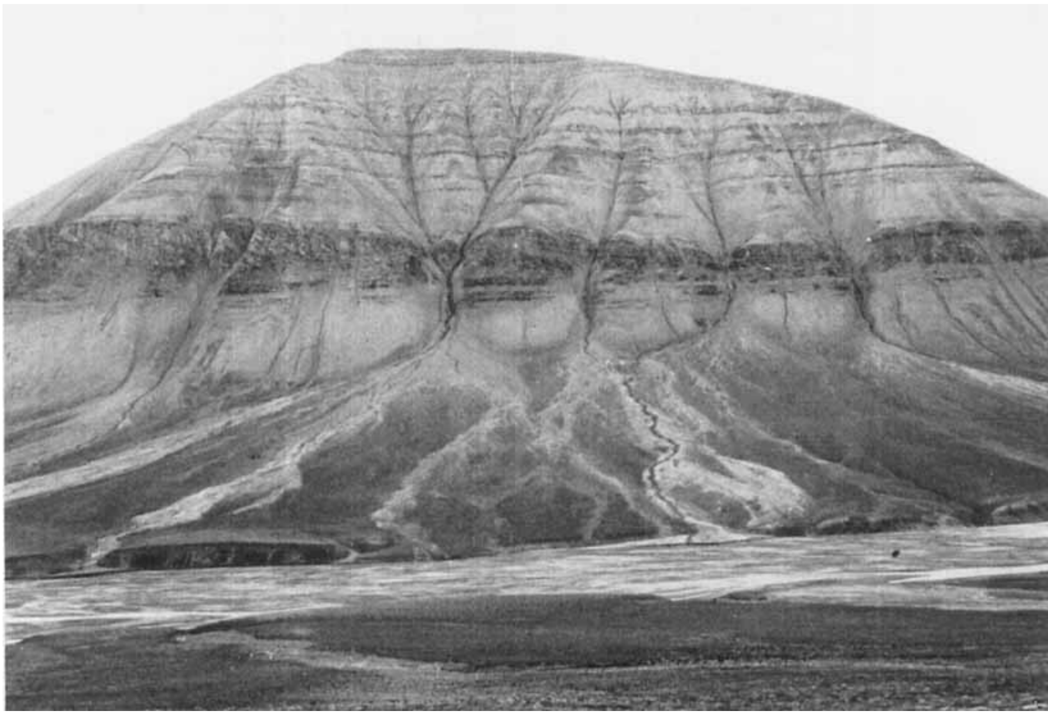


Fig. 23. Local thickening of the Barentsøya Formation by intense reverse faulting and imbrication in the upper, dark, cliff-forming part of the Botneheia Member in Vendomdalen. Viewed from the north.

omdalen (Fig. 2) the dark, cliff-forming, easily recognizable Botneheia Member is doubled, locally tripled in thickness, due to intense internal imbrication (Fig. 23). Structural data from Vendomdalen are presented in Fig. 24. More recent observations both east and west of the present study area support the idea that a regional décollement zone exists at this stratigraphical level and that it extends all the way to the east coast of Spitsbergen (Haremo & Andresen 1988, in press); Andresen, Haremo, Swensson & Berg unpublished).

Other evidences of deformation recognized along the Billefjorden Fault Zone

Normal faults

Normal faults with minor (0.1–1 m) offset have been recognized locally in the Janusfjellet Subgroup. The majority of these faults are planar, steep to vertical, and strike slightly west of north. Where relative age-relationship can be stated, these normal-faults post-date the reverse faults described above.

Small-scale, listric post-sedimentary normal faults (0.2–1 m offset) have also been observed in the Janusfjellet Subgroup. Locally, offset with as much as 20 m has been reported on these faults (Gabrielsen pers. comm. 1986). It is uncertain if these faults post- or pre-date the earlier described reverse faults. They may represent young faults associated with landslides. Gabrielsen (pers. comm. 1986) also reported local development of listric syn-sedimentary normal faults with offsets in the order of 1–5 m within the Janusfjellet Subgroup along the Billefjorden Fault Zone in the Adventdalen area.

Cleavage and joints

Cleavage is generally not a characteristic deformation structure in the mapped area. In competent sandstones cleavage has not been recognized at

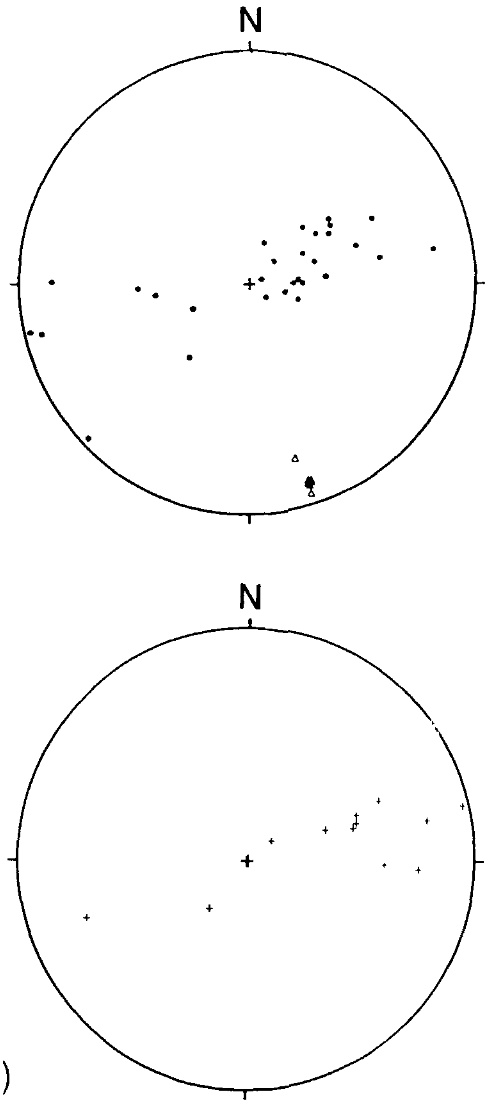


Fig. 24. Structural data from the Lower Décollement Zone in Vendomdalen. a) Bedding readings (poles to bedding, circles) and fold axis (triangles). b) poles to reverse faults. Schmidt equal area projection.

all. However, in the less competent shales of the Janusfjellet Subgroup a spaced cleavage (1–10 cm) is developed locally, often related to zones of localized deformation. Cleavage orientation is generally steep (>45°), dipping both southwest and northeast.

Joints with spacing in the range of 15 cm to 1 m are also recognized. These are not restricted to shaly units, but have also been recognized in the more competent sandstones. Joint orientation is highly variable, but generally steep (>60°). Calcite-filled extensional joints (0.1–1 mm) are found locally throughout the Janusfjellet Subgroup.

Slickensides/stylolites

Although neither slickensides nor stylolites are characteristic deformation structures in the study area, both structures have been observed, in particular associated with the Skolten and Tronfjellet Anticlines. Slickensides with a well developed quartz-fibre mineral lineation are frequently observed in the Festningen Sandstone, close to the hinge-line of the anticlines (Fig. 25). Plunges are highly variable whereas east–northeast/west–southwest trends dominate. Movement direction is often difficult to decide, but indications of both reverse and normal movement have been observed. Slickensides have also been observed on reverse faults in the Janusfjellet Subgroup, where they cut through silty and sandy units.

Bedding-parallel stylolites have only been observed on bedding planes in the Festningen Sandstone and in the upper sandy part of the Wilhelmøya Formation (Fig. 26). As in the case of the slickensides, their appearance is closely related to the Skolten and Tronfjellet Anticlines.

Dolerite dykes

A dolerite dyke swarm cutting rocks belonging to the upper part of the Agardhfjellet Formation has been recognized at one locality in the Advent River about 2 km west of the Adventelva Duplex. The dykes weather light brown and contain rectangular plagioclase phenocrysts (0.2–10 mm), giving the rock a distinct glomero-porphyric texture. The matrix consists mainly of plagioclase, quartz and iron rich calcite, the latter probably a replacement-product of iron rich amphiboles.

Discussion

The structural development recognized in the Mesozoic sequence along the Billefjorden Fault Zone shows that deformation is dominated by compressional tectonics involving the entire Mesozoic sequence. As is seen on the map and cross-sections (Figs. 2 and 3), Tertiary rocks are

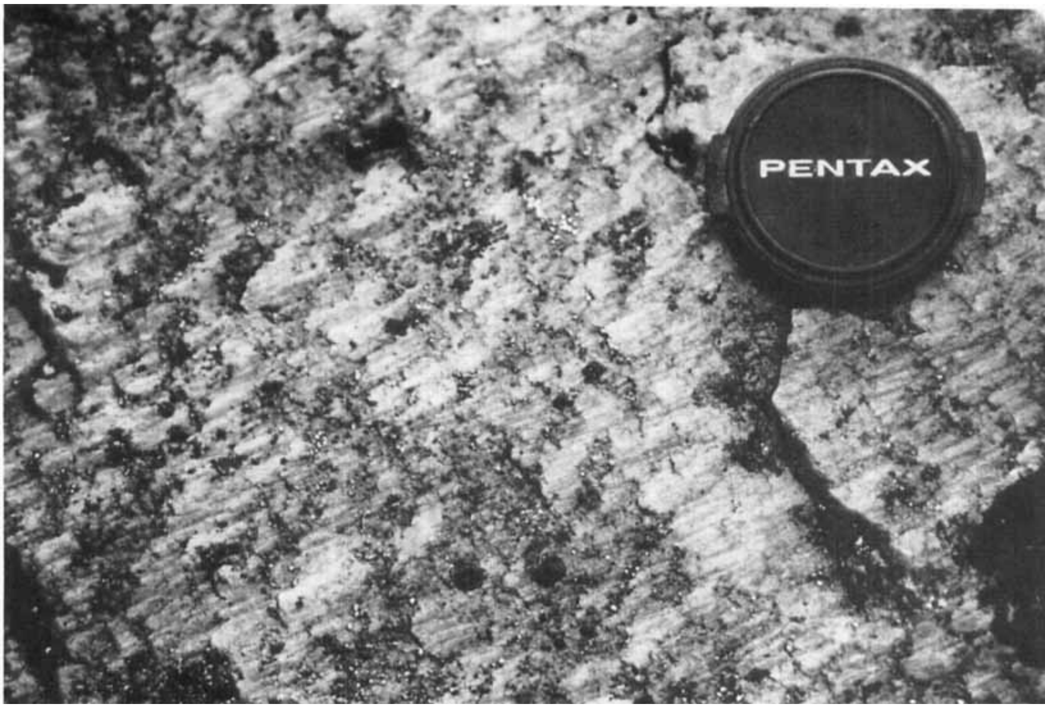


Fig. 25. Quartz fibre lineation developed in the Festningen Sandstone on the western flank of the Skolten Anticline at Aasgaardfjellet.

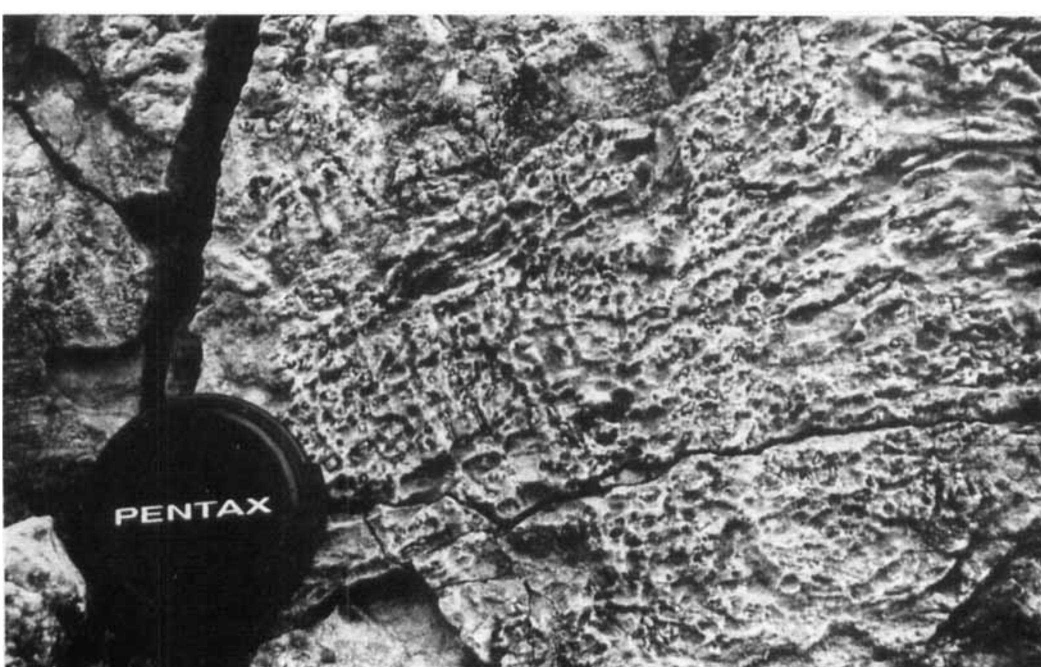


Fig. 26. Festningen Sandstone with bedding parallel stylolites on the western flank of the Skolten Anticline at Aasgaardfjellet.

also involved, suggesting that the deformation is related to the West-Spitsbergen Orogeny of supposed Eocene age (e.g. Harland & Horsfield 1974; Eldholm et al. 1987).

The structural evidences presented in the present paper and in Haremo & Andresen (1988, unpublished) imply a shortening of the Festningen Sandstone of minimum 3 km and possibly as much as 10 km across the Billefjorden Fault Zone. Data from outside the study area suggest that Tertiary shortening associated with décollement thrusting extends eastward all the way to the east coast of Spitsbergen (Andresen et al. 1988, unpublished; Dypvik et al. 1990; Haremo & Andresen 1988, in press). A linkage toward similar décollement zones developed in the West-Spitsbergen Foldbelt to the west is also suggested (Bergh & Andresen 1990; Bergh et al. 1988; Dallmann & Maher 1989; Faleide et al. 1988; Haremo & Andresen 1988, in press; Nøttvedt et al. 1988). These data emphasize the compressional nature and the significant lateral extent of Tertiary deformation on Spitsbergen, contradicting the traditional view that the West-Spitsbergen Orogeny represents a transpressive orogen with wrench tectonic structures (Lowell 1972; Kellogg 1975; Harland & Horsfield 1974; Steel et al. 1985).

The Tertiary thin-skinned structures associated with the Billefjorden Fault Zone are interpreted to be controlled by one or more high angle reverse faults located in the pre-Mesozoic basement (Fig. 3). The west-facing Juvdalskampen Monocline, involving the entire Mesozoic sequence, is the only indication of this thick-skinned deformation in the study area. However, east dipping steep reverse faults observed in the Permian strata at Gipshuken (Ringset 1988; Ringset & Andresen 1988) support this interpretation.

The position and geometry of these steeply dipping reverse faults are somewhat uncertain. Based on data from Billefjorden, we consider the steep reverse faults to be reactivated normal-faults associated with the western margin of the Carboniferous Billefjorden Trough (Haremo & Andresen unpublished).

This combined thin-skinned/thick-skinned structural style developed along the Billefjorden Fault Zone is most likely responsible for the thinning of the Janusfjellet Subgroup observed on Juvdalskampen (Figs. 2 and 3). In Fig. 21 a model in which thick-skinned reverse faulting pre-dates thin-skinned décollement thrusting explains the observed thinning of the Janusfjellet Subgroup by a regional out-of-sequence thrust, cutting down-section in the transport direction.

It is suggested that the Juvdalskampen Monocline, involving the entire Mesozoic sequence, is unrelated to the Tronfjellet Anticline. Such an interpretation is supported by the non-parallelity between the structures as seen from the map (Fig. 2). This may suggest the existence of two sets of lineaments in the pre-Mesozoic basement, cutting each other with an acute angle. The change in fold geometry recognized across Adventdalen (Figs. 2 and 3) may be related to a northward interference of these lineaments.

The along strike variation in structural development and geometry recognized along the Billefjorden Fault Zone suggests the existence of transfer zones or oblique/lateral ramps climbing Reindalen(?) (Haremo & Andersen unpublished). The southward appearance of thrust faults above the Festningen Sandstone level may indicate that movement on the Upper Décollement Zone across Adventdalen partly is transferred to a higher stratigraphical level. The possible complex geometry which has been observed across Reindalen along the Skolten Anticline may suggest the existence of a similar transfer zone in this area. It is furthermore speculated that such a southward climb of thrusts is related to the pre-deformational Late Cretaceous/Early Tertiary uplift and southward tilting of Spitsbergen (Haremo & Andresen unpublished).

Evidence of Mesozoic extensional movements along the Billefjorden Fault Zone is minor only. The dramatic thinning of the Janusfjellet Subgroup across Juvdalskampen which previously has been interpreted as the result of syn-depositional Late Jurassic extensional tectonics (Parker 1966, 1967; Harland et al. 1974) is rejected. These authors claimed that during Late Jurassic sedimentation, an elevated area existed across the Billefjorden Fault Zone in the Juvdalskampen area (Harland et al. 1974, their Figs. 3 and 4). We have found no evidence, structural or sedimentologic, supporting the existence of such an elevated area. Paleontological data from Juvdalskampen show that the thinned Janusfjellet Subgroup is made up of the lower part of the Agardhfjellet Formation and the upper part of the Rurikfjellet Formation. This is in agreement with our model, explaining the thinned sequence by an out-of-sequence thrust, cutting downsection in the transport direction (Fig. 21). It is inconsistent with the model of Parker (1967) and Harland et al. (1974) which is mainly based on the assumption that the thinned Janusfjellet Subgroup on Juvdalskampen is due to the lack of the Agardhfjellet Formation in this area.

North of the study area, in Flowerdalen, the Janusfjellet Subgroup rests directly upon the Triassic Sassendalen Group (Major & Nagy 1972). Harland et al. (1974) explained this thinning by repeated extensional movements along the Billefjorden Fault Zone throughout the Mesozoic. We argue, based on this and other studies (Dypvik et al. 1990; Haremo & Andresen in press; Haremo et al. unpublished), that neither structures documenting extensional movements nor facies changes (to be expected in the Harland et al. (1974) model) are recognized across the Billefjorden Fault Zone in Flowerdalen. We suggest instead that the thinned Mesozoic sequence east of the Billefjorden Fault Zone in Flowerdalen is due to Tertiary compressional tectonics representing the northward continuation of the out-of-sequence thrust recognized at Juvdalskampen.

Conclusions

The structural evolution along the Billefjorden Fault Zone in the area between Kjellströmdalen and Sassendalen/Adventdalen is mainly the result of Tertiary compressional tectonics. The structural development is characterized by a combined thin-skinned/thick-skinned structural style, suggesting a Tertiary shortening of at least 3 km, possibly as much as 10 km across the Billefjorden Fault Zone. Décollement zones are developed in the Janusfjellet Subgroup (Late Jurassic/Lower Cretaceous) and in the Sassendalen Group (Lower/Middle Triassic). Reverse faulting involving the pre-Mesozoic basement and probably being related to reactivation of pre-Mesozoic, steep, normal faults has played an important role in the structural development of the study area.

Evidence of Mesozoic extensional tectonics is minor only. The observed thickness variations in the Mesozoic sequence are explained by Tertiary compressional tectonics, and are not a result of syn-sedimentary tectonic activity.

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