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The Billefjorden Fault Zone, Spitsbergen

the long history of a major tectonic lineament

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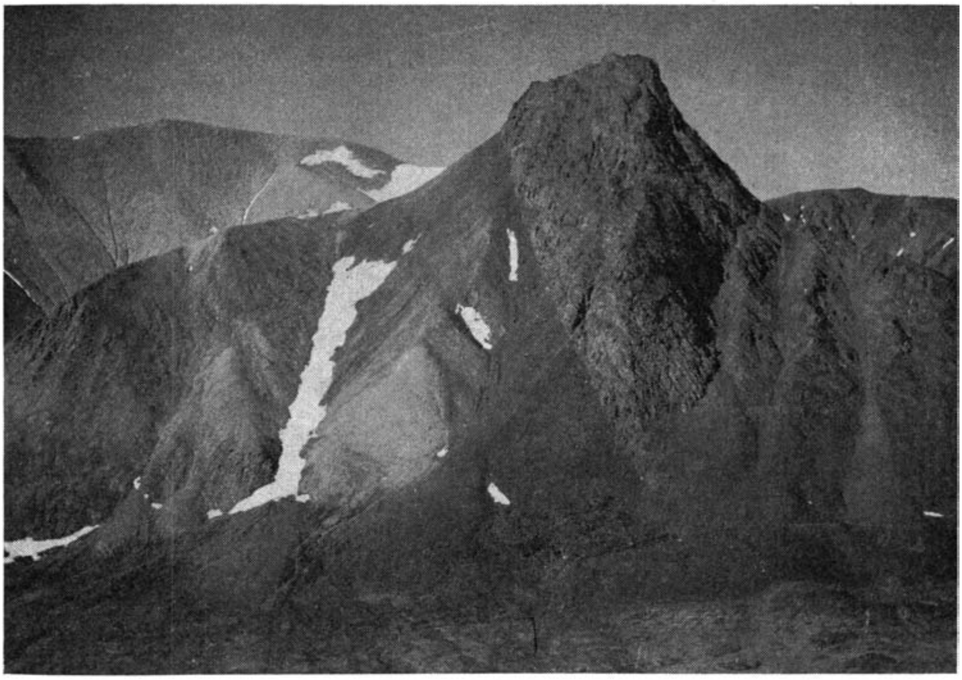
Aerial view south along the fault line to the south-west shore of Austfjorden. Devonian rocks occupy the right foreground and Hecla Hoek the left foreground and the islands. The exhumed basal Carboniferous peneplain is seen in the higher mountain tops to the left and in the distance the overlying Carboniferous and Permian strata extend over the width of the Fault Zone.
Oblique air photograph by B. Luncke *1936, reproduced by kind permission of Norsk Polarinstitutt.*



Balliolbreen Fault seen from south with Hecla Hoek metamorphic rocks to east (right) and Old Red Sandstone to west. The photograph (H 207.12 by W. H. Harland *1953) was taken from the foot of Sentinelfjellet, across Alandelva to Odellfjellet.*

Contents

Abstract

The Billefjorden Fault Zone, already well known as a key element in the structure of Spitsbergen, is described throughout its inferred length of at least 250 km. Stratigraphic, structural and geophysical studies during the last few years make it possible to document a long history of activity, and for some periods with stratigraphic precision. There are indications of a substantial Precambrian structural feature which later controlled the main Caledonian (Ny Friesland) orogeny. Associated with, and possibly causing the late Caledonian (Svalbardian) folding and thrusting, a sinistral strikeslip movement along the fault is inferred for which a displacement of at least 200 km and possibly up to 1000 km or more is suggested. No substantial strikeslip thereafter is demonstrable. Movements with gravity faulting controlled sedimentation through Carboniferous time. Permian to Lower Tertiary sedimentation reflects relatively little diastrophism. This includes slight new warping in mid-Triassic time and more noticeable faulting with dolerite intrusions about the Jurassic/Cretaceous boundary. Intense Cenozoic orogenic activity 50 km to the west was accompanied by renewed fault movements along the zone, these being mainly gravity faulting which produced minor graben. Some superficial thrusting in the south of this zone (and possibly traces of dextral strike slip in the north) accompanied the major compression, graben formation and dextral strike slip in the west (West Spitsbergen Orogeny). The sequence of movements provides a key to crustal movements generally in the Arctic-North Atlantic region and demonstrates some evidence by which this ancient plate margin may be recognized.

I. Introduction

Spitsbergen is traversed by several major fault systems trending approximately N—S. These are shown in relation to the general geology in Fig. 1. The Billefjorden Fault Zone or Lineament was early recognised as an important structural feature and this has never been seriously doubted. It traverses the length of Wijdefjorden, is exposed between Austfjorden and Billefjorden, cuts southwest Bünsow Land and thence across Sassenfjorden along Flowerdalen and south in the direction of the coast at Kvalvågen — a total distance of more than 250 km.

1. History of investigations

The earliest geological maps of north Spitsbergen (e.g. Nordenskiöld 1863) show Wijde Bay (Wijdefjorden) separating the crystalline rocks to the east from the red sandstone to the west along its total length of 110 km.

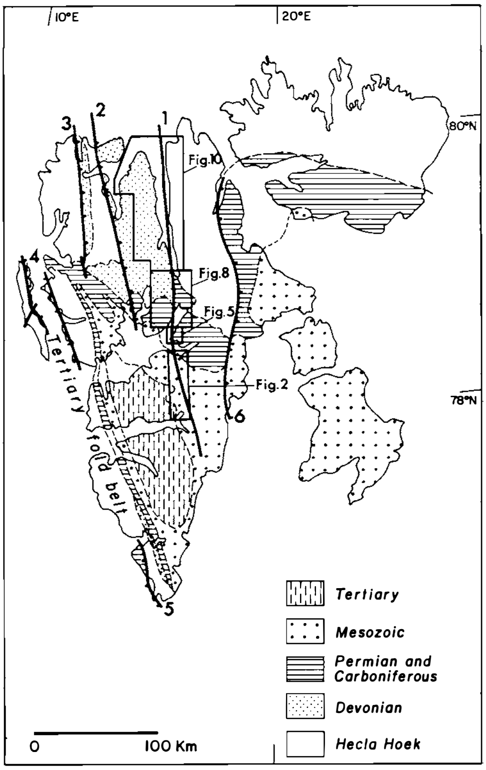


Fig. 1. *Outline geological map of Spitsbergen showing major fault zones: 1 Billefjorden, 2 Bockfjorden, 3 Raudfjorden, 4 Forlandsundet Graben, 5 Hornsund, 6 Lomfjorden, and locations of figures 2, 5, 8 and 10.*

Nathorst, in the first comprehensive account of the geology of Spitsbergen (1910) showed a fault defining the east coast of Wijdefjorden and extending 10 km SSE towards Bünsow Land separating "Urgebirge" from "Karbon und Perm", and another fault just south of Wijdefjorden separating "Devon" and "Karbon und Perm". This was shown to cross Billefjorden nearly to the shore of Bünsow Land. This map shows the most obvious course of the fault zone we are considering as the eastern fault of the Devonian graben.

De Geer (1909), who considered especially the relationship between structure and geomorphology, also showed these two faults on his map. However he continued the first fault line (without giving evidence) through to southeast Bünsow Land; and the other fault, west of Billefjorden, he extended to the north of Dickson Land, so defining a minor Carboniferous graben. He also mapped several Tertiary faults bounding the fjord coastlines; one of these runs inland SSE along Flowerdalen.

Hagerman (1925) first described the fault zone structures in Kjellströmdalen in the southern area.

A small geological and survey party comprising J. Brough, R. H. S. Robertson and E. E. Mann of the Oxford Expedition in 1933 reconnoitred Dickson Land (Glen 1934). The results were not published but they made observations on the main thrust fault (Balliolbreen Fault) subsequently privately communicated to us.

Little new was published about this Fault Zone between the wars. Orvin (1940) reviewed available knowledge in his outline of the geology of Spitsbergen with a 1:1,000,000 map which has since become the authority. He used the name "Inland Fault" and showed an anastamosing fault system extending the full length of Wijdefjorden, through Billefjorden, and south to a point east of Van Mijenfjorden, a total distance of 260 km.

Official publications of the Norsk Polarinstitutt added little to Orvin's work in this connexion until 1972 when two important maps with their accompanying memoirs appeared. The 1:100,000 geological sheet Adventdalen, although printed in 1964 for internal use, was not in fact published and made generally available until the accompanying memoir (Major and Nagy 1972) was published. This sheet and its memoir include the critical area where the Lineament continues south through Mesozoic rocks. Similarly the 1:500,000 map (Flood, Nagy and Winsnes 1971) depicts the course of the Fault Zone south of latitude 78° 35′ as far as Kjellströmdalen (77° 55′).

The above official maps and descriptions clearly owe much to the unpublished contribution of the Amoseas exploration of this area in 1960, 1961 and 1962 when, for the first time, helicopters were used for systematic geological exploration.

A. EARLIER CAMBRIDGE STRUCTURAL STUDIES

During the 1949 expedition from Cambridge to Billefjorden McWhae focussed his study of the Carboniferous and Permian rocks on this part of the zone (McWhae 1953).

McWhae (1953) distinguished a complex of several faults and flexures

(A to I) of late Devonian and post Permian age and discussed their history. Of these the Balliolbreen reverse fault (as noted in 1933 by R. H. S. Robertson and privately communicated in 1950/51) is the most dramatic. For a short stretch it shows crystalline Hecla Hoek rocks on the hanging wall and folded Old Red Sandstone on the footwall, and nowhere is Hecla Hoek found immediately to the west of this line nor Old Red Sandstone to the east. Harland (1959) further discussed the mechanics of this late Caledonian boundary fault. Although the possibility for strike slip along such a long straight fault zone was considered neither author could at that time find supporting evidence for it, whereas dip-slip slickensiding and compressive structures abounded.

B. CAMBRIDGE STRATIGRAPHIC STUDIES

Stratigraphic studies by most of us were made on rocks of different ages on a regional basis and many of these incidentally overlapped the Fault Zone. Other studies tended to be bounded by the Fault Zone which forms a natural demarkation line between different areas. Therefore for a period, with the exception of geophysical work, interest in the Fault Zone was incidental to broader stratigraphic investigations.

C. CAMBRIDGE GEOPHYSICAL STUDIES

Total magnetic field data were collected from around the Fault Zone (in surveys undertaken in 1962, 1964 and 1965). In general the only conspicuously magnetic rocks appear to be certain basic igneous rocks and amphibolites within the Hecla Hoek. These do not throw much light on the nature of the fault itself because the obvious magnetic rocks do not help to delineate the major structure but largely reflect the pattern of basic igneous activity.

However, gravity surveys were made around Wijdefjorden, Austfjorden and along the shores of Isfjorden and the adjoining fjords. These surveys were somewhat restricted in extent partly because of difficulty of terrain and access, and mainly because of limitation of resources. This restriction must limit any interpretation. The gravity data were obtained on two independent surveys: that around Isfjorden in 1962, and the other, in north Spitsbergen in 1967. All the standard reductions have been applied including terrain corrections, to yield Bouguer anomalies. Because of an initial lack of gravity ties the two surveys have not been reduced to a common datum: the Isfjorden survey is related to a zero value at Longyearbyen war memorial, while the north Spitsbergen survey is related to a zero at Biskayerhuken. It was subsequently established that Biskayerhuken has a Bouguer anomaly of $+63.8 \pm 2.4$ milligals with respect to the Norges Geografiske Oppmåling station in the basement of the official residence of Sysselmannen, and gravity at the Longyearbyen war memorial is 1.16 milligal greater than at the N.G.O. station. The north Spitsbergen survey comprised a coastal network of stations some 5 to 10 km apart into some of which were tied detailed traverses. The Isfjorden survey comprised 13 separate traverses, all of which were tied to Longyearbyen by means of helicopter transits. In both surveys stations were about 500 metres apart along traverses.

Table 1

Field Work related to Billefjorden Fault Zone by Cambridge Spitsbergen Expeditions

Year of Expedition	
1938	W. B. Harland reconnoitred Carboniferous and Permian stratigraphy east of fault belt and mapped Lemströmfjellet Fault to the north. (McCabe 1939; Harland 1941).
1949	J. R. H. McWhae investigated fault belt in south Dickson Land and Bünsow Land and W. B. Harland worked on Hecla Hoek stratigraphy to east of Billefjorden Fault Zone. (Harland 1950).
1951	W. B. Harland (leader) visited structures in O.R.S. around Vatnedalen and Andredalen on the west coast of Wijdefjorden, and in Hecla Hoek along the east coast of Wijdefjorden between Stubendorffbreen and Tryggvebreen. (Harland & Hollin 1953).
1953	W. B. Harland (leader) visited Balliolbreen Fault and reconnoitred its continuation north to Austfjorden, with O.R.S. to the west and Hecla Hoek to the east. (Harland 1954).
1954	M. B. Bayly (leader) surveyed Reinbokkdalen and other areas along the east coast of Wijdefjorden. (Bayly 1955).
1955	P. F. Friend and B. Moore continued investigations of the 1953 expedition, especially O.R.S., in the area west of N.Mittag-Lefflerbreen and S. Austfjorden. M. B. Bayly, C. B. Wilson and D. Masson-Smith extended the geological and topographical survey of Ny Friesland. (Edwards 1956; Harland 1957).
1957	P. F. Friend (leader) continued investigations of 1953 and 1954, around west of Austfjorden and in Vestfjorden, in connection with studies of O.R.S. (Friend 1958).
1958	Party A, leader P. F. Friend, extended work in Devonian areas of N. Dickson Land and in Andrée Land, based on camps along west coast of Wijdefjorden. (Friend 1959). Party B, leader D. J. Gobbett, worked at the head of Billefjorden for a little time.
1959	W. B. Harland, P. F. Friend, and others carried out structural and stratigraphical studies in N. Andrée Land, in Mimerdalen and west Billefjorden, in continuation of work of 1958 expedition. (Harland 1960b). D. J. Gobbett worked on Carboniferous and Permian stratigraphy in the central Fault Zone area.
1961	Party C, leader R. A. Gayer, surveyed and mapped the stratigraphy and structures of the Hecla Hoek of N. Ny Friesland, working from camps along the east coast of Wijdefjorden. Party E, leader J. L. Cutbill, investigated Mesozoic sections near Deltaneset, and post-Devonian rocks around Dicksonfjorden and Billefjorden. Party B, leader P. F. Friend, worked in Andrée Land and Dickson Land. (Friend 1962; Harland 1962).
1962	Party A, leader R. A. Gayer, continued Gayer's 1961 studies and those of earlier expeditions on the structure and stratigraphy of Hecla Hoek rocks in N.W. Ny Friesland, mapping 1:20,000 around Mosselbukta and Femmilsjøen. Party D, leader P.F. Friend, continued structural studies in Devonian rocks on the eastern boundary belt in Mimerdalen, Ålanddalen and Jäderindalen, and on the Grønhorgdalen belt in Hugindalen, Grønhorgdalen and Universitetsbreen. Party E, leader J. L. Cutbill, continued investigation of Carboniferous and Permian rocks in Dickson Land and Bünsow Land. Party F, leader S. H. Buchan, mapped in S. Dickson Land and studied Triassic sections in 3 areas, including S. Dickson Land (x5) and between Sassendalen and De Geerdalen (x4). Party G, leader F. J. Vine, made a series of gravity traverses along the shores of Isfjorden and the adjoining fjords. S. F. Reddaway and W.B. Harland (Party H) carried out preliminary magnetometer studies in central Spitsbergen. (Harland 1963a,b).

(contd)

Table 1 (contd)

Year of Expedition	
1963	Party A, leader R. A. Gayer, continued investigations on the Hecla Hoek of N.W. Ny Friesland, between Mosselbukta and Femmilsjøen, and extended south to the Cookbreen area for correlation of structures and stratigraphy. Party E, leader J. L. Cutbill, examined the structures and stratigraphy of Carboniferous and Permian rocks in Dickson Land (between Pyramiden, Triungen and Odellfjellet), and around Ebbadalen, Brucebyen, and N. Bünsow Land. Party F, leader J. R. Parker, based at Vindodden (Sassenfjorden), studied stratigraphy and structure of the Billefjorden Lineament. Detailed mapping of Flowerdalen was continued from 1962 and 9 further Triassic sections measured in Sassendalen. (Harland 1964, 1965a).
1964	Party A was formed of two groups. R. H. Wallis sledged from Billefjorden and then by small boat worked northwards up the west coast of Ny Friesland and W. B. Harland set off by small boat working south from Polhem and joined him for combined studies between Austfjordnes and Sørbreen. In one season the whole western strip of Ny Friesland was reconnoitred. Party B, leader D. E. T. Bidgood, extended magnetometer studies to north Spitsbergen, begun by Harland and Reddaway in 1962. Party E, leader D. W. Holliday, continued studies begun in 1963 by Holliday (Party E) on Mid-Carboniferous stratigraphy and sedimentation in N. Billefjorden, at Odellfjellet, Trikolorfjellet, Pyramiden, Ebbadalen, and N. Bünsow Land. Party F, leader J. R. Parker, investigated stratigraphy and structure along the Billefjorden Lineament between Kjellströmdalen and Adventdalen. (Harland 1965b).
1965	Parties A and B, leader R. H. Wallis, extended structural and stratigraphic studies in north Ny Friesland, especially at Sørbreen and Austfjorden. K. Howells, continuing the work of 1962 and 1964, carried out Proton Magnetometer traverses in the Wijdefjorden area. Party D, leader P. F. Friend, completed reconnaissance of Old Red Sandstone outcrops, including S. Dickson Land, and in Austfjorden. Party E, leader D. W. Holliday, continued the work of 1963 and 1964 and completed his investigation of Mid-Carboniferous rocks of central Spitsbergen, based first at Brucebyen (working both sides of Adolfbukta and Nordenskiöldbreen) then at Skottehytta (working both sides of Petuniabukta). (Harland and Wallis 1966).
1966	K. Howells (leader), P. I. Maton and others continued the geophysical investigation of geological structures in the fjords and coastal areas of central west Spitsbergen including the south side Isfjorden (from Adventfjorden to Grönfjorden and the entrance to Sassenfjorden). (Howells 1967).
1967	R. H. Wallis (leader) and D. G. Smith worked on the stratigraphy and structural evolution of Hecla Hoek rocks, based at Mosseldalen (Wijdefjorden). P. I. Maton (leader Party B) carried out a gravimeter survey and extended magnetometer surveys of 1964, 1965 and 1966, in N. Spitsbergen (including N.W. Ny Friesland and Wijdefjorden). (Wallis and Harland 1968).
1968	Party B (leader P. I. Maton) continued gravity surveys in north and west Spitsbergen, and N.W. Isfjorden. (Maton & Harland 1969).
1972	Party C (leader W. G. Henderson with C. A. Croxton) investigated "Indre Billefjord" concession of Store Norske Spitsbergen Kulkompani. (Harland, Henderson & Smith 1973).

D. LATER CAMBRIDGE TECTONIC STUDIES

Subsequently we have come to the conclusion that the Billefjorden Fault Zone is the line of major transcurrent movement in mid-Palaeozoic time. Moreover, while engaged in various aspects of the structure of different areas and stratigraphy of different groups of rocks, we found that our combined observations as they affected this zone documented a long geological history. We therefore attempt a synthesis, believing it to be of general interest.

It might be added that most subsequent Cambridge work has been located mainly to the west or to the east of this lineament. Such work, however, has served to emphasise the importance of this line in the tectonic history of Svalbard.

2. Contributions to this paper

Specific contributions by the authors have been made according to the areas and disciplines in which they have worked as follows: Precambrian rocks of Ny Friesland by W.B.H. & R.H.W.; Old Red Sandstone by P.F.F.; Carboniferous and Permian rocks by J.L.C., D.J.G. & D.W.H.; Mesozoic rocks by J.R.P.; geophysical survey by P.I.M.; tectonic history by W.B.H. and with assistance of D.J.G. in final drafting of the paper. The results reported were not necessarily the work of the authors, who have drawn on all work available in Cambridge. Moreover the work was only made possible by assistance in the field and organization in Cambridge. The many to whom we are indebted for this are listed in the separate expedition accounts included in the references. This field work is summarized in Table 1. Much of the work referred to here has been published, as will be indicated by the references in the text, and some is new. However, the data obtained in the course of independent studies, when related, contribute to our understanding of the Billefjorden Fault Zone.

3. Nomenclature of lineament

We refer to the linear structure that is the subject of this paper generally as the Billefjorden Lineament which includes the whole length of the fault zone whether or not fault structures are exposed. The most important exposed fault of this complex is the Balliolbreen Fault and every individual fault in this complex has for convenient refence in this paper been given a name. The name Billefjorden is not used for any individual structure but is used for the whole complex of faults which we refer to as the Billefjorden Fault Zone (abbreviated B.F.Z.).

4. Stratigraphic framework

It so happens that the Billefjorden Lineament runs through the central areas of Spitsbergen where the most complete stratigraphic sequence in Svalbard is exposed. Indeed for such a small area the completeness of the stratigraphic record during the last billion years gives a remarkably continuous stratigraphic control which is the justification for attempting this detailed account of the structure and its history.

The stratigraphic sequence is summarized in Table 2 where only the briefest outline is given, its main purpose being to show the relationships of the many names needed for description.

A. STRATIGRAPHIC AGES

The stratigraphic scheme with inferred ages is compiled from many sources and the following recent publications refer either directly to the evidence for the ages quoted or review the evidence given in earlier publications.

Palaeocene: Ravn 1929; Vonderbank 1970.
Mesozoic: reviewed Harland 1973a.
  e.g. for:
  Albian: Nagy 1970;
  Cretaceous and Jurassic: Parker 1967;
  Triassic: Buchan et al. 1965; Tozer & Parker 1968.
Permian and Carboniferous: Cutbill & Challinor 1965.
Devonian: Friend 1961; Friend, Heintz and Moody-Stuart 1966.
Pre-Devonian: Harland, Wallis & Gayer 1966; Harland & Gayer 1972; Fortey & Bruton 1973.

B. STRATIGRAPHIC NOMENCLATURE – NEW PROPOSALS

After a long history of competing miscellaneous and inconsistent nomenclatural schemes, the last decade or two have seen the emergence of a coherent system of supergroups, groups, formations, members and beds. Table 2 emphasises those parts of the sequence especially related to the Lineament so that, for instance, the immense stratigraphical complexity of the Upper and Middle Hecla Hoek is presented only at group level. Nevertheless the table gives a reasonable survey of the stratigraphy of Svalbard, to which we have variously contributed.

Now, after an interval of several years since accepting or proposing the constituent parts of the scheme in separate papers, we have second thoughts on the rank of some names which we believe will be welcomed by colleagues elsewhere. We therefore propose a slight revision as follows for convenience in tabulation and for memory.

The original "Aucella shales" (Hagerman 1925) were divided into two formations Agardhfjellet and Rurikfjellet because of the break between them representing a hiatus from Mid Tithonian to part Valanginian with faulting and igneous intrusion (Parker 1967). Nevertheless in many outcrops it is not always easy to distinguish them especially in the west where deformed. At the same time the name Janusfjellet was used informally for Aucella shale and we have hitherto applied it as a subgroup to include the two constituent formations. We now, for simplicity in tabulation accept the proposal to demote Janusfjellet to the rank of formation (as in Major and Nagy 1972) and Agardhfjellet and Rurikfjellet as defined by Parker (1967) to the rank of member as now shown in Table 2.

Table 2

Stratigraphical succession along the Billefjorden Fault Zone

GROUP	FORMATION	MEMBER/BED	DOMINANT LITHOLOGY	AGE	RELATION TO LINEAMENT
VAN MIJEN-FJORDEN 1,500	Aspelintoppen 500–600		dominantly continental silts with coal measures and a few marine beds	Palaeocene	Outcrop almost entirely to west
	Battfjellet 200				
	Gilsonryggen 300				
	Sarkofagen 200				
	Basilika 20–130				
	Firkanten 110–120		productive coal seams		few outliers to east
ADVENTDALEN 1,250	Carolinefjellet 270–850	Schönrockfjellet Zillerberget	marine siltstones	Albian	
		Langstakken	marine sandstones		
		Innkjegla	marine shales & siltsts.	Aptian & ? older	
		Dalkjegla	marine sandstones		
	Helvetiafjellet 50–100	Glitrefjellet Festningen	continental sandstones	?Barremian	
	Janusfjellet 400–440	Rurikfjellet	marine shales with *Aucella*	Valanginian-Hauterivian	thins along Lineament
		Agardhfjellet		Bathonian-Volgian	eroded along narrow zone east of Flowerdalen Fault

(contd)

Table 2 (contd)

GROUP	FORMATION	MEMBER/BED	DOMINANT LITHOLOGY	AGE	RELATION TO LINEAMENT
		(Brentskardhaugen Bed	thin conglomerate)	Toarcian	eroded along narrow zone east of Flowerdalen Fault
KAPP TOSCANA 250	De Geerdalen 190		continental sandstones siltstones	?Karnian — ?Pliensbachian	
	Tschermakfjellet 63		through to marine shales	Ladinian	
SASSENDALEN 750	Botneheia 260		marine shales	Ladinian —Anisian	slight thickening to east across Lineament.
	Sticky Keep 120—310	Kaosfjellet Iskletten	marine siltstones marine shales	Spathian —Smithian	
	Vardebukta 140—290	Siksaken Selmaneset	marine siltstones marine shales	Dienerian —Griesbachian	
TEMPEL-FJORDEN 400	Kapp Starostin 381—459	Vöringen	+ cherty sandstones limestones brachiopods	«Svalbardian» cf. late Permian	thickens east of Anservika flexure.
GIPSDALEN 1500	Gipshuken 211—353		gypsum & dolomite	Artinskian	
	Nordenskiöldbreen 422—906	Tyrrellfjellet	limestones	Asselian- Sakmarian	
		Kapitol & Cadellfjellet Minkinfjellet	limestones	Late Carboniferous Moscovian	Overlaps and oversteps to the west of B.F.Z.
	Ebbadalen 281—700		limestones, shales & gypsum	Bashkirian	East of Cheopsfjellet Fault
BILLEFJORDEN 400	Svenbreen 127—200	Hultberget Sporehøgda	sandstone	Namurian	
	Hörbyebreen 189—265	Hoelbreen Triungen	shales & coal	Dinantian? ?Tournaisian	

(contd)

Table 2 (contd)

INFORMAL COLLECTIVE NAME	SUPER GROUP	GROUP	FORMATION	DOMINANT LITHOLOGY	AGE	RELATION TO LINEAMENT
OLD RED SANDSTONE		ANDRÉE LAND 4500	Mimer Valley in S 1000m	red and grey siltstones sandstones & conglomerates	Eifelian, Givetian, and late Emsian.	Outcrop along length of Lineament but only to west of it.
			equivalent to Wijde Bay 500	grey siltstones & sandstones	Eifelian and Givetian	
			Grey Hoek 1000m in N	grey siltstones	Givetian — late Emsian	
			Wood Bay 2900m	red siltstones & red or grey sandstones	Early Eifelian — Early Siegenian	
		RED BAY 0—2000		red sandstones and conglomerates	Gedinnian	Exposed only 50km or more to west
		SIKTEFJELLET 0—1500		grey sandstones & cgls.	Gedinnian — late Silurain	

(contd)

Table 2 (contd)

INFORMAL COLLECTIVE NAME	SUPER GROUP	GROUP	FORMATION	DOMINANT LITHOLOGY	AGE	RELATION TO LINEAMENT
UPPER HECLA HOEK	HINLOPEN-STRETET 1 Km	OSLOBREEN 1200 POLARISBREEN 830		limestones and dolomites tillites and shales	Llanvirn — Early Cambrian Varangian	outcrops only down dip 25 or more km to east
MIDDLE HECLA HOEK	LOMFJORDEN 5.7 Km	AKADEMIKERBREEN 1350—2490 VETERANEN 3790		limestones & dolomites greywackés & quartzites	Late Riphean	
LOWER HECLA HOEK	STUBENDORFF-BREEN 11.5 Km	PLANETFJELLA 4750	Vildadalen 3250 Flåen 1500			outcrop along Lineament but only to east of it.
			Sørbreen 250 Vassfaret 600	quartzites		
		HARKERBREEN 4100	Bangenhuk 2000	acid feldspathite		
			Rittervatnet 350	with tilloid		
			Polhem 900	quartzites		
		FINNLANDVEGGEN 2700	Smutsbreen 1200	psammites & pelites with marbles		
			Eskolabreen 1500	feldspathites & amphibolites	? 1000 Ma	

Conversely we promote Kapp Toscana to the rank of group from that of formation and its constituent members (De Geerdalen and Tschermakfjellet) as defined by BUCHAN et al. (1965) to the rank of formation, as the table now shows; the Brentskardhaugen Bed, however, remains as a bed. We could justify these changes by pointing to the great importance of the De Geerdalen rocks which are of distinctive facies and very different from the Tschermakfjellet strata. But the main reason is to avoid the anomaly of a gap there in the sequence of groups which often leads to oversights in tabulation (e.g. HARLAND 1969b p. 834).

We do not, however, wish to demote the Kongressfjellet subgroup to the rank of formation (as in MAJOR and NAGY 1972) because the Botneheia formation is a very distinctive one and is capable of subdivision into members.

Finally we take this opportunity to attempt to tidy the Old Red Sandstone nomenclature a stage further. We propose a new group name: Andrée Land, to refer to the formations that outcrop there and in Dickson Land. These are the Wijde Bay, Grey Hoek and Wood Bay Formations (FRIEND, HEINTZ and MOODY-STUART, 1966); and the Mimer Valley Formation which unit is now given formation status, in place of group status which has been used recently (ALLEN, DINELEY and FRIEND, 1968; FRIEND, 1973); its subdivisions (Planteklǿfta, Planteryggen, Fiskeklǿfta and Estheriahaugen (FRIEND, 1961)) are now given member status. The underlying Red Bay rocks have already been allocated the status of group, as have the Siktefjellet rocks.

C. THE STRATIGRAPHIC SEQUENCE
(Table 2)

The table does show clearly some points which need emphasis. The sequence divides naturally into the younger and the older rocks (HARLAND 1961), namely post-Devonian and pre-Carboniferous.

The younger rocks represent a relatively concordant platform sequence generally totalling less than 6 km, which except in the West Spitsbergen Orogen is mostly flatlying. Facies, except in the Carboniferous rocks, are fairly uniform throughout the area. These platform rocks cover the main fault and reflect its presence beneath by minor faulting and folding.

The older rocks comprise the late tectonic Old Red Sandstone separated from the platform sequence by the Svalbardian movements and transcurrent faulting so important for the story in this paper. The main Caledonian movements precede the Old Red Sandstone and define the Hecla Hoek geosyncline of some 18 km or more thickness with regional metamorphism and intrusive tectonism in the lower Hecla Hoek. The Old Red Sandstone outcrops only to the west of the Lineament and what lies beneath it is unknown. The Hecla Hoek sequence listed in the table outcrops only to the east of the Lineament. In contrast the platform rocks generally occur both east and west of the Lineament — indeed they overstep the Late Devonian fault structures.

5. Published maps

The most comprehensive topographic map of the area traversed by the Lineament is the Norsk Polarinstitutt 1/500,000 map, sheets N.W. and S.W. (1968). Larger scale maps are the 1/125,000 map of the central and north-eastern area (HARLAND and MASSON-SMITH 1962), and the 1/100,000 maps of the southern area, Norsk Polarinstitutt sheets C9 (1950, C9G 1964) and C10.

The only general geological maps of the area are: 1/1M (ORVIN 1940) in one sheet; the S.W. sheet of the 1/500,000 Norsk Polarinstitutt map (FLOOD, NAGY and WINSNES 1971); the 1/100,000 map of Adventdalen (MAJOR and NAGY 1972); and the 1/100,000 map of Bünsow Land (Gee, HARLAND and MCWHAE 1953). In addition are those published in individual papers, e.g. FRIEND and MOODY-STUART (1972, of Andrée Land) and HARLAND (1959, of Ny Friesland).

6. Plan of this paper

The fault structures and immediate stratigraphical relations are described first, from south to north, and then the more general or regional evidence bearing on the history of fault movements is considered in stratigraphical order from the top down. The analysis thus proceeds from the better known. By this arrangement, moreover, we attempt to separate observation from inference. The succeeding resumé interprets the history in chronological order, with reference to a wider setting. Running through the heart of Spitsbergen, this tectonic synthesis epitomizes not only the later movement history of this part of the Arctic, but also the nature of stratigraphical and structural evidence that delineates an ancient plate boundary.

II. Areal description of structures

It is convenient to consider the evidence related to the Fault Zone in five areas from south to north. These are named the southern area, Southwest Bünsow Land, the central area, the Austfjorden area and the Wijdefjorden area. The coastal areas of Billefjorden and Sassenfjorden are treated together for the gravity survey. This account is supported by Tables 2 and 3 relating the stratigraphic names used and giving details of individual faults.

1. The southern area
(SE of Van Mijenfjorden to Sassenfjorden, Figures 2, 3 and 4)

The southern area exposes mainly Mesozoic rocks. Their structure has been described by LIVSHITS (1965), PARKER (1966 & 1967), and MAJOR & NAGY (1972). We distinguish two post-Triassic episodes of deformation here: Albian

Table 3

Data on faults within the Billefjorden Fault Zone

NAME OF FAULT WITH SYNONYMS	RELATIONSHIPS IN PLAN	RELATIONSHIPS IN SECTION	EVIDENCE FOR AGE
Ålandvatnet Fault	Trends 350°	Downthrow 100—200m E Dips E	Cuts Ebbadalen Fm.
Anservika Flexure (JLC) Flexure K (McWhae (1953	Trends 350°	Dips W at 70° Downthrow 100m W increasing northwards and decreasing southwards. Passes down into a vertical fault in the Nordenskiöldbreen Formation	Affects Kapp Starostin Formation. Jur.-Cret. dolerite sill restricted to W. Thus Cret. or younger.
Arctowskifjellet Fault (Major & Nagy 1972)	Trends N—S. Probably a continuation of Gattytoppen Thrust	Dips W at c. 30° Downthrow E. Thrust fault. (Parker 1966)	Cuts Carolinefjellet Formation. Probably Tertiary in age.
Balliolbreen Fault Fault B (McWhae 1953)	Trends 355° O.R.S. restricted to West; Hecla Hoek to East. Interpreted as a transcurrent fault. Joins Odellfjellet Fault to N. and Svenbreen Fault to S.	Dips East 60—72° Reverse fault movement late.	Affects Mimer Valley Group overlain by Triungen Member. Thus U. Devonian. No late movement.
Bulmanfjellet Fault	Trends 350°	Dips E at 50° Downthrow 1km E.	Affects Mimer Valley Group.
Cheopsfjellet Fault Fault D (McWhae 1953)	Trends 350° Forms W. border of Billefjorden Trough. Ebbadalen Fm restricted to east. Continued southwards by Pyramiden Fault.	Dips East at high angle at south end. Downthrow 1200m + E at Odellfjellet 100m E at Cheopsfjellet. No throw at Ferdinandbreen. Small downthrow to W after Svenbreen Fm. and before Minkinfjellet Member deposited near Elsabreen.	Several movements during and just post Ebbadalen Fm (500m on Odellfjellet) Movement during Minkinfjellet Member (400m on Odellfjellet & Cheopsfjellet). Late movement — cuts Matthewbreen beds of Odellfjellet Member. Major movements post-Permian.
Cookbreen Fault	Trends 350° Faults down Svenbreen Fm into Hecla Hoek	High angle Downthrow W.	Post Svenbreen Fm.

(contd)

Table 3 (contd)

NAME OF FAULT WITH SYNONYMS	RELATIONSHIPS IN PLAN	RELATIONSHIPS IN SECTION	EVIDENCE FOR AGE
Cowantoppen Fault	Trends 350° Runs northwards into an anticline	Vertical. S of Gipsdalen no downthrow but to N downthrow increased to 200m W Sporehøgda Member against Kapitol Member.	Cuts Kapp Starostin Fm.
Ebbabreen Fault Fault H (McWhae 1953) Ragnarbreen Fault	Trends 330°	Steep dip to W. Downthrow up to 200m W	Cuts Svenbreen Fm. Overlain by Ebbadalen Fm.
Faraofjellet Fault	Trends N—S Northern continuation of Svenbreen fault?	Downthrow 200m E	Cuts Triungen Member.
Ferdinandbreen Fault	Trends N—S	Downthrow 200m E decreasing rapidly to N and also to south.	Cuts Svenbreen Fm but not Minkinfjellet Member i.e. Bashkirian movement.
Fleksurfjellet Anticline Eastern line (Major & Nagy 1972)	Trends 340°	Sharp anticline overthrust to E. in its southern part.	Affects Carolinefjellet Fm Probably Tertiary in age.
Flowerdalen Fault (Parker 1966, Major & Nagy 1972)	Trends 346°	Dip W at high angle Downthrow W 350 m	Botneheia Fm & Tschermakfjellet Fm of Kapp Toscana Group thicken to E.? Mid Triassic movement. Agardhfjellet Member of the Janusfjellet Formation, Kapp Toscana Group and upper part of Botneheia Formation eroded in narrow zone to east of fault at end of Jurassic times.
Gattytoppen Thrust (Major & Nagy 1972)	Trends 340°	Dips SW at low angle Throw 150m to E	Cuts Carolinefjellet Fm. Probably Tertiary.
Gipsdalen Fault Poss. a continuation of Terrierfjellet Fault	Trend 010°	Downthrow W	Cuts Ebbadalen Fm but not Minkinfjellet Fm.
Gipshuken Fault Flexure Zone D (McWhae 1953)		Vertical Downthrow 200m W decreasing to 0m northwards.	Cuts Kapp Starostin Fm.

(contd)

Table 3 (contd)

NAME OF FAULT WITH SYNONYMS	RELATIONSHIPS IN PLAN	RELATIONSHIPS IN SECTION	EVIDENCE FOR AGE
Gizehfjellet Flexure Develops into Wordiekammen and Lövehovden Flexures in the south.	Trends 335°	Downthrow 500m W	Cuts Tyrrellfjellet Member Post Permian.
Karnakfjellet Fault	Trends 357°	Downthrow 400m E	Cuts Sporehøgda Member Prob. pre-Ebbadalen Fm.
Kinanderfjellet Fault	Trends 012°	Downthrow 150m W	Cuts Hoelbreen Member Overlain by Kapitol Member Pre-Gzhelian, prob. pre Ebbadalen Fm.
Lemströmfjellet Fault (HARLAND 1941)	Trends 354° Hörbyebreen Fm restricted to W. Poss. continuous with Ragnarbreen Fault	Downthrow 500m W	Cuts Sporehøgda Member. Prob. movement post Hörbyebreen and pre Svenbreen Members also post Svenbreen.
Lövehovden Flexure southern development of Gizehfjellet Flexure	Trends 010° — 330°	Downthrow 380m W	Affects Tyrrellfjellet Member.
Munindalen Fault Fault J (McWHAE 1953) (STENSIÖ 1918)	Trends 340°	Dips E at 45° Downthrow to W. 1Km.	Cuts Mimer Valley Group. Overlain by Kapitol Member.
Odellfjellet Fault	Trends 357° Northern continuation of Svenbreen Fault	Downthrow 100m W	Cuts Hoelbreen Member.
Petuniabukta Flexure	Trends N—S Associated with Pyramiden Fault	Downthrow up to 1000m E	Affects Tyrrellfjellet Member.
Pyramiden Fault Fault D (McWHAE 1953)	Trends N—S Southern continuation of Cheopsfjellet Fault	Downthrow up to 1000m E	Affects Tyrrellfjellet Member.
Ragnardalen Fault	Trends N—S May join with Lemströmfjellet Fault to N	Downthrow 40m W Dips E to vertical	Cuts Ebbadalen Fm Overlain by Minkinfjellet Member.
Sentinelfjellet Fault	Trends 355°	Downthrow 100m E	Cuts Triungen Member.

(contd)

Table 3 (contd)

NAME OF FAULT WITH SYNONYMS	RELATIONSHIPS IN PLAN	RELATIONSHIPS IN SECTION	EVIDENCE FOR AGE
Skolten Anticline Arnicadalen Fault Zone (LIVSHITS 1965) Western Line (MAJOR & NAGY 1972)	Trends 345°	Sharp Anticline overthrust to E	Affects Carolinefjellet Fm. Probably Tertiary in age.
Svenbreen Fault	Trends 355°	Downthrow E in Carboniferous later Downthrow 300m W — inferred from base of Permian across Bertilbreen	Post Billefjorden Group pre Ebbadalen Fm. and during Minkinfjellet Member. Later cuts Tyrrellfjellet Member
Terrierfjellet Fault (WILSON 1958)	Trends N—S	Downthrow large in Hecla Hoek Downthrow 20m W in Carboniferous	Pre Carboniferous — large throw in Hecla Hoek. Later movement post Ebbadalen Fm pre-Minkinfjellet Member
Triungen Fault	Trends 350°	Downthrow E	Cuts Triungen Member
Wordiekammen Flexure	Trends 010°—330°	Downthrow 200m W	Affects Tyrrellfjellet Member
Yggdrasilkampen Fault	Trends 350° Southern continuation of Svenbreen Fault	Downthrow E 800—900m	Cuts Kapitol Member movements post-Billefjorden pre Ebbadalen (400m): Minkinfjellet Member (300—400m) Later movement (100m)

strata (the youngest rocks exposed within most of the fault zone) are affected by the latest movements while the earlier structures are post-part Volgian and pre-part Valanginian.

The earlier structure was an anticline running NW from Fleksurfjellet to Sassenfjorden, east of Flowerdalen (PARKER 1966). In the south this is an open structure (width 3 km, amplitude 350 m) asymmetrical with a more gentle western slope. To the north a single main fault or narrow fault zone develops on the western limb — the Flowerdalen Fault, downthrow 350 m, and here the structure is about 6 km wide with an amplitude of 500 m.

On Fleksurfjellet and Marmierfjellet these structures are seen to affect rocks from the Sassenfjorden Group (Triassic) to the Agardhfjellet Member (Upper Jurassic) of the Janusfjellet Formation but not to affect the Rurikfjellet Member (Lower Cretaceous) of the Janusfjellet Formation. However, this anticlinal structure persisted as a positive submarine feature during the deposition of the Rurikfjellet Member.

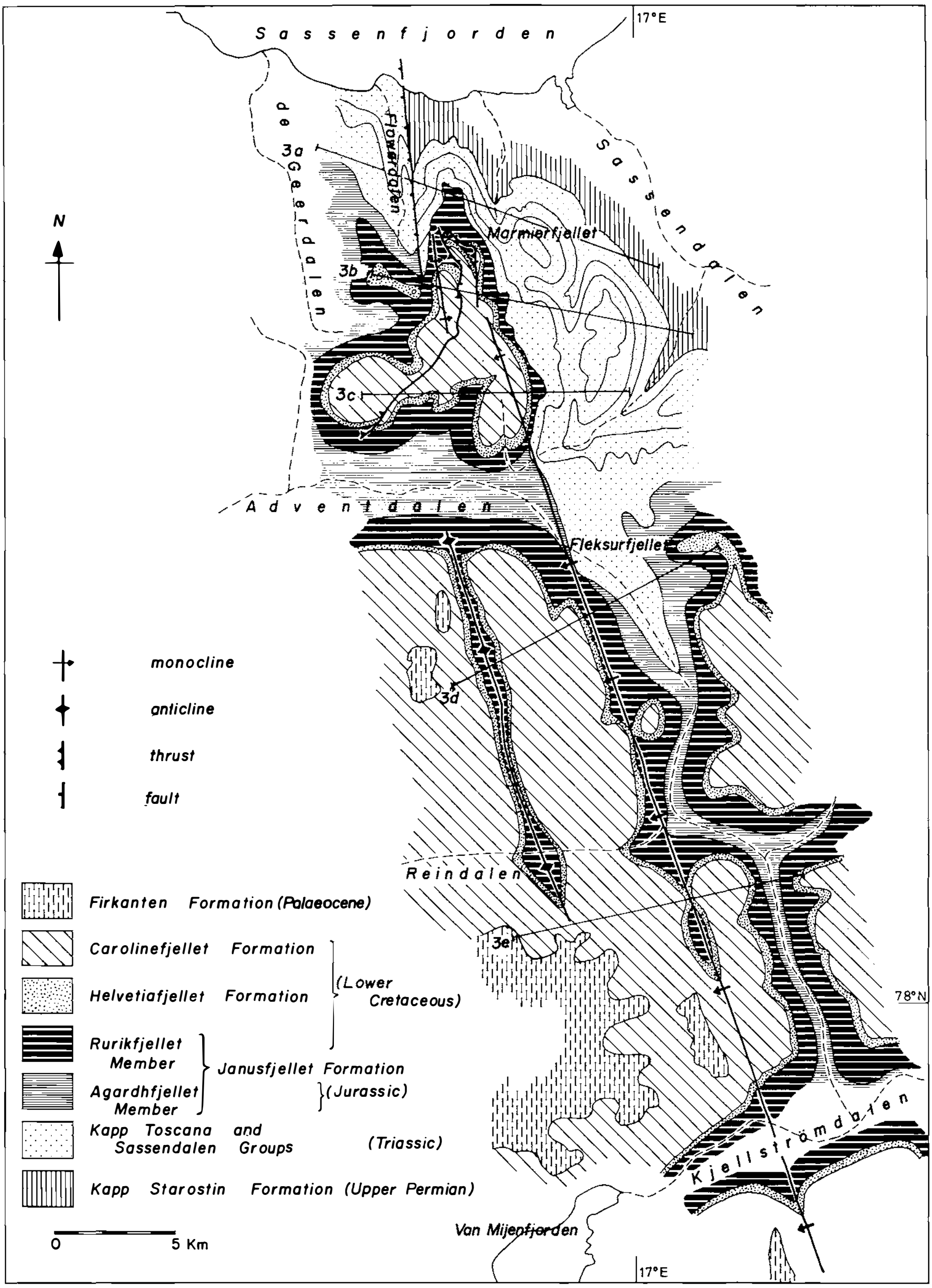
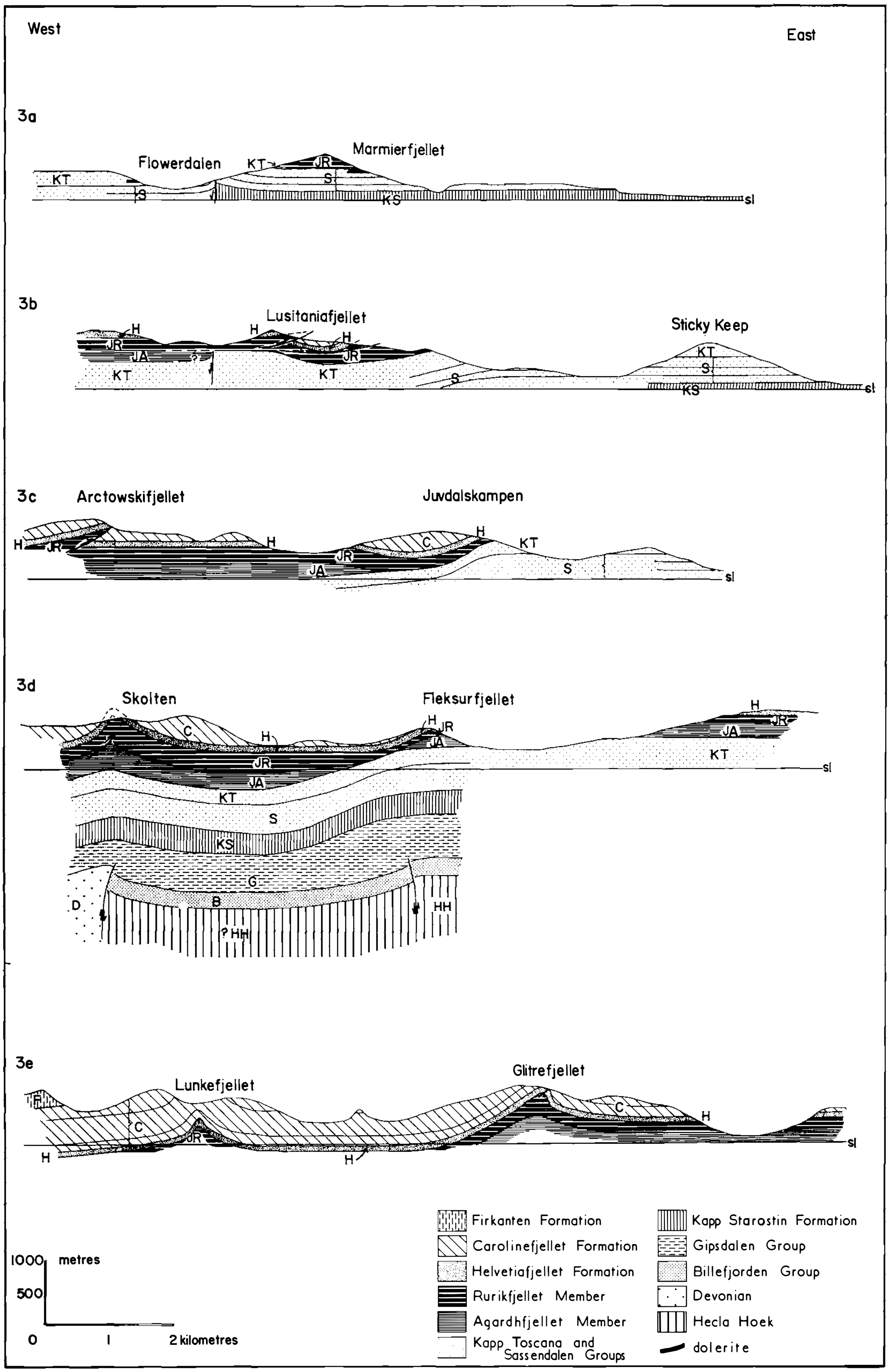


Fig. 2. *Billefjorden Fault Zone; southern area. Geological map of the Sassenfjorden—Kjellströmdalen area of the Billefjorden Lineament showing location of cross sections in Figure 3.*

Fig. 3. *Cross sections through the Billefjorden Lineament in the Sassenfjorden—Kjellströmdalen area. Location of cross sections shown on Figure 2. sl = sea level. Cross section 3d shows an interpretation of the deeper structure.* →


West
East
3a
Flowerdalen
Marmierfjellet
KT
S
JR
KS
sl
3b
Lusitaniafjellet
Sticky Keep
H
JR
JA
KT
S
KS
sl
3c
Arctowskifjellet
Juvdalskampen
H
JR
JA
C
KT
S
sl
3d
Skolten
Fleksurfjellet
C
H
JR
JA
KT
S
KS
G
B
D
HH
?HH
sl
3e
Lunkefjellet
Glitrefjellet
F
C
JR
H
sl
1000 metres
500
0
1
2 kilometres
Firkanten Formation
Carolinefjellet Formation
Helvetiafjellet Formation
Rurikfjellet Member
Agardhfjellet Member
Kapp Toscana and Sassendalen Groups
Kapp Starostin Formation
Gipsdalen Group
Billefjorden Group
Devonian
Hecla Hoek
dolerite

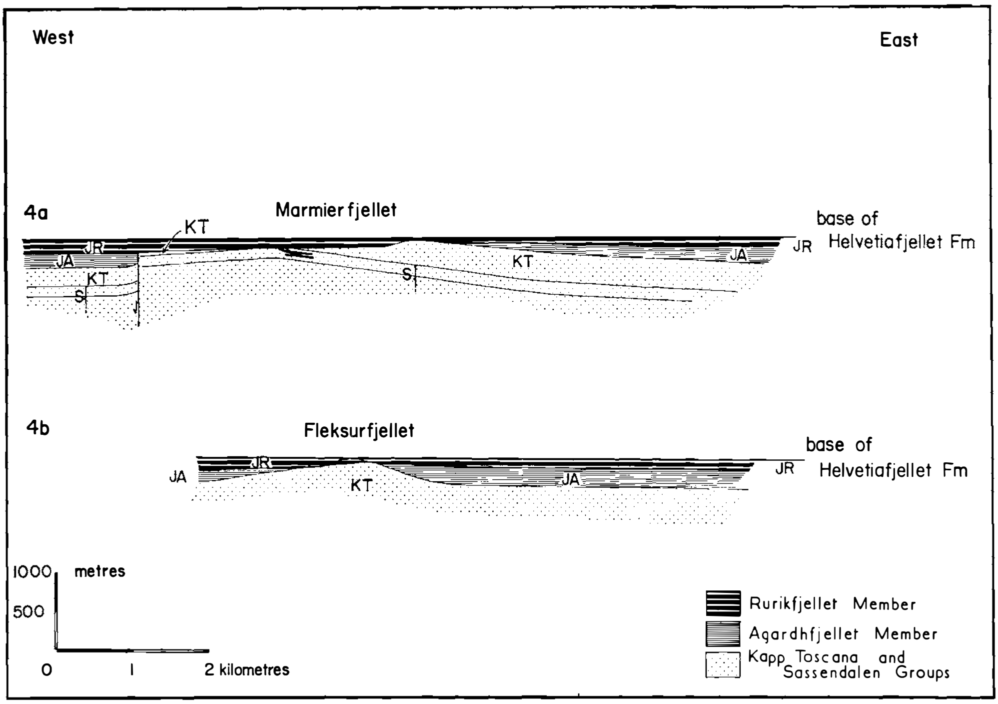


Fig. 4. *Cross sections showing the pre Tertiary structure in the Sassenfjorden—Kjellströmdalen area. Datum is base of the Helvetiafjellet Formation. Cross section 4a is composite, based on cross sections 3a—3c; the dolerites on Marmierfjellet are shown diagrammatically only. Cross section 4b is based on cross section 3d.*

The later group of structures diverge from Marmierfjellet southwards, one persisting along the line of the Flowerdalen Fault and the other lying to the east along the eastern limb of the pre-existing anticline.

The eastern line of structures consists of a westward facing monocline which passes to the south into a sharp anticline in places with an overthrust western limb. This line of structures is seen from Fleksurfjellet southwards to Kjellströmdalen and may continue to Kvalvågen. In the south the effective western downthrow is 100 m but this increases to 600 m in the north.

The western line (Arnicadalen fault zone of Livshits 1965) consists of the eastwardly directed Gattytoppen Thrust in the north, which further south passes into a sharp anticlinal structure cut by a high angle reverse fault persisting to the south of Reindalen. The effective eastern downthrow increases from 220 m in the south to 250 m in the north.

The effect of these two divergent asymmetrical anticlinal structural lines is to create a synclinal structure running parallel to the earlier anticline. The structures appear within the area of westward regional dip without any net displacement to east or west across the fault zone.

The later structures are seen to affect the Cretaceous and Palaeogene rocks (Major and Nagy 1972).

2. Southwest Bünsow land
(Figures 5 and 6)

Bünsow Land exposes Permian and Carboniferous strata and its southwest corner contains dolerite intrusions assumed to be of Mesozoic age, probably post-Lower Volgian and pre-Valanginian (Parker 1966). In the absence of post-Permian strata Mesozoic and Cenozoic events cannot be distinguished except from Palaeozoic events or by postulating an approximately Berriasian age for a dolerite sill.

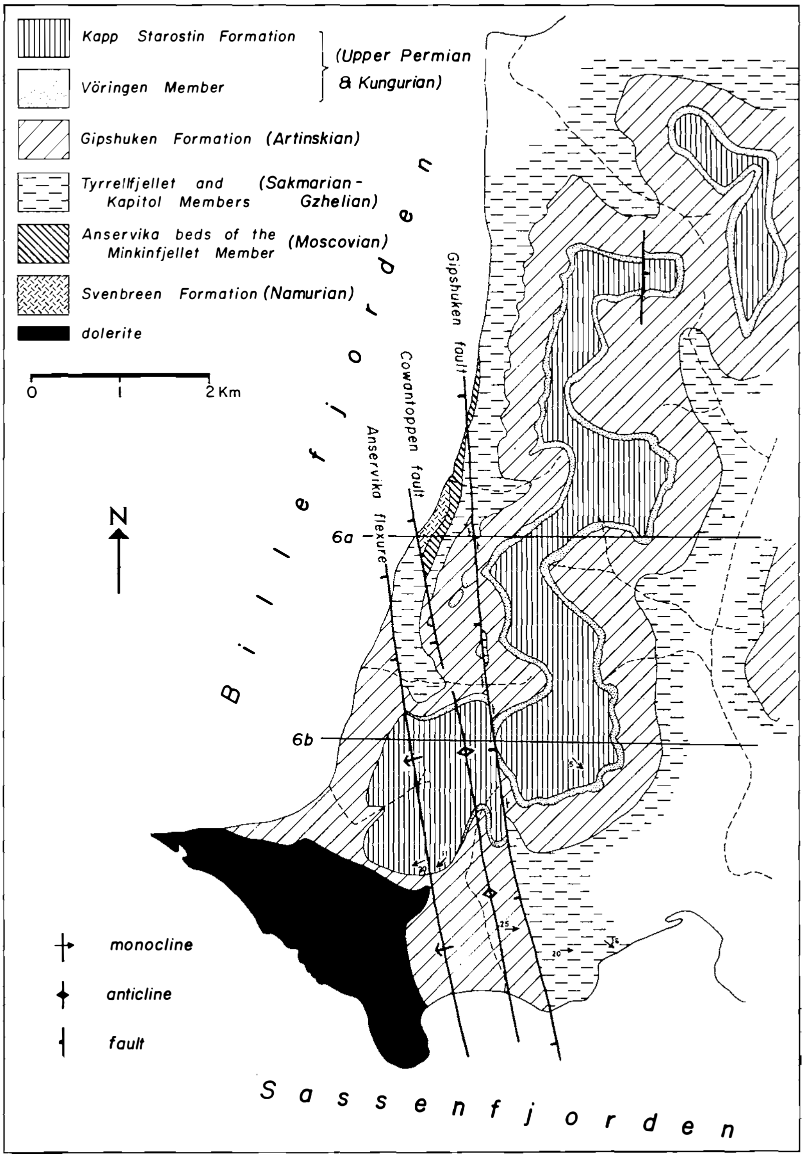
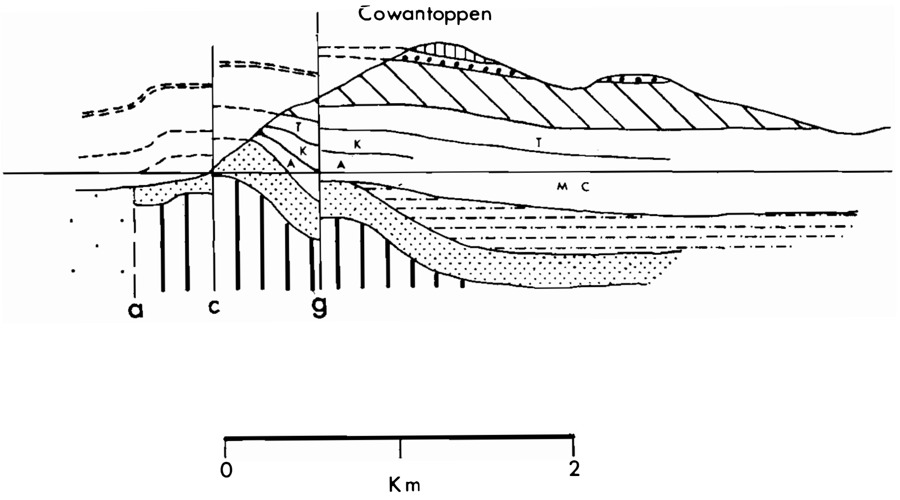
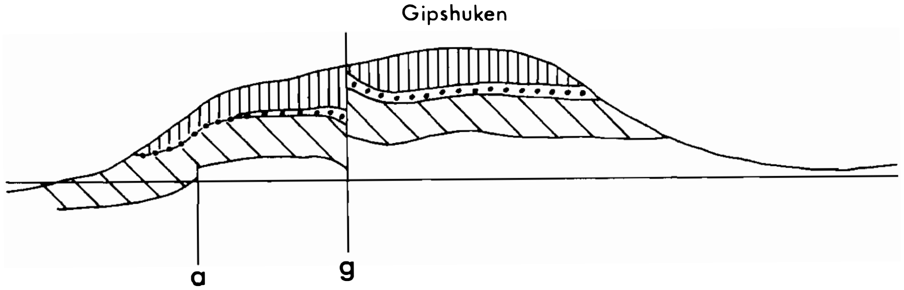


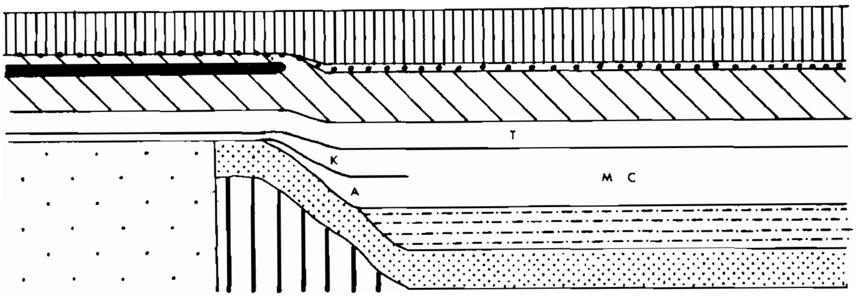
Fig. 5. *Billefjorden Fault Zone; geological map of southwest Bünsow Land, showing lines of sections in Figure 6.*


Cowantoppen
T
K
A
K
T
A
M C
a
c
g
0
2
Km

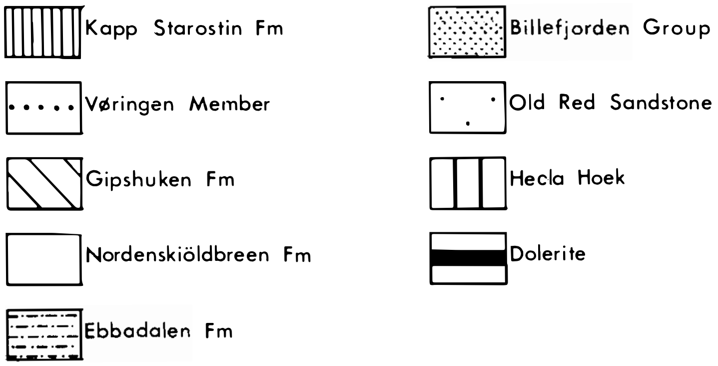
6a


Gipshuken
a
g

6b


T
K
A
M C

6c


Kapp Starostin Fm
Vøringen Member
Gipshuken Fm
Nordenskiöldbreen Fm
Ebbadalen Fm
Billefjorden Group
Old Red Sandstone
Hecla Hoek
Dolerite

A system of three parallel faults and flexures throws the Kapp Starostin Formation (Upper Permian) about 500 m down to the west. The faults are almost vertical and at its north end the Anservika flexure in the Gipshuken and Kapp Starostin Formations passes down into a vertical fault in the Nordenskiöldbreen Formation (Fig. 6b). The displacement along each structure varies markedly but the total displacement across the three structures is nearly constant. This west-facing system is superimposed on an older, broad, east-facing flexure whose presence is shown by rapid eastwards increase in thickness in the lower half of the Kapp Starostin Formation and in the Minkinfjellet and Cadellfjellet Members (Moscovian to Orenburgian) of the Nordenskiöldbreen Formation (Fig. 6a and c).

A large dolerite sill is intruded into the Gipshuken Formation (Artinskian) and terminates to the east on the line of the Anservika flexure which itself is in the western part of the later Palaeozoic flexure. Another intrusion is found 7–8 km to the east on Templet on the top of the Kapp Starostin Formation. This apparent transgression to a stratigraphically high horizon suggests that the dolerites were intruded prior to the formation of the main west-facing structures and the difference in the level of intrusion was controlled by the east-facing flexure in the Kapp Starostin Formation (Fig. 6c). This indicates that the westfacing structures are Cretaceous or younger.

3. Gravity traverses around Billefjorden, Sassenfjorden and Isfjorden (Fig. 7)

Subsequent work on the gravity survey, carried out by Party G of C.S.E. 1962 (Leader F. J. VINE), by K. HOWELLS revealed in the Bouguer anomaly profiles various gradients upon which were superimposed shorter wavelength local anomalies. The uneven distribution of data led to the use of a trend surface analysis approach in order to reveal the overall regional gradient. It also necessitates the caution that the following remarks may require amendment in the light of further gravimetric survey in the area.

The sixth degree surface was assumed to be the regional gravity gradient as it provided a better fit for each traverse than the other surfaces. This is shown in Fig. 7, and it can be seen that there is a fall of about 44 milligals between the north-east end of Tempelfjorden and the south-west coast of Adventfjorden.

The regional gravity gradient is relatively steep over Tempelfjorden and Sassenfjorden and flattens out over Isfjorden and Nordfjorden. The trend of the gradient approximates to the trends of the major geological structures of the area. Over Tempelfjorden and Sassenfjorden it is mainly NW to SE, but swings around to N —S at the mouth of Billefjorden where the southward continuation

Fig. 6. *Sections across southwest Bünsow Land. Location of cross sections 6a and 6b shown on Figure 5; 6c is a reconstruction at the end of the Permian showing the Gipsvika Flexure. a Anservika flexure, c Cowentoppen fault, g Gipshuken fault. Subdivisions of the Nordenskiöldbreen Formation shown as follows; A Anservika Beds, C Cadellfjellet Member, K Kapitol Member, M Minkinfjellet Member, T Tyrrellfjellet Member.* ←

of the eastern boundary fault of the Devonian graben intersects Gipshuken and the coast at Vindodden on the south side of Sassenfjorden (Fig. 7). Similarly, over Ekmanfjorden, the trend swings round to a NNW—SSE direction over the Ekmanfjorden fault which is the southward continuation of the Breibogen and Rabotdalen—Hannabreen faults, which form the western margin of the Devonian graben. The ENE—WSW gravity trends over the area between Billefjorden and Ekmanfjorden parallel the trends of the outcrop margins of the Carboniferous, Permian and Triassic deposits in this area, with the negative gravity gradient to the south-west coinciding with an increase in thickness of the Upper Palaeozoic and Mesozoic sediments, as shown for the Carboniferous and Permian deposits by CUTBILL and CHALLINOR (1965). The overall negative gradient is probably a result of the increasing thickness of lighter sediments towards the centre of the Spitsbergen Mesozoic and Tertiary sedimentary basin. The regional gravity trend is in the same direction as the line of dislocation which DE GEER (1909) postulated to extend from Brøggerhalvøya to Yoldiabukta. It is also the trend of the edge of the Mesozoic and Tertiary sediments in this area.

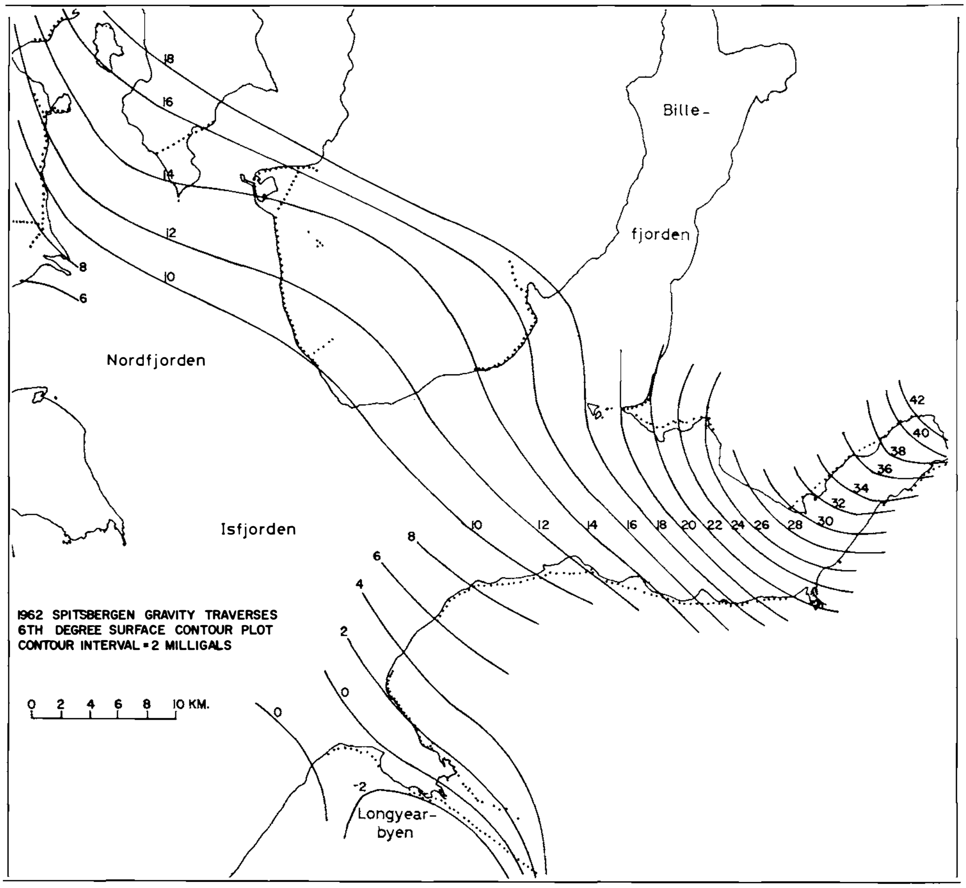


Fig. 7. *Sixth degree surface of gravity field of Isfjorden.*

Kurinin (1965) estimated the weighted mean density of the Upper and Middle Palaeozoic, Mesozoic and early Cenozoic sediments to be 2.69 gm/cc and the mean density of the Hecla Hoek metamorphics to be 2.72 gm/cc. From our work 2.76 gm/cc is a more probable average density for the Hecla Hoek rocks. Using the formula for the attraction of an infinite slab, a density contrast of 0.07 gm/cc and an anomaly of 44 milligals gives the thickness of sediments overlying the Hecla Hoek as 15 km at Adventfjorden, if it is assumed that the Hecla Hoek is close to or at the surface at the eastern end of Tempelfjorden. From stratigraphic evidence, this thickness of overlying sediments is unexpectedly large. Orvin (1940) showed a thickness of only 3.5 km of sediments in a geological cross-section 30 km to the south of Adventfjorden. Moreover our own figures of exposed thickness for Carboniferous and Permian strata well north of Adventdalen would fall between about 1.5 and 3.0 km.

Kurinin sampled the overlying sediments from the Isfjorden area and, except for early Cenozoic rocks whose density was 2.61 gm/cc, their mean densities were close to 2.72 gm/cc, varying from 2.67 gm/cc to 2.79 gm/cc. Hence, it is probable that further density contrasts are to be found within the Hecla Hoek rocks and possibly within the overlying sediments. Density contrasts of 0.1 gm/cc and 0.15 gm/cc give thicknesses of 10.5 and 7 km respectively.

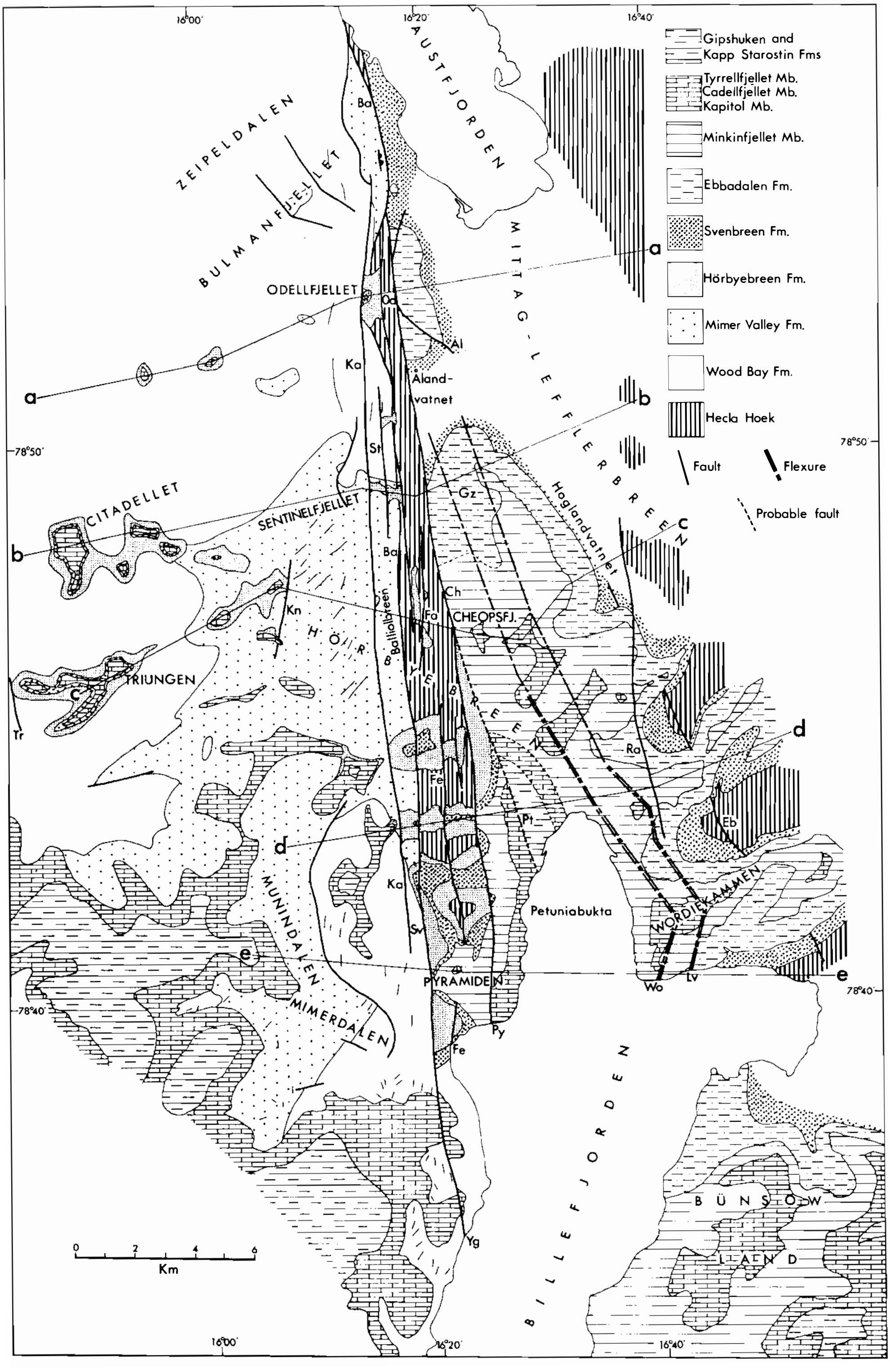
Thick local development of gypsum (density 2.2—2.4) rather than of anhydrite (density 2.7—3.0) in the late Carboniferous or early Permian strata would go some way to reducing the figures to be explained. More than 1 km thickness would be needed to explain the whole amount. Hecla Hoek metamorphic rocks range from amphibolites and marbles of density around 2.8 or 2.9 to quartzites of density around 2.7, which is not significantly different from the density of the Old Red Sandstone (see Table 4), and great thicknesses of these strata are known so that alternative possibilities are easy to envisage and it is a challenge to apply this density contrast constraint to the interpretation of deeper stratigraphy.

The steep gradient over Tempelfjorden cannot be explained entirely by increasing post-Caledonian sediment thickness to the west. Devonian sediments are not known to the east of the Balliolbreen Fault. Hence it is probable that the steep gradient is due to density contrasts within or at the base of the Hecla Hoek, and may even be partly due to crustal thinning over the southward extension of Ny Friesland (see Section III below).

4. The central area

(Figures 8 and 9)

This area exposes the whole width of the Fault Zone and shows the fault relationships of rocks of different ages from Precambrian to Permian. Figure 8 shows the distribution of the faults and Table 3 gives details of their movements. They will be described briefly below in relationship to the Balliolbreen Fault Line.


16°00'
16°20'
16°40'
78°50'
78°40'
Gipshuken and Kapp Starostin Fms
Tyrrellfjellet Mb.
Cadellfjellet Mb.
Kapitol Mb.
Minkinfjellet Mb.
Ebbadalen Fm.
Svenbreen Fm.
Hörbyebreen Fm.
Mimer Valley Fm.
Wood Bay Fm.
Hecla Hoek
Fault
Flexure
Probable fault
AUSTFJORDEN
ZEIPELDALEN
BULMANFJELLET
ODELLFJELLET
MITTAG-LEFFLERBREEN
Ålandvatnet
CITADELLET
SENTINELFJELLET
Hoglandvatnet
TRIUNGEN
CHEOPSFJ.
HÖRBYEBREEN
Balliolbreen
MUNINDALEN
MIMERDALEN
Petuniabukta
WORDIEKAMMEN
PYRAMIDEN
BILLEFJORDEN
BÜNSOW LAND
Ba
Od
Al
Ka
St
Gz
Kn
Ch
Fa
Fe
Pt
Ra
Eb
Sv
Py
Wo
Lv
Yg
Tr
a
b
c
d
e
0
2
4
6
Km

A. THE BALLIOLBREEN FAULT LINE

The Balliolbreen Fault is the most conspicuous structure in the area. On Odellfjellet highly tectonized metamorphic Hecla Hoek rocks of the Harkerbreen Group on the east are faulted against Lower Old Red Sandstone (Wood Bay Formation). On Sentinelfjellet the Hecla Hoek is faulted against the Mimer Valley Formation (uppermost Givetian). The Balliolbreen Fault is overlain by the unfaulted Triungen Member (Tournaisian) of the Billefjorden Group. Thus the age of the Balliolbreen Fault and the adjacent eastern boundary fold belt must be younger than uppermost Givetian and older than Tournaisian — that is, of Late Devonian age. South of Sentinelfjellet this contact is obscured by Carboniferous cover or ice. The Fault is high angle (60—70 degrees), eastward dipping and reverse, but the greater part of its reverse nature is due to post Early Carboniferous tilting. Dip-slip slickensiding is abundant in Old Red Sandstone rocks adjacent to the Fault, e.g. east Bulmanfjellet. An estimate of

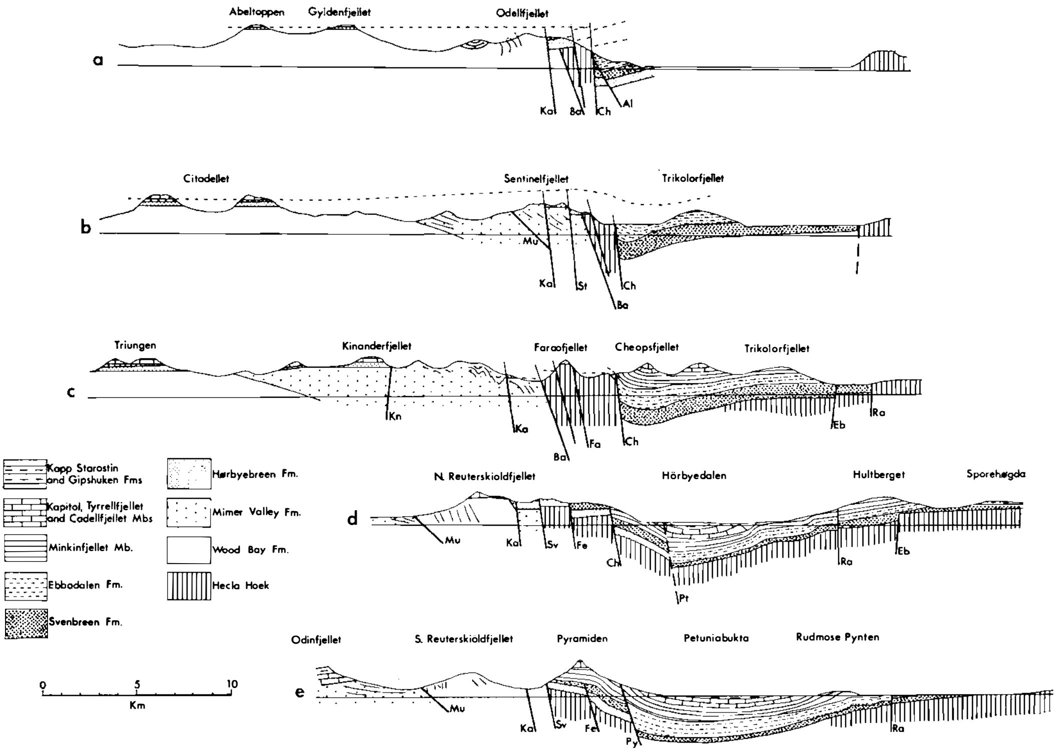


Fig. 9. *Sections across the central area of the Billefjorden Fault Zone. Positions of sections and faults identified on Figure 8.*

Fig. 8. *Billefjorden Fault Zone; central area. Geological map showing lines of cross sections a—e in Figure 9. Faults, Ål Ålandvatnet; Ba Balliolbreen; Ch Cheopsfjellet; Eb Ebbabreen; Fa Faraofjellet; Fe Ferdinandbreen; Ka Karnakfjellet; Kn Kinanderfjellet; Od Odellfjellet; Pt Petuniabukta; Py Pyramiden; Ra Ragnardalen; St Sentinelfjellet; Sv Svenbreen; Yg Yggdrasilkampen. Flexures, Gz Gizehfjellet; Lv Løvehovden; Wo Wordiekammen. Thin lines in Wood Bay and Mimer Valley Formation areas are strike lines where strata dip more than 30°.*

←

the minimum thickness of Old Red Sandstone rocks brought down along the Fault gives it a minimum dip-slip of 2000 m. The Fault is not offset by cross structures in any part of its outcrop on land. The apparent southern continuation of the Balliolbreen Fault, the Svenbreen Fault, also separates Old Red Sandstone in the west from Hecla Hoek in the east. However Middle Carboniferous movements on this Fault gave a downthrow to the east and later movements a downthrow to the west (Table 3).

Further south along this line, the Yggdrasilkampen Fault shows both Middle Carboniferous and later movements down-east (Table 3).

B. STRUCTURES EAST OF THE BALLIOLBREEN FAULT LINE

East of the Balliolbreen Fault lies the Billefjorden trough, bounded by inward facing faults and flexures (Cutbill and Challinor 1965). Old Red Sandstone rocks are everywhere absent and Carboniferous rocks rest unconformably on the Hecla Hoek. The thick Carboniferous sequence has been affected by block faulting. A number of faults more or less parallel to the Balliolbreen Fault were active during Carboniferous time and later (Fig. 9). Thus the Dinantian Hörbyebreen Formation is found only to the west-facing Lemströmfjellet Fault and to the west of Billefjorden. This predates the formation of the Billefjorden trough. The Namurian Svenbreen Formation is thin or missing on the East Dickson Land Axis (Cutbill and Challinor 1965) and is succeeded by the Bashkirian Ebbadalen Formation only in the developing trough to the east of the east-facing Cheopsfjellet and Pyramiden Faults. The younger Minkinfjellet Member extends westwards to the line of the Balliolbreen Fault (Svenbreen Fault).

The east side of the Billefjorden trough shows west-facing structures. The Ebbabreen Fault, which predates the Ebbadalen Formation, throwing down the Svenbreen Formation against Hecla Hoek to the east, can be traced NNW to Hoglandvatnet. The Ragnardalen Fault (downthrow about 40 m west) cuts the Ebbadalen Formation but does not affect the Minkinfjellet Member. Further to the east the Terrierfjellet Fault predates the Minkinfjellet Member, downfaulting to the west the Ebbadalen Formation. This Fault is probably continuous with the Gipsdalen Fault which shows similar relationships. Later west-facing flexures affected the Nordenskiöldbreen Formation on Wordiekammen and continued NNW as the more gentle Gizehfjellet Flexure (Fig. 9)
.

C. STRUCTURES WEST OF THE BALLIOLBREEN FAULT LINE

West of the Balliolbreen Fault is a belt of deformed Old Red Sandstone rocks up to 4 km wide and extending 50 km between Billefjorden and Austfjorden. Both the fold belt and the Balliolbreen Fault run parallel to the strike of the Hecla Hoek rocks of Sentinelfjellet and Odellfjellet and to their regional strike in Ny Friesland.

Axial traces in the folds trend from 340° to 080°, with a mode at 020° (Fig. 8). The folds with 060° to 080° trends are open folds crossing tighter folds with the

020° trend. There is no systematic relationship between axial trend and proximity to the Fault. No fold can be traced for more than 5 km, but the more persistent folds have variable trends giving sinuous outcrops.

The majority of folds are asymmetrical, with steep westward limbs which are in some cases overturned. Major anticlines with this asymmetry are exposed in the Billefjorden and Austfjorden regions, deforming up to 1500 m of sediment. These folds plunge north and south respectively towards a central region of Mimer Valley Formation outcrop in which the folds have no marked plunge. Fracture cleavage occurs in fine-grained sediments throughout the belt. It is usually steeply dipping and varies in strike between 360° and 030°, but in parts of the west coast of Billefjorden a gently dipping set of cleavage surfaces dominates the bedding of the inverted western limb of a major asymmetric anticline.

Two major faults occur in the Old Red Sandstone rocks. The Munindalen Fault trends 340° and dips 45° east, with a downthrow of about 1 km to the west. The Bulmanfjellet Fault traverses east Bulmanfjellet, at 350° dipping 50° west with a downthrow of about 1 km to the east. Both Faults cut folds in the neighbouring rocks. Two minor E—W trending faults, apparently with dextral strike-slip components, occur in central Mimerdalen. Two other major N—S trending faults affect both the Old Red Sandstone and the Carboniferous rocks. One of these, the Kinander Fault, crosses the upper end of Hörbyebreen and throws down the Hörbyebreen Formation 150 m to the west against the Lower Devonian Wood Bay Formation. The Kapitol Member of the Nordenskiöldbreen Formation (Upper Carboniferous) lies undisturbed over the Fault.

The other, the Karnak Fault, runs almost straight from southwest of Bulmanfjellet south to Mimerdalen. Its southern half is largely ice-covered. It throws down the Sporehøgda Member of the Svenbreen Formation at least 400 m to the east and is probably also pre-Upper Carboniferous in age.

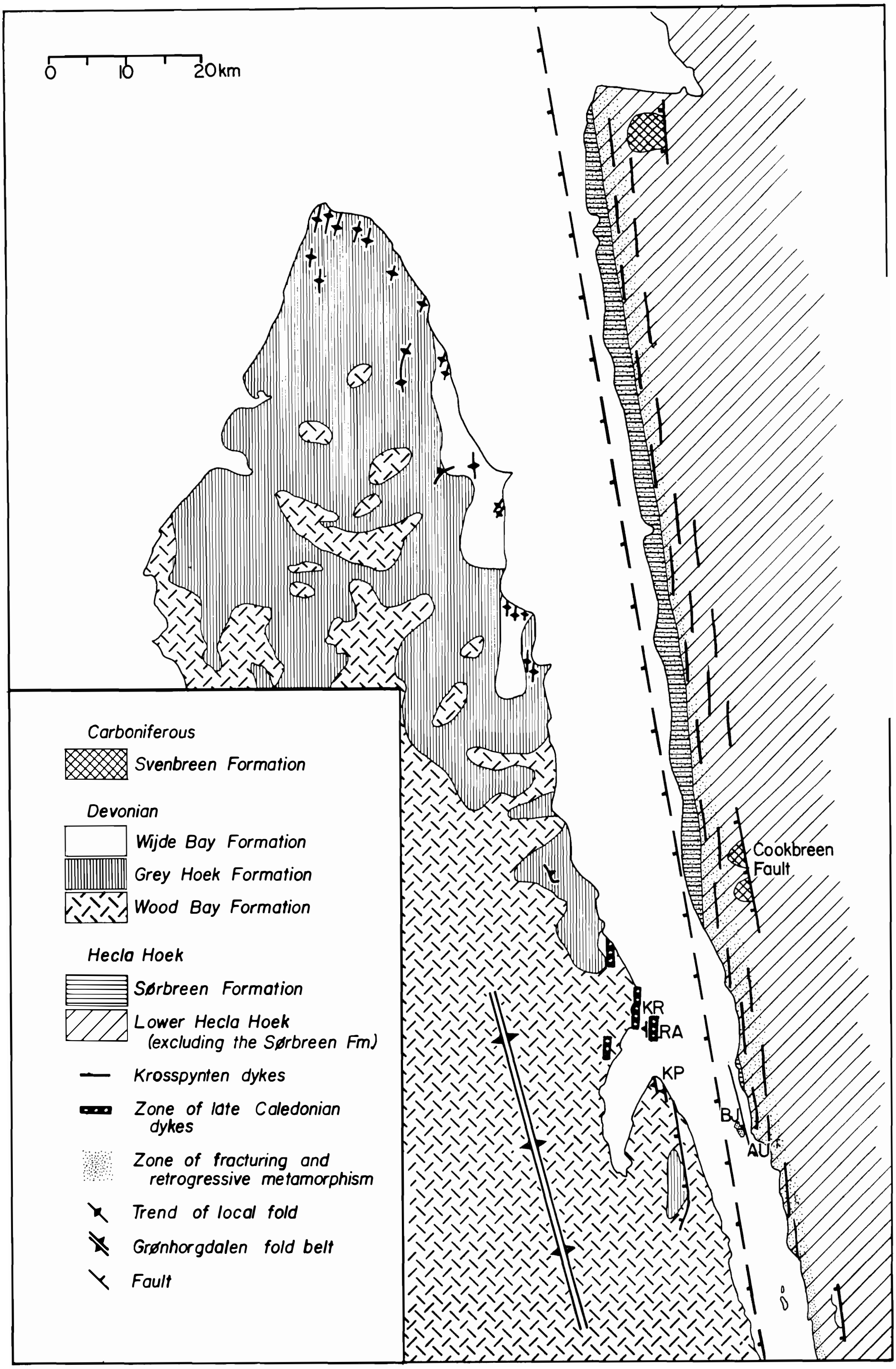
5. The Austfjorden and Wijdefjorden areas

(Figures 10 to 15; Table 4)

Wijdefjorden and its major branch, Austfjorden, constitute an almost straight waterway running N —S for more than 100 km and varying in width from 5 km at the south to 20 km in the north. It has long been regarded as a major fault line because its eastern shores (Ny Friesland) are formed of metamorphosed Hecla Hoek rocks while softer Old Red Sandstones outcrop throughout its length on the west. Because of the presumed erosion of the Fault Zone to below sea level much evidence as to its nature is indirect.

A. SURFACE CONFIGURATION

The west coast of Ny Friesland follows the strike of the Hecla Hoek rocks very exactly for 65 km (i.e. most) of its northern course (Fig. 10). Extrapolation of the Balliolbreen Fault Line would suggest that it follows this coastline very closely. It is likely that the Hecla Hoek rocks are more resistant to erosion than the Old Red Sandstone, and that most of Wijdefjorden has therefore been eroded from Old Red Sandstone.


0 10 20km
Cookbreen Fault
KR
RA
KP
BJ
AU
Carboniferous
Svenbreen Formation
Devonian
Wijde Bay Formation
Grey Hoek Formation
Wood Bay Formation
Hecla Hoek
Sørbreen Formation
Lower Hecla Hoek (excluding the Sørbreen Fm)
Krosspynten dykes
Zone of late Caledonian dykes
Zone of fracturing and retrogressive metamorphism
Trend of local fold
Grønhorgdalen fold belt
Fault

Hydrographic survey (Figs. 11a & b) also suggests such a course for the Fault. In Austfjorden a narrow trench is evident in the bed of the fjord. This trench may mark the trace of a fault over-deepened by glacial scour. The trench, and hence presumably the Fault, is close to the western shore at the southern end of Austfjorden, and gradually crosses to the eastern side of Wijdefjorden by Bangenhuken, whence also the trench gives way to a steep west-facing step. Our first hydrographic survey (1965) was conducted in connexion with proton magnetometer traverses and resulted in a series of echo profiles. Subsequently the Norsk Polarinstitutt has conducted a more complete hydrographic survey of the fjord.

B. EXPOSED STRUCTURES

The Balliolbreen Fault runs along the southwestern shore of Austfjorden and enters the fjord just north of Zeipeldalen. If its strike is extrapolated northwards it impinges on the continuous Hecla Hoek outcrops of Ny Friesland near Austfjordnes, 20 km north. This raises the possibilities of some slight offset by sinuous fault surface, crossfaulting or by thrust faulting.

The southern part of Austfjorden is a continuation of the Mittag-Lefflerbreen depression lying east of the Balliolbreen Fault, and west of the Lemströmfjellet Fault (HARLAND 1941). The latter cuts the Sporehøgda Member of the Svenbreen Formation and has a downthrow of 500 m to the west. The only demonstrated movement is post-Sporehøgda Member, but as the fault appears to mark the eastern limit of the Hörbyebreen Formation there was most probably earlier movement along it. The Lemströmfjellet Fault may be continuous with the Ragnardalen Fault.

Hecla Hoek rocks of the eastern coastal belt of Austfjorden and Wijdefjorden show a marked degree of shattering and retrogressive metamorphism. Extensive shatter zones and joint sets (more or less normal to the fjord) affect all the formations, but do not extend more than 4 km eastwards. Between the fjord-side and the axial belt of Lower Hecla Hoek rocks, 8 km inland, all formations have suffered late retrogressive metamorphism which is not found to the east of the axial belt.

On Bjørnnesholmen, an island in Austfjorden, Lower Hecla Hoek rocks are intensely shattered and show an anomalous strike of 030°, contrasting with the general 350° strike of Ny Friesland. This strike swing, in very steeply dipping rocks, could result from a dextral strike-slip movement under relatively small overburden and is therefore probably a late structure.

Along the western coast of Austfjorden the Old Red Sandstone dips fairly gently until, in the immediate vicinity of Kapp Petermann, and on Ræstadholmen, steep westward dips occur. Just east of Kapp Petermann a normal fault down-throwing 100 m west extends for 12 km to the South.

Fig. 10. *Billefjorden Fault Zone; geological map of Austfjorden and Wijdefjorden. AU Austfjordnes;* ← *BJ Bjørnnesholmen; KP Kapp Petermann; KR Krosspynten; RA Ræstadholmen.*

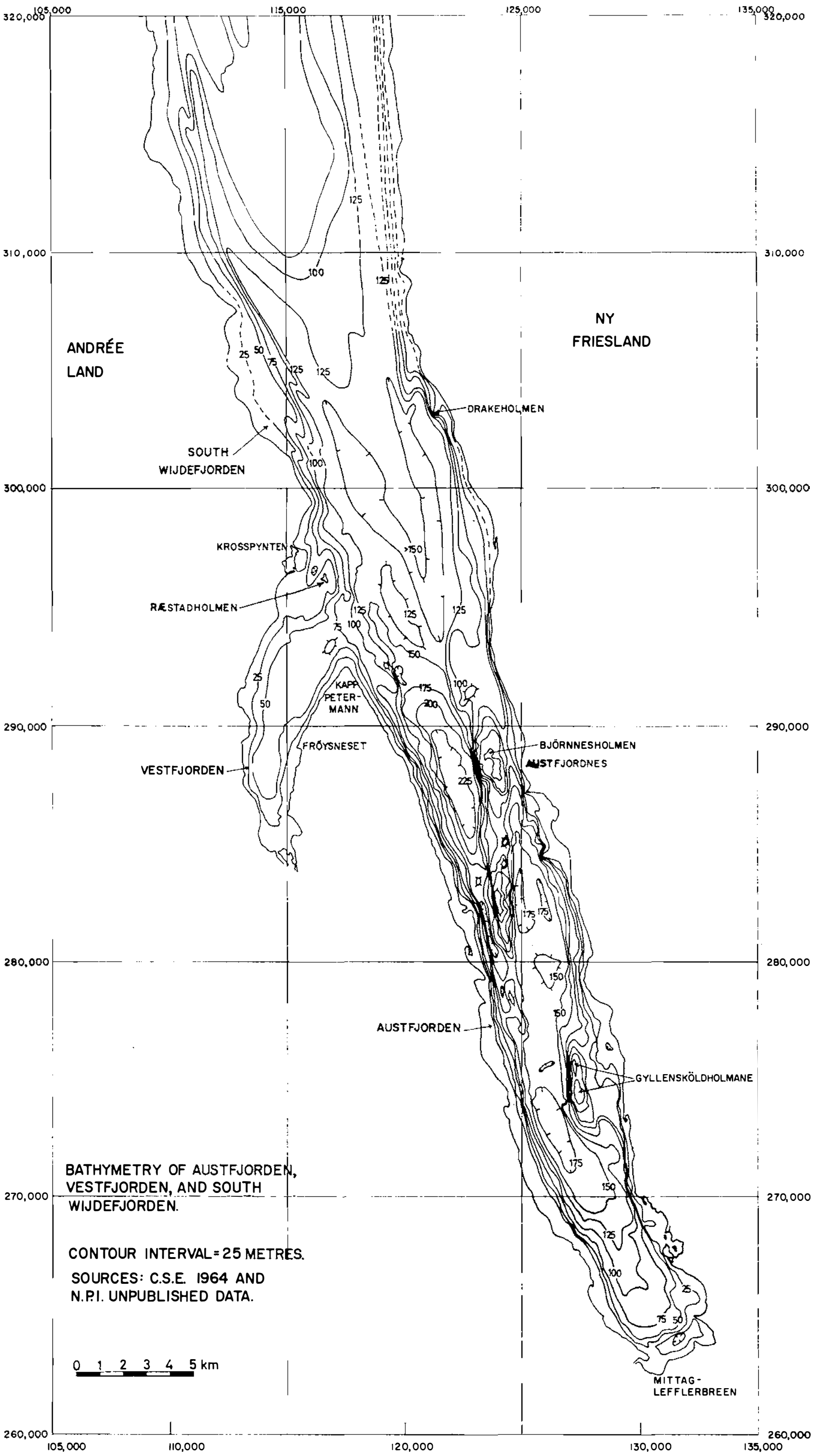
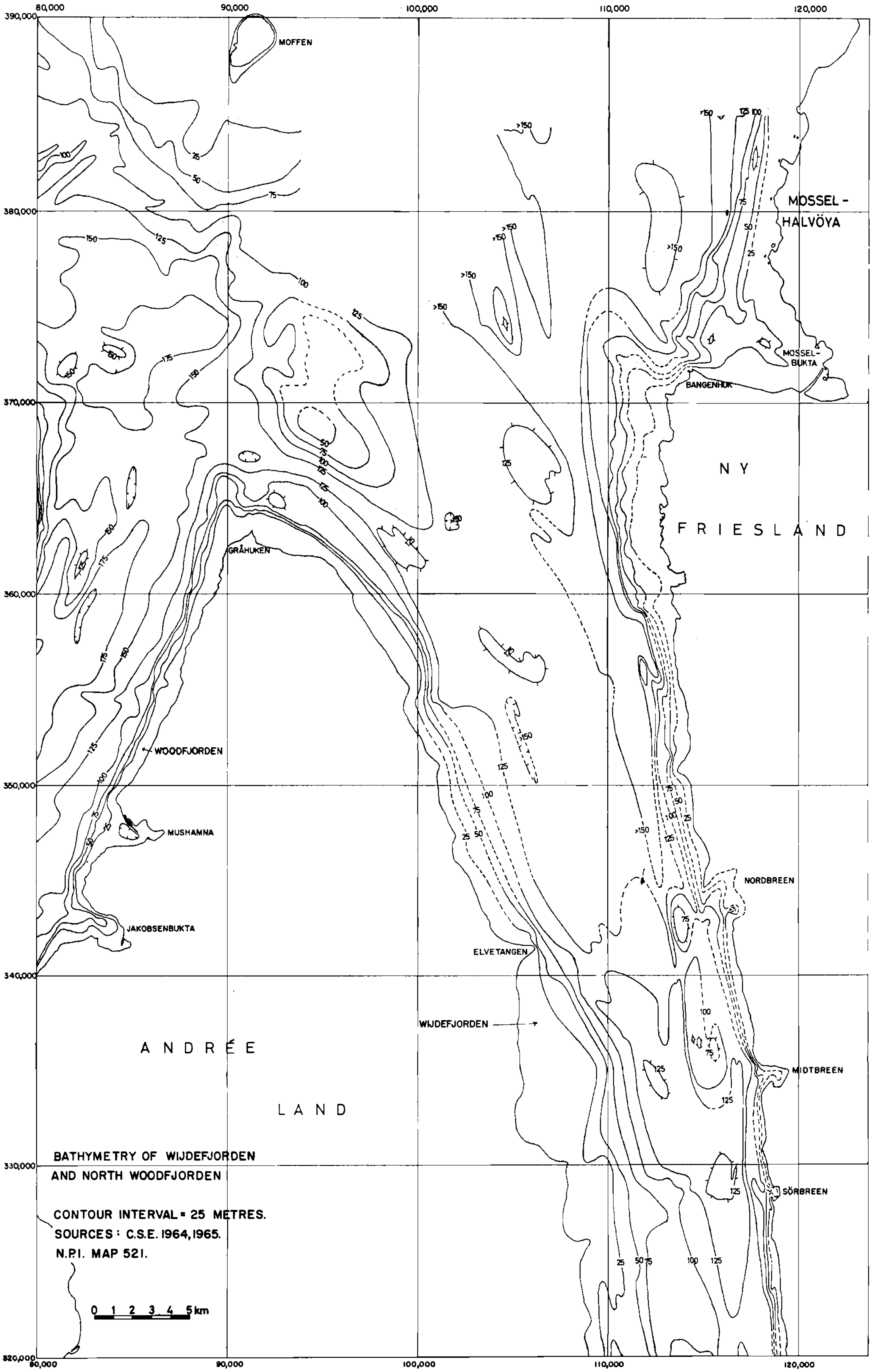


Fig. 11a. *Bathymetry of Austfjorden.*

Fig. 11b. *Bathymetry of Wijdefjorden.* →



BATHYMETRY OF WIJDEFJORDEN
AND NORTH WOODFJORDEN

CONTOUR INTERVAL = 25 METRES.
SOURCES: C.S.E. 1964, 1965.
N.P.I. MAP 521.

About Krosspynten and on Ræstadholmen are a group of narrow (1 m) monchiquite (lamprophyre) dykes trending N—S. These dykes cut the Kapp Kjeldsen division of the Wood Bay Formation (Lower Devonian). Radiometric age determination gives a Middle Carboniferous age of 309 ± 5 Ma (GAYER et al. 1966).

C. GRAVITY FIELD OF NORTH SPITSBERGEN

(Figs. 12–15; Table 4)

A gravity survey was undertaken in 1967 from motorboats at shore stations and along traverses up some valleys. The Bouguer anomalies over north Spitsbergen (Fig. 12) are expressed relative to an arbitrary zero at the Cambridge built hut at Biskayerhuken, which is situated near the western margin of the Devonian graben.

Since the anomaly amplitudes have been related to a local base which is just outside the margin of the anomaly causing mass, the Bouguer anomaly values observed will approximate to those actually caused by the said mass. That is to say, there will not be a large background or regional level included in the observed values. The two dominant features of this field are the E—W trend over Reinsdyrflya and Liefdefjorden, and the N—S trend adjacent to Wijdefjorden. Between these areas, over northern Andrée Land, the isogals trend north-east before swinging south into the Wijdefjorden trend. Lack of data north of Mosselbukta renders the contouring in this disturbed area somewhat speculative.

From zero at Biskayerhuken the Bouguer anomaly falls at approximately 1 mgal km^{-1} to the south: this gradient decreases to about 0.5 mgal km^{-1} in mid Woodfjorden. Minus 35 mgal is approximately the maximum amplitude that the anomaly attains, and in southern Woodfjorden it occurs quite close to the western margin of the Devonian graben.

In northern Andrée Land the —30 mgal isogal trends north-east to an area east of Ægirfjellet whence it turns southwards. This embayment in the Bouguer minimum coincides with the outcrop of the Wijde Bay Formation (FRIEND 1961), and hence presumably with the maximum present day thickness of the Devonian sediments. This coincidence may be relevant to the assessment of the regional and residual anomalies.

Around Mosselbukta a relatively steep gradient af 1.5 mgal km^{-1} decreases the anomaly eastwards from —8.4 mgal at Bangenhuken to —29.5 mgal at the head of Mosseldalen (Fig. 13). A less steep gradient of 0.9 mgal km^{-1} decreases the anomaly southwards from Bangenhuken. Lack of further easterly traverses from Wijdefjorden creates serious problems for any interpretation. However, it is probable that the easterly negative gradient observed in Mosseldalen persists to the south across Ny Friesland, reflecting the decrease in metamorphic grade and the upward progression through the Hecla Hoek sequence towards the east.

To the west of the Wijdefjorden fault, traverses in Forkdalen and Zeipeldalen show that the Bouguer anomaly decreases to levels between —26 and —34 mgal within the outcrop of the Devonian rocks.

The Bouguer anomaly profile between Bangenhuken and the head of Mossel-

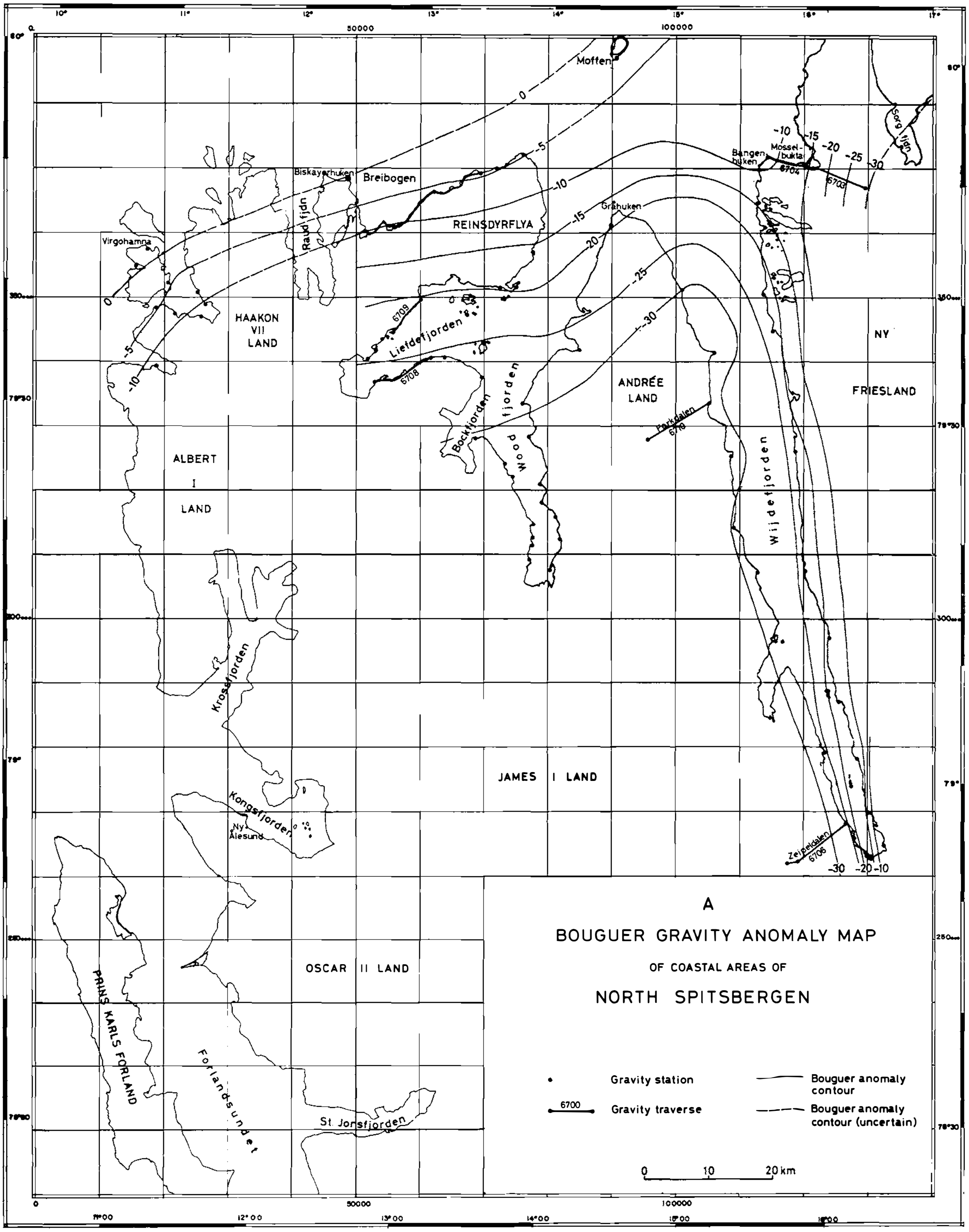


Fig. 12. *Bouguer gravity anomaly map of Austfjorden and Wijdefjorden.*

dalen is shown in Fig. 13. A gradient of 1 mgal km^{-1} decreases the anomaly eastwards from Bangenhuken and steepens to 3 mgal km^{-1} in Mosseldalen. This change of gradient occurs near the boundary between the Harkerbreen and Planetfjella Groups of rocks.

Figure 14 illustrates the Bouguer anomaly in Forkdalen. Short (c. 3 km) wavelength variations of amplitude c. 31.5 mgal occur near the head of the valley but give way to a general level of -27.5 ± 1 mgal for about 5 km. The

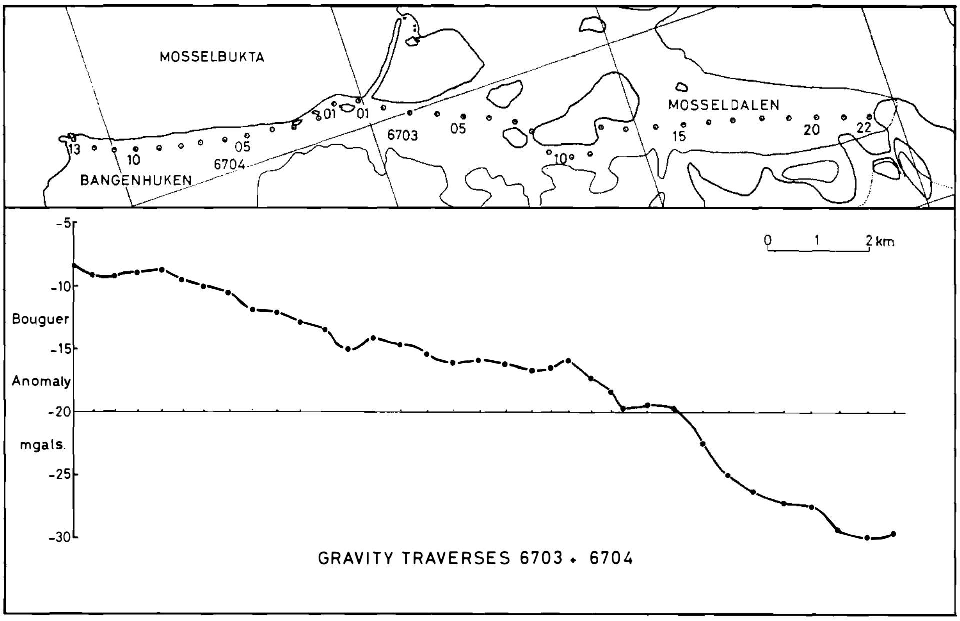


Fig. 13. *Bouguer anomaly profile between Bangenhuken and the head of Mosseldalen, showing the decrease in anomaly level as the Hecla Hoek succession is ascended.*

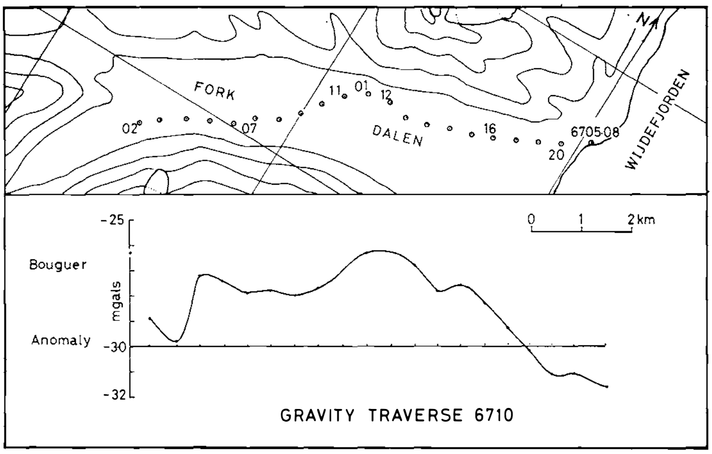


Fig. 14. *Bouguer anomaly profile along Forkdalen. Figures 14 and 15 indicate variations in the Bouguer anomaly within and towards the eastern limit of the Devonian outcrop in Andrée Land.*

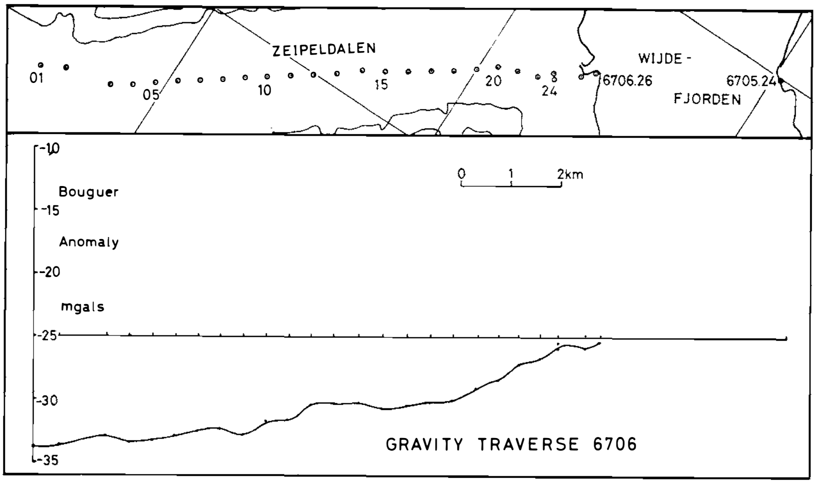


Fig. 15. *Bouguer anomaly profile along Zeipeldalen.*

level then decreases to about —32 mgal on the western shore of Wijdefjorden. Thus the positive gradient increasing the anomaly to the east of this latitude is restricted to within and east of Wijdefjorden.

The Bouguer anomaly in Zeipeldalen (Fig. 15) increases from —33.8 mgal at only 0.2 mgal km^{-1} from the head of the valley. The gradient increases to 2.0 mgal km^{-1} near stations 11 and 19. From the latter station the anomaly rises to —25.3 mgal on the western shore of Wijdefjorden, and then —10.9 mgal on the eastern side of the fjord. The gradient across the fjord is thus at least 4.3 mgal km^{-1}.

Since the anomaly field in the present study is incompletely defined a "semi-quantitative" approach to the interpretation is attempted by the use of limiting depth formulae (BOTT and SMITH 1958; SMITH 1959), taken together with a qualitative consideration of the geological implications of the observed anomalies.

The densities of Devonian and Hecla Hoek rocks have been measured by KURININ (1965), and by D. MASSON SMITH, and MATON. Table 4 summarises the values adopted for the following interpretation.

Limiting depth formulae may be applied to the Bouguer anomaly along Zeipeldalen and across Wijdefjorden (Fig. 15). The maximum depth so obtained for the Devonian basement is 4.1 km. It is assumed in this calculation that the anomaly at station 6705.24, in front of Stubendorffbreen, represents the maximum value at which the anomaly levels off to the east, neglecting any effects from negative density contrasts in that direction. If the anomaly at 6705.23 is projected north-north-west on to the profile of Fig. 15, the limiting depth becomes 4.2 km, which at least provides some measure of consistency.

This depth compares with a stratigraphic thickness of 1.5 km for the lower Wood Bay rocks which outcrop locally. This thickness may be doubled by folding: thus the gravity field suggests that a considerable thickness of other rocks could occur between the Wood Bay Formation and the Hecla Hoek surface west of the Wijdefjorden fault. This intervening formation could be a conglomeratic one analogous to the Siktefjellet and Red Bay groups in the west of the graben.

Table 4

North Spitsbergen rock densities

System	Group	Group Density gm.cm^{-3}	Weighted Formation density gm.cm^{-3}
Devonian	Wijde Bay	2.717	2.720
	Grey Hoek	2.730	
	Wood Bay	2.727	
	Red Bay	2.713	
	Siktefjellet	2.709	
Hecla Hoek	Upper and Middle Hecla Hoek	2.713	2.713 (Kurinin, 1965).
	Planetfjella	2.719	2.761
	Harkerbreen	2.761	
	Finnlandveggen	2.825	
	Metadolerites	3.094	

III. Regional gravity evidence for fault structure

In conclusion the Bouguer anomalies indicate that despite the small measured density contrast between Old Red Sandstone and Hecla Hoek rocks there is a sharp anomaly of about —25 mgals amplitude. This anomaly must be caused, at no great depth, by the above contrast along Wijdefjorden.

The eastern boundary fault appears to have a present day throw of at least 4 km in the south and between 4— 7 km in the north, although the structural complexity of the Hecla Hoek to the east of Wijdefjorden defies interpretation in terms of simple bodies.

Further south the anomaly is less sharp; that is, the gradients are less steep, and this may be attributed to the cover of Carboniferous and younger strata which extend east of the Balliolbreen Fault Zone and thin to the east. Here also the total amplitude of the anomaly increases, approaching —45 milligals. The known densities and stratigraphic thicknesses of the Carboniferous and younger strata are probably insufficient to generate an anomaly of this amplitude. It thus appears that the Old Red Sandstone graben may be assumed to extend south of its outcrop beneath the younger sedimentary basin of south Spitsbergen.

IV. Regional stratigraphic evidence for fault activity

A relatively complete sequence of Carboniferous to early Tertiary rocks outcrops across the centre and south of the Fault Zone. In the north, the Old Red Sandstone only occurs west of the Fault and the pre-Devonian (Hecla Hoek) rocks occupy most of the area to the east of the Fault line. Altogether along this zone there is evidence relating to most episodes in the geological history of Svalbard. In this section the stratigraphical evidence relating to the Billefjorden Fault Zone is reviewed in order of increasing age.

1. Recent activity

(Fig. 16)

The nearest major recent fault activity from seismic evidence lies on the Spitsbergen Fracture Zone in the ocean 200 km to the west of the Billefjorden Fault Zone (HORSFIELD and MATON 1970). However, three if not four earthquakes recorded by the World Wide Seismic Standard Network (WWSSN) in the epoch 1962—1969 lie along or not far from the Billefjorden Lineament. These are shown on Fig. 16 based on work prepared without this particular zone in mind (HORSFIELD and MATON).

In Northwest Spitsbergen, at Bockfjorden, there are well known Quaternary volcanoes and small present day hot springs along a NS line parallel to and 50 km to the west of the Billefjorden Zone. The volcanic rocks, with olivine nodules (GJELSVIK 1963), suggest a mantle origin. Moreover, the position of N.W. Spitsbergen in relation to the oceanic spreading ridges and the transform fault now evident in the Spitsbergen Fracture Zone suggest that, as the ocean between Spitsbergen and Greenland has spread, the hottest zone of the underlying mantle has moved relatively to the west. More properly Spitsbergen with the Barents Shelf and Europe has moved to the east with respect to the spreading zone.

Hot springs give evidence of high geothermal gradients related to volcanic activity and/or faulting related to the lithosphere movements suggested above. In this connexion it is not so well known that there is a hot spring in Tempelfjorden, about ½ km from the shore 2 km E.N.E. of Kapp Murdoch, which can now be located in the winter by a hole in the sea ice. The only published reference to this known to us is from the Scottish Spitsbergen Syndicate expeditions (ALLAN 1941). "These springs [S. of Bockfjorden and Woodfjorden] follow an important line of fault near Wood Bay and their continuation to the S.E. may be indicated by a curious basin at the inner end of Tempel Bay, where, despite proximity to a large glacier, abnormally high temperatures have been recorded together with absence of freezing when Ice Fjord was covered with winter ice, and a state of commotion in the water noted." There is no obvious fault in the

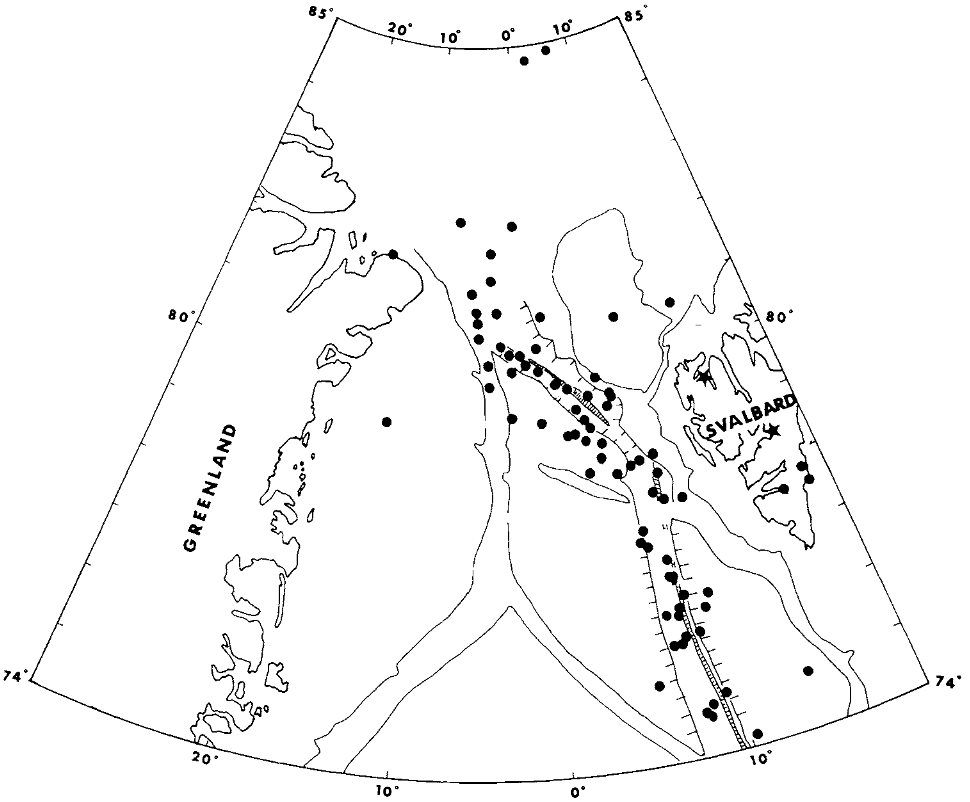


Fig. 16. *Present day tectonic activity around Svalbard.* ● *are World Wide Standardized Seismograph Network (WWSSN) earthquake epicentres (1962–1969) after* Horsfield & Maton *(1970).* ★ *are volcanic and/or hydrothermal centres.*

mountains to the north of this spring so that if it lies on a fault it could cut the Pre-Carboniferous basement and belong to the system of faults east of the Billefjorden Lineament, possibly running parallel to it and 18 km to the east. Therefore the whole of the corner of the Barents Shelf occupied by Spitsbergen may be regarded as still to some extent underlain by hotter mantle which cuts the corner in extending from the Norwegian sea to the Gakkel Ridge in the Eurasian Arctic basin.

2. Topographic surface

The dissected erosion surface inferred from the concordance of summit heights throughout Spitsbergen is not noticeably dislocated across the Fault Zone. This erosion surface truncates all the main deformation structures in Spitsbergen including those of post-Palaeocene or post-Eocene age. Its age before uplift and dissection was probably therefore later Tertiary.

The present fjord and valley patterns developed on this surface do not appear to have any relationship to underlying structures, with the noteworthy exception of Wijdefjorden and to a lesser extent Billefjorden, both of which run perpendicular to the apparent consequent slope thus displaying structural control by the Billefjorden Lineament.

3. Mid-Cenozoic deformation structures

It has long been known that the west coast of Spitsbergen exposes a Tertiary fold belt. More recently its structure has been elucidated in different segments e.g. in Hornsund by BIRKENMAYER (1960, 1972), in Brøggerhalvøya by CHALLINOR (1967), and in Oscar II Land by HARLAND and HORSFIELD (1974). This belt comprises along its eastern boundary what appears as a vertical monoclinal western limb of the broad central Spitsbergen basin. This is overturned, thrust and folded, and analysis of the structure further west shows that a complex orogenic belt extends right to the west coast and beyond. The West Sptisbergen Orogeny is a major tectonic event in the history of Svalbard. One limit to its age can be fixed as post-Van Mijenfjorden Group, i.e. probably post-Palaeocene. A later limit is more difficult and will not be argued here except to suggest from the probable age of the late orogenic Forlandsundet (Tertiary) graben, from ocean floor magnetic anomalies and from geomorphic evidence that it was largely accomplished within Eocene to Oligocene time.

The movements causing the West Spitsbergen Orogeny were the result of compression coupled with dextral strike slip which transported the Barents Shelf (with Spitsbergen) from a position north of Greenland (HARLAND 1965c). This was in effect an early zone of transpression (HARLAND 1972), an interpretation admirably elucidated by LOWELL (1972).

The Billefjorden Lineament lies well to the east of the eastern front of the West Spitsbergen Orogen. The Lineament runs approximately N—S to NNW—SSE. The Orogen front runs SE through Kongsfjorden (so hardly affecting the NW corner of Spitsbergen) and then swings round to a SSE course approximately parallel to the Lineament in the south. In central and southern Spitsbergen the distance between them is approximately 50 km.

Within the platform cover rocks for at least 100 km east of the orogen, minor folds and thrusts are evident and these are conspicuously concentrated within the Billefjorden Fault Zone and another fault zone still further to the east, running from Lomfjorden to Agardhbukta (Fig. 1). These structures generally show overthrusting towards the east but they are variable.

4. Palaeocene and Cretaceous strata

A. EARLY PALAEOGENE DEPOSITS

The central basin of Spitsbergen is occupied by the Van Mijenfjorden Group comprising six formations (1500 m with coal measures) whose age, long suspected as Palaeocene from molluscan evidence (RAVN 1929), has been confirmed as such from foraminifera (VONDERBANK 1970). However, the possibility that the uppermost strata may be Eocene cannot be excluded. Nevertheless this gives a probable older limit to the West Spitsbergen Orogeny since the youngest formation (Aspelintoppen) is deformed by it.

The Firkanten Formation rests unconformably on Lower Cretaceous strata overstepping different members of the Carolinefjellet Formation (Aptian-Albian), suggesting a regional warping with uplift increasing northwards. No evidence has been reported of any local discordance along the Billefjorden Fault Zone during this hiatus. The oldest Cenozoic formation sets a younger limit to the late Cretaceous movements.

Russian workers (LIVSHITS 1965) have claimed that the Fault Zone represents the eastern edge of the Tertiary sedimentary basin. We know of no evidence to support this view. Two or three outliers of Firkanten Formation occur just to the east of the Lineament west of Hyllingebreen and further south (FLOOD et al. 1971).

B. LOWER CRETACEOUS DEPOSITS

Because the Carolinefjellet Formation is truncated and overstepped by the overlying Firkanten Formation we are ignorant of movements that could have modified the thicknesses or facies of the upper members of the Formation. The three lower members are depicted by NAGY (1970 p. 78) as apparently identical in thickness on both sides of the Lineament in Kjellströmdalen (from sections measured at Trollstedet and Dalkjegla). Nevertheless NAGY shows an apparent increase in sand/shale ratio northwards and this coupled with an increasing rate of sedimentation in Albian time is perhaps the first suggestion of increasing mobility with uplift towards the north that was subsequently effected in Cenozoic time.

The underlying Helvetiafjellet Formation does not vary in any local way across the fault zone.

The Rurikfjellet Member of the Janusfjellet Formation, however, thins to a third of its usual thickness over a postulated submarine ridge along the Fault belt.

5. Cretaceous — Jurassic structures

It has been established that the (Jurassic-Cretaceous) Janusfjellet Formation is divisible into two units, the Rurikfjellet and Agardhfjellet Members, separated by an unconformity (PARKER 1967 and as redefined in this paper). This unconformity represents the period of time which includes the Jurassic-Cretaceous boundary and the intrusion of some, possibly all, of the late dolerite sills in Spitsbergen (TYRRELL and SANDFORD 1933; GAYER et al. 1966; PARKER 1966, 1967; SPALL 1968). On the other hand volcanic activity continued elsewhere in Svalbard into Barremian times (NATHORST 1910; HAGERMAN 1925; and reviewed PARKER 1967, and HARLAND 1973 a.). During this interval the earlier major anticlinal and fault structures of the southern area were formed along the Fault Zone. Later differential erosion of these structures took place along the Fault Zone to remove 540 m of Agardhfjellet Member and Kapp

Toscana Group down to the Mid-Triassic Botneheia Formation. The evidence in Flowerdalen (PARKER 1966) suggests that the dolerites were intruded prior to folding.

6. Jurassic to Permian strata

We have no evidence for differential movements during the period of Lower Jurassic erosion and possible emergence, but this would be difficult to detect because in the Fault belt the erosion associated with the Jurassic-Cretaceous movements cuts down into Triassic strata. The Triassic Sassendalen and Upper Permian Tempelfjorden Groups both extend across the Fault belt with no marked regional change in facies or thickness. However, some movement along the Fault belt is evident from a detailed study of thicknesses. During Middle Triassic time, movement with downthrow to the east is suggested by a sedimentary thickening of 55 m eastwards across the belt; and the lower part of the Tempelfjorden Group shows a rapid increase in thickness of 120 m across the belt, also eastwards.

7. Post-Permian structures north of Sassendalen

Structures cutting the base of the Permian in the central area involve displacements of about 1000 —1500 m and are thus much larger than the structures described from further south.

The down-west group of structures (Terrierfjellet Fault and Wordiekammen, Løvhovden and Gizehfjellet Flexures) are faced by a down-east group (Yggdrasilkampen, Cheopsfjellet, and Pyramiden Faults) which together produce an elongate basin about 40 km long with a maximum width of 20 km. These structures are associated with subsurface sliding and brecciation at the base of the Cadellfjellet member and are thus regional tensional features.

From the relative structural unity of the Permian-Triassic strata it would seem unlikely that these displacements have occurred before late Jurassic. Thus two periods of potential movement suggest themselves. Either the period of late Jurassic-Cretaceous folding and faulting, or a period of faulting after the mid-Tertiary compressive episode. We have no critical evidence to decide between these hypotheses.

Circumstantial evidence suggests a Tertiary age as follows: (I) a similar graben in southeast Ny Friesland cuts dolerite sills, (II) a mid to late Tertiary graben structure is known in Forlandsundet, (III) a suitable tectonic environment obtained for renewed uplift of Ny Friesland and graben formation with the uplift of the Tertiary peneplane in mid or late Tertiary time (HARLAND 1969a).

8. Late Carboniferous and Late to Middle Moscovian strata

During later Carboniferous time there was a local thickening from 30 to 300 m down-east across the Fault belt in the upper Minkinfjellet and the Cadellfjellet Members. We have no evidence for any faulting; limestone deposition dominates and almost no terrigenous clastics were deposited. Fusuline studies make it possible to detect further movements with slight erosion and transgression on the East Dickson Land axis to the west of the Fault belt during this period. After an early transgression in late Moscovian time further uplift on the axis occurred prior to the Orenburgian and again prior to the Permian, the depth of erosion being a matter of tens of metres in each case.

9. Early Moscovian and Bashkirian strata

There was considerable uplift and erosion along the narrow East Dickson Land axis. Over 600 m of the Billefjorden Group and an unknown amount of Old Red Sandstone were removed. In a basin east of the Balliolbreen Fault line rapid sedimentation occurred (GEE, HARLAND and McWHAE 1953; CUTBILL and CHALLINOR 1965; HOLLIDAY and CUTBILL 1972). On the west edge of this basin rocks total over 1200 m in thickness, but in any one place not more than 1000 m (Fig. 9). The basin thins very rapidly to the east so that on Minkinfjellet, 20 km east of the Balliolbreen Fault line, there are only 250 m of strata, and the Minkinfjellet Member has overstepped the Ebbadalen Formation.

Altogether this involves a differential movement across the Balliolbreen Fault line of over 1500 m down-east.

The rocks deposited at this time are now bounded on the west by the Cheopsfjellet Fault. Direct evidence indicates only two distinct periods of movement on this Fault during Middle Carboniferous time, one prior to the Minkinfjellet Member and one before the late Moscovian transgression on to the Dickson Land block.

The present distribution of sediments seems to result from continuous warping with sedimentation on the downthrow side, with the faults moving only at the specific times indicated, for there is little evidence of banking of sediments against the faults (HOLLIDAY and CUTBILL 1972). The result was a thick sequence of *Sabkha* deposits separating thick elastic sediments close to the downwarp from thin limestones further east (HOLLIDAY 1966, 1967 and 1968).

10. Namurian strata

The Svenbreen Formation was deposited in a subsiding basin apparently centred on the Fault belt. The unit thins from 200 m just east of the Balliolbreen Fault line to 40 m 20 km east; and less certainly to the west. During the deposition of the Hultberget Member, conglomerates started to appear from the west and these foreshadowed the structural sedimentation pattern of the Middle Carboniferous.

11. Viséan and Tournaisian strata

The basal beds of the Carboniferous sequence (Triungen Member of the Hörbyebreen Formation) appear to be derived from the east, but the overlying shales were probably deposited very widely with little variation and there is no evidence for movement during Tournaisian and early Viséan times. A large late Viséan movement occurred on a fault in the east part of the main belt, leaving the Hörbyebreen Formation preserved in a basin west of the Fault. This formation thins westwards below the Svenbreen Formation and, projecting beyond Triungen, would have vanished about 35 km west of the Fault. The north and south continuations of this basin are unknown. Similar structural relations do not occur again during Carboniferous and Permian time.

12. Pre-Gipsdalen Group faulting

A number of faults, e.g. Karnakfjellet, Triungen, Kinanderfjellet, which do not cut the Gipsdalen Group, are difficult to date because the base of the Gipsdalen Group varies from Lower Bashkirian to Upper Carboniferous. The effect of these movements was to preserve the Billefjorden Group in a basin east of a major fault. This fault was west of that delimiting the Hörbyebreen Formation so there is a strip in the centre of the Fault belt in which the Billefjorden Group is very thick (Fig. 8). Within the basin several minor structures were formed before the deposition of the Ebbadalen Formation, e.g. Ebbabreen Fault.

13. Tournaisian to Palaeocene sequence

The sequence of strata ranging in age from 345 Ma to 54 Ma, in spite of the variations across the Fault Zone as listed above, nevertheless represents a relatively continuous platform sequence approximately 6 km thick. Visible angular discordances within the succession are quite exceptional.

This sequence has very many similarities with that of Ellesmere Island, Axel Heiberg Island and other islands of the NE Queen Elizabeth Island. The similarities are entirely consistent with (and indeed require) the generally accepted hypothesis that the whole of Spitsbergen was during that time span placed close to N. Greenland and the Canadian Arctic.

The mid-Tertiary movements evidenced by the West Spitsbergen Orogeny above were primarily a dextral transcurrence (with orogenic transpression) that brought the Barents Shelf including Spitsbergen to its present position (HARLAND 1965c, LOWELL 1972).

14. Post-Svalbardian unconformity

Throughout Spitsbergen there is a very marked pre-Carboniferous erosion surface which is an unconformity with marked overstep and some overlap. The overlapping rocks range from Billefjorden Group (Tournaisian) to Upper Carboniferous, whilst the overstepped rocks range from Mimer Valley Group (Givetian) to Finnlandveggen Group of the Lower Hecla Hoek. However, we have no evidence for differential erosion over the belt. In general this seems to be a remarkably smooth surface, so we may regard it as recording post-Devonian deformation.

15. Svalbardian (Late Devonian) structures

VOGT (1928, 1938) gave the name Svalbardian to the conspicuous earth-movements of Late Devonian age. The precise age limits are post-uppermost Givetian (Mimer Valley Group) and pre-Tournaisian (Billefjorden Group).

In eastern Dickson Land, the belt of deformed Middle Old Red Sandstone rocks immediately west of the Balliolbreen Fault (Fig. 8) provides the only direct structural evidence of this period of deformation.

The trend of the fold axes (Fig. 8) has a mode which crosses the trace of the Balliolbreen Fault, as well as the trend of the Hecla Hoek basement, at an angle of about 20°. Some of the fold axes are strongly sinuous. After this folding had occurred, the Balliolbreen Fault fractured the Old Red Sandstone sediments, crossing the folds obliquely, and some folds were partially rotated. The establishment of the obliqueness of this superficial stress pattern strongly supports the hypothesis of sinistral strike slip at this period along this zone. Accepting this hypothesis, the minor faults in Mimerdalen would be interpreted as shear fractures conjugate to the Balliolbreen Fault, and the large Munindalen and Bulmanfjellet Faults would be oblique strike-slip faults.

Westward overturning or asymmetry of most of the folds, and local development of dip-slip slickensiding, provide evidence for the existence of a stress couple resulting in westward downthrow, parallel to the dip of the Balliolbreen Fault. A similar sense of dip-slip has already been recorded for the Balliolbreen Fault, with a magnitude of at least 2000 m. Thus eastern Dickson Land provides evidence for two components of Late Devonian movement along the Billefjorden Fault Zone: (I) sinistral strike-slip, (II) dip-slip with westward downthrow.

Eastern Andrée Land provides little information which adds to the understanding of the Billefjorden Fault Zone. Our estimate of the position of the Fault Zone in the northern sector would place it in Wijdefjorden, at least 6 km from the easternmost Old Red Sandstone exposure. In the Vestfjorden area a syncline is exposed, with a similar size and asymmetry to the structures exposed further south in Dickson Land.

16. Devonian sedimentation

The Early Devonian, and possibly Late Silurian, Siktefjellet Group and the Red Bay Group are the earliest sedimentary units found resting on the surface of regionally metamorphosed rocks (metamorphism dated at 400 ± 20 Ma GAYER et al. 1966). Both units are thick and include coarse conglomerates. They are separated by a major unconformity. These earliest, late-orogenic sediments are only exposed west of Woodfjorden.

Along the Lineament the earliest exposed Old Red Sandstone sediments belong to the Early Devonian Wood Bay Formation. Although little is known about regional variations in its thickness, the pattern of its sedimentation is better known than that of the younger Devonian units. There is no evidence for fault movement along the Lineament at any period during Devonian sedimentation.

Palaeocurrent data from the Wood Bay Formation (FRIEND and MOODY-STUART 1972) indicate a general flow direction northwards, parallel to the Lineament, except in the lowest (Kapp Kjeldsen) faunal division in easternmost Dickson Land, where the mean flow directions trend more westerly. Sandstones low in the Wood Bay Formation, in contrast with the sandstones of similar age in the west, are rich in orthoclase typical of metamorphic (e.g. Lower Hecla Hoek) rocks.

The mean grain-sizes for sections over the whole outcrop area indicate a consistent coarsening of sediment towards the Lineament. There is, however, a remarkable lack of coarse-grained detritus; apart from vertebrate fragments and intraformational clasts, the largest pebble recorded is 2 cm in diameter and clasts greater than 5 mm are very rare. The proximity of these relatively fine grained sediments to the Ny Friesland mountains which must at that time have been uplifted encouraged us, from different evidence, to postulate a different palaeo-position before substantial post- Wood Bay Formation sinistral transcurrent movement (HARLAND 1969b p. 829; FRIEND and MOODY-STUART 1972).

The youngest Devonian sediments of Dickson Land (Mimer Valley Group) include conglomerates which have been interpreted as the results of uplift and erosion associated with the Svalbardian movements.

17. The Old Red Sandstone graben

If late Devonian sinistral strike-slip movement along the Lineament be accepted this leads to an interpretation of the Old Red Sandstone graben as a late Devonian structure and not as an original sedimentary trough. On this hypothesis (HARLAND 1969b, FRIEND and MOODY-STUART 1972), the western side of the graben shows a normal deposition slope with extensive basal conglomerates, whereas the faulted eastern margin transected the fluvial plain which originally extended eastwards with northwards flowing rivers. The Lineament therefore does not represent the eastern margin of the original sedimentary apron.

The Ny Friesland mountains were at that time far to the south. How far to the south is a matter for conjecture. From sedimentary structures within the graben a minimum of 200 km is suggested (FRIEND and MOODY-STUART 1972). From wider regional stratigraphic comparisons an even larger figure (say 1000 km) has been suggested (HARLAND 1969b) though part of this may result from parallel movements on faults further West.

18. Ny Friesland Orogeny and the main Caledonian unconformity (Silurian)

The Ny Friesland Orogeny was the main Caledonian phase, compared with which the Svalbardian folding can be considered as a posthumous movement

(HARLAND 1959). It intensely deformed and metamorphosed the Lower Hecla Hoek rocks (Stubendorffbreen Supergroup) that border the Fault Zone to the east. The youngest fossiliferous rocks involved in this deformation are Llanvirn (FORTEY and BRUTON 1973). It seems clear that the main tectogenesis culminated in Silurian time, as is supported by radiometric determinations of the last metamorphic event of 419—436 Ma, and by the post-tectonic granite 385—406 Ma (GAYER et al. 1966). Structure and stratigraphy suggest that, in the west of Ny Friesland, 15 km or more of strata have been removed by erosion. The oldest overlying rocks are the Hörbyebreen Formation of Lower Carboniferous age.

No undoubted Old Red Sandstone is recorded from any locality in Svalbard east of the Billefjorden Fault Zone. A claim was made for Old Red Sandstone in Nordaustlandet (e.g. SANDFORD 1926 p. 637) but this rock was later re-interpreted (SANDFORD 1963 p. 17) as Sveanor tillite.

The style of deformation in Ny Friesland is marked by N—S lineation and in an extraordinarily constant strike, especially just to the east of the line of the Billefjorden Fault Zone. This implies E—W compression and N—S as well as vertical extension where the main tectonic culmination exposes the deepest rocks 5 km east of the inferred Fault Zone.

The N—S extension is demonstrated generally by boudinage, and in the only conglomerate known in the sequence, i.e the tilloids of the Harkerbreen Group (Rittervatnet Formation, HARLAND, WALLIS and GAYER 1966; and GAYER 1969). When only the boudinage was known to us the degree of extension beyond 50 to 100% was not clear and it was interpreted as axial extension due to extreme E—W compression (HARLAND 1959). This explanation had more than one shortcoming and when extension in the Rittervatnet Formation as well as in other rocks elsewhere in Svalbard was discovered it was clearly inadequate and an alternative explanation was proposed, namely: a shear zone resulting from transpression (HARLAND 1971).

This unexpected situation in which rocks low in the geosyncline were much attenuated (as well as metamorphosed and otherwise deformed), whereas rocks higher up in an apparently continuous stratigraphic sequence were neither metamorphosed nor intimately deformed, is explained by the fact that the older rocks are exposed along the shear zone to the west of Ny Friesland and immediately east of the Lineament whereas the younger rocks are only preserved further east where there is no evidence for such a shear zone.

19. Pre-Caledonian sedimentation

A. HECLA HOEK EAST OF THE LINEAMENT

Hecla Hoek rocks of Ny Friesland have indeed been described as an apparently continuous sequence of strata (HARLAND and WILSON 1956; HARLAND, WALLIS and GAYER 1966). A thickness of about 18 km begins with about 12 km of metamorphic rocks with a high volcanic content which are exposed along the line of the Billefjorden Lineament and a steep average dip to the E brings in successively younger rocks to the east through 4 km clastics (Veteranen Group), about 2 km carbonates (Akademikerbreen Group), 1 km Varangian tillites and shale (Polarisbreen Group) and 1.2 km largely carbonates Early Cambrian to Llanvirn (Oslobreen Group).

The Lower Hecla Hoek is a typical eugeosynclinal facies especially in its lower part where basic volcanics alternate often with acid (continental) detritus (HARLAND, WALLIS and GAYER 1966). This can best be explained as forming at a continental-oceanic margin. On this evidence of a linear belt of such facies it is argued here that the Billefjorden Lineament began at such a margin.

Whether the basic volcanics represent an initial divergent plate margin at the birth of an ocean is beyond the scope of this paper to discuss. Certainly, however, by the time of the Caledonian orogeny the locus of this geosynclinal belt had to some extent determined the orogenic pattern between convergent plates.

The Ny Friesland Hecla Hoek sequence is similar in many respects to the sequence in Central East Greenland. This similarity applies more particularly to the Middle and Upper Hecla Hoek both in terms of the general sequence, thicknesses and facies — though not in the detailed divisions of formations, members and beds. On this basis original close proximity is not suggested but it is difficult to imagine two such similar sequences developing except within the same geosyncline and subject to the same broad tectonic history. It is on this basis that a position for Ny Friesland in Hecla Hoek times south of rather than north of Peary Land has been proposed (e.g. HARLAND 1965c et seq.).

B. PRE-CALEDONIAN ROCKS WEST OF THE LINEAMENT

Pre-Devonian rocks to the west of the Lineament are separated by 50 km or more of later rocks and, while almost certainly they correspond in age with parts of the type Hecla Hoek sequence, they differ in several respects both from the Hecla Hoek of Ny Friesland and that between the different outcrops from N to S along the west coast.

In *Northwest Spitsbergen* only rocks of Lower Hecla Hoek type occur and radiometric ages indicate a possible thermal event apparently synchronous with stable Upper Hecla Hoek sedimentation. The sequence has been compared with the Hagenfjord rocks in N.E.Greenland, where they are said to be affected by late Precambrian Carolinidian diastrophism.

In *Central West Spitsbergen* more than 1 km of clastic sediments overlies up to 4 km of tillites of Varangian age and these are underlain by dolerites, limestones quartzites, and volcanics totalling 4—5 km. There is the possibility that this sequence may extend through late Ordovician, and a Silurian age for the youngest sediments cannot be ruled out. Moreover the facies are altogether more clastic, reflecting a mobile movement in early Palaeozoic time. Both facies and age range seem to contrast with the Ny Friesland sequence and so also with East Greenland sequence. The suggestion has been made (HARLAND 1972) that extensive sinistral transcurrence along the Billefjorden Fault Zone brought these two areas much closer together than they were originally, Ny Friesland being transported from a position well south of Peary Land to join with Central West Spitsbergen which could have been formed north of Peary Land.

C. PALAEO-POSITIONS

The minimum sinistral displacement of 200 km accepted by us could go some way to increasing the distance between the different sequences and so alleviate apparent anomalies, especially when the undoubted E—W compression is also restored.

A much larger displacement (e.g. 1000 km) along the Billefjorden Fault Zone would not only serve to separate these contrasting sequences but would also enable them each to be placed nearer the sequences in Greenland with which they compare most closely. On this hypothesis the Ny Friesland sequence would be related to that in Central East Greenland with which it compares most closely and the western sequences might be placed in one or more positions nearer north Greenland. There is general agreement that both east and west sequences, in approximately their present relationship within Spitsbergen, moved from a position off North Greenland during Cenozoic time by dextral transcurrent and transpressive movements localized 100 km and more to the west of the Billefjorden Fault Zone.

D. THE NAME "HECLA HOEK"

If the hypothesis of large horizontal displacement be taken seriously, either entirely along the Billefjorden Fault Zone or cumulatively along it and other faults, then it might be better for the time being to restrict the name Hecla Hoek to the sequence where it was originally defined in N.E. Svalbard. Then, until the identity and relationships of the older western rocks has been determined, they could be referred to as West Spitsbergen basement complex. In due course the older rocks of Spitsbergen may *either* all be established as Hecla Hoek *or* as belonging to two or more sequences originally far distant, of which one is the Hecla Hoek sequence (sensu stricto) east of the Billefjorden Fault Zone. By uncritically using the name Hecla Hoek to mean any rocks subject to Caledonian diastrophism or older, a valuable name may cease to have any distinctive meaning other than an association with Svalbard.

V. Tectonic synthesis

The attempt that follows is necessarily speculative. Qualifications and alternatives will be minimised in order to present a coherent story and because the limited nature of the evidence on which we base our story has already been made plain. Indeed, our account of the earliest history witnessed by currently exposed rocks is extremely tentative and is put forward to stimulate further research.

1. Late Precambrian and early Palaeozoic history

The oldest rocks deep in the Hecla Hoek geosyncline lay at least 18 km below fossiliferous Lower Cambrian and an age of 1000 Ma would not be unreasonable (HARLAND and GAYER 1972). We do not know on what basement the oldest Hecla Hoek rocks in the Billefjorden Fault Zone were formed. Some sedimentation suggests a predominantly N—S trough. Indeed it is not impossible that a N—S fracture system developed in the early stages of the eugosyncline if it were not already there. This would account for the substantial basic volcanic component in the two lowest groups (Finnlandveggen and Harkerbreen), together forming the lowest 7 km of the known geosynclinal succession throughout a length of more than 130 km. Throughout this and the overlying 5 km of the Planetfjella Group, sediments were probably derived from a neighbouring tectonic land with an abundant acid volcanic component. This may have been a land (?geanticline) undergoing subaerial erosion to the east. If so, the boundary of the Ny Friesland geosyncline might have approximated to the Billefjorden Fault Zone or a parallel structure forming a hinge between uplift and subsidence.

It is possible that the region to the west and north underwent an early diastrophism and metamorphic phase (the Carolinidian deformation) and so does not represent the relatively undisturbed succession of Ny Friesland. In this way the contrast also evident in East Greenland (between central and northern successions) would be paralleled.

In late Precambrian to Palaeozoic (Upper Hecla Hoek) time succesions in Ny Friesland, south Spitsbergen, Nordaustlandet, and central East Greenland are very similar and suggest tectonically stable conditions. It is not clear whether they extended to northwest Spitsbergen or northeast Greenland. At the same time sedimentation in Central West Spitsbergen certainly reflects a tectonically more mobile environment.

HARLAND (1960a) and HARLAND, WALLIS and GAYER (1966) have commented on the correlation of the Hecla Hoek successions of Hornsund and Ny Friesland, and the difficulty of correlating Ny Friesland and northwest Spitsbergen. These difficulties can to some extent be resolved (HARLAND 1965c) by some pre-Svalbardian (post-Hecla Hoek pre-Devonian) sinistral strike-slip displacement. Such a hypothesis of sinistral displacement could be argued by the contrast between the basement complex of central west Spitsbergen and the Hecla Hoek of Ny Friesland.

2. Caledonian history

A. THE NY FRIESLAND OROGENY

Given an already N—S trend to the margin of the western geanticline, the E—W compression of Late Ordovician to Silurian time accentuated the N—S lineation by folding. Moreover axial N—S extension that accompanied this compression or more probably transpression (HARLAND 1959, 1966, and 1971) was considerable. This must have involved a degree of strike-slip, at least locally along the margins of this line. The effect of the Orogeny was to accentuate to an extreme degree the N—S structural trend. If the Billefjorden Fault Zone had not been active earlier it must certainly have been determined by this Orogen as an axially placed strike-slip fault zone.

At some earlier stage during the diastrophism N—S basic dykes were formed probably along the same line and a later series of late dolerite dykes trending N—S were intruded. This has been explained in terms of local transtension in an overall transcurrent shear system (HARLAND 1971). Otherwise a most unlikely episode of extension between compression phases would need to be postulated.

At a late stage in the diastrophism, seen especially in the later Hecla Hoek rocks, i.e. in the superstructure, a number of oblique sinistral faults were developed — possibly a response to combined N—S extension and incipient sinistral shear stress. These could relate to the late orogenic phase of granite emplacement. Small scale fractures and retrogressive metamorphism observed along the Fault Zone could belong to this time.

Further implications of these points will be considered in a study of the Geology of Ny Friesland (in preparation by W. B. HARLAND and R. H. WALLIS).

B. DEVONIAN UPLIFT AND SEDIMENTATION

The lack of coarse sediments in the Wood Bay Formation first suggested that no structure existed at the surface along the line of the Billefjorden Fault Zone in Lower Devonian times, but rather that the Ny Friesland mountains were at that time contributing sediment to a broad plain from a position well to the south of their present situation east of the Old Red Sandstone graben (HARLAND 1969b).

Palaeocurrent and grain size distribution studies of the Wood Bay Formation (FRIEND and MOODY-STUART 1972) confirmed this hypothesis with a detailed model of rivers flowing north and carrying metamorphic detritus (of Lower Hecla Hoek type) from high ground to the south which probably occupied both sides of the Lineament. The mountainous source area, allowing for isostatic adjustment, probably uplifted 9 km relative to the area of sediment accumulation to the north.

C. SVALBARDIAN MOVEMENTS (LATE DEVONIAN)
(Fig. 17)

Towards the end of Middle Devonian times a differently orientated pattern of earth movements developed. A sinistral N—S shear in the basement may have led to folding *en echelon* of the Old Red Sandstone along the Billefjorden Fault Zone and these folds were later cut by the major transcurrent Balliolbreen Fault. In order to account for the evidence of strong E—W compression, folding, fracture, cleavage, thrusting, and dip slip slickensiding to the west of this zone, it has been argued that a compressive component developed (transpression), intensely but locally, on the fault where its sinuous shape brought this about (HARLAND 1969b).

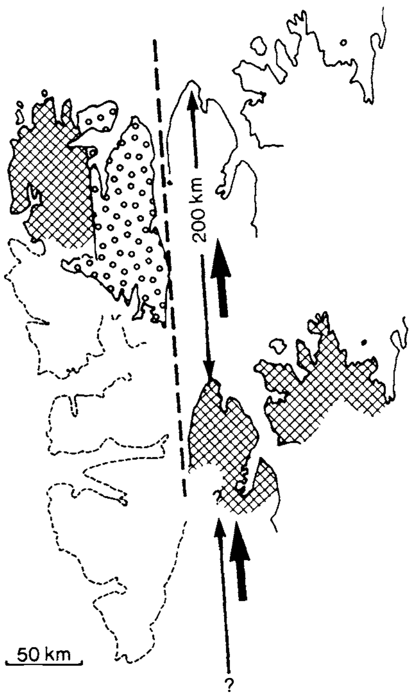


Fig. 17. *Diagram showing suggested major strike-slip movement of the Billefjorden Fault Zone. 200 Km is estimated minimum movement. Pre-Devonian rocks are cross-hatched; those to the east of the BFZ belong to the Hecla Hoek sequence. Devonian (Andrée Land Group) rocks are shown by regular circle ornament.*

The hypothesis of at least 200 km strike slip explains the contrast between the rocks on each side of the Fault as seen now. The fracturing and retrogressive metamorphism in the Hecla Hoek rocks could have occurred early in the movements when Ny Friesland was far south of Andrée Land. At that time the structures seen in Dickson Land were transmitted from the basement to the cover where Old Red Sandstone was deformed against itself. Then at a later stage the strike slip fault cut through the cover and juxtaposed Hecla Hoek from the south against Old Red Sandstone.

If a larger figure (e.g. approaching 1000 km) be considered for Svalbardian transcurrence it might all have been accomplished along the Billefjorden Fault Zone. On the other hand it might have been distributed amongst other faults as well so that a complex anastomosing shear zone between major plates operated (Harland 1972). It would be possible to allow a figure of 200 km or a little more for the Billefjorden Fault Zone and achieve the additional transport suggested by broader stratigraphic comparisons in a series of other faults, some exposed in West Spitsbergen and others now beneath the sea and rejuvenated in Tertiary time as the Spitsbergen Fracture Zone.

If a larger figure, e.g. around 1000 km, be considered for Svalbardian sinistral strike slip it does not alter the Andrée Land Group sedimentary interpretation. The immediate mountains to the south were Lower Hecla Hoek type — i.e. regionally metamorphosed of amphibolite facies but not the actual rocks of Ny Friesland which would have lain still further south.

D. INITIATION OF NY FRIESLAND BLOCK

This late Caledonian strike slip faulting initiated the western edge of the Ny Friesland Block, and this line then acted as a hinge to counter any imbalance between the Hecla Hoek, tectonized with late-tectonic granites to the east, and the Devonian strata to the west.

By Carboniferous time general isostatic balance between the segments of lithosphere to east and west of the Lineament is demonstrable stratigraphically. However along the Lineament itself subsidence continued with the formation of smaller Carboniferous basins. This could either be the result of slight continuing movement along the Fault Zone with net extension (transtension) or the effect of continuing uplift of the newly emplaced granites in Ny Friesland with transfer of asthenosphere from the peripheral zone.

3. Carboniferous to Palaeocene history

The span of time from the end of Devonian to Mid-Tertiary in Svalbard was one of relative tectonic inactivity between major orogenic-type movements of the Mid-Palaeozoic and Mid-Tertiary diastrophism. But just because for this time there is a relatively full stratigraphical record, it is possible to identify minor movements generating structures of platform type.

It is worth noting that Spitsbergen typifies the western Arctic in having no important late Palaeozoic orogeny. The Carboniferous instability essentially diminished till widespread stable conditions obtained in Permian time. In this respect there is a progressive change as late Palaeozoic movements became more important southwards towards the Armorican and Appalachian systems (cf. HARLAND 1961).

The similarity of the platform sequence from Carboniferous through Palaeocene strata with that of the Canadian Arctic islands has already been referred to. Indeed that sequence may be regarded as an extension of the Sverdrup Basin, initiating when the Svalbardian movements brought the parts of Spitsbergen together north of Greenland and breaking up in post-Palaeocene time when the whole of Svalbard moved thence to its present position by dextral transform faulting.

4. Mid-Cenozoic history

The central west and southwest coasts of Spitsbergen expose a N—S Tertiary orogenic belt and it is relevant here only to state that considerable folding and thrusting is evident, with a dextral transpressive component, and that the whole is cut by faults, some of which have been inferred to be major N—S faults with dextral strike-slip (HARLAND 1965c, CHALLINOR 1967). Such dextral movement is an essential element in the interpretation of the final opening of the Arctic, as first implied by TAYLOR (1910) and later with a different scheme by WEGENER (1915). This was considered in more detail in 1964 when one of us postulated a specific sequence of movements (Harland 1965c). Many other accounts have used such a dextral strike-slip and transpressive hypothesis (J. T. Wilson 1963; Harland 1966, 1967; Friend 1967; Lowell 1972).

To the east of this Tertiary orogen the platform area was warped to give the present synclinal structure of the south of Spitsbergen.

The Billefjorden Fault Zone crosses this platform parallel to and 50—100 km to the east of the Tertiary fold belt. Being a time of further tectonic activity, further movements took place along this Zone. To the north it is impossible to date much of the faulting as being other than post-Carboniferous or post-Permian but it is possible that much of the faulting there was a response to Tertiary stresses.

In the southern sector of the Fault Zone, predominantly vertical movements in the underlying rocks resulted in the formation of the synclinal structure in the Mesozoic and Tertiary cover. The superficial structures of the Fault belt in this area are almost certainly related to the major compression of the west coast belt, this being transmitted through the strata overlying the Janusfjellet Formation shales, which acted as a decollement horizon, and being localized along the Fault belt due to the thinning of the Janusfjellet Formation over the late Jurassic structure. It is not impossible that some of these folds and thrusts contain a small strike-slip component. If so this might well accommodate a little dextral displacement along this lineament when the main dextral transcurrence and transpression was operating further west. The only positive evidence for dextral strike-slip in this lineament is from Bjørnnesholmen in Wijdefjorden.

Some faulting in the central area that leads to reversed graben formation along the Billefjorden Fault Zone could match a similar graben formation in Eastern Ny Friesland along another N—S fault zone. It has been suggested that the Cenozoic heating of the lithosphere that caused ocean spreading in the North Atlantic and Arctic basins affected Spitsbergen in somewhat the same way, but expanded the lithosphere in the axial region of Ny Friesland causing further uplift and renewed compensatory subsidence along its periphery (Harland 1969a).

The date of some of these fault movements is post-Palaeocene. A younger limit to their age might be possible in the Forlandsundet area but the age of the late-tectonic Tertiary rocks there is not yet known with certainty.

A. DIFFERENTIAL UPLIFT TO THE NORTH

Superimposed on the above differential motions was a general uplift of northern Svalbard that was eroded to sea level and resulted in a mid-Cenozoic peneplain cutting down to Hecla Hoek and Old Red Sandstone rocks in the rocks in the north while preserving Palaeocene in the south. This uplift was also the result of mantle expansion along the zone of ocean spreading that cut across beneath the N.W. corner of the Barents Shelf (Harland 1969a).

5. Late Cenozoic history

A. RENEWED UPLIFT

The peneplain was in due course uplifted unequally as seen from the surface joining the summit heights (HARLAND 1969a). A broadly domed surface resulted and was dissected first by rivers and then by ice.

B. TRANSCURRENT FAULTING

By this time the whole Barents Shelf was behaving as a stable platform and the zone of mobility had transposed to the west of the Billefjorden Fault Zone even beyond where volcanity and thermal springs mark the Bockfjorden Fault Zone (Fig. 1). The inferred major dextral transcurrent movement with seismic effects now observable along the Spitsbergen Fracture Zone (HORSFIELD and MATON 1970), continues to separate this north-west corner of Europe (represented above sea level by the whole of Spitsbergen) from Greenland and America. It would thus seem that stresses generated in the mantle are by now relieved off the west coast of Spitsbergen and that the Billefjorden Fault Zone is for the time being almost dormant. But some seismic activity persists (Fig. 16).

VI. Conclusions

1. Summary of successive tectonic regimes

(1) Basic volcanism in early Hecla Hoek times suggests an early phase of crustal extension and/or high mantle heat flow during late Precambrian times at, say, 1000 Ma ± 200. This might relate to an early oceanic phase (Proto-Iapetus).

(2) Basic rocks alternate with, and are then superceded by, acid strata with a volcanic and possibly an arkosic contribution. This indicates at least proximity to a continental or island arc.

Both (1) and (2) show the area to have been a mobile zone characteristic of a lithosphere plate margin. The later Hecla Hoek rocks of miogeosynclinal facies to Mid-Ordovician age suggesting long slow subsidence are only preserved down dip, say 10—15 km to the east of the Lineament.

(3) Around Silurian time the main Caledonian deformation (Ny Friesland Orogeny) took place and at first resulted in E—W compression and crustal thickening. This could correspond to the final closure of the Iapetus Ocean.

(4) Normal compression was succeeded by N—S extension as evidenced by boudinage. This could correspond to a change of motion from compression to sinistral transpression along the axial zone of the Billefjorden Lineament which served to localize it.

(5) A late stage in the Ny Friesland orogeny showed local N—S basic dyke emplacement. This has been interpreted as local transtension in a dominantly transcurrent regime.

(6) During Early and Middle Devonian time no evidence of faulting is known but uplift to the south and west provided detritus for deposition with continued subsidence.

(7) Interruption in the sedimentary record between Mid Devonian (Givetian) and earliest Carboniferous (Tournaisian) corresponded to the Svalbardian movements. Folding and thrusting appear to be secondary transpression effects of dominantly sinistral transcurrence along the central and northern areas. This episode corresponded to substantial transcurrence of two major plates, with or without minor plate slivers between them (HARLAND 1972).

From internal evidence a movement of at least 200 km is probable. From external evidence 100 to 1000 or more km is possible. Such a large displacement might have been accommodated either all along the Billefjorden Lineament or have been distributed amongst more than one fault.

(8) No further strike slip is recorded from strata ranging in age from Tournaisian to Palaeocene that cover the Fault Zone in the southern part of Spitsbergen. Nevertheless vertical movements are recorded in these strata which by comparison with the preceding history were relatively minor. They do, however, indicate that the Lineament continued to exercise some control in the basement.

(9) Carboniferous sedimentation was closely controlled by vertical movements as, for instance, the Pyramiden conglomerates. At the same time (? Late Moscovian or early Bashkirian) N—S lamprophyre dykes indicate localised E—W extension.

(10) Relative stability with only widespread rather than localized disconformities continued through Permian to the end of Jurassic time.

(11) Basic intrusive and volcanic activity began in post part Volgian time and continued intermittently over a broad area of Svalbard into Barremian time. At the same time faulting and some folding took place along the Lineament.

Intrusion was in no way limited to the Lineament, nevertheless some considerable sills seem to have been controlled in their emplacement by the Fault Zone.

(12) Late Cretaceous upwarping was also general rather than localized as was Palaeocene subsidence.

(13) In Mid-Cenozoic time the main E—W compression and transpression of the West Spitsbergen Orogen, transmitted somewhat independently through cover and basement, appeared as minor folds and faults in the platform strata especially localized along the Lineament.

(14) Some extension or compensatory subsidence with renewed minor graben formation may have accompanied a general mantle expansion.

(15) Thereafter transcurrent motion transferred still further west to the Spitsbergen Fracture Zone and Spitsbergen itself behaved as a block with differential uplift, erosion, renewed uplift and dissection to the present.

2. History of horizontal motions

Substantial horizontal displacement between the major plates (Greenland on the west and the Barents Shelf on the east), whose junction is the Billefjorden Lineament, was active in pre-Carboniferous time but not later.

A. PRE-CARBONIFEROUS MOTIONS

We cannot be sure of a period of extension to accompany the early Hecla Hoek volcanics. E—W compression followed by sinistral E—W transpression, possibly followed by sinistral N—S transcurrence, was accomplished by the end of Silurian time. The magnitude of the motions concerned is difficult to estimate. The opposing plates were probably converging on and closing the Iapetus Ocean so that hundreds of kilometres or more were involved.

Renewed sinistral late Devonian transcurrence is estimated here as ranging from about 200 km to 1000 km.

B. POST-DEVONIAN MOTIONS

Until the West Spitsbergen Orogeny no horizontal motion exceeding a few tens of metres can be demonstrated. In mid-Cenozoic time the dextral motion which brought Svalbard 1000 km from N. of Greenland by a major transform motion was largely accomplished 100 and 200 km to the west of the Lineament, for example, in the West Spitsbergen Orogeny (transpressive) and Spitsbergen Fracture Zone (transcurrent). Small amounts (a few tens of metres in each case) of E—W motion is demonstrable in later faults and folds along the Lineament and only similar small magnitudes of strike-slip could be concealed there if it occurred at all.

3. History of vertical motions

(Fig. 18; Table 5)

Vertical motions are generally far less extensive and slower than horizontal motions but they are recorded in sedimentary history often with precision. At least the net subsidence is measured by stratal thickness and rough estimates can be made of the magnitude if not the details of net uplift. On this basis Figure 18 has been constructed to show the history of the Lineament. It is inevitably a composite figure generalizing and compiling data from along the length of the Fault Zone and so cannot be expected to depict accurately the motions at one locality. Moreover, the magnitude of the post-Devonian movements that are shown there was about one order less than the pre-Devonian movements.

Table 5

Relative magnitudes of vertical motions along the Billefjorden Lineament

figures have been simplified;

sources: Pre-Devonian: Harland & Gayer 1972
Devonian: Friend & Moody-Stuart 1972
Post-Devonian: Harland 1969

AGE approx in Ma	EVENT deposition of rock unit d erosion or elevation e		TOTAL VERTICAL MOTION in Km	AVERAGE RATE OF VERTICAL MOTION m/Ma	DIFFERENTIAL MOTION Km West	East
~1050	Stubendorff Supergroup	d	12+	60 –200	?down	
~850	Veteranen Group	d	3.8	76 –200	?down	
~800	Akademikerbreen Group	d	2.5	20	?0	
675	Polarisbreen Group	d	0.8	8	?0	
570	Oslobreen Group	d	1.2+	12.5	?0	
475	Ny Friesland Orogeny	e	18+	225+	up	18+ up
395	Old Red Sandstone	d	4	138	0	
359	Svalbardian movements	e	2	136	up	up
345	Billefjorden Group	d	0.4	14		0.22 down
317	Gipsdalen Group	d	1.5	25	0.6 up	1.0 down
255–8	Tempelfjorden Group	d	0.4	12.7		0.24 down
225	Sassendalen Group	d	0.75	13		0.05 down
167	Agardhfjellet Formation	d	0.25	12	0	
146	vulcanicity, uplift etc.	e	0.0 –0.5	0 –30		0.54 up
130	Rurikfjellet Formation	d	0.17	14	0	
118	Helvetiafjellet Formation	d	0.05–0.1	12.5	0	
112	Carolinefjellet Formation	d	0.27–0.77	22.5–64	0	
100	upwarping	e	0.3–1.5	8 –43	?	
65	Van Mijenfjorden Group	d	1.5	150	0	
~54	West Spitsbergen Orogeny	e	6.5–8	175	?up	?down
~20	upwarping	e	1 –1.5	62		
0						

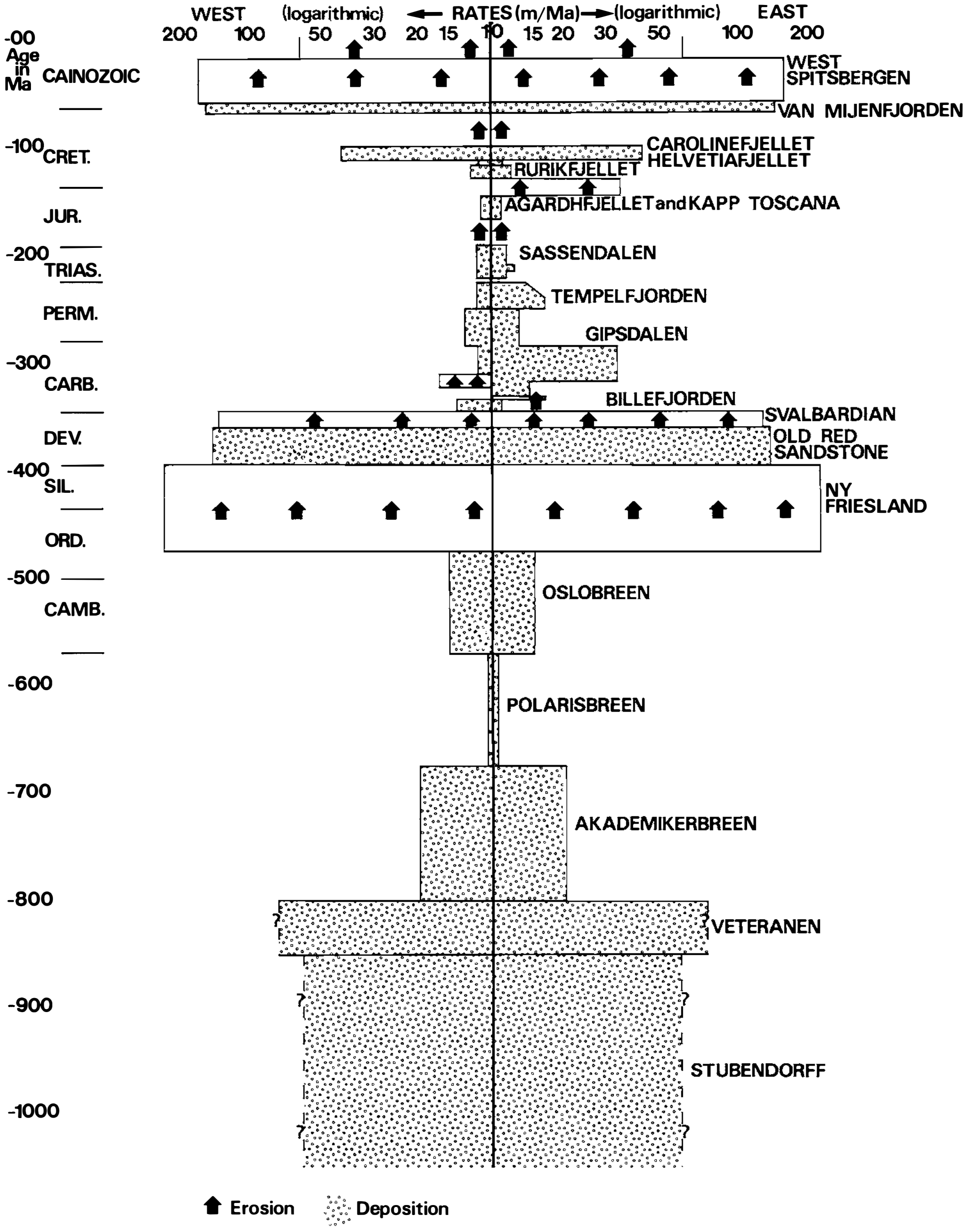


Fig. 18. *Rates of vertical movement through time along the Billefjorden Fault Zone.*

Table 5 calls for some explanation and comment. The figures calculated in the fourth and fifth columns are averages and even so may be in error by at least a factor of 2 for the following reasons. (1) A slight change in the geochronometric figures would affect the result significantly. In particular the first three figures are based on two calculations made by HARLAND and GAYER (1972) which were in turn based on estimates of rate of sedimentation. (2) Although thicknesses of strata can be measured accurately the thicknesses vary along the length of the Lineament and are generalized here. (3) The vertical motion is not directly derivable from thickness of sediments formed or removed but bears a minimum relation to it.

Therefore only the major differences in the average rate of inferred vertical motion are significant. Indeed we recognise two levels of activity distinguished in these figures by an order of magnitude. If, not averages, but actual rates of motion were known it seems probable that two or more orders of magnitude in rate of motion would separate the two rates which may simply be distinguished as high (orogenic) and low (platform). The high rates apply to Lower Hecla Hoek, Ny Friesland Orogeny, Old Red Sandstone, Palaeocene sedimentation and uplift. This last increase in mobility is already heralded in later Carolinefjellet (Albian) deposition.

The amount of differential movement along the Fault in late Cenozoic time, as evident in the southern area, increases from south to north. It is impossible to test whether such a contrast continues to the north because a much greater movement took place at about the same time with the effect of uplifting North Spitsbergen in relation to South Spitsbergen and so exposing Hecla Hoek and Old Red Sandstone in the north, where the Pre-Carboniferous peneplain is about 1 km above sea level, to the main sedimentary basin where it is at least 3 km below sea level. This has a profound effect on the appearance of the outcrop map of Svalbard and has been explained as upwarping resulting from the expansion of the mantle at the time of opening of the North Atlantic and Eurasian Arctic basin (HARLAND 1969a).

4. Billefjorden Lineament as recurrent mobile zone between lithosphere plates

In conclusion we have a plate margin with a known history of 500 Ma and probably with evidence going back about 1000 Ma. The latter part of the history is reasonably well documented in a detailed stratigraphic record. The larger earlier movements are not so clearly determined.

Initially the Lineament may have been the site of a Precambrian ocean margin, presumably one that originated in a divergent zone. With more evidence Caledonian tectonic phases can be plotted beginning with normal orogenic compression, possibly leading through a late phase of transpression with local axial elongation and finally with some transcurrent motion. This sinistral transcurrence (with local transpression and transtension) was resumed in late Devonian time.

Thereafter demonstrable strike-slip motion ceased and post Devonian history reflects a number of minor vertical motions in the detailed stratal record of a relatively stable platform sequence till at least Palaeocene time. Throughout this time the Billefjorden Fault Zone in the basement appeared to localise motion in the cover so that at times "posthumous" fault movements recurred in Carboniferous and late Jurassic — early Cretaceous time.

This effect became more pronounced when the West Spitsbergen Orogeny in Mid Cenozoic time affected the whole area again with a localization of folding and thrusting along this zone but the main compressive and transcurrent movements, dextral transpression (LOWELL 1972), were concentrated along the west coast of Svalbard about 100 km to the west. By later Cenozoic time a phase dominantly of dextral transcurrence had transferred still further west to the Spitsbergen Fracture Zone about 200 km west of the Billefjorden Lineament which, by now, appears to have become almost inactive.

Thus we have a history of early plate margins and much tectonic mobility becoming relatively but not absolutely immobile, characterized by a wide variety of minor platform deformation phenomena all localized above this major Fault Zone.

A major transcurrent fault zone of this magnitude might resurrect at any time if appropriate stresses were to operate across it again, because the contrasts in the lithosphere composition on either side will retain it as a zone of weakness and potential renewed fracture. The earthquakes need closer investigation.

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