



SKRIFTER NR. 167

The geological development of Svalbard during the Precambrian, Lower Palaeozoic, and Devonian

Symposium on Svalbard's geology
Oslo, 2–5 June, 1975



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A symposium on The geological development of Svalbard during the Precambrian, Lower Palaeozoic, and Devonian, took place 2–5 June 1975 at the Norwegian Academy of Sciences, Drammensveien 7B, Oslo, Norway. The permission to use these premises is greatly acknowledged.

Twenty-six scientists doing geological work in Svalbard and other Arctic regions took part in the Symposium, which also included a number of interested guests. The papers presented should be of interest to a much wider audience than the few scientists present; they are therefore published in this Norsk Polarinstitutt Skrifter Nr. 167. Although these papers appear at a very late date, their contents is still of great value, adding significantly to the existing knowledge of Svalbard.

I want to express my gratitude to the authors having waited patiently for a very long time to have their papers printed. Thanks are also extended to everybody having assisted with the arrangement of the Symposium and the preparatory work for this publication.

Rolfstangen, January 1979

THORE S. WINSNES
Chairman of the Symposium

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Introductory address by the director of Norsk Polarinstitut, TORE GJELSVIK, at the opening of the Symposium :

On behalf of Norsk Polarinstitut I wish you all a most cordial welcome to this Symposium. In particular I should like to greet our foreign guests from the Greenland Geological Survey, the Geological Survey of Canada, SEVMORGEO including the Arctic Geological Institute of the Soviet Union, and the U.S. Geological Survey. I thank you for your willingness to submit papers on the specific topics we have suggested. My gratitude is also extended to Dr. Anna Siedlecka from the Geological Survey of Norway, for her readiness to introduce us to the geology of the Varanger Peninsula, thus closing the geological Arctic Circle.

This meeting has two objectives. First, to enable geologists working in Svalbard in fields related to the chosen subject, to report on, discuss and try to solve problems of common interest. Secondly, to stimulate and strengthen the geological cooperation between the Arctic circumpolar nations.

I would also like to thank you for coming on such short notice. Invitations could not be sent out until our geologists were safely back from their Antarctic expedition in mid-February, and until we could be sure of a good Soviet-russian participation. The short notice prevented participation by some of those we had hoped to see. I would like to particularly mention Mr. W. B. Harland of Cambridge University who, together with his students and scientific colleagues, has given such important contributions to the geological research in Svalbard.

Finally, welcome to all those of you having come without a special invitation. Although the symposium is especially called for geologists having worked in Svalbard, we realize that science knows no national and institutional borders. We are happy, therefore, that you have wanted to join us.

For the reasons mentioned and because of economic limitations, it would be impossible for us to arrange a symposium for a big crowd; it could not be officially or widely advertised. I do think, however, that this symposium will prove that we are a group of just the right size to have a good working session.

Is there an early Precambrian granite-gneiss complex in northwestern Spitsbergen?

By M. G. RAVICH

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Abstract

Rocks of near-contact zones of granitoid intrusions in northwestern Spitsbergen which by previous investigators of this region have been assigned to the gneiss-magmatic complexes of regional metamorphism of the early Precambrian, are described. In connection with a detailed study of these rocks it has come out that they are represented mainly by biotitic hornfels and hornfelsified schists (with cordierite and sillimanite), having formed in the process of contact metamorphism from phyllites of the lower part of the Hecla Hoek Group. In this connection their affiliation to the granite-gneissose complex of early Precambrian age is made an object of doubt in the paper. The assumption is put forth that the rocks described are contactmetamorphic, having formed in connection with an intrusion of Caledonian granitoids.

Introduction

KRASIL'SČIKOV (1973) states: "... the pre-Riphean basement of the Caledonides outcrops in the northeast part of the archipelago, in western Ny Friesland, in southwestern (Hornsund) and in northwestern Spitsbergen. Early pre-Riphean formations are unknown in Svalbard. The oldest rock-type in the archipelago is probably granite-gneiss (of the deeply eroded outcrop of the basement) formed by the pre-Riphean ultrametamorphism of biotite and garnet-biotite paragneisses, relics of which are preserved as skialiths in granite bodies." (p. 84).

KRASIL'SČIKOV concluded this on the basis of traverses (scale 1:100,000) in the Krossfjorden area which had been carried out by ABAKUMOV in 1969 (ABAKUMOV 1976). He traversed the area bounded by 78°57' and 79°20' N, and 11°40' and 12°20' E, including exposures on the flanks of the Kollerbreen, Fjortende Julibreen, Conwaybreen, and Kronebreen, and on the shores of Krossfjorden and Kongsfjorden in north-western Spitsbergen.

ABAKUMOV described and mapped three formations at a scale of 1:100,000. The Signehamna and Generalfjella Formations, first described by GEE and HJELLE (1966), are of Riphean age (lower part of the Hecla Hoek). They consist of greenschist facies metamorphics with a total thickness of 8000 m and are common in Spitsbergen. The third formation, the Kollerbreen Formation, is at least 3500 m thick, and is assigned to the Lower-Middle Proterozoic (ABAKUMOV 1976). According to ABAKUMOV the Formation consists mainly of metasomatic migmatite and various gneisses (biotite, biotite-garnet, biotite-amphibole, biotite-sillimanite, biotite-cordierite, etc.). Numerous photographs and sketches show that the gneisses usually occur in metasomatic granite, sometimes as units 100–250 m thick, and contain skialiths and occasionally zenoliths. Massive granite (anatectite) is subordinate and occupies the central part of the Formation.

A narrow zone (2–4 km) of the Kollerbreen Formation trending approximately north–south, extends for more than 40 km in the eastern part of the area. This zone is associated with a fracture zone, and its contacts with other formations seem to be mainly tectonic though in places gradational contacts may be observed where other formations occur within the fracture zone.

In ABAKUMOV's opinion the Kollerbreen Formation formed under amphibolite facies regional metamorphism. The increase in metamorphic grade was accompanied by an increase in plagioclase feldspathization and plagiomigmatization under the amphibolite facies conditions, followed by ultrametamorphism. This resulted in partial melting, the formation of bodies of heterogeneous plagiogranite, and then homogenous plagiogranite and anatectite. When the anatectites attained a certain critical state, they were intruded into higher levels. The second stage was the potassium metasomatism and granitization and was mainly associated with earlier melts.

The Svalbard archipelago is situated in an area of Caledonian folding which influenced later tectonism (KRASIL'SČIKOV 1973). Caledonian structures usually involve both the Riphean and older basement rocks. Because of the close association between Riphean greenschists and the older crystalline basement rocks, in situ observations are necessary for interpretation in the Kongsfjorden and Krossfjorden area. Dating may be misleading; for example absolute age determinations of the so-called Lower Proterozoic basement gneiss yielded 420–440 m.y., indicating that they formed during the Caledonian orogeny (KRASIL'SČIKOV 1973).

The author, who had been studying the crystalline basements of ancient platforms for twenty years, visited the Kollerbreen Formation exposures from a field camp on the Krossfjorden coast with ABAKUMOV in July 1972. Three sequences were studied across strike in the upper reaches of the Kollerbreen,

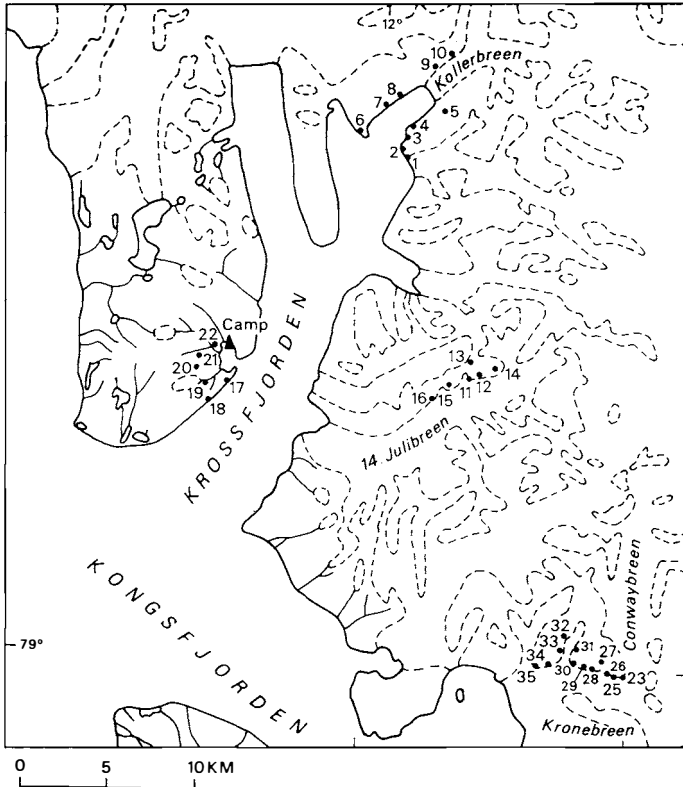


Fig. 1. Map of the study area in northwestern Spitsbergen.
•10: Numbers of outcrops in M. G. RAVICH's routes in 1972.

Fjortende Julibreen, and Kronebreen totalling about 25 km of outcrops of the Kollerbreen Formation. About 100 samples were collected. Petrographic and chemical study of these samples, as well as additional absolute age determinations, provided evidence for a new interpretation of the composition and origin of the Kollerbreen Formation.

In the field the "migmatites" and "gneisses" associated with granite bodies appear to be contact-metamorphosed, originating from the two-mica schists and phyllites of the Signehamna Formation. The so-called metasomatic granite and granite-gneiss that seem to predominate in the Kollerbreen Formation may have formed during injection of granitic melt; after crystallization the country rock shattered to a varying degree under stress resulting in poorly defined banding and quasi-gneissosity. Therefore the rock formerly called metasomatic granite and granite-gneiss is actually shattered plagiogranite and granite (granodiorite), while migmatite is actually schistose hornfels injected by aplitic and quartzose material. Gneiss is hornfels that retained the schistosity of the original phyllite and in places has a gneissose habit.

Along the glacier flanks there is a noticeable repeated alternation of granite bodies, concordant with contact metamorphosed phyllite, with units 150-

250 m thick of hornfels and quasi-migmatite. Discordant granitoid dome-like bodies (several hundred metres in size) were also observed. Contacts of the concordant granites and dome-like granitoid bodies with the contact metamorphosed phyllite are always sharp. But contacts of the two-mica schist and phyllite with the contact metamorphosed rock and quasi-migmatite (where the contacts have not been cut by faults) are always obscure and gradational.

The description of rocks of the Kollerbreen Formation will be given below.

Petrography

The following rock types occur in the area under study:

A. Country rocks

1. Two-mica schist with intercalations of chlorite-mica schist, quartzite and marble, making up the Signehamna Formation below.

2. Marble with intercalations of mica schist and quartzite constituting the Generalfjella Formation above.

B. Granitoids

3. Granite and plagiogranite approaching granodiorite.

C. Contact metamorphic rocks

4. Biotite hornfels and contact metamorphosed schist (in places with cordierite and sillimanite).

5. Contact calciphyre.

COUNTRY ROCKS

A two-mica schist is wide-spread throughout the area. Generally it appears rather uniform, though the percentages of the constituent minerals vary considerably: fine flaked muscovite — 15 to 50%, fine flaked biotite — 10 to 25%, quartz — 30 to 70%, and all the remaining minerals (almandine, spinel, plagioclase, tourmaline, hematite, sometimes apatite) — 3 to 6%. Although the latter form a minor component, they are very interesting and are of a quite different genesis.

The schist matrix consists of 0.03–0.1 mm slightly recrystallized quartz grains. These are interwoven into aggregates of light-brown biotite and colourless muscovite, which are strongly crimped and elongated along schistosity. The micas seem to have developed from the cement of the original siltstone. In the schist, up to 2–3% of fine fragments and skeletal grains of garnet (0.1 to 2.3 mm in size) are observed. Sometimes the garnets form piles of contiguous euhedral crystals as if cemented with biotite, and sericite flakes associating ore minerals. The garnet has an obvious clastic nature and is a typical terrigenous mineral bearing no relation to the paragenetic schist association. Much rarer rounded grains (0.1–0.2 mm) of green spinel occur in the same way. But bluish-greenish columnar crystals of tourmaline, up to 0.1–0.2 mm in length, scattered throughout the rock obviously have a neogenic habit, possibly due to neighbouring (at a distance of several hundred metres) granite bodies. Fine apatite crystals are

minerals of the same category. The genesis of the fine (0.1–0.2 mm) isometric lathes of plagioclase, sometimes up to 2–3%, is not clear. From their crystalline forms they also appear to be neogenic.

In some schists chlorite is almost as abundant as biotite. Other schist varieties usually resemble laminated quartzite due to the small amount of mica. Finally the third schist group consists mainly of carbonate and mica flakes, while quartz grains constitute only 33% of the rock. All of the schist varieties are cut by quartz veins varying from thread-like to 5–10 cm thick. The number of quartz veins increases towards the granite bodies.

Marble and quartzite predominate in the upper part of the sequence of Riphean metasediments. Marble is characterized by wonderful homogeneity and fine-grained structure. It consists of equidimensional 0.1–0.2 mm isometric grains of calcite, often with cross-shaped polysynthetic twins. By contrast, quartzite is impure and generally contains 70–90% of quartz, as well as albite (An_{5-6})¹, muscovite and biotite (5–10% in all) and rare calcite grains. Recrystallized and shattered quartz grains 0.05–0.1 mm in size of irregular shape, sometimes indented, make up the matrix. Irregular albite lathes are no more than 0.1–0.2 mm in size. Mica flakes, less than 0.05 mm in length, are uniformly scattered throughout the rock and do not form aggregates.

Rare schist intercalations in the quartzite differ from the typical two-mica schist described above. They contain far less quartz (no more than 30%), and micas make up more than half of the rock (up to 60%). Greenish-brown biotite flakes predominate over colourless muscovite flakes. Mica flakes reach 1 mm in length and they are mixed with chlorite, forming independent aggregates together with ore minerals.

In summary, the metasediments of the area are characterized by a simple association of minerals: quartz, albite, chlorite, muscovite and biotite, which are associated with the muscovite-chlorite and biotite-chlorite subfacies of greenschist facies of regional metamorphism.

GRANITOIDS

The metasedimentary rocks described above form the country rock around intruded granites associated with the narrow fracture zone crossing the whole area along its eastern edge.

In sections along the flanks of the three glaciers, heterogeneous granitoids, different for each glacier, are exposed. In exposures of Kollerbreen, plagiogranite similar to granodiorite in composition, is developed. It is practically devoid of orthoclase, and contains andesine, and relatively low percentages of quartz. In the rock along the Fjortende Julibreen, there are outcrops of leucocratic granite, with oligoclase and orthoclase present in almost equal quantities and biotite subordinate. And, last, in the Conwaybreen (Kronebreen) glacier rocks, granite is exposed with a predominance of oligoclase-andesine over orthoclase and with increased biotite content.

¹ Here and everywhere in the work the composition of plagioclase has been determined with the aid of the Fedorov universal stage.

Plagio-granite is usually represented by fresh medium to fine-grained slightly cataclastic rock. It has a hypidiomorphic granular texture, close to prismoid granular.

Zoned plagioclases are predominant (50–60%); prismatic lathes, 1 to 3 mm in size, are composed of andesine (An_{37-39}) in the centre and oligoclase (An_{25}) along the edges. 1–2 mm long biotite flakes with sinuate edges are evenly distributed composing 15–17% of the whole rock. Now and then biotite is replaced by chlorite. Quartz grains, 1–2 mm in size, fill interstitial spaces among other minerals. The quartz content is no more than 20–22%. Irregular blotchy blue and green tourmaline grains (1–2 mm in size) with sinuate edges, make up 2–3% of the rock.

Similar plagiogranites form relatively small homogenous bodies of a “spider” shape, surrounded by quasi-migmatite of different patterns formed by intrusion of plagiogranite into contact-metamorphosed mica schist. The plagiogranite here forms 10 to 50% of the migmatite. As a result several rock-types develop: (a) banded migmatite where plagiogranite neosome forms vague 1 cm to 1 m thick intercalations; (b) plagiogneiss where plagiogranite penetrates mica schists with still finer intercalation, enriching them primarily in andesine; (c) feldspathized schist where plagiogranite constitutes no more than 10% of the rock volume and forms very thin lenses of aplitic material. The composition of the intruded material is close to that of the larger plagiogranite bodies. For example the An content of the plagioclase does not decrease below An_{32-35} in both cases.

The Fjortende Julibreen rock-type is a typical leucocratic *granite*, frequently porphyritic and noticeably cataclastic. Medium- and fine-grained granite has a hypidiomorphic granular, slightly porphyritic texture, transformed by cataclasis, resulting even in partial recrystallization of the rock. This mainly affects quartz grains which form lenticular banded aligned aggregates producing a granulitic texture.

The granite is rather simple and constant in composition: almost equal amounts of oligoclase and orthoclase constitute together 65–70%, quartz 25–30%, and biotite in association with subordinate muscovite totals 3–5%. Along with weakly altered orthoclase lathes (1–3 mm in size), substantially saussuritized and sericitized, fine lathes 1–2 mm albite-oligoclase (An_{14-15} to An_{19-20}) make up the rock matrix. Between the feldspar lathes there are irregular quartz grains (1–1.5 mm) with sinuous outlines and strong undulatory extinction. A granitic porphyritic structure is caused by larger (5–6 mm) orthoclase lathes. Mafic minerals are exclusively represented by closely associated fine (1 mm) biotite and muscovite flakes. Muscovite is always developed after biotite and associated with appearance of fine-grained opaque aggregates along the cleavage planes of mica. Fine (0.5 mm) apatite crystals are uniformly distributed over the whole rock.

Cataclasis is manifested not only as partial recrystallization of granite, but with intense saussuritization and albitization of plagioclase, as albite (An_5) occur with oligoclase (An_{15}). Muscovitization of biotite belongs to the same category of secondary processes. The granitic texture changed during intense

cataclasis, often attaining a gneissose habit and so this rock-type may be formally called gneissose granite or even "metasomatic" granite.

Similar granite not only forms separate small massifs, but more often forms small (10–100 m in width) irregularly formed bodies with sinuous outlines elongated in one direction and oriented quite concordantly with their enclosing strongly contact-metamorphosed schists and quartzite. Examination of the rocky glacier flanks for a distance of more than 1 km reveals that no less than 60% of the outcrops is composed of granite while less than 40% consists of contact-metamorphosed rock.

These intrusions of granite indicate, firstly, the plastic state of the enclosing rock and, secondly, the high mobility and the rather low viscosity of the granite melt. The enclosing phyllite, schist, interbedded quartzite and marble are laterally "impregnated" with granite melt and therefore they are not only contact-metamorphosed, but transformed into a gneiss or migmatite-like rock. This situation arises in fracture zones at a relatively shallow erosion level when the formation of homogeneous granite bodies is prevented by the physical state of the country rock and accounts for the penetration of the fine intrusions of mobile granite melt into the country rock.

Granite in the upper Kronebreen and Conwaybreen differs from the two above-described granite types in exhibiting (a) cataclasis and gneissification; (b) increased deuteric alteration of plagioclase in particular; (c) formation of new sillimanite (fibrolite) and very small relict fragments of garnet; (d) considerable predominance of oligoclase over orthoclase; (e) a higher (for normal granite) content of biotite (sometimes chloritized) with zircon inclusions. This granite outcrops for a distance of more than 2 km; it is very homogeneous and contains comparatively rare skialiths of contact-metamorphosed schist as 100–150 m thick units, and hornfels xenoliths, 10–20 cm to 2 m in size. These occur within granite bodies as discontinuous chains constituting up to 10% of the granite volume.

A typical rock type is cataclastic and gneissified fine-grained, sometimes porphyritic, granite. Its texture is hypidiomorphic granular, broken down by cataclasis. A porphyritic texture is caused by 5–10 mm lathes of orthoclase and, rarely, plagioclase. The porphyritic phenocrysts cut the pseudo-gneissosity of the rock and have not been subject to cataclasis. The granite is typically composed of: oligoclase (An_{19-23}), 30–40%; orthoclase, 10–20%; quartz, 25–30%; biotite 10–20%; sillimanite (in places) 2–6%; locally detrital garnet — up to 1%; accessories, zircon and apatite and rare needles of rutile.

Cataclasis mainly affects quartz, 1–2 mm long flexuous-flattened grains of which have indented outlines and a strong undulatory extinction. 1–2 mm plagioclase lathes are intensely sericitized, but not broken down. Finer tabular orthoclase grains are materially kaolinized. The granite owes its gneissic appearance to aligned banded-chain-shaped aggregates of biotite flakes (up to 2 mm in length), including many rutile needles and small zircon grains with pleochroic haloes. Biotite flakes are sometimes decolourized, and often chloritized. Sillimanite forms separate banded aggregates of fine needles of fibrolite type, which cut the rock as 1–2 mm thick discontinuous lenses and

chains. Fine (0.1 mm) fragments of garnet grains, occur locally and sometimes form aggregations up to 1 mm in size. There is no doubt about the relict nature of the garnet, and the sillimanite aggregates appear to have originated from inclusions of mica schist.

Individual exposures are made up of massive and weakly cataclastic, medium- or fine-grained, slightly porphyritic biotite-granite. Biotite-granite forms more homogeneous intrusions devoid of gneissosity and xenoliths of country rock. These more massive granites differ little in composition from cataclastic and gneissified ones. For example, oligoclase (An_{20-21}) accounts for 40–50%; orthoclase, up to 20%; quartz, 30%; biotite, 8–12%. However, inclusions of sillimanite and garnet do not occur, and xenoliths of country rock are very rare.

As a whole, in spite of the differences in composition and structure of granitoid rocks in the area, they are the products of crystallization of granite melt which was capable under those geological conditions of active intrusion and caused partial contamination and contact metamorphism of the country rock.

CONTACT-METAMORPHIC ROCKS

During the crystallization of the granite melt, the adjacent Riphean phyllitic schist, quartzite and marble (the lower Hecla Hoek succession) were affected in various ways. As well as the thermal influence of the melt and the addition of volatic components from the melt to the country rock forming hornfels and contact-metamorphosed schist, fine injection of the melt into the country rock followed by crystallization occurred, resulting in the formation of a quite different migmatite-like rock. The contact processes were complicated by the fact that they occurred both in the exocontact zones and internally, in skialiths and xenoliths of country rock. As a result a rather complex group of rocks has formed, which, if studied without due attention, may be identified as gneiss, migmatite, metasomatic granite, etc., varying enormously in composition. As a consequence АВАКУМОВ (1976) described pseudo-ultrametamorphic rock of high-grade regional metamorphism instead of true contact-metamorphic rock.

A gradation from normal two-mica schist to contact-metamorphic rock was observed on the Kollerfjorden coast almost along its contact with the glacier. At a distance of 1.5–2 km from the granite body a standard two-mica schist is exposed. This contains no plagioclase, and consists of almost 50% fine (0.03–0.1 mm) weakly recrystallized silt-size grains of quartz, the other 50% being composed of biotite and muscovite aggregates with fine garnet fragments. The occurrence of fine tourmaline crystals is the only indication of the proximity of the granite body.

At approximately 1 km from the granite intrusion, the schist undergoes fundamental changes. It is essentially recrystallized, although the size of grains does not change and 0.1 mm isometric quartz grains still make up no less than half of the rock. The main characteristic feature is the appearance of tabular zoned plagioclase grains (0.2 mm in size), composing 20–25% of the rock. A reverse zoning, with andesine (An_{33}) cores and more basic rims (An_{40}) is frequent. The neogenic character of plagioclase, associated with

contact metamorphism, is obvious. This rock type is technically not a plagiogneiss. Another indication is the gradual disappearance of muscovite which is almost completely replaced by biotite (up to 30% of the rock volume) under contact metamorphism. Recrystallization of the rock is manifested, firstly, by the typical hornfels texture owing to the formation of more isometric granules of salic minerals, losing their detrital habit, and secondly, by a uniform distribution of 0.1–0.2 mm long well formed biotite flakes over the whole rock, instead of the former banded aggregates in the mica schist.

Further alteration of mica schist under contact metamorphism is shown in an increase of plagioclase content. Its An-content (An_{35-37}) is identical to that of the adjacent plagiogranite, and its volume increases to 40–50%. Fibrolite aggregates appear in amounts reaching 7–8%; muscovite, confined to 2–3%, gradually disappears; isolated cordierite grains and more abundant fine tourmaline crystals appear. The process of schist recrystallization becomes more active and, as a result, dimensions of the mineral grains especially quartz, increase up to 0.2–0.3 mm and they lose their detrital texture. The rock gradually loses its schistosity and becomes more compact resembling fine-grained plagiogneiss in appearance, although biotite flakes do not form banded aggregates, but are uniformly distributed.

Less than 200–300 m from the intrusion, the mica schist turns into a typical hornfels, with a matrix composed of equidimensional 0.2–0.4 mm isometric quartz granules and tabular grains of andesine (An_{30-32}) totalling 70–75%. The content of biotites become larger (up to 1 mm in length), bright-brown and aligned; the indistinct schistosity of hornfels is a characteristic feature. Fine (0.2–0.4 mm) granules of cordierite and thin aggregates of fibrolite (up to 10%) increase steadily. Apatite, tourmaline and ore minerals often total 2–3%. Recrystallization of the rock manifests not only in increase of grain size, but the rock loses its heterogeneous lepido- and granoblastic texture, and attains a typical hornfels texture. The relict schistosity is not pronounced, typical gneissification is absent and hornfels is represented by a rather dense, uniform rock, very different from gneiss, although it was formerly associated with plagiogneiss in composition. The gradation from two-mica phyllite-like schist to hornfels, observed over many hundreds of metres, rule out its identification as plagiogneiss of amphibolite-facies regional metamorphism (АВАКУМОВ 1976).

The gradation from the phyllitic schists of the Signehamna Formation to contact hornfels applies for the whole area and not just in places where hornfels is faulted against the country rock. Some tens of thin sections of the contact rock have been studied to verify that it is not regionally metamorphosed. The dense, rather uniform contact rock is as a rule extremely fine-grained, slightly cataclastic and resembles gneiss slightly, especially when it is not impregnated with granite veins; but even this is closer to migmatite and its nebulitic varieties.

In the contact zone varieties prevail with relict schistosity largely owing to aligned biotites. Biotite flakes, sometimes with torn rims, are variable in size — from 0.1 to 0.8 mm long; and the nearer the rock to granitoid, the lighter the

colour of the biotite. Muscovite flakes are preserved only in samples taken at a distance from the contact, and gradually disappear towards the contact. The contact rock is enriched in cordierite and fibrolite, and they are almost constantly present in the so-called skialiths and xenoliths of the former mica schist.

Contact-metamorphosed mica schist is characterized by a rather constant composition: plagioclase — 30–40%; quartz — up to 40%; biotite — 15–25%; the remaining minerals (cordierite, fibrolite, garnet) account for 2–10%. The plagioclase composition of the contact metamorphosed schist is not constant and depends on that of the adjacent granitoid. For example, in the Kollerbreen contact-metamorphic rock, plagioclase is represented only by andesine varying from An_{30} to An_{40} ; in the Fjortende Julibreen rock, plagioclase belongs to acid oligoclase (An_{14-18}); and in the Kronebreen contact rock it is represented by typical oligoclase (An_{22-28}). The second peculiarity of the contact metamorphosed schist is the presence of fragments of relict garnet, which never form good crystals; temperatures of contact metamorphism were probably not high enough for recrystallization of terrigenous almandine garnet.

These salient features confirm that (a) contact-metamorphic rock has formed from mica schist and therefore it retains relict garnet, and (b) processes of contact metamorphism are closely associated with the intrusion of granite melt, and therefore the composition of the plagioclase is related to that of the adjacent granite intrusion.

Concluding the description of contact-metamorphic schist it is necessary to emphasize that, where cataclasis is more developed, biotite is partially replaced by chlorite with formation of thin ore aggregates; plagioclase is essentially sericitized; and quartz grains have strong undulatory extinction and tend to form flattened and thin lenticular aggregates of granulitic texture.

Typical hornfels occurs mainly as skialiths within the granitoid. It is characterized by the appearance of orthoclase, when skialiths occur in two-feldspar granite; but when included in plagiogranite, hornfels is devoid of orthoclase. Orthoclase-bearing hornfels does not contain cordierite. Generally this rather uniform rock is represented by biotite hornfels of the following composition: plagioclase 30–50%; orthoclase 0 to 25% (content of plagioclase decreases respectively); quartz 30–35%; biotite 15–20%; accessories include rare crystals of tourmaline and ore minerals. Isolated occurrences of hornfels retain fine (0.05 mm) fragments of garnet, with chlorite rims. The hornfels is a very dense fine-grained rock, and isometric grains do not exceed 0.2–0.3 mm in size. Quartz has a strong undulatory extinction and plagioclase is sericitized. Plagioclase composition in hornfels depends on that of the enclosing granitoid as was described for contact metamorphic schist.

In highly cataclastic hornfels varieties, intense sericitization of plagioclase and chloritization of biotite has taken place. Although 0.2–0.4 mm long bright-brown biotite flakes are elongate, they are uniformly spaced in the rock. The rock texture is typical hornfels (pavement) because of the equidimensional isometric grains of adjacent silic minerals. Only orthoclase

segregations, larger and more irregular than the main granular matrix, look neogenic.

The small (up to 1 m in size) xenoliths in the granite vary in composition. Common biotite hornfels, sometimes with orthoclase inclusions, predominates. Another rock-type containing hornblende, garnet, and sphene, occurs in xenoliths. This is rather unusual in composition: sericitized andesine (An_{34-35}) 30%; quartz 15%; orthoclase 15%; green hornblende 17%; garnet 6%; sphene 3%; biotite 7%; ore minerals 4%. Its texture is heteroblastic, locally idioblastic or granoblastic. Only segregated orthoclase and brown well formed biotite flakes may be neogenic. All other minerals, though recrystallized, may have relict nature; this especially is true for 1 mm skeletal grains of garnet and ore minerals. This rock-type may be derived from a basic rock, accounting for the occurrence of sphene, garnet and amphibole along with rather basic plagioclase. Amphibole rarely occurs in the contact rock in this area, or in the mica schists of the Signehamna Formation.

Individual interlayers of *quartzite* within mica schist were also subject to contact metamorphism, especially if they occur in skialiths. This hornfels is distinguished by the abundance of quartz (60–75%) in fine (0.1–0.2 mm) isometric grains, resulting in a pavement texture. A second feature is a rather acid persistence of fine grains of cordierite and bundles of fibrolite (altogether 15–20%). Another feature is a leucocratic composition, biotite flakes constituting less than 5%.

Marble interlayers undergo pronounced contact metamorphism, especially when they are included in skialiths. In the contact calciphyre the percentage of preserved carbonate reaches 20–50%, giving way, first of all, to diopside (30–60%) and then to wollastonite (6–8%), and amphibole (3–10%). In the calciphyre, sphene is always present, but in small amounts (1–2%). Diopside forms 1–2 mm prismatic crystals, constituting the matrix. Wollastonite aggregates and more irregular prismatic grains of amphibole are included in these rocks. Isometric granules of calcite (0.3–0.6 mm in size) fill the interstices. Fine (0.1–0.2 mm) sphene crystals occur throughout.

Thin veins of aplite and thicker veins of pegmatite cut both granite and country rock. Migmatite-like rock occurs in skialiths and in exocontacts of granite bodies.

Palaeosome of pseudo-migmatite is contact-metamorphic mica schist, while neosome is represented by plagiopelite or leucocratic normal granite, and sometimes even by conformable quartz veinlets.

Extremely finely-layered migmatite, taken for gneiss in the field, outcrops along the Kollerbreen. Study in thin sections shows that 5–6 mm thick layers of palaeosome consist of fine-grained cordierite-biotite-hornfels (with high content of biotite up to 30% and equal quantities of fine 0.2–0.3 mm granules of cordierite and quartz — 10–15% each), while neosome consists of indented 2–3 mm grains of quartz 80% and sericitized lathes of andesine (An_{32-33}) — 20%. Generally, the composition of the neosome is identical to that of the granite, in which the skialiths occur.

Study of composition of numerous skialiths, granite bodies, and the country

rock reveals that contact-metamorphics of the anthophyllite-cordierite subfacies of the amphibolite facies are distributed in the study area within the narrow subfracture zone. Four paragenetic mineral associations of this subfacies occur.

- (1) muscovite-biotite-plagioclase-orthoclase-quartz;
- (2) biotite-amphibole-plagioclase-orthoclase-quartz;
- (3) diopside-amphibole-wollastonite-quartz;
- (4) muscovite-biotite-cordierite-plagioclase-quartz.

The last association is most widespread, while the second is least common. Association of diopside with wollastonite and amphibole is only typical of contact-metamorphic marble. Almandine is not inherent in the anthophyllite-cordierite subfacies, where it is generally replaced by cordierite. Sillimanite is also not characteristic of this subfacies, but its low-temperature variety, fibrolite, can occur when the original rock was oversaturated with alumina.

The anthophyllite-cordierite subfacies of the amphibolite facies is typical of contact-metamorphic rocks, while all other subfacies of amphibolite facies, staurolite-kyanite, sillimanite-almandine and almandine-diopside-hornblende are characteristic of regional-metamorphic rock. These three subfacies do not occur in the study area. Therefore, these metamorphic rocks are not part of a granite-gneiss or gneiss-migmatite complex of ultrametamorphic genesis, as was formerly put forward by KRASIL'SČIKOV (1973).

Chemical composition

The chemical composition of the three main granitoid types of the area, Table 1, confirms the above conclusions made in the petrography.

Granitoid from outcrops along the Conwaybreen (sample 23-v) are represented by a normal two-feldspar granite, with a composition practically unaffected by cataclasis and gneissification. The leucocratic nature of granite from the Fjortende Julibreen outcrops (sample 14-ž) is reflected by the low percentage of mafic components (especially FeO and MgO) and the high content of alkalis. Na₂O and K₂O constitute up to 8%, closely approaching the alkali content in granosyenite. Plagiogranite from outcrops along the Kollerbreen (sample 10), approaches the composition of granodiorite, differing greatly from the two rocks described above. The SiO₂ content is low at 64.7%. The FeO and MgO contents are twice as high compared with granite; and there is a low content of K₂O, only 2.2%, reduced from approximately 4% in a normal granite.

It is interesting to compare the chemical compositions (Table 1) of phyllite surrounding granites of the lower part of the Hecla Hoek succession with hornfels xenoliths formed within granite. The influence of contact metamorphism on phyllite of the lower Signehamna Formation in the Kollerfjord area is especially clear. At 1.5 km from the plagiogranite body of the Kollerbreen, two-mica schist (sample 3) almost unaffected by contact-metamorphism is exposed; it consists of almost equal amounts of aggregates of sericite, fine

Table 1
Chemical compositions of rocks of the North-Western part of Spitsbergen.

Ord. Nos.	Sample Nos.	Name of rock	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	CaO	MgO	MnO	K ₂ O	Na ₂ O	P ₂ O ₅	H ₂ O	Total
			Granitoid												
1.	10	Plagiogranite (transitional to granodiorite)	64.70	0.68	16.67	0.07	4.83	3.81	2.19	0.09	2.20	3.55	0.27	0.86	99.92
2.	14-ž	Leucocratic granite	72.35	0.16	14.27	0.12	1.92	1.23	1.02	0.04	4.88	3.38	0.09	0.63	100.09
3.	23-v	Gneissified granite	73.19	0.36	13.20	—	3.10	1.20	1.63	0.04	3.78	2.65	0.11	0.77	100.03
			Country rock (phyllite)												
4.	3	Two-mica schist (phyllite)	65.21	0.90	16.75	0.33	4.92	0.70	2.66	0.08	4.63	1.70	0.08	2.01	99.97
5.	16	Mica schist (meta-argillite)	67.25	0.78	15.11	0.35	4.09	1.01	2.11	0.09	4.34	2.61	0.12	1.84	99.70
			Contact metamorphosed rock												
6.	5-a	Contact — metamorphic schist	69.57	0.74	13.72	0.52	4.17	1.14	2.19	0.06	3.64	2.22	0.09	1.68	99.74
7.	31	Contact — metamorphic schist	67.09	0.80	14.97	0.27	4.97	1.55	2.50	0.07	2.45	4.00	0.15	1.02	99.84
8.	9	Schistose hornfels	69.48	0.63	13.38	—	5.55	2.16	2.05	0.09	2.90	2.54	0.11	0.98	99.87
9.	14-v	Schistose hornfels	66.77	0.82	14.70	0.10	5.71	2.34	2.67	0.08	2.60	3.00	0.12	1.22	100.13
10.	28-b	Schistose hornfels	67.13	0.74	14.28	0.24	6.35	1.79	2.42	0.19	2.59	2.71	0.09	1.65	100.18
			Hornfels xenoliths in granite												
11.	32-v	Hornfels in granite	72.29	0.66	11.69	0.69	4.85	0.28	2.68	0.08	2.53	2.19	0.12	1.84	99.90
12.	23-g	Hornfels in granite	74.01	0.59	11.67	0.36	4.13	1.72	1.81	0.07	1.64	3.08	0.10	0.84	100.02

flaky biotite, and silt-size granules of quartz with a small addition of fine feldspar granules. It is characterized by a high content of K_2O (4.63%) and Al_2O_3 (16.75%) and a low content of SiO_2 (65.21%) and Na_2O (1.70%). At a distance of 1 km from the granite body the schist (sample 5-a) undergoes marked changes in composition: SiO_2 rises by 4.36%, while Al_2O_3 and K_2O are lower by 3% and 1%, respectively. The contents of other elements varies slightly within the range of 0.5%. Further changes take place at a distance as near as 200 m from the granite body, when two-mica schist (phyllite) is replaced by typical schistose hornfels (sample 9), resembling gneiss in texture. Compared to moderate contact-metamorphic schist (sample 5-a), typical hornfels (sample 9) differs by having a higher content of FeO and CaO, while content of K_2O decreases by 0.74%, Fe_2O_3 — by 0.52% and Al_2O_3 — by 0.34%. The contents of some other elements also decreases.

The change from almost unaltered mica schist to typical hornfels, as represented by samples 3, 5-a, and 9 is accompanied by relatively small changes in composition shown by higher contents of silica and iron and lower contents of potassium and alumina. However, there is no reason to believe that typical mica schist and contact rock, are gneiss and crystalline schist respectively, or that the contact rock is older. The same is true for hornfels of all other areas which are close in composition (analyses of samples 31, 14-v, 28-b), as they also formed during contact metamorphism of phyllite.

Hornfels xenoliths (samples 32-v and 23-g) occurring within granite, differ from mica schist. They are characterized by an increase in SiO_2 content (by 6–7%), a decrease in Al_2O_3 (by 4–5%), and a lower percentage of alkalis, especially K_2O (by 1.5–2%); the content of alkalis is equal to that of the surrounding granite.

The similarity in chemical composition between mica schist and so-called “gneiss”, “migmatite” and “crystalline schist” (that are really hornfels) establishes that they are closely related, and not separate sequences of older Proterozoic rock. All the changes taking place in mica schist near granite bodies are easily explained by the mobility of silica and alkalis and by the increase of thermal gradient in the country rock during the crystallization of granite magma.

Physical properties of the rock

Measurements of density, differential measurements of velocity of propagation of longitudinal elastic fractuaction, magnetic susceptibility and specific electric resistance were carried out for three rock groups: phyllite, hornfels and granite. Sonic measurements were carried out on 4–6 mm thick plates. The error of density measurements is ± 0.001 g/cm³. Relative error of measurements of magnetic susceptibility and of electrical resistivity is not more than 10%.

The comparison of physical properties of three samples, 3–phyllite, 5-a–contact metamorphosed phyllite, and 9–schistose hornfels, is of principal interest. In phyllite (Fig. 2), in spite of its microscopic homogeneity, boundaries of uniform fields of magnetic susceptibility, of sonic speed, and, to a lesser

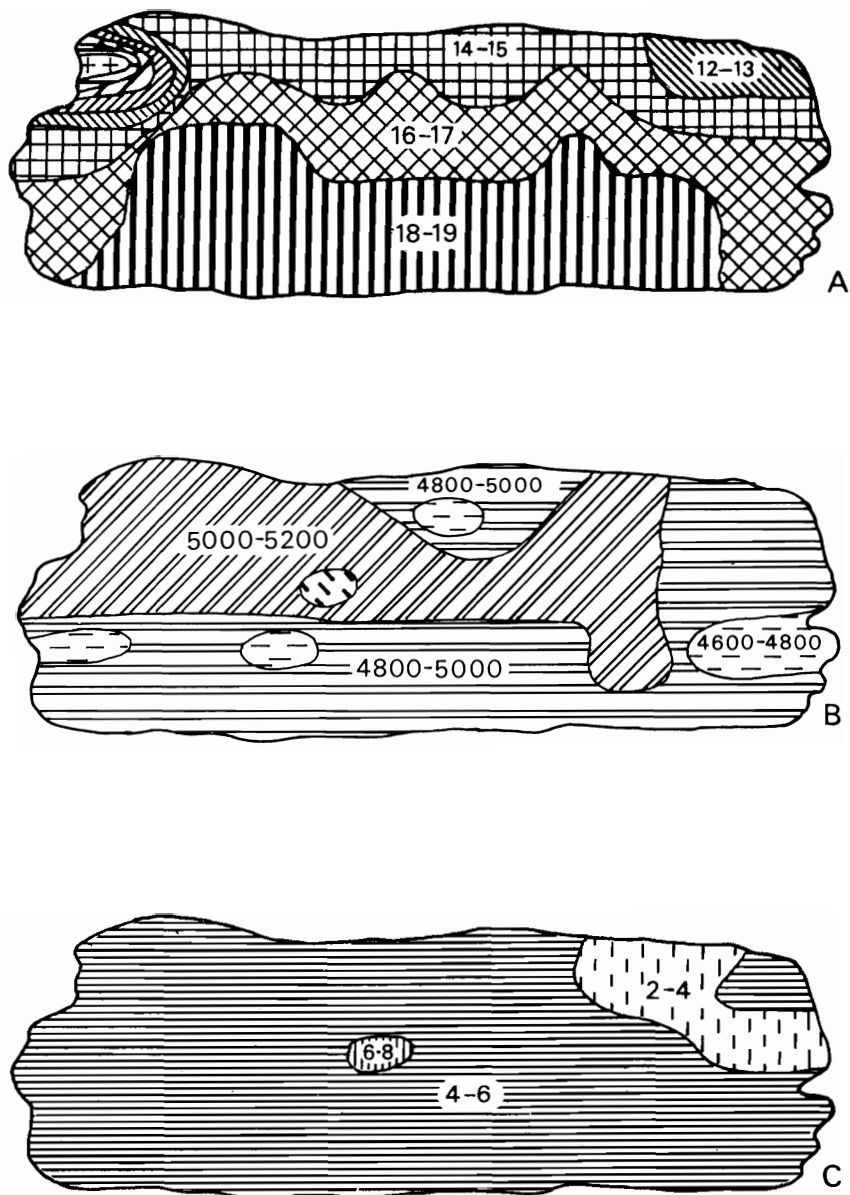


Fig. 2. Specimen 3 ($\frac{1}{2}$ of natural size): A — distribution of velocity of longitudinal elastic waves (m/sec); B — distribution of magnetic susceptibility (10^{-6} CGSM); C — distribution of electric resistance (10^5 ohm/cm).

degree, of electrical resistivity are defined, and they coincide with the direction of schistosity. This may indicate that the given distribution is syngenetic to phyllitization or related to earlier sedimentary texture processes. In the contact-metamorphosed phyllite (Fig. 3) the boundaries change their configurations but still coincide with the direction of schistosity. This is most pronounced for the distribution of magnetic susceptibility. Magnetic susceptibility generally decreases compared with normal phyllite, while electrical

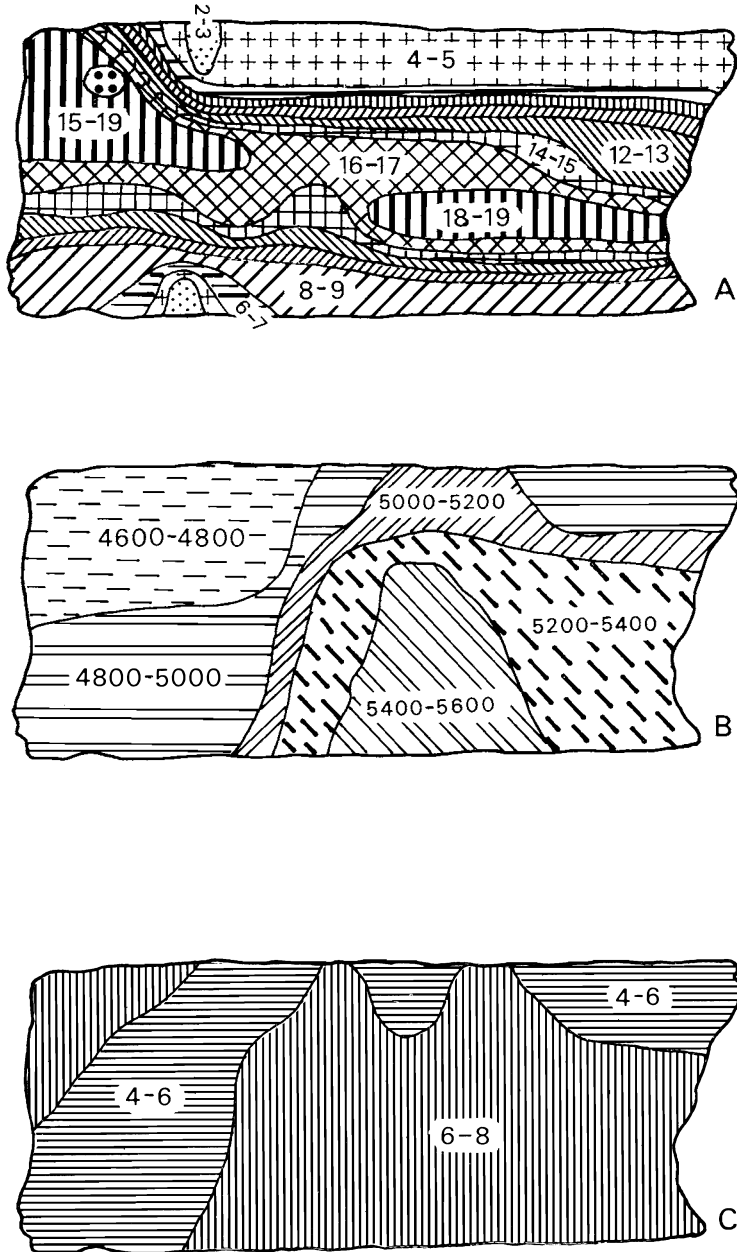


Fig. 3. Specimen 5-a ($\frac{1}{2}$ of natural size): A — distribution of velocity of longitudinal elastic waves (m/sec); B — distribution of magnetic susceptibility (10^{-6} CGSM); C — distribution of electric resistance (10^5 ohm/cm).

resistivity and speed of propagation of longitudinal ultrasonic waves show an increase. Further changes in all these physical properties take place in hornfels (Fig. 4), though their boundaries, as a whole, continue to coincide with the direction of schistosity of the primary phyllite. In this case, the magnetic susceptibility falls, while sonic speed and electric resistance rise. These changes are associated with recrystallization of phyllite during contact metamorphism,

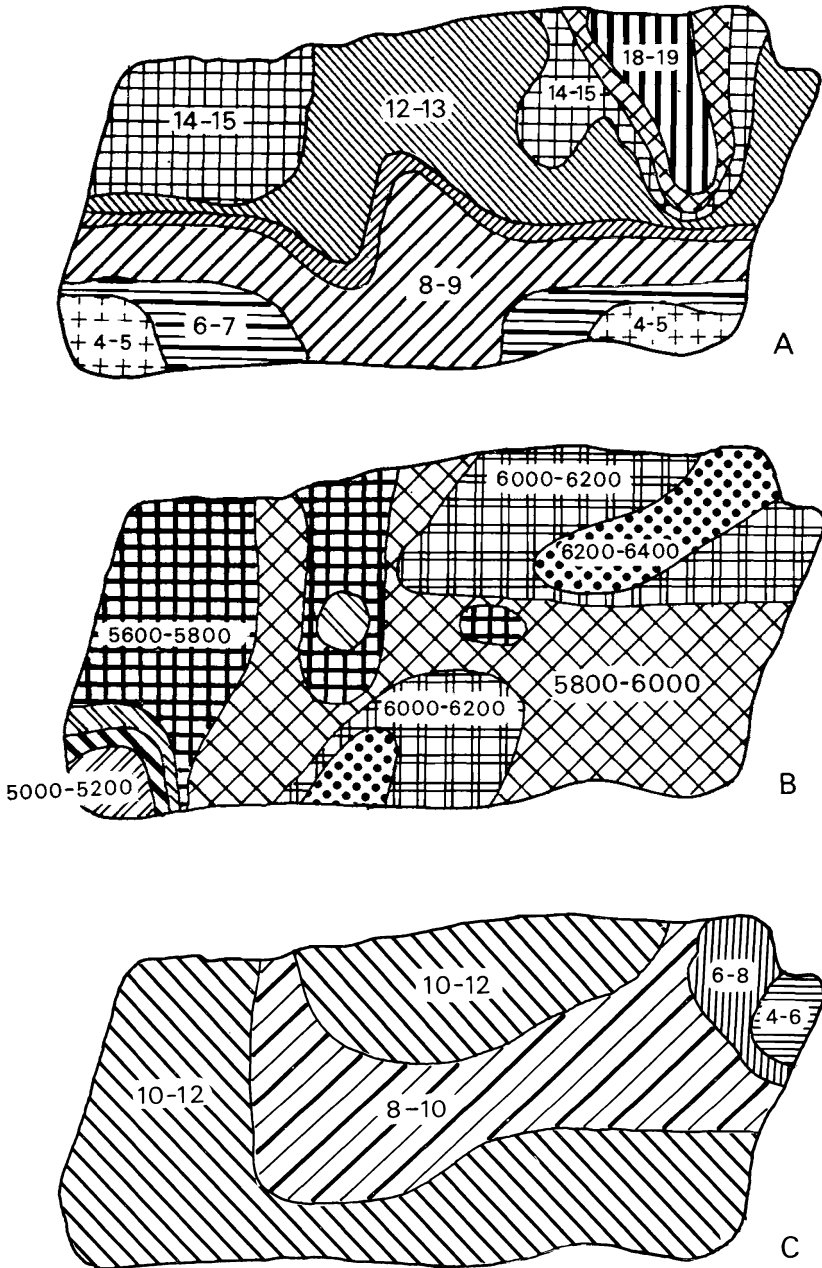


Fig. 4. Specimen 9 ($\frac{1}{2}$ of natural size) : A — distribution of velocity of longitudinal elastic waves (m/sec); B — distribution of magnetic susceptibility (10^{-6} CGSM); C — distribution of electric resistance (10^5 ohm/cm).

causing improvement of acoustic contact between grains and raising the electric resistance. Recrystallization leads to ruptures of boundaries of uniform fields in hornfels, and therefore field boundaries in places cut the primary schistosity.

In spite of changes shown above, the majority of physical properties of the three samples, collected from a section along the flanks of Kollerfjorden and

the Kollerbreen are so similar, that their close consanguinity is doubtless. The density of the three samples varies from 2.704 to 2.780, decreasing with increase in metamorphic grade. Magnetic susceptibility decreases from $14-19 \times 10^{-6}$ CGSM to 8–18, while electric resistance increases throughout the complete recrystallization of phyllite to hornfels.

The distributions of these physical properties for the adjacent granite massif (Fig. 5) are given for comparison with the phyllite-hornfels series described above. Granite is characterized by uniform physical parameters and by the absence of differentiation within a sample. Its magnetic susceptibility is much lower and its electric resistance measurements are higher than those of phyllite and even hornfels. Granite is a massive uniform rock, in contrast to phyllite and hornfels, where schistosity causes variation in physical parameters.

Conclusions

Results of careful geological, petrographical, chemical, and petrophysical studies carried out on rocks of the Krossfjorden area by the author during the summer of 1972 and subsequently in 1974, prove that there is no Early Precambrian Kollerbreen Formation, more than 3500 m in thickness, made up of metasomatic granite, migmatite and gneiss, in the study area. In this region the Signehamna and Generalfjella Formations are developed with a total thickness of 8–9 km, representing the lower Hecla Hoek Succession probably of Riphean age. Mica schist (phyllite) with horizons of marble in the upper part prevail.

A fracture in the Krossfjorden area controlled intrusion (during the Caledonian tectogenesis) of granitoid, probably of rheomorphic mass. Crystallization of the magma resulted in the formation of granite and granodiorite bodies and metamorphism of phyllite into hornfels, sometimes with fine injection of granite magma, resulting in a migmatite-like rock. These schistose hornfels and migmatite-like rocks have been taken by previous workers for a regional-metamorphic gneiss-migmatic complex.

K-Ar absolute age determinations on this quasi-gneiss-migmatite rock yield values in the range of 420–440 m.y. (KRASIL'ČIKOV 1973). Additional determinations have been carried out in the laboratory of the Radium Institute of the Academy of Sciences of the USSR. The results obtained are given below.

Sample No.	Rock	K%	Ar $\frac{\text{cm}^3}{\text{g}} \cdot 10^{-5}$	Age, m.y.	Note
23-I	Hornfels (gneissic)	3.34	6.12	430	The upper Conway-breen
14-ž	Leucocratic granite	4.36	6.89	380	The flank of the Fjortende Julibreen

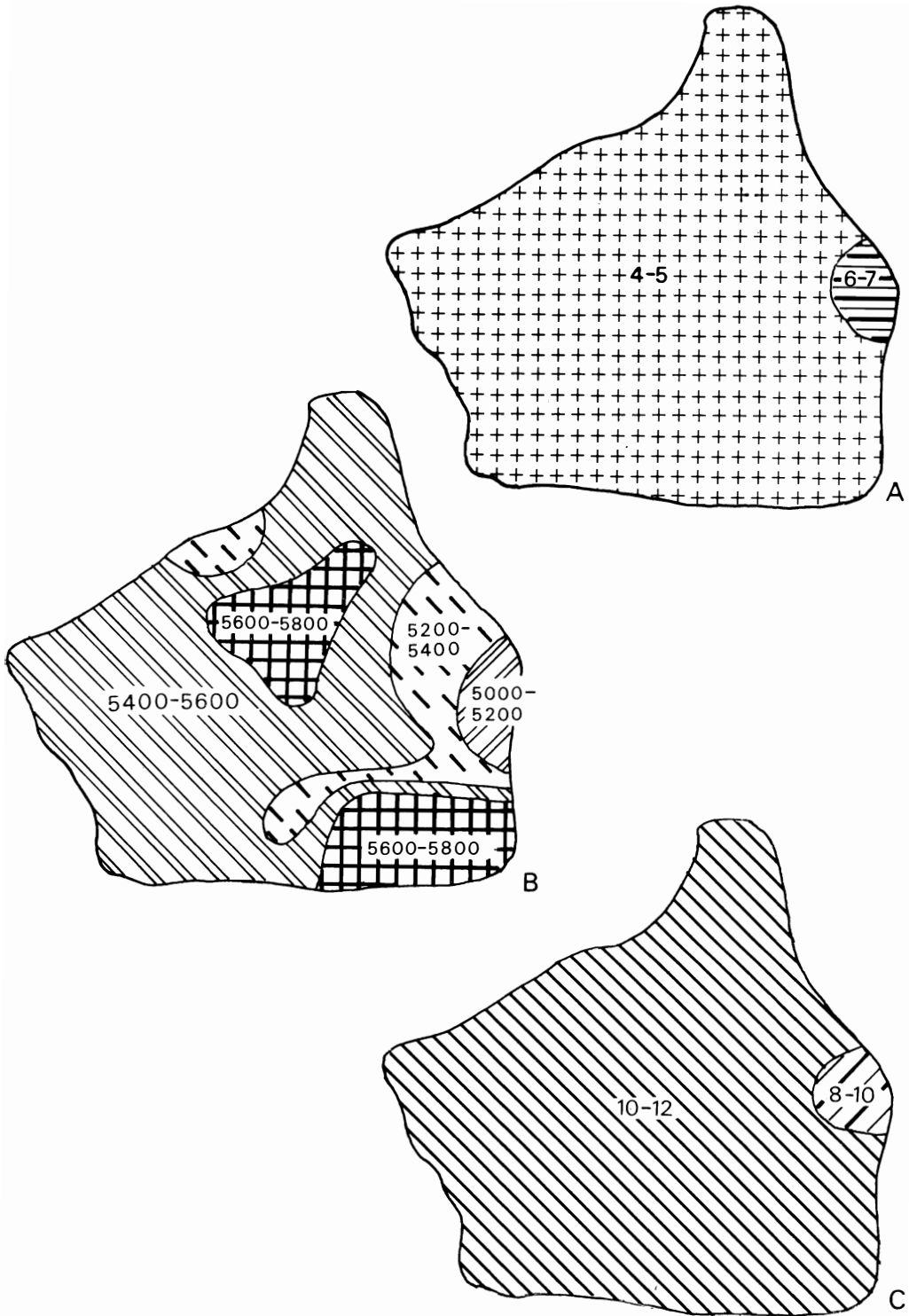


Fig. 5. Specimen 14-z: A — distribution of velocity of longitudinal elastic waves (m/sec); B — distribution of magnetic susceptibility (10^{-6} CGSM); C — distribution of electric resistance (10^5 ohm/cm).

Thus, the hornfels is 50 m.y. older than granite. This can only be explained by an initial Riphean age for the hornfels which was not quite rejuvenated during contact metamorphism.

Although the band of contact-metamorphic rock traced by the author extends for only 40 km across strike, we think that the age of the so-called Early Precambrian gneiss-migmatite complexes of Spitsbergen should be reviewed, as they may prove to be not the oldest crystalline rocks of the region, but contact-metamorphics associated with Caledonian magmatism. The Caledonian geosyncline of Spitsbergen may have been initiated not within the Early Precambrian basement of Karel tectogenesis, but within the Riphean basement of Baikal tectogenesis, like all the other Phanerozoic geosynclines. The Caledonian tectono-magmatic processes in Svalbard with wide spread development of deep rheomorphism, and associated plutonic activity and metamorphism, is characteristic.

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Peculiar features of regional metamorphism of northwestern Spitsbergen

By S. A. АБАКУМОВ

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Abstract

Main features of geology and metamorphism of northwestern Svalbard are presented in the paper. On the basis of a lithological analysis, a division into formations of the metamorphic successions of the Precambrian is produced. The crystalline complex of the base of the Caledonides and the main geosynclinal complex have been recognized. The main regular patterns and succession of the metamorphic transformations have been exposed. A petrographic study and an analysis of paragenetic associations have made possible a conclusion on the superposition of the Caledonian metamorphism on to the crystalline rocks of the basement. The latter circumstance has predetermined a vast development of processes of ultrametamorphism in Caledonian time.

Introduction

For several years the author has carried out geologic-petrographic studies of metamorphic rocks in north-western Spitsbergen. The study area is bounded to the south by Kongsfjorden and to the east by a major fault extending southward from Raudfjorden along Monacobreen and Isachsenfonna. The area includes Amsterdamøya and Danskøya. The author was a member of the Spitsbergen Expedition of the Research Institute of the Geology of the Arctic. Study of mineral composition and the metamorphic history of the area were the main aims of the investigation.

Previous work

North-western Spitsbergen has attracted the attention of geologists from different countries since the second half of the last century; works by HOEL, HOLTEDAL, SCHENK, FREBOLD and HARLAND are well known. The results of earlier investigations are reviewed in recent papers by GEE and HJELLE (1966), HJELLE and OHTA (1974) and KRASIL'SČIKOV (1973).

GEE and HJELLE (1966) proposed a stratigraphic standard section for slightly metamorphosed Upper Proterozoic rocks in the Krossfjorden area. They recognized a migmatite complex at the base of the section, overlain by three rock units totalling a thickness of 7000–7500 m.

The Nissenfjella Formation was reported from the western coast in the Nissenfjella area. It is composed of “pelites” with numerous thin amphibolite lenses and bands. The lower part of the formation contains granitized rocks which increase in quantity northward. The upper boundary is drawn at the disappearance of amphibolites.

The Signehamna Formation outcrops in Mitrahalfvøya and on the shores of Krossfjorden. It is represented by a homogenous “pelites” with subordinate “psammites” and quartzites.

The Generalfjella Formation caps the sequence of metamorphic rocks in the area and is composed mainly of marbles and quartzites.

Later HJELLE (in: HJELLE and OHTA 1974) rejected the previously recognized Nissenfjella Formation and considered the whole complex of deeply metamorphosed rocks to be equivalent to the Signehamna and Generalfjella Formations, granitized and migmatized during the Caledonian orogeny.

KRASIL'SČIKOV (1973) in his papers on the tectonic reconstruction of the archipelago in the pre-Cambrian–Lower Paleozoic, recognizes a crystalline basement of the Caledonides (presumably of Lower-Middle Proterozoic age) and a major geosynclinal complex composed of Upper Proterozoic rocks. Thus, at present there are two opposing concepts of north-western Spitsbergen geology.

Geological setting

The analysis of previous investigations and the evidence obtained by the author allow a new interpretation of the main features of north-western Spitsbergen geology.

A granite-gneiss complex, made up of both granitized and migmatized Riphean rocks and relics of deeply metamorphosed Caledonide basement, lies at the base of the metamorphic rock section. Because of the present lack of knowledge it is impossible to map separately formations within this complex.

The granitized Riphean rocks are represented at the base by the Nissenfjella Formation, overlain by the Signehamna and Generalfjella Formations.

Igneous rocks consist of the Pre-Riphean dyke complex of basic rocks and Caledonian granitoids, among which synorogenic and post-orogenic intrusions may be distinguished.

It has been suggested that a southward plunge of fold axes, one of the main structural features of the region, accounts for the exposure of deeper parts of the section in the northern areas. In my opinion block tectonics resulted in the juxtaposition of rocks of different metamorphic grade, and are of great importance.

Stratigraphy

The scheme proposed by GEE and HJELLE (1966) is used as a basis for stratigraphic subdivision. However, while sedimentary layering of the slightly metamorphosed upper subdivisions (the Signehamna and Generalfjella Formations) is evident, the mapping of the underlying Nissenfjella Formation presents some difficulties. This is because the metamorphic facies boundary does not coincide with and thereby obscures the stratigraphic boundaries. A stratigraphic succession in the area follows:

- A — Pre-Riphean complex, the basement of Caledonian geosyncline
- B — Riphean geosynclinal complex:
 - 1 — Nissenfjella Formation
 - 2 — Signehamna Formation
 - 3 — Generalfjella Formation

Geologic and petrographic characteristics are given below for each stratigraphic unit.

PRE-RIPHEAN COMPLEX, THE BASEMENT OF CALEDONIAN GEOSYNCLINE

This is largely a granite-gneiss complex underlying more than half of the study area. To the south, in the Conwaybreen area, it consists of a N-S trending band 4 km wide gradually widening northward. In the Kollerbreen area this band is 10–11 km wide and in the Magdalenefjorden–Raudfjorden area and farther north the granite-gneiss complex reaches 20–27 km in width. It extends for more than 100 km from south to north.

This complex consists largely of biotite, biotite-garnet, biotite-amphibole, and biotite-cordierite in places with sillimanite gneisses and plagiogneisses. Thin bands of skarns occur throughout the section. Amphibolite lenses and bands occur in the western part of the complex (Danskøya, Amsterdamøya and southern coast of Bjørnefjorden).

Magnesian and aluminous rocks may be distinguished by their different bulk chemical compositions. Magnesian rocks occur in Danskøya and Amsterdamøya and on the peninsula between Bjørnefjorden and Magdalenefjorden. Aluminous rocks occur in the eastern exposures of the granite-gneiss complex.

Biotite gneisses, biotite-amphibolite, clinopyroxene-amphibolite, and biotite-garnet gneisses and plagiogneisses dominate in the magnesian-rich parts. Bands of carbonate rocks are represented by salitic pyroxene and marbles with diopside, scapolite, garnet, and wollastonite assemblages. Feldspar amphibolites, concordant with the surrounding rocks, occur as thin lenticular

beds. Highly aluminous rocks such as biotite-cordierite gneisses and plagiogneisses are not characteristic of this complex. They occur as scarce, thin beds pinching out along strike.

In the eastern part of the granite-gneiss complex, biotite-cordierite gneisses and occasionally plagiogneisses with low sillimanite content predominate, with biotite gneisses and plagiogneisses. Biotite-garnet gneisses occur in the form of thin lenticular beds. Diopside, olivine, wollastonite, prehnite, muscovite, with spinel, gummite and phlogopite are the most common minerals in the marble. Amphibolite bodies, and beds containing scapolite and amphibole gneiss units are absent in this part.

The base of the sequence is unknown. This contact with overlying succession in many places is a tectonic one. In particular this is true of the eastern contact. To the south in the Conwaybreen area the boundary follows a glacier, and granitoids are cataclastic in the contact zone. The western contact of the granite-gneiss has a complex pattern. Lines of granite intrusions occur along the contact zone, and cataclastics are common. This extensive zone of cataclasis can be traced along the eastern coast of Smeerenburgfjorden. Farther south a big intrusion of the Hornemantoppen granite is related to this zone. In the Kollerfjorden, Mayerbreen, and Tinayrebreen areas there are small intrusions of synorogenic granites probably also related to this tectonically weakened zone. There is no evident unconformity and a transitional zone up to 100 m thick impregnated with granitoid material lies adjacent to the contact.

In Amsterdamøya, Danskøya and Hoelhalvøya, granitized and migmatized rocks of the Nissenfjella Formation cannot be distinguished from the Pre-Riphean complex.

RIPHEAN GEOSYNCLINAL COMPLEX

The Nissenfjella Formation lies at the base of an up to 3 km wide outcrop of the Riphean complex. It consists mainly of biotite plagiogneisses with porphyroblastic texture and banded gneisses. There are occasional lenses of feldspar amphibolite. A characteristic feature is the complete absence of highly aluminous rocks. The amount of granite gradually increases from south to north, and on Hoelhalvøya layered migmatites are present. The Nissenfjella Formation can be recognized in the south-western Danskøya. The contact with the overlying Signehamna Formation on the west coast of Spitsbergen follows a thrust of N–S trend and eastward dip.

The Signehamna Formation occurs on the shores of Krossfjorden where it forms the limbs of a complex synclinal structure. In southern Mitrahallvøya the formation is composed of mica-quartz-albite slates, and in the northern part of mica-quartz-garnet-albite slates. In Kong Haakons Halvøy and Kapp Guisnez, quartzite beds are intercalated with the slates. The outcrop on the northern coast of Kongsfjorden consists of mica-quartz-albite slates. There is a stratigraphic contact with the overlying Generalfjella Formation.

The Generalfjella Formation is exposed in central Mitrahallvøya where it forms the core of a syncline. It also outcrops as a northward narrowing band from Blomstrandhallvøya in the south, to Tinayrebreen in the north. Another small

outcrop forms a N-S trending band at the mouth of Conwaybreen pinching out at Fjortende Julibreen. The formation consists of marbles of varying colours.

PETROGRAPHY. METAMORPHISM, AND ULTRAMETAMORPHISM

Marbles making up the Generalfjella Formation are inequigranular and fully recrystalline. Their composition is close to that of limestones with a significant magnesium content. The content of dolomite ranges from 5 to 25 per cent. Thermal studies of the marbles gave a temperature of carbon combustion of 550°, corresponding to greenschist facies metamorphic conditions (BLJUMAN et al. 1974).

There are two dominant rock types in the Signehamna Formation, mica-albite-quartz and mica-oligoclase-quartz-garnet slates. Numerous determination of refractive indices and orientation of optical indicatrix showed that the former contain only albite (An_{10}) while the latter include acid oligoclases (An_{9-14}). The garnet is close in composition to almandine (83%). Tourmaline is a notable accessory mineral. The mica-albite-quartz association corresponds to the greenschist facies recognized by ESKOLA. Albite-epidote-chlorite is considered to be a diagnostic association of the greenschist facies. However, TURNER (1951) points out that an abrupt change in plagioclase composition from albite (An_{0-7}) to oligoclase (An_{15-20}) takes place with increase in metamorphic grade of pelites to the almandine zone. This change in plagioclase composition marks the high temperature limit of the greenschist facies. SUDOVIKOV (1964) claims that with the appropriate K content and relatively high temperatures, chlorite may be completely replaced by biotite, and calcic epidote by muscovite.

Another characteristic association, oligoclase-almandine-biotite, corresponds to the epidote-amphibolite facies (SUDOVIKOV 1964). Thus, mica-albite-quartz is characteristic of high temperature subfacies of the greenschist facies and of a zone transitional to epidote-amphibolite facies. This reflects the increase in grade of metamorphism with depth. The mica-olig-Qtz-garnet association is characteristic of a higher grade of metamorphism marked by the appearance of garnet in the epidote-amphibolite facies.

The Nissenfjella rocks are also characterized by an oligoclase, biotite-garnet paragenesis. Almandine is the predominant component of garnet (53-69%) but the grossularite content (21-36%) is significant. Biotite pleochroic from dark brown with a greenish tint to colourless grey, with a molecular Fe content from 40 to 50%, is common. This mineral association also corresponds to the epidote-amphibolite facies. Thus, the rocks composing the lower part of the Riphean geosynclinal complex formed under the conditions of greenschist and epidote-amphibolite facies metamorphism. A gradual increase in metamorphic grade occurs through the section and the boundary between the two facies is the Signehamna Formation.

Gneisses and plagiogneisses building up the Pre-Riphean granite-gneiss complex have a distinct granoblastic texture and gneissic or flow-structure. Usually they are equigranular and homogeneous with an average grain size of

1.5 to 2.0 mm. In eutaxitic varieties minerals are differentially distributed in the form of quartz-feldspar bands and bands enriched with mafic minerals.

Plagioclase is represented by two minerals. Acid-intermediate andesite (An_{20-40}) is most common. Labradorite (An_{50-70}) is less common and occurs mainly in Amsterdamøya and Danskøya and in northern Albert I Land. Potash feldspar occurs in the form of orthoclase-anorthoclase. Amphiboles are usually represented by hornblende with a molecular Fe content of 27 to 65%. Cumming tonite is less common. Biotite in the rocks free from granitization is duller in colour with a molecular Fe content of 40 to 50%. Cordierite occurs as both fresh unaltered grains, and partly or completely converted into pinitite. Molecular Fe content of the cordierite is usually 18 to 20%. Garnet is usually unaltered euhedra with rare intensely chloritized varieties. Almandine (58–72%) predominates, the pyrope content slightly higher (17–25%) compared to the Riphean rocks.

Paragenetic associations of the above listed minerals are stable only under the conditions of the amphibolite facies. Specific conditions due to partial melting commencing at this stage effect mineral assemblages. Taking this into account, SUDOVIKOV (1964) divided the amphibolite facies into two sub-facies; relatively low temperature staurolite-kyanite and higher temperature sillimanite-almandine. Rocks not affected by ultra-metamorphism belong to the former, and those which have undergone melting, to the latter.

An important feature of regional metamorphism in north-western Spitsbergen is that the transition to rocks of the granite-gneiss complex is accompanied by an abrupt increase in metamorphic grade.

A remarkable feature of the region is an exposure of the Caledonian ultra-metamorphism zone, an area of early melting and magma generation with inclusions of unmelted substratum. The volume of the melt accounts for the injection and movement of the bulk of the rock. Migmatite formation, granitization, and anatexis were widely operative.

Bearing in mind the different usage of the above terms by many geologists it is appropriate to explain their meaning as they are used in the present paper. According to READ (1949) granitization is the transformation of solid metamorphic rocks into granitic rocks without passing the magmatic stage. The refinement proposed by NIGGLI (1949) that granitization incorporates recrystallization accompanied by pneumatolytic-hydrothermal processes is also appropriate in this case. The term “anatexis” is taken to mean granitization accompanied by melting.

The most homogeneous parts of the melting zones could have been injected into higher beds during partial melting.

Migmatites of different morphological types are widespread. White, light-grey granitoids occur as vein material. The paleozones consist of plagiogneiss and gneiss of the Pre-Riphean complex. Lenticular bodies, up to 0.8–1.0 m, of pegmatitic granitoid often occur in the vein material. Thus, the melt which gradually developed in the zone of ultrametamorphism, is thought to have been injected in the form of vein material.

There is a gradation between migmatites of different types. They are all

related by a granite migmatite vein material. The process of granitization accounts for the appearance of closely related granite migmatites. Migmatization took place simultaneously or before granitization, resulting in the formation of layered migmatites. During their formation granite migmatites were in a plastic state and were able to move over short distances. This is shown by the numerous xenoliths and skialiths of plagiomigmatites, plagiogneisses and gneisses in the matrix. Granite migmatites forming at a centre (perhaps a central part of a partial melting chamber) were withdrawn from migmatites, skialiths and similar relics in the matrix. They became lighter and more homogeneous and the gneissic structure disappeared. This occurred when the bulk of the rock melted and could be injected. Synorogenic anatectic granites formed on solidification. The presence of garnet, sillimanite and cordierite, minerals not characteristic of normal igneous rocks, provides support for this view.

Local processes of sodium and solitic-potash metasomatism were operative during the final stage. In general porphyroblasts were not oriented, indicating their formation subsequent to the main stage of deformation.

The petrographic study of the granitoids indicates zonal features and an effect on rocks of the Pre-Riphean basement. In all the granitoids, plagioclase has an extremely, constant composition of basic oligoclase (An_{25-30}). Potash feldspar is usually represented by orthoclase-anorthoclase and, only in granites of Miethebreen, by lattice-twinned microcline. A characteristic feature is the predominance of plagioclase over potash feldspar. Only rocks subjected to silicic-potassium metasomatism, is the composition close to that of typical granites. Newly formed biotite is bright redbrown in colour and its molecular Fe content ranges up to 60–65%. Another feature characteristic of most of the granitoids is the presence of two generations of plagioclase, potash feldspar, and biotite. The combination of successive stages of plagioclase formation resulted in the coexistence of plagioclases with different anorthite content in a given rock. The geological study of migmatites and granite migmatites and their relationships suggest at least two stages of migmatization and granitization. The Caledonian metamorphism effected the earlier migmatized and granitized rocks. The formation of the latter took place under the conditions of at least amphibolite facies.

The highest grade of metamorphism in Caledonian time appears to be reflected by ultrametamorphism characterized by the formation of plagioclase with a composition not higher than that of oligoclase-acid andesite. It follows that the presence in plagiogneisses of a plagioclase matrix with a composition of basic andesite and even labradorite argues for Pre-Caledonian conditions of metamorphism higher than those of the Caledonian metamorphism.

Conclusions

The area includes two main rock-units, the Pre-Riphean basement and the Riphean geosynclinal complex.

Two units differing in lithology of enclosing rocks are recognized in the Pre-Riphean basement.

The Riphean complex is mapped as a single syncline closing in central Lilliehöökreen.

The basement complex is polymetamorphic in origin. Metamorphic events of Pre-Caledonian time predated the widespread processes of ultrametamorphism in Caledonian time. Petrographic studies show the superposition of low temperature associations of amphibolite facies on high temperature Pre-Caledonian facies.

Two stages of migmatization — Pre-Caledonian and Caledonian — have been recognized.

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Aspects of the geology of northwest Spitsbergen

(Some results of recent Norsk Polarinstitut expeditions)

A. HJELLE

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Abstract

The geology of the crystalline rocks in the Caledonian orogenic zone of northwest Spitsbergen is summarized.

Late Precambrian supracrustals were regionally metamorphosed during late Proterozoic or early Paleozoic. Migmatization commenced in (mid?-) Silurian, and as the last plutonic event a batholithic post-tectonic granite was intruded in late Silurian.

Several deformation episodes are distinguished, corresponding to various stages of metamorphism and intrusion.

Introduction

GENERAL

This paper summarizes some of the main results from the geological work of the Norsk Polarinstitut expeditions 1963–75, considering the areas of map sheets A4, A5 and A6. Reports and publications concerning these areas, from the following geologists, are used without separate references in the text: T. v. AUTENBOER, D. G. GEE, T. GJELSVIK, A. HJELLE, Y. OHTA and E. TVETEN. The approximate areas visited by the various geologists are indicated in Fig. 1.

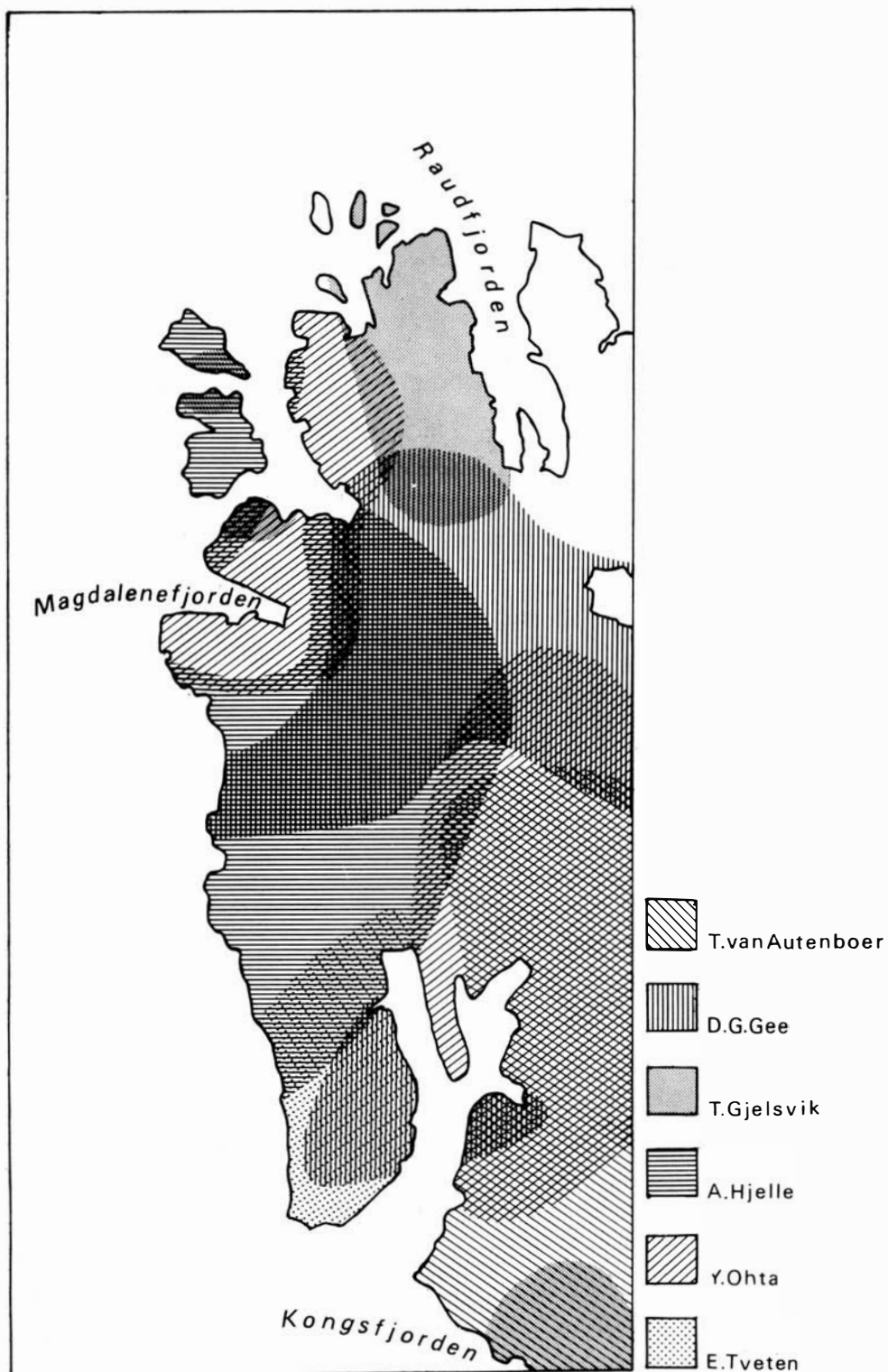


Fig. 1. Areas visited by various geologists of Norsk Polarinstitutt expeditions 1963-75.

GEOLOGICAL SETTING

The described area is situated north of Kongsfjorden, in the Caledonian orogenic zone of northwest Spitsbergen. Fig. 2. The areas adjacent to Kongsfjorden are largely made up of metasediments, whereas various gneisses, migmatites, and granites prevail in the central and northern parts.

The southern area (east and west of Krossfjorden-Möllerfjorden).

Except for a zone of migmatite and grey granite in the easternmost part, marbles and a monotonous sequence of pelitic rocks with subordinate quartzite and psammite make up most of the area. The calcareous rocks are overlying the main pelitic sequence both east and west of Krossfjorden. Primary sedimentary structures indicate a general lack of inversion.

The western transition area (between Førstebreen and Sjettebreen).

Though the rocks here have much of the character of metasediments as in the southern area, they are generally more gneissic, and to the north and east they grade into migmatites and syntectonic weakly foliated granite rocks.

The eastern transition area (west of the northern part of Monacobreen).

Conditions are much the same as in the western transition area, except that mobility and granitisation increase towards the west. The metasediments are separated from the Downtonian beds in the east by more or less continuous faults along Monacobreen and Isachensfonna.

The central and northern gneiss and migmatite area (mainly north of Sjettebreen to Louetbreen).

The bulk of the rocks are various gneisses, and migmatites with pelitic and psammitic restites, and with minor scattered marble bands. Grey granitic rocks occur widely, as dykes or more or less dome shaped bodies. As a last plutonic event the Horneman granite was intruded in the central part of the area. This red granite appears to have affected the regional structures considerably, however, in small scale it cuts the surrounding rocks discordantly.

Stratigraphy

In the central and northern gneiss and migmatite areas plastic deformation and differential melting have modified the supracrustal rocks to such an extent that it is impossible to establish all but scattered local stratigraphic sections. The best preserved beds occur in the metasedimentary areas of Kongsfjorden and Krossfjorden.

The absence of (macro-) fossiliferous strata and of rocks which are typical in Vendian or upper Riphean beds in Spitsbergen as tilloid rocks, oolitic rocks or black sulphurous limestone, suggests the beds in Kongsfjorden and Krossfjorden to be pre upper Riphean, or, in the terms established for Ny Friesland and Olav V Land, pre Akademikerbreen Group (HARLAND et al. 1966).

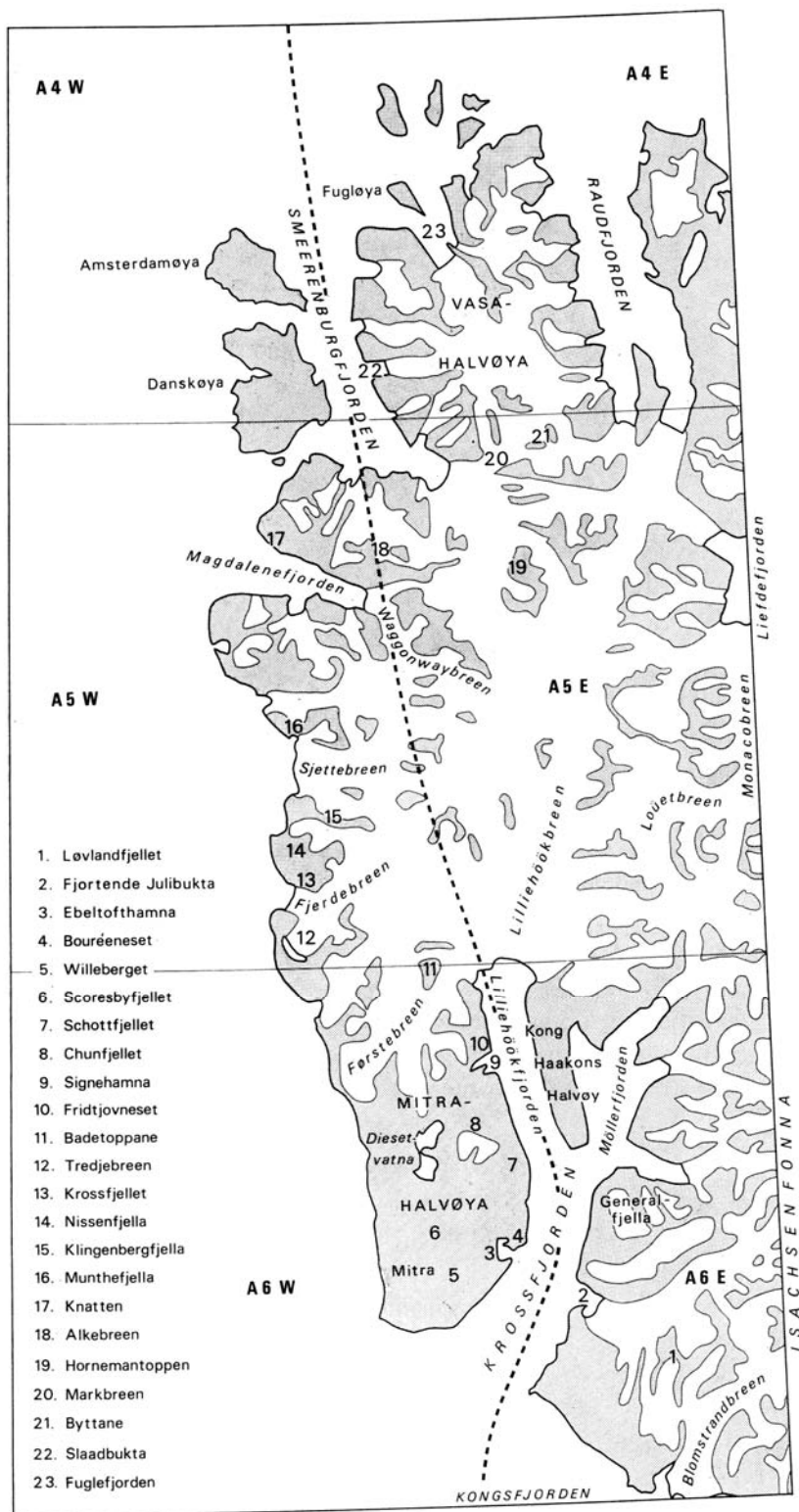


Fig. 2. Location map for names used in the paper. The map also shows the division used in Fig. 7.

In Ny Friesland several of the formations below the Akademikerbreen Group contain calcareous beds which might be possible correlatives to the calcareous beds described below:

Kingbreen Formation	}	Veteranen Group
Kortbreen Formation		
Lower part of Vildadalen Formation		Planetfjella Group
Rittervatnet Formation		Harkerbreen Group
Upper part of Smutsbreen Formation		Finnlandveggen Group

Considering the lithology and thickness of the beds north of Kongsfjorden, the author prefers to correlate the calcareous rocks to the lower part of the Vildadalen Formation and the various pelitic, psammitic and gneissic rocks below to the Flåen and Sørbreen formations below the Vildadalen formation. Variations in thickness and sedimentary facies due to different primary positions of Ny Friesland and the west coast area in a main geosyncline, or due to relative north-south movements would introduce considerable uncertainties in the correlation.

In the beds around Krossfjorden, the metamorphic grade generally decreases upwards and southwards and the lowest grade lithologies broadly coincide with the highest stratigraphic levels exposed.

The youngest beds occur east of Krossfjorden (the Generalfjella Formation, GEE and HJELLE 1966) (Fig. 3):

- >400 m of light grey or buff marbles intercalated with subordinate phyllite
- C. 1300 m of phyllite with subordinate marble and quartzite
- C. 800 m interbanded marble, phyllite and quartzite

In the northern slope of Generalfjella, c. 3 km northwest of the main (855 m) peak, a section is obtained from the upper part of the Generalfjella Formation:

- > 150 m banded marble
- 60 m yellow, banded marble
- 30 m blue marble
- 12 m yellow weathering marble
- 2 m phyllite
- 7 m blue marbe
- 40 m silver grey phyllite
- 30 m marble, weathering blue in upper part, yellow in lower part
- 40 m silver grey phyllite, weathering dark red brown
- 25 m marble with thin quartzite bands at base
- 60 m silver grey phyllite, weathering dark red brown
- 10 m red stained quartzite
- 20 m dark grey quartzite
- 10 m (+ ?) silver grey phyllite
- ? m psammitic quartzite with subordinate phyllite

Further east, just north northwest of the 855 m peak another section is measured, which apparently belongs to a lower part of the Generalfjella than the section above (a fault with downthrow to the west divides the two sections):

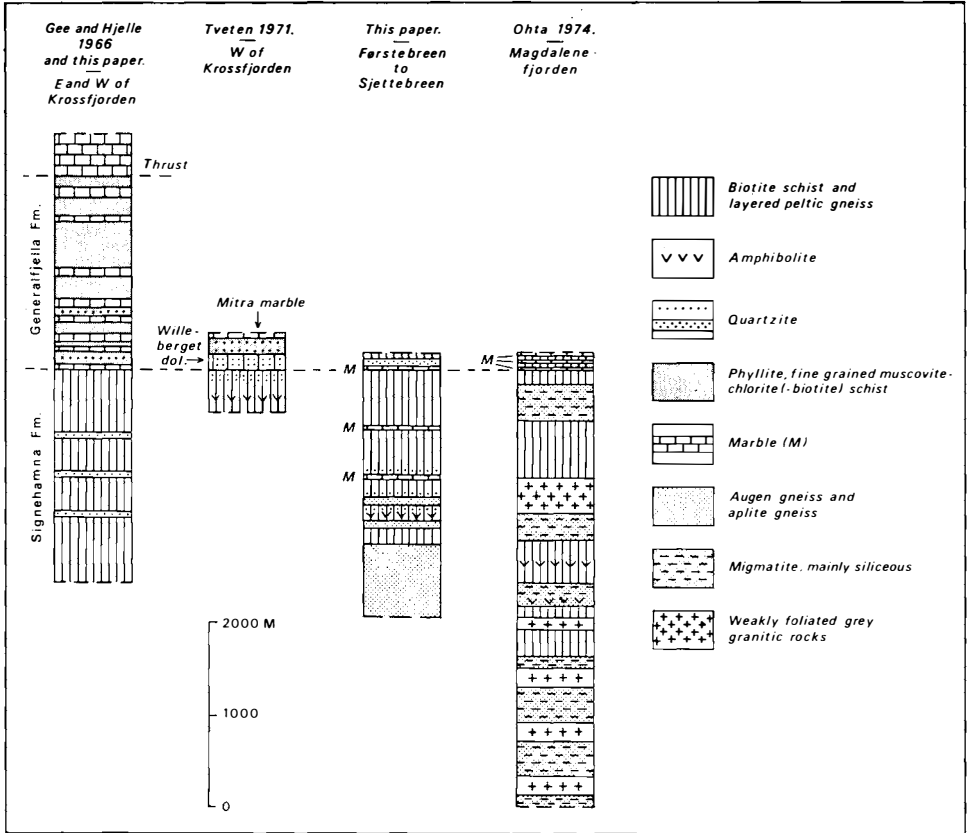


Fig. 3. Local stratigraphical correlation, northwest Spitsbergen.

- ? m thinly interbanded blue grey and white phyllite and quartzite
- 20 m phyllite
- 30 m psammitic phyllite
- 15 m buff marble
- 2 m blue marble
- 15 m buff marble
- 3 m silver grey phyllite and green sandstone
- 0.5 m brown sandstone
- 3 m purple grey phyllite
- 3 m black quartzite
- 3 m white quartzite and green phyllite, finely interbanded
- 1.5 m conglomerate and grey quartzite in the upper part, green and red purple phyllite in the lower part
- 7 m grey and white quartzite
- 3 m grey psammitic phyllite
- 0.5 m grey quartzite
- 5 m phyllite, purple red at top, green at base
- 3 m quartzite
- c. 15 m black quartzite and phyllite (in scree)
- c. 130 m silver grey phyllite with quartzite lenses
- 10 m blue banded marble
- 3 m phyllite
- 5 m quartzite

2 m buff banded marble
4 m dark blue marble
0.5 m buff quartzite
4 m blue grey marble
1 m silver grey phyllite
3 m banded blue grey and buff marble
0.5 m quartzite
0.5 m silver grey phyllite
10 m blue grey banded marble

West of Krossfjorden apparently only the lower 400 m of the formation occurs:

50 m of calcite marble (= Mitra Marble, TVETEN 1971)
C. 230 m of mica schist and phyllite, with some quartzite. Chloritoid occurs in the uppermost part and some minor inpersistent dolomite beds in the lower part.
C. 150 m of dolomite marble (= Willeberget Dolomite), with c. 1 m of quartzite in the middle part.
Max. 3 m of local basal conglomerate beds, recorded at two places in the southern part of Mitrahalvøya. The conglomerate consists of 7 or 8 beds, each 30–50 cm, of poorly sorted quartzite pebbles, 1–25 cm² in section.

Below this conglomerate, calcareous beds are fewer, and mica schist becomes predominant:

GEE and HJELLE 1966, this paper (Signehamna Formation)

2000–2500 m of pelitic and psammitic schist with subordinate quartzites and a few minor marble horizons. Small concordant lenses of amphibolite is recorded from two places.

TVETEN 1971

700 m of mica schist, with subordinate quartzite and minor amphibolite. Chloritoid occurs locally in the uppermost part.

Several occurrences of sedimentary structures, as quartzite to pelite transitions and current/ripple bedding, suggest that the beds around Krossfjorden in general are not inverted. Three observations, c. 2 km apart, in southern Mitrahalvøya, indicate a direction of transport from east to west. In the northern part of Mitrahalvøya and at the eastern shores of Lilliehöökfjorden, the lower mainly pelitic rocks of the Signehamna Formation makes up most of the area. A wedge of mica schist also extends towards the NNW from the inner part of Blomstrandbreen. A c. 25 m marble+quartzite horizon which crops out at various places from Førstebreen to Fjerdebreen suggests the pelitic sequence to extend almost to Nissenfjella, although the pelitic rocks are more gneissose in the northern part. Calcareous lithologies are very scarce in the inland area between Magdalenefjorden and Førstebreen.

Approaching the areas north and east of Sjettebreen, the general structure becomes more complex and the lithologies give evidence of a much greater mobility during deformation. Permeation of quartz-feldspathic material as schlieren or dyke and sheets increases towards the migmatites, and an estimation of the thickness of the Nissenfjella Formation (below Signehamna For-

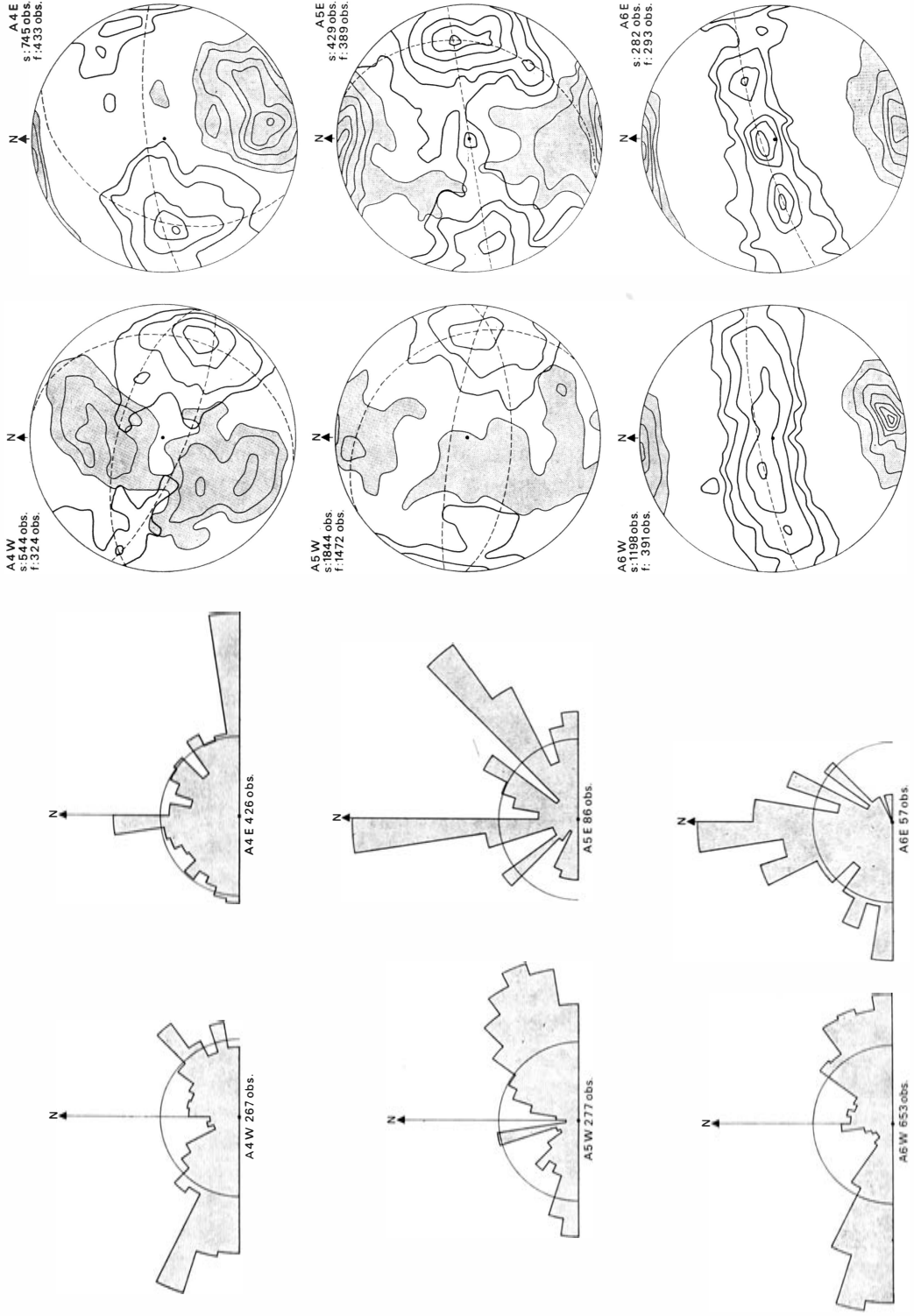


Fig. 4.

mation, GEE and HJELLE 1966) is doubtful. However, with an average plunge of mesoscopic fold axes of c. 12° (Fig. 4), at least some of the rocks at the south side of Sjettebreen could represent stratigraphic levels c. 2000 m below the Signehamna Formation. The rocks include pelitic schists, amphibolites and various feldspathic lithologies. Amphibolites are most conspicuous in the western parts of Nissenfjella and Klingenbergfjella. The feldspathic rocks are mainly concordant, foliated aplites, and soda feldspar augen-gneisses transitional to mica schist.

North of Sjettebreen the south plunge of folds alternates with north plunge, and areas occur with apparently younger beds as the calcareous and pelitic beds in outer Magdalenefjorden, in parts of Danskøya and Amsterdamøya, and in western Vasahalfvøya. These beds are tentatively correlated with the lower part of the Generalfjella Formation and the upper part of the Signehamna Formation.

In outer Magdalenefjorden distinct layers of marble occur in Knatten, at the northern entrance of the fjord:

100 m layered gneiss
10 m consisting of 3 marble layers in layered and fine grained biotite gneiss
80 m siliceous layered gneiss
100 m pelitic layered gneiss
50 m siliceous migmatite
several m unexposed
several m of layered gneiss with one marble layer

In the western part of Vasahalfvøya the estimated thickness of the meta-sediments is about 5200 m, presuming there is no repetition by folding or thrusting.

In the area north of the innermost part of Kongsfjorden, a marble occurs, which in its southern part is in fault contact with mica schist, and in the north wedge out in grey granite and migmatite. Occurring as an isolated body in the migmatite, this marble is thought to belong to the lower part of the Generalfjella Formation. North of the marble a persistent zone of marble relics continues northwards in the migmatite as far as to the north coast. That all these marbles represent associated sediments of about the same horizon is testified to by their continuity on a regional scale parallel to the regional strike of the non-migmatitic metasediments. If a $5\text{--}10^\circ$ south plunge of the mesoscopic fold axes in this area (Fig. 4) also is representative of the regional folding, successively lower beds should be expected northwards. The persistency of the north-south trending zone of marble relics could be explained as being parts of a regional synclinorium of considerable amplitude, probably more than 5000 m.

←

Fig. 4. *Right: Mesoscopic fold axes and associated lineations (shaded, f) and bedding, layering and gneissosity (thick lines, s). Mainly bedding in the south (map sheet A6), mainly layering and gneissosity in the north (A4 and A5). Stereographic projection, lower hemisphere. Contours for 1, 2, 4, 6, 8, 10% per 1% area.*

Left: Directions of vertical or almost vertical joints ($\geq 70^\circ$) Demicircle indicates 5% of total counts. For location of the areas see Fig. 2.

The mica schist and associated quartzites close to the head of Liefdefjorden underlie marble beds, which are in fault contact to the Downtonian beds in the east. The lithology of the schist-quartzite-marble unit, and the mutual stratigraphic relations, suggest a correlation with the upper part of the Signehamna Formation and the lowermost part of the Generalfjella Formation.

Thus rocks of possible similar stratigraphical position occur both along the west coast and in the eastern inland area as moderately deformed metasediments, as gneisses, or as more or less digested relics in migmatitic rocks. It is therefore likely that in northwest Spitsbergen differences in metamorphism or in degree of deformation do not necessarily establish relative age.

In general pelitic schists are the far most common sedimentary relics in the gneiss migmatite areas, and it is suggested that the bulk of the relics originated from the Signehamna and Nissenfjella Formations, and that calcareous rocks from the lower part of the Generalfjella Formation were preserved only in synclinal areas.

Petrography

METASEDIMENTS

Calcareous rocks.

The main upper marble in the Generalfjella Formation continues south of the present map area to Blomstrandhalvøya. The calcareous rocks here are mostly rather inhomogeneous and the grain size varies from less than 0.1 mm to 10 mm, occasionally veins occur with large calcite crystals up to 10 cm (SIGGERUD 1963). The fine grained varieties often show signs of being on the point of recrystallization, i.e. larger crystals starting to grow in the fine grained groundmass. Extensive brecciation has taken place with subsequent calcite cementation. Some quartz can be found as layers of well rounded grains 0.1–0.3 mm.

In Generalfjella the marble units are rather homogeneous in texture, but the weathering colour varies considerably from light grey and yellow through buff to greyish blue and dark blue.

West of Krossfjorden the Willeberget Dolomite probably represents the lowermost part of the Generalfjella Formation. This dolomite generally possesses an even grained polygonal texture 0.06–0.1 mm, indicating a complete recrystallization. Occasionally quartz rich layers occur of a maximum thickness of 2 cm. The most completely recrystallized parts of the dolomite almost invariably contain small amounts of evenly distributed muscovite, and some 0.5 cm clusters containing quartz, minor actinolite and feldspar, and occasionally minor enstatite. The upper part of the dolomite which is flexure folded and imbricated, is of grey colour with accessoric magnetite, whereas the lower part, which often shows plastic deformation, is mainly of yellow or reddish colour, and contains accessoric hematite and pyrite.

The apparently uppermost lithologic unit west of Krossfjorden is the Mitra Marble which crops out in the southwestern Mitrahalvøya, and which constitutes the mountain Mitra. This calcite marble is of light grey colour, and

often contains reddish calcite veins. The texture is generally somewhat deformed polygonal, with grain size 0.1–0.3 mm. Occasionally breccias accompany overturned folds of c. 20 m wavelength, the breccia fragments being up to 20 cm across, with white or reddish matrix. A weak foliation occurs in the marble due to a minor content of carbonaceous matter between the calcite grains. Up to c. 5 vol. % of recrystallized quartz is present.

In the gneiss and migmatite areas the calcareous beds are frequently split into numerous layers, pods, and lenticles which are very variable in thickness and persistence downstrike. The layers are commonly boudinaged and less than 10×100 m in outcrop. In the migmatites the calcareous inclusions are the most angular and show the greatest tendency to retain an orientation parallel to the original compositional banding. A banded nature of many of the marbles, with intercalated quartz rich horizons and pelitic schists, suggests that these rocks were primary limestones with sandy or silty, and pelitic beds. The siliceous layers are commonly accompanied by layers rich in diopside, wollastonite, vesuvianite and grossularite. Other common skarn minerals are scapolite, sphene and hematite. Some marbles have small graphite flakes among the calcite grains.

Some boudinaged layers of marble in mica schist and layered gneiss are recorded to continue into areas of highly mobilized rocks as migmatite and syntectonic granite where they wedge out, leaving pockets, layers or vague zones enriched in skarn minerals. From a small island southeast of Fugløya and from the western entrance of Fuglefjorden, a skarn paragenesis of forsterite, phlogopite, and spinel is recorded accompanying marbles in granitic migmatite.

East of the innermost part of Magdalenefjorden, in the migmatite area along Miethebreen, a shear zone occurs with skarn lenses associated with a pink aplite, which possibly is related to the red post-tectonic Hornemantoppen granite further east.

In the migmatite areas agmatitic or lensoid amphibolites may be followed for several kilometres, e.g. in the Magdalenefjorden–Danskøya area and in western Vasahalvøya. Frequently the amphibolites here are found in the vicinity of marble bands and migmatite may give evidence of amphibolitisation of marble.

Phyllite.

The occurrence of phyllite is mainly confined to the Generalfjella Formation in the southern area. The phyllite rocks are fine to medium grained with a greenish or greenish grey or silvery grey colour, and with a lustrous, sheeny look. The relatively low grade appearance of these rocks in outcrop is confirmed by microscopic examination; the main constituents are quartz, sericite, chlorite, and in the lower part of the formation occasional minor biotite. With increasing amount of quartz the phyllite grade into psammite and quartzite.

In Mitrahelvøya parallel oriented crystals of chloritoid are recorded in an intensely folded quartz-muscovite-chlorite phyllite which is thought to belong to the lower part of the Generalfjella Formation. The chloritoid occurs in the vicinity of thrust or fault zones, and the amount varies from 3 to 30 vol. %

Optical data from one thin section: α 1.714–1.720, β c. 1.72, γ 1.73; pleochroism α yellow green, β and γ pale blue green; mainly optical positive. Hematite occurs in small quantities in the phyllite.

Mica schist.

From the upper part of the Signehamna Formation and downwards biotite-muscovite schist gradually becomes more common and garnet begins to appear. Most of the rocks in the middle and lower part of this formation are schistose muscovite-biotite-quartz assemblages with small amounts of chlorite and plagioclase, and are usually garnetiferous. The original compositional banding and a weak graded bedding structure is sometimes preserved in the more psammitic schists.

In Mitrahalvøya the most common type of schist contains 50–70 vol.% mica with variable proportions of muscovite and biotite. Muscovite, which is the major platy mineral, is examined in 10 specimens. $2V_x$ vary between 43 and 45°, and N_γ from 1.594 to 1.600. The structure is always of the 2M type.

Biotite show some variation, the most common pleochroism is: α colourless β and γ red brown. The red colour is most pronounced in biotites from the northern part of Mitrahalvøya. N_α varies between 1.56 and 1.58, independent of geographical distribution. Two generations of biotite can be distinguished in some thin sections, the older one being somewhat chloritized.

The garnets in the Mitrahalvøya schists show no visual zoning, and there is little variation in the refractive index. Ten garnets from various schists sampled north of the garnet isograd show $N = 1.797$ – 1.808 ; in the same garnets the unit cell (a) varies from 11.588 to 11.595 Å, with a/N 6.41–6.45, average 6.43.

Feldspar always occurs in small amounts, especially south of the garnet isograd. Almost all feldspar recorded are acid plagioclases, two thirds of these show an An content lower than 15%. Only a few schists contain potassium feldspar, mainly as clastic grains in quartzitic schists, in three cases post-tectonic orthoclase crystals were seen as joint fillings.

In general the amount of newly formed feldspar increases towards the north, 9 vol.% is found in a mica schist c. 5 km NNW of Signehamna. In the same schist 1–3 cm crystals of andalusite accompanied by some sillimanite are found bordering a head size pocket of post-tectonic quartz.

Only few details are recorded from the schists at the eastern shore of Krossfjorden and from Kong Haakons Halvøy, however in lithology and general appearance they seem to closely resemble the schists of the Signehamna Formation.

Although the mica schists retain their NNW–SSE trend and their clearly metasedimentary appearance in the area between Mitrahalvøya and Sjetebreen (Fig. 5, upper part), the texture of the rocks is generally coarser, and segregation and recrystallization of felsic material commence. Interfingering with, and transition to feldspar porphyroblastic schists occur, the newly formed feldspar being most abundant in schists with impure quartzite and psammitic bands.

GNEISS AND MIGMATITE

Layered gneiss

In composition and mesoscopic structure, the layered gneiss lies between mica schist with quartzitic beds, and migmatite. Sporadic lenses and pockets of igneous textured feldspar rich rocks, less than half a meter across, occur sometimes in the mica schist as forerunners to layered gneiss. Although the layered gneisses are characterized by a generally higher proportion of feldspar than the mica schist group, the main part of them are of pelitic composition (Table 1). Compared to the mica schist and its felsic bands, the grain size has increased in the layered gneiss and the difference in composition between the mafic and felsic bands is smaller. The development of plagioclase porphyroblasts, both in the light bands and, more scattered, in the pelitic bands, and the common occurrence of accessory cordierite and sillimanite, are other characteristic features.

The layered gneiss has developed a distinct axial plane gneissosity, and most of the detailed primary compositional banding is usually lost. However, some granitic layers are evidently feldspathized beds of primary impure quartzites, and alternations of layered gneiss and migmatite is considered to reflect the original lithological successions of sediments. The thickness of the individual layers within the gneiss varies from a few cm. to several metres, the thickest can be followed along the strike direction for more than one hundred metres. The felsic layers are commonly boudinaged.

Typical compositions of layered gneiss are (C. modal %):

Mafic layer: Quartz 30, K-feldspar 5, plagioclase 35 (An 20–30),
biotite 25, garnet + cordierite + sillimanite 2.

Felsic layer: Quartz 35, K-feldspar 15, plagioclase 35 (An 10–20),
Biotite 10.

Corundum is found locally in layered gneiss east of Waggonwaybreen, and corundum and spinel in some of the gneisses east of Smeerenburgfjorden. The plagioclase porphyroblasts apparently developed during the last stage of the formation of the layered gneiss. Both the biotite and the plagioclase porphyroblasts are arranged parallel to the plane of gneissosity. In the west coast area, between Førstebreen and Tredjebreen, transitions from mica schist with quartzitic beds to layered gneiss occur frequently. While the lithology of the various layers clearly depends on the primary composition of the mica schist beds, the degree of recrystallization and mobilization is related to both the intensity of deformation and the stratigraphical depth as well as to the primary composition. Depending on the initial lithology, various layers occur:

Dark biotite mica schist → dark pelitic gneiss with more or less scattered plagioclase porphyroblasts.

Pure quartzite → coarse glassy quartzite, mainly less than 10 cm thickness.

Feldspathic quartzite → coarse K-feldspar – quartz – albite gneiss with minor muscovite, sometimes of pegmatitic appearance and with weak foliation.

Table 1.

Modal and chemical analyses.

Modal and chemical analyses of northwest Spitsbergen rocks

	1	2	3	4	5	6	7	8	9	10
Quartz	36.7	28.7	36.2	26.0	33.8	26.5	22.9	0.9	34.0	32.8
K-feldspar	22.4	16.4	22.0	4.0	3.7	2.1	2.2	-	22.1	25.5
Plagioclase	33.4	43.9	31.9	33.8	34.1	54.3	54.6	28.5	31.0	27.9
Muscovite	-	-	4.7	-	0.8	-	1.4	-	4.1	2.5
Biotite	6.7	10.6	3.2	36.0	26.5	14.4	12.2	-	7.4	8.1
Chlorite	0.4	-	0.9	-	-	2.4	0.1	2.4	0.4	1.2
Amphibole	-	-	-	-	-	-	5.3	60.7	-	-
Epid.zois.	-	-	0.3	-	0.6	-	-	0.3	0.8	1.2
Apatite	-	-	-	-	-	-	0.2	-	tr.	0.3
Sphene	-	-	0.1	-	-	-	0.4	3.1	-	tr.
Ore mins.	0.5	0.3	0.5	0.3	0.4	0.3	0.7	4.2	0.2	0.4
Garnet	-	-	-	tr.	tr.	-	-	-	-	-
	100.1	99.9	99.8	100.1	99.9	100.0	100.0	100.1	100.0	99.9
%An in plag.	(15)	(15)	(21)	(35)	(15)	(20)	(30)	(50)	(33)	(30)

SiO ₂	70.70	69.15	70.80	67.70	65.15	63.60	64.71	56.70	69.99	69.93
TiO ₂	0.22	0.47	0.18	0.73	1.20	1.15	0.59	2.48	0.40	0.38
Al ₂ O ₃	17.00	16.32	16.24	14.44	15.32	16.97	16.18	11.87	15.24	15.31
Fe ₂ O ₃	0.61	1.21	0.29	2.39	2.57	1.91	0.43	4.24	0.09	1.20
FeO	1.29	2.52	1.22	3.23	4.74	4.31	3.97	2.31	2.79	1.56
MgO	0.19	0.69	0.34	1.86	1.74	1.74	2.14	5.77	0.87	0.81
CaO	1.53	1.61	1.31	3.68	1.88	2.68	4.24	9.53	2.57	2.45
Na ₂ O	3.45	3.31	3.50	2.89	3.21	3.35	4.33	2.10	3.44	3.31
K ₂ O	4.20	4.68	4.03	2.05	2.48	3.00	2.85	1.60	3.49	3.90
H ₂ O	±0.36	±0.75	n.d.	±0.80	±1.58	±1.64	+0.74	±2.21	+0.79	+0.74
P ₂ O ₅	0.05	0.12	0.06	0.11	0.07	0.08	0.07	0.30	0.08	0.05
	99.60	100.83	97.97	99.88	99.94	100.43	100.25	99.11	99.75	99.64

No.inTable	NP No.	°N	°E	Location	Type
1	66 YO 217	79°40.8'	11°24.5'	SW Vasahalvøya	Fine grained gneiss, felsic
2	66 YO 52-3	79°45.8'	11°17.4'	NW "	Layered gneiss, felsic
3	66 Hj 26 B	79°45.3'	10°40.8'	W Amsterdamøya	Migmatite, felsic part
4	66 YO 178-2	79°43.1'	11°12.7'	W Vasahalvøya	Fine grained gneiss, mafic
5	66 YO 449	79°41.3'	11°14.5'	" "	Layered gneiss, mafic
6	66 YO 394	79°43.6'	11°22.4'	" "	Migmatite, mafic part
7	66 Hj 82 A	79°43.1'	10°51.5'	N Danskøya	Amphibole bearing gn.
8	66 YO 284-2	79°45.1'	11°13.9'	NW Vasahalvøya	Amphibolite
9	64 Hj 230 A	79°38.9'	10°47.7'	SW Danskøya	Granitic dyke
10	64 G 118	79°39.0'	11°39.8'	S Vasahalvøya	Hornemantp.monzogranite

Nos. 1-6, and 8 analyst Y.Ohta

No.7 analyst J.Røste, Norges geologiske undersøkelse

Nos.9 and 10 analyst P.R.Graff, Norges geologiske undersøkelse

Pelitic feldspathic quartzite → homogeneous augengneisses of quartz diorite composition.

Some felsic layers are evidently mobilized, and granitic material penetrate the surrounding layers concordantly parallel to the axial plane gneissosity. Occasionally dark micaceous lenses are entirely enclosed by granitic materials

Migmatite.

When the degree of feldspathization and mobilization increases in the layered gneiss, the granitic material begins to penetrate the layers and the axial planes of the small folds discordantly, and the layered gneiss grade into migmatites with pygmatic granitic veins. As the mobility increases away from the layered gneiss, the migmatite metastere loses orientation related to earlier structures and the compositional banding becomes more diffuse.

The inclusions in the migmatite, on the basis of composition, fall into four groups — pelitic, psammitic, calcareous and amphibolitic. Pelitic masses are most typical in the migmatites and particularly in those not containing marbles. They are usually smaller than other coexisting masses and less angular.

Amphibolite masses (other than those of skarn association) are frequently found in migmatites in the vicinity of the marble bands e.g. in Danskøya. They give evidence of amphibolitization and induced foliation prior to migmatization. Hornfelsing of these masses is usual, with margins of c. 1 cm around the xenoliths. Pyroxene-bearing xenoliths are found on each side of the Horneman granite north of Markbreen.

The migmatite inclusions occur in different stages of assimilation, some rotated and some in parallel orientation with the country rock. Several observations of transitions from agmatitic to streaky and nebulitic migmatite, suggest a progressive assimilation process, beginning with formation of agmatitic migmatite by injection and ending with nebulitic migmatites and ghost granites.

The metastere often shows preservation of folds and a penetrative schistosity which clearly developed prior to the superimposition of hornfelsed margins during migmatization. This suggests that the migmatization occurred after the first main folding and regional metamorphism of the rocks.

The migmatites are rarely homogeneous over long distances, and discontinuous masses of mica schist, layered gneiss, amphibolite and marble occur as bodies, c. 10–200 m across within the migmatite. Persistent rocks as amphibolite and marble show least alteration, and they may be followed as broken up and boudinaged layers for several hundred metres.

There is much variation in the composition of the metatect, particularly adjacent to the metastere, however the bulk composition is that of a granite or granodiorite. Although locally massive, these rocks often possess a flow structure and are somewhat foliated concordant to the metastere throughout most of the area. Biotite is the dominant mafic constituent of the metatect, however when the migmatite has originated from gneisses with amphibolitic and/or calcareous layers hornblende frequently occurs together with biotite.

Typical compositions of migmatites are (C. modal %):

Metaster (pelitic type): Quartz 25, K-feldspar 5, plagioclase 45 (An 20–30) biotite 20, garnet, sillimanite, cordierite 2.

Metatect: Quartz 30, K-feldspar 17, plagioclase 40 (An 15–25), biotite 8.

AMPHIBOLITE

Concordant lenses and bands of amphibolite have been recorded at four places in Mitrahalvøya:

- (1) C. 0.5 km west of nordre Diesetvatnet, as several lenses of c. 0.5 m thickness, extending c. 30 m north-south. The main minerals of the amphibolite are hornblende, quartz, biotite, chlorite, magnetite and apatite. Plagioclase (An 10–20) occurs as an accessory mineral. The hornblende has $Z \wedge c = 18^\circ$ and $2V = 74^\circ$ with pleochroism: α brownish yellow, β and γ blue green. $N_y = 1.635$.
- (2) C. 0.5 km north of Fridtjovneset, Signehamna. This amphibolite contains garnet; the hornblende closely resembles that in No. 1.
- (3) In the coastal section north of Bouréeneset. Several small lenses of less than 0.3 m thickness.
- (4) Below the marbles on the northwest side of Scoresbyfjellet. Only one small amphibolite lens was recorded.

Further north amphibolites are recorded both in supracrustal rocks and in gneisses and migmatites. Conspicuous amphibolites occur in the western and central Nissenfjella, as lenses and boudins up to 5 m across. These bodies in general conform to the regional structure, and their planar structures are parallel to the lithologic layering and/or gneissic structure of the enclosing rocks, and like them they are locally folded. Foliation is highly developed in the thinner amphibolite layers, whereas in the central parts of some of the larger bodies it is indistinct and the rocks appear massive. Where the amphibolite shows little sign of shearing, garnet is stable; retrogression of both garnet and biotite is to be found in most outcrops.

Similar amphibolite and amphibolite interbanded with pelitic rocks are also contained in the core of the antiform in the eastern part of Hestekoen, close to the head of Liefdefjorden.

The zone of amphibolite-bearing migmatites and gneisses extending N to NNW from the central Magdalenefjorden area is, as a whole, slightly oblique to the general NNE structural trend of the surrounding rocks, however it broadly coincides with the regional trend of the marbles and amphibole-bearing biotite gneisses in Danskøya and Amsterdamøya, which probably represent the trend of the strata of the pre-metamorphic rock sequence. This suggests that the amphibolites originated from rocks which primarily paralleled the supracrustal beds as volcanics or as sills, and that the NNE layering and foliation which is especially well developed in the pelitic rocks, to a great extent were produced by tectonic transposition and intercalation during in-

tense deformation. The absence of primary structures in the amphibolite makes it difficult to decide whether the amphibolite were originally dykes or extrusive sheets.

GRANITIC ROCKS

Weakly foliated grey granitic rocks

Nearly all of these rocks are closely associated with the migmatites. The contact against the surrounding rocks are mainly gradual, although there are some cross-cutting contacts. In Danskøya, vaguely defined dome structures are observed with central bodies of massive or weakly foliated granitic rocks containing small and scattered shadowy metaster. The granitic rocks grade outwards into migmatite and layered biotite gneiss. This evidence shows that these granitic rocks could be considered as “mature” migmatites with a high metatect/metaster ratio.

Several granite bodies of considerable size occur within or adjacent to the migmatite area:

- (1) The granite east of Krossfjorden–Möllerfjorden. Fault contact occurs against the metasediments in the west, and transitional or intrusive contacts towards the eastern migmatite. The composition is mainly monzogranitic.
- (2) The dyke-like granite extending northwestwards from Sjettebreen. This is a monzogranite of medium grained homogeneous texture, and with a faint foliation. Inclusions are rare except when approaching the migmatite in the north and east. Although migmatite varieties occur within this granite, its general appearance shows many similarities with a discordant dyke intrusion.
- (3) The granitoid at the middle and inner part of Magdalenefjorden. With a relatively low content of potassium feldspar, this is rather a granodiorite with transitions to quartz diorite than a granite. The texture is medium grained homogeneous, and the main body occurs as a wedge-shaped intrusive layer sub-concordant with the surrounding gneisses, the eastern contact being relatively sharp. A faint, but consistent foliation also conforms with the structures outside.
- (4) The plagioclase porphyroblastic granite east of Smeerenburgfjorden, extending southeastwards from Slaadbukta. This is a monzogranite with transitions to granodiorite, and the body fills the core of a synform structure in the metamorphic rock. In the northwest sharp drag contacts occur against the metamorphic rock, while in the east the granite has a sharp but intrusive contact to layered gneiss.

Below the general characteristics of the weakly foliated grey granitic rocks are summarized:

- (a) All locations of these rocks are in or adjacent to the migmatite area.
- (b) The lithology of the granites is about the same as in the migmatite metatects and the granitic layers in the layered gneiss.

- (c) Structurally the granitic rocks conform to the flow structures and the open folding of the migmatites rather than to the older tight isoclinal folding of the gneisses.
- (d) The contact to migmatite is mainly transitional, to the gneissic rocks often sharp, with signs of emplacement by liquid intrusion.

Thus, the observations suggest a late tectonic origin of the granites, closely connected in time to the development of the migmatites. The granites may be regarded as the mobilized and homogenized end products of the granitization of the regional metamorphic rocks.

Grey granitic dykes

The grey granitic dykes occur in two generations:

- (1) When approaching the migmatite areas, e.g. just south of Sjettebreen, aplite and pegmatite sheets begin to appear in the metasediments. Their positions are mainly concordant, but some are also discordant to the foliation. These latter cut only the metasediments, and belong to the initial stage of agmatitic migmatites. However, dyke activity is recorded as far southwest as in Schottfjellet and Chunfjellet (west of Lilliehöök-fjorden), where scattered light coloured intrusive sheets of 3–5 m thickness traverse the mica schist obliquely with a c. 40° dip towards the ENE. The dykes are folded, and were emplaced before the main folding ceased in this area.
- (2) The predominating dykes are those cutting both metasediments, grey late tectonic granites and the migmatites. They include aplitic and pegmatitic varieties and are almost wholly confined to the migmatite area. These rocks occur as sharp cut dykes, and only in rare cases are they transitional to larger masses of grey granite. The bulk of the dykes, therefore, was intruded after the consolidation of the migmatites.

The granitic dykes referred to below, all belong to group (2).

The composition of the aplitic dykes is rather homogeneous, monzogranitic to granodioritic, with a compositional range between that of the migmatite metatect and biotite gneiss, i.e. it approximates the bulk composition of a common migmatite (Table 1, No. 9). The aplitic dykes also show compositional similarities with the Hornemantoppen monzogranite.

The pegmatite veins and dykes are usually more felsic than the aplitic ones, and when the two types occur together, the pegmatites are commonly the younger.

In Danskøya and Amsterdamøya a comparison between the directions of dykes and joints, showed a close similarity, indicating a preferred direction of intrusion parallel to the jointing. Of the pegmatite dykes in this area only c. 10% exceeded 1/2 m in width, compared to c. 65% of the aplitic dykes.

Regarding the granitic dykes in northwest Spitsbergen as a whole, there appear to be two main maxima of directions: c. 90° and c. 150°, i.e. close to the directions of cross joints which could be expected in the mainly N or NNE trending zone of gneiss and migmatite.

Post-tectonic red granite.

The Hornemantoppen granite forms an outcrop pattern, elongated NNW, within the migmatite area. The granite is centered on Hornemantoppen, and the ridges and nunataks cover an area of more than 150 km², including ice cover. Within this area are found a variety of granites, the far most conspicuous of which is a coarse to medium grained red granite.

Except for minor border facies rocks the pluton is a nonfoliated mass of monzogranite composition, and the rocks contrast markedly with the adjacent metamorphic rocks into which they are intruded. The texture is hypidiomorph equigranular, sometimes with transitions into more porphyritic varieties. Potash and plagioclase feldspars occur in about equal proportions, potassium feldspar as large grains up to 2 cm across, plagioclase somewhat smaller. The remainder is mostly made up of aggregates of quartz grains and more or less evenly distributed biotite, which is the dominant ferromagnesian mineral in all examined specimens. In five specimens of the most common variety, the modal range was: Quartz 29.7–34.7%, potassium feldspar 24.6–26.3%, plagioclase (oligoclase-andesine) 21.7–30.0%, biotite 4.1–8.1%, and chlorite 0.7–2.6%. The plagioclase grains are often strongly decomposed to sericite and zoisite. When fresh, the plagioclase often shows grading from andesine cores to sodic oligoclase rims, with strong oscillatory zones. Considering the volume, an average plagioclase composition of An 25–35 is common. Fine grained, late crystallized potassium feldspar and quartz fill interstices between other minerals.

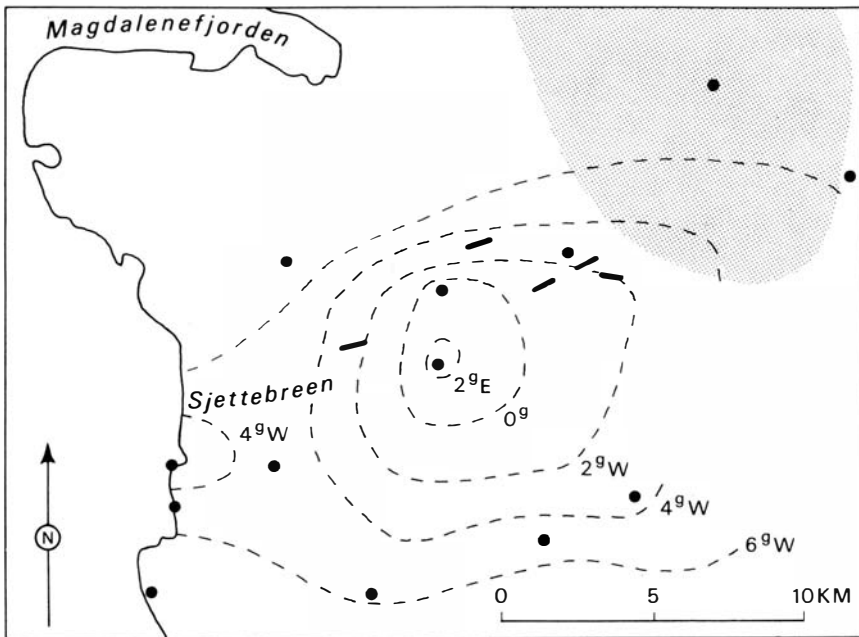


Fig. 5. Magnetic declination west of Sjettebreen. (Difference in mirror compass bearings and directions, measured on 1: 100 000 map, Magdalenefjorden, Norsk Polarinstitut 1969). Dots: observations. Thick lines: basic post-tectonic dykes. Shaded: post-tectonic granite.

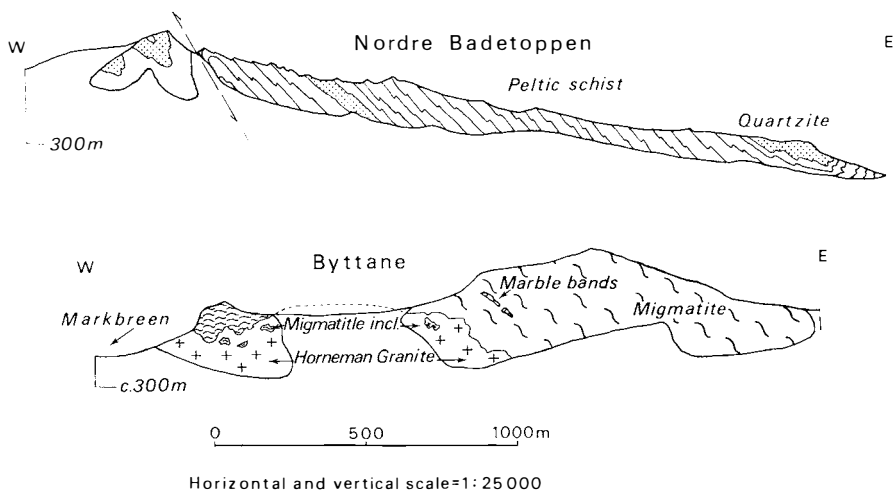


Fig. 6. Typical outcrops, in the mica schist area (upper) and at the northern part of the Hornemantoppen granite (lower).

Common accessory minerals are sphene, apatite, pyrite and magnetite. The magnetite content is generally slightly higher than in the surrounding metamorphic rocks, and a magnetic anomaly in central Albert I Land, with amplitude c. 200 γ might be related to this higher magnetite content (ÅM 1975). Magnetic declination measurements made by the author in 1975 show anomalies east of Sjettebreen (Fig. 5). In this area, several post-tectonic basic dykes are recorded, and it is considered possible that the anomaly arises from basic differentiation products from which the dykes might have originated, and which is related to the Hornemantoppen granite intrusion. The basic dykes are nowhere seen to cut the granite.

Away from the contact area, the granite contains a very subordinated amount of xenoliths, typically rounded and smaller than $\frac{1}{2}$ m in diameter. Closely associated with the granite are pink or red aplites and pegmatites which are recorded from several contact areas. The latter are typically quartz potassium feldspar intergrowths with muscovite and occasional pyrite. The aplites contain up to 10% biotite. Pink aplitic veins and dykes, up to 3 m thick, occur also further west, in the east side of Alkebreen. The dykes, which seem to be closely connected with the emplacement of the granite are associated with skarn masses.

The Hornemantoppen granite is eroded to a depth of more than 700 m to form a rugged topography. The depth of erosion indicates that the granite is probably not a laccolith or sill-like mass. The foliations marginal to the granite, which generally dips away from the contact, seem to confirm the suggestion of a batholithic intrusion.

The main part of the granite is intruded only into migmatitic lithologies, and occurs as a north-south elongated core in the migmatite domain. In all examined localities the granite contacts were unshered and intrusive, the granite prominently cutting the foliation in the migmatite rocks. Hornfelsing

in the contact rocks has not been observed, however it should not be expected either, considering the near magmatic conditions during migmatitisation.

In the northern part of the granite, north of Markbreen, where the roof zone is well exposed (Fig. 6, lower part), large xenoliths and large projecting masses of horizontally foliated migmatite gneiss from the overlying country rock are immersed in granite, and apparently the surface of erosion here corresponds closely with the roof of the batholith.

As the Hornemantoppen granite injected after the main metamorphism and deformation, it could (1) have been emplaced into a pre-existing major antiform or (2) have arched the metamorphic rocks into an antiform, which was possibly initiated already during the formation of the migmatite. The homogeneity of the granite and the relative lack of xenoliths seem to favour the latter. So do also the fact that the migmatite — metasediment contact south-west of the granite roughly parallels the granite contact rather than the general north-south structures in this area. Bowing aside of the country rock is also suggested by occasional anomalous north plunges of fold axes north of the granite, and a general increase in the dip of the migmatites as the contact with the granite is approached from the west.

Structure

In general the structure is dominated by the main northwest anticlinorium of NNW–SSE trend. This main open structure is apparently superimposed on most other structures, and therefore considered to be of relatively late origin, and related to migmatite upwelling and the emplacement of the Hornemantoppen granite.

The simplest structures appear in the areas of relatively well preserved metasedimentary rocks, particularly those east and west of Krossfjorden, Lilliehöökfjorden and Möllerfjorden. Three main phases of deformation are distinguished here: F0 early, weak folding of north-south trend, with a corresponding schistosity, S0 parallel to the original compositional banding of the metasediments. S0 is folded by small isoclinal folds, F1, which themselves are refolded by ubiquitous minor isoclinal folds, F2, with a common wavelength and amplitude of c. 1 dm to c. 5 m. In suitable lithologies, as pelitic rocks, this folding causes development of a pronounced axial plane schistosity, S2. These minor folds are closely related to large open synforms and antiforms with wavelengths up to a few kilometres, and with axial trend paralleling those of the smaller folds.

In general the F1 and F2 folds and corresponding lineations are homoaxial, with a shallow S to SSE plunge (Fig. 4), but local plunge culminations occur in the Krossfjorden area, around Ebeltoftthamna and Fjortende Julibukta, which might indicate a later, F3 deformation of possible NW–SE trend. The relatively simple pattern of homoaxial F1 and F2 folds are more or less persistent throughout the areas of metasediments, in the coastal area northwards to Sjettebreen, in the east as far north as to Liefdefjorden.

With increasing mobilization the patterns become more complex:

- (1) Layered gneiss forms in close connection with development of the F2 axial plane schistosity and gneissosity of tight minor isoclinal folds with subhorizontal axes. The F2 trend is, at least northwards from Sjettebreen, around NNE–SSW. As in the southwest, the F2 folds are here both of large open and minor tight type. A well exposed large open F2 synform appears in Knatten, at the northern entrance to Magdalenefjorden. In the western Nissenfjella a comparison of limb lengths indicates that the minor F2 folds are related to a major antiform closing up westwards.

The F1 and F2 folds are largely confined to the metasediments and their corresponding inclusions in the migmatites, however, adjacent to prominent masses of inclusions, also the mobile phase of the migmatite may show a foliation paralleling the lithological banding and/or the S2 of the metasedimentary inclusions.

- (2) Migmatite and late tectonic granitic rocks develop in connection with a late, open folding, trending NW–SE to W–E with relatively steep plunges of axes, which deforms the S2 schistositities and gneissositities. Whereas the granitic layers in the layered gneiss preferably develop as concordant seams at the crests of small F2 folds, the migmatites and late tectonic granitic rocks are frequently contained in the cores of F3 open antiforms and dome-like structures. A foliation or plastic flow structure may exist in these rocks, independent of the structure of the xenoliths and therefore clearly induced during and after the F2 folding, which is preserved in the xenoliths. Thus the migmatization occurred after the F2 folding of the metasediments.

Most of the fault lines have a NNW–SSE trend, less pronounced are directions around WNW–ESE and WSW–ENE. A major fault with downthrow to the east occurs along Monacobreen and the eastern shore of Raudfjorden, separating the metamorphic rocks from the eastern lower Devonian beds. Further south small occurrences of lower Devonian rocks have also been found at Lovénøyane (GJELSVIK 1974), and in Løvlandfjellet, north of Blomstrandbreen, in all localities preserved due to faults of NNW trend. Devonian strata are not found west of Krossfjorden-Lilliehöökfjorden, however, a red weathered soil which occurs in the upper part of Mitra may suggest a pre-Devonian erosional surface here.

The majority of the faults on the north side of Kongsfjorden are not directly related to the local folding, but parallel the main anticlinorium axis, with a general downthrow to the west, suggesting a relationship to this regional structure. In the Generalfjella synform the upper marble beds are in tectonic contact with pelites of lower stratigraphic position, the thrust plane having a 30° easterly dip near Krossfjorden.

West of Krossfjorden, Lilliehöökreen, and the Hornemantoppen granite, thrust faults with easterly dip prevail, which are themselves often horizontally

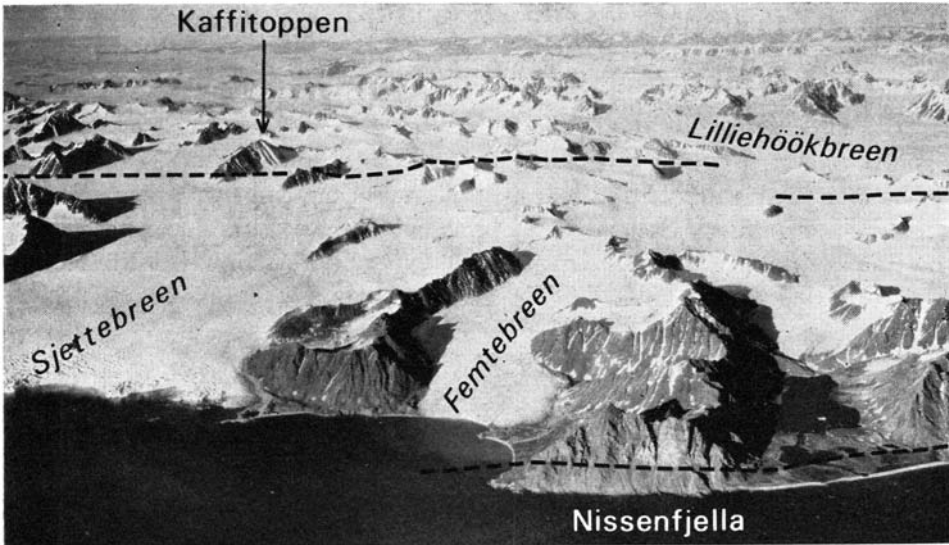


Fig. 7. From the coastal area south of Sjettebreen, looking towards the ENE. Major thrust lines are indicated.

displaced by faults of SW to NW trend. These have often also caused dislocations of the granite and the post lower Devonian faults.

From Lilliehöökbreen to Smeerenburgfjorden an important reverse fault zone passes NNW–SSE through the migmatites (Fig. 7). It contains mylonitic as well as brecciated material. The occurrence of skarn masses and pink aplitic veins in this zone east of Magdalenefjorden, indicates a correlation to the emplacement of the postorogenic granite in the east.

In Mitrahalvøya many faults are apparently directly related to folding, mainly F2, and folds overturned to the west are often thrust along the axial planes, both in mesoscopic and megascopic scale.

It is concluded that three main groups of faults may be distinguished in north west Spitsbergen:

- (1) Related to folding, mainly F2, early Palaeozoic? Mainly NNW strike.
- (2a) Related to anticlinorium development, late Palaeozoic? Mainly NNW strike.
- (2b) Related to the Hornemantoppen granite emplacement, late Silurian. Mainly NNW strike.
- (3a) Occurred after the granite emplacement, post late Silurian, mainly strike NW to NE.
- (3b) Post early Devonian (reactivated older faults), strikes around NNW.

In the western areas (A4W, A5W, and A6W) well defined maxima of joints have developed around W–E, with steep northerly dip of joint planes. The pronounced directions about normal to the F1 and F2 trends suggest cross joints, the general northerly dip of joint planes being due to the regional southerly plunge of the fold axes.

In the eastern areas (A4E, A5E, and A6E) the most distinct joint maxima occur around N to NNW. This direction parallels the main anticlinorium and the main fault lines, including the post lower Devonian faults near Kongsfjorden. These longitudinal joints are considered to be tension joints developed along the main anticlinal structure.

The major longitudinal and oblique fault directions can easily be recognized in the geomorphology of northwest Spitsbergen; coastal lines, fjords, glaciers, and mountain ridges frequently run in the NNW, WSW or WNW directions.

Metamorphism, absolute ages

When approaching the axial part of the main anticlinorium, increasingly deeper tectonic sections are exposed. The regional southerly plunge of fold axes has a similar effect towards the north. The result is a relatively narrow belt of gneissic and granitic rocks in the southern extension of the anticlinorium, widening towards the north. In general the metamorphic grade appears to correlate with tectonic depth and the lowest grade rocks occur at the southwestern flanks of the anticlinorium, in the actual area particularly in the synclines east and west of Krossfjorden. The metamorphic grade there increases from the chlorite zone in the beds above the upper marble unit to the garnet zone below the marble. The garnet crystallized here during the F1 deformation phase. This was later overprinted during and after the F2 formation under conditions stable to biotite and with decomposition of garnet. Chloritoid-bearing schists which occur locally in Mitrahelvøya in areas of thrusting and faulting, appear to be related to the immediate tectonic setting rather than being of regional metamorphic significance.

In the gneiss areas, cordierite-garnet-biotite paragenesis is common and sillimanite develops very often. Of the thin sections of gneiss and migmatite rocks examined from Danskøya and Amsterdamøya, about $\frac{1}{3}$ contain cordierite and sillimanite together, indicating a relatively low pressure amphibolite facies. Assuming $P_{H_2O} = 5 \text{ Kb}$ and a Ab/An ratio of 2.9, a value frequently met with in the biotite gneisses, the composition of a common migmatite mobilisate falls close to the cotectic line in the Ab-Or-Q diagram, suggesting that initial anatexis could have taken place at $P_{H_2O} = 5 \text{ Kb}$ and a temperature of 650–700°. Also the occurrence of wollastonite indicates that the metamorphism took place at a temperature where at least partial melting of pelitic and psammitic rocks would be expected.

In the Kennedybreen area, east of Smeerenburgfjorden, mesoperthite mantled with oligoclase were found in biotite gneiss containing diopside-wollastonite skarn. The perthite is supposed to be formed through replacement of sericitized calcic plagioclase by potash feldspar in the upper amphibolite facies, under extremely wet conditions.

Kyanite is only found in two rocks from Danskøya; the mineral shows inversion to muscovite from grain margins, and may indicate local pockets of high pressure or small relics of older deepseated (Archean?) basement rocks. Special parageneses are corundum-spinel-cordierite-biotite in clots in

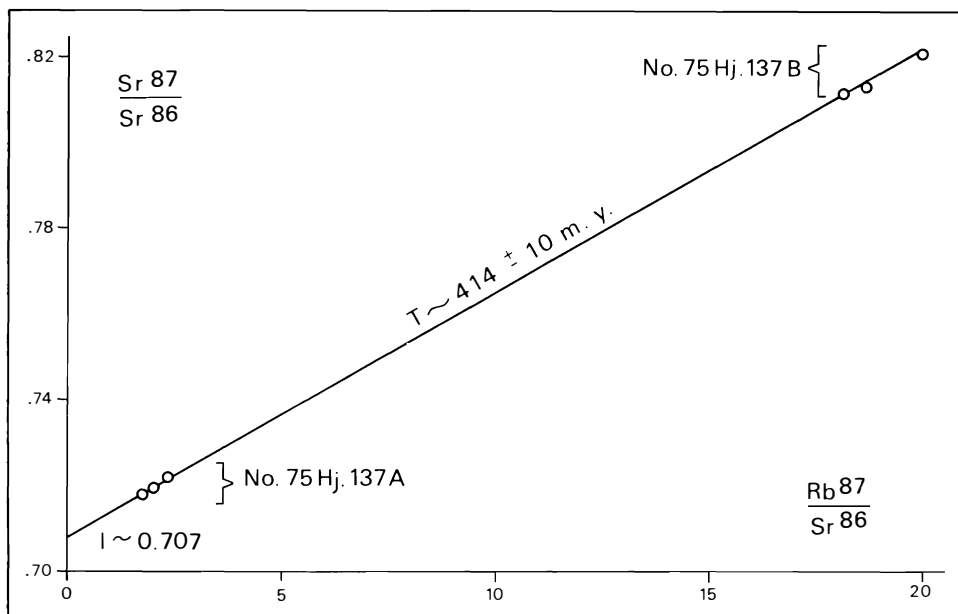


Fig. 8. Preliminary Rb/Sr isochron for the Hornemantoppen granite. No. 137 A: Coarse grained main granite. 137 B: Aplitic granite. Location: Nunatak 1 km south of Hornemantoppen ($79^{\circ}32.4'N-11^{\circ}46.0'E$). Analyst: A. RÅHEIM, Geological Museum, University of Oslo.

some pelitic rocks on the eastern side of Smeerenburgfjorden and spinel-fayalite-clinohumite-phlogopite and wollastonite-diposide-scapolite in the Fuglefjorden area. In Danskøya and Amsterdamøya a diopside-wollastonite-vesuvianite-hematite paragenesis is common in skarn rocks. In the Magdalenefjorden area diopside and garnet are common skarn minerals; vesuvianite and wollastonite are observed in some relatively thick skarn layers, and antophyllite was found at one locality. East of Magdalenefjorden, skarn lenses with hedenbergite, grossularite, epidote, pyrite, chalcopyrite and magnetite occur in a relatively late (F4) shear zone associated with the intrusion of the post-tectonic Hornemantoppen granite. Shear zones elsewhere around the granite, commonly show chloritization.

The post-tectonic granitic dyke activity in northwest Spitsbergen is closely related to the formation of the migmatites, and granitic dykes are scarcely found outside the gneiss-migmatite areas. This suggests that migmatization has occurred slightly before the intrusion of the dykes. The apparent K/Ar ages of granitic dykes, which centres around 400 m.y. (GAYER et al. 1966), thus may give an indication of the age of the migmatitization.

One preliminary Rb/Sr age determination has been carried out on Hornemantoppen granite material. Two types of granite from the same locality were analyzed, aplite granite and coarse grained granite with interfingering and transitional contacts (Fig 8).

The published absolute ages from northwest Spitsbergen give no evident support for an early Precambrian complex in this part of Svalbard. Although rapid changes might occur in metamorphism and structure in the Kongs-

fjorden area, no basal conglomerate or undisputable depositional unconformities have yet been established between the Precambrian metasedimentary strata and the highly metamorphic rocks. Thus, information up to now suggests that all metamorphic rocks in northwest Spitsbergen took part in the Palaeozoic metamorphism and deformation.

The main phases of the tectonic development and metamorphism in northwest Spitsbergen are summarized below:

- F0 Geosynclinal deposition, weak regional recrystallization. Greenschist facies. Late Precambrian.
- F1 Regional metamorphism, tight isoclinal folding, development of cleavage. Upper amphibolite facies. Late Proterozoic or early Palaeozoic. (Early Caledonian phase).
- F2 Main recrystallization, tight isoclinal folding, development of layering and gneissosity. Upper to lower amphibolite facies. (Mid?-) Silurian?
- F3 Migmatization, emplacement of syntectonic grey granitic rocks. Weak open folding, initiation of main anticlinorium. Lower amphibolite facies. (Local development of lowermost granulite facies east of Smeerenburgfjorden). Probably Silurian.
- F4 Intrusion of post-tectonic Hornemantoppen granite. Block movements, mylonitization. Lower amphibolite to greenschist facies. Late Silurian. Dislocation lines were reactivated in Devonian.

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The Hecla Hoek ridge of the Devonian Graben between Liefdefjorden and Holtedahlfonna, Spitsbergen

By TORE GJELSVIK

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Geological setting

The ridge to be discussed in this paper, is located in the west part of the main Graben, between the two westernmost N–S trending faults on ORVIN's 1940 map. It is bordered to the west by the NW Spitsbergen Hecla Hoek block, but a large glacier system — the Isachsenfonna and its northern continuation, Monacobreen — covers the border zone in a width of several kilometres all the way between Holtedahlfonna and Liefdefjorden. To the east follows the major Devonian system of Spitsbergen, preserved in the deepest part of the Graben.

Field work

In the early part of the geological exploration of Svalbard, during the 19th century, only the coastal region was investigated; most of the interior was geologically speaking unknown land. In 1911 OLAF HOLTEDAHL made the first geological traverse of the area, pulling sledges from Krossfjorden to Woodfjorden, and from there to Ekmanfjorden. Unfortunately, the weather was most adverse, and he was not able to observe very much. Nevertheless, he noted the inland continuation of the Hecla Hoek system from Kongsfjorden and the existence of Red Bay conglomerate and marble in the highest part of

the plateau (HOLTEDAHL 1914, 1926). During the English-Norwegian-Swedish Expedition 1939, mostly Devonian formations were studied, and investigation was made around the major fault zone at Vonbreen at the head of Woodfjorden (FØYN and HEINTZ 1943). JOHN PRESTON in 1957 explored the southern part of the ridge, from Holtedahlfonna to Eidsvollfjellet, and produced a geological map (1959). From 1959 to 1964 members of the Cambridge University Spitsbergen Expedition, headed by W. B. HARLAND, worked in the northern part of the ridge. Most of the work, by D. G. GEE, took place to the north of Liefdefjorden, and only preliminary reconnaissance studies near the coasts were carried out to the south. Some information on their work is given by GEE and MOODY-STUART (1966).

Members of the Svalbard expeditions of the Arctic Geology Institute of Leningrad (now SEVMORGEO) have worked in the area since 1962. Some accounts of their work on the volcanic rocks and the Devonian ones have been published (BUROV 1965; BUROV and MURASOV 1967), but no details of their studies of the Hecla Hoek stratigraphy, lithology, and structure have been known to me.

During some weeks of the summer seasons 1962–64, and on a snowscooter expedition in the spring of 1967, then accompanied by T. S. WINSNES, I was able to cover in a preliminary way most of the grounds of the ridge. My administrative duties, however, have, until recently, prevented me from microscopic and other laboratory studies of my samples. Therefore, only a general outline of the geology of the ridge will be presented here, and I shall only treat the Hecla Hoek sequence and the contact relationship to the Devonian rocks.

Lithology of the Hecla Hoek rocks

The following rock types, in various stages of metamorphism, are present:

1. Granite, migmatite, and gneiss.
2. Pelitic rocks, ranging from phyllite (greenschist facies), micaschist, and transition rocks to gneiss. Proper quartzite is rare, and occurs mainly in thin, inconsistent beds in the metasediments or as relics (paleosomes) in the granitic rocks.
3. Carbonate rocks, mostly pure calcite marble, sometimes also interbedded dolomite. In most cases banded or bedded types are found, the thickness of individual layers ranging from 1 mm to a few dm. In some areas massive, dense, or fine-grained marbles are found, in other areas coarse- to medium-grained ones. Fossils have nowhere been detected, neither have proper algal structures been observed, so far.

The maximum thickness of exposed strata in the ridge is in no place more than approximately 800 m, of which gneiss and/or pelitic strata make up 500 m. If my structural and stratigraphic considerations are correct, the total thickness of the exposed sequence in the investigated area may not exceed one kilometre.

Distribution, stratigraphy, and petrology of rocks

Gneiss and migmatite occupy a broad, central part of the ridge from Fortunafjellet (a little to the north of Eidsvollfjellet-Falsenslottet) and northwards to Keisar Wilhelmhøgda, just south of Liefdefjorden (map, Fig. 1). The granitic rocks are very similar to the “grey granite”, or migmatite, which is found in the NW corner of Spitsbergen (and to the west of Monacobreen), consisting of plagioclase (oligoclase), quartz, and micas. Plagioclase occurs also as scattered, rectangular porphyroblasts. Dark schlieren of biotite-ghosts or partly digested inclusions of pelitic rocks are ubiquitous. Angular and often rotated relics of schists and impure quartzite are commonly observed (Fig. 2). These features are all typical for the “grey granite” and migmatites in NW Spitsbergen. In various places gneisses containing a fair amount of FeMgCa silicates, such as hornblende, diopside, garnet, and even olivine, are found, partly near the contact of marble layers, but also elsewhere. In other localities sillimanite-bearing gneiss is observed. In one or two samples, cordierite or its alteration products are seen. These mineral assemblages indicate that the metamorphic grade during regional metamorphism and metasomatism reached upper amphibolite facies. Retrograde mineralization has taken place in zones of strong tectonization, such as thrustplanes. Here the highgrade minerals, along with plagioclase, have been altered to hydroxyl-bearing minerals: serpentine, antigorite, micas, chlorite, epidote, etc.

On the northern slope of the Keisar Wilhelmhøgda towards Liefdefjorden, coarse-grained, somewhat gneissified schists with dikes of granite are found in



Fig. 2. *Migmatite from the Bockfjorden area. Note variety and relationship of fragments, indicating rotation, feldspar porphyroblasts and biotite-rich schlieren in the matrix.*

the lower part of the succession. Higher up they grade into schists of lower metamorphic facies, without taking on the greenish colour typical of the low-grade schists on the north side of Liefdefjorden. The metasediments on Lernerøyane in the fiord are in a transition stage of schist and gneiss, and contain minerals of amphibolite facies. A few thin beds of coarse-grained marble and numerous schlieren of sulphide-bearing reaction skarn lenses are intercalated.

In the northern half of the ridge, marble layers of two types are found. On the east flank, a minimum of 200 m thick, coarse- to medium-grained (3–2 mm) marble layers are capping the gneiss zone, sometimes with a thin transition zone of micaschist or reaction skarn rocks. In the Keisar Wilhelmhøgda this marble layer is seen to bend over the gneiss, and is also found, mostly in thinner layers, at places along the west flank. Northwards from here, the coarse-grained marble layer, stepwise faulted down, can be followed rather continuously to the western Lernerøyane. On the east flank north of Keisar Wilhelmhøgda the coarse-grained marble cannot be traced continuously, but it outcrops again near the south shore of Liefdefjorden.

Impure beds of this marble formation contain the same minerals as those of the reaction skarn in the gneisses. Small lenses, often boudine-shaped, of dark green rocks occur, mostly consisting of diopside and hornblende with minor amounts of iron sulphides and trace amounts of other metal sulphide. In the gneiss, at some tens to one hundred metres, stratigraphically, below the main marble formation, small and rather discontinuous bands of the same marble are found.

The marble layer is not always fully concordant with the underlying gneiss. In most places the discordance is small and may be interpreted as a result of the difference in competence of the pelitic and carbonate beds during the orogeny. On the north side of Fränkelbreen (the northernmost tributary glacier of Vonbreen), a major discordance is located at the footwall of the marble layer. It is parallel to the adjacent main fault separating the ridge from the major Devonian Graben, and may thus be of post Hecla Hoek age.

The other type of marble, always fine-grained and mostly massive, is found only in the outer part of the west flank, an exception being some thin beds of dolomite and calcite marbles capping the central part of the northern slope of Keisar Wilhelmhøgda. When observable, the lower contact of the fine-grained marble is always a thrust zone, the underlying rock being either the coarse-grained marble or gneiss, in one place also Devonian rocks (Fig. 1, prof. C–D). Its upper contact always appears to be a weathered erosion surface, grading upwards into the typical Red Bay conglomerate. In these northern exposures, the fine-grained marble is a bluish grey, massive rock, frequently showing a criss-cross network of white veins, which mostly consists of calcite with subordinate quartz. In places, the marble contains thin beds of chlorite-muscovite schists.

The grain size of this marble formation ranges from submicroscopic to approximately 0.1 mm. This feature, coupled with the mineral composition of the intercalated metapelites, indicates a metamorphic grade of greenschist facies.



Fig. 3. Photo from NW: Falsenslottet (left), marble on top (white), biotite schist (black) below and in the foreground. Eastern shoulder of Eidsvollfjellet (right) with the same beds. Fault located under the white talus on the left side of the saddle between the two mountains. The contorted structure of the marble beds behind the saddle are located in the south extension of the fault zone.

In the southern part of the investigated area, comprising Eidsvollfjellet, Falsenslottet, and the two southwards running series of exposures, Snøfjella in the west and Dovrefjell-Wiechertfjellet in the east, only marble formations and pelitic schists make up the metamorphic sequence. The marbles consist of alternating calcite and dolomite beds of thicknesses ranging from a few mm to several tens of metres.

The rocks of Eidsvollfjellet-Falsenslottet mostly dip to the WSW and ENE due to folding with axis plunging to the SSE. Marbles occupy the higher parts, and are in the north wall underlain by black pelitic schists with scattered plagioclase porphyroblasts in some layers (Fig. 3). The marbles are medium- to coarse-grained, the schists are of high greenschist to low amphibolite facies. Despite the possibility of displacements by ENE running faults in between, the position of the Falsenslottet marble seems to match pretty well that of the eastern marble zone to the north.

In the axial part of Snøfjella, tightly folded, fine-grained marbles occur. They mostly dip to the WSW and are succeeded by typical Red Bay conglomerate with a transition zone of weathered marble and a sedimentation breccia of marble. In the northern part the strike bends to the NW because of SE dipping fold axes. A fairly thick bed of schists runs across the northernmost nunataks of Snøfjella in a SE direction. It is tightly folded and squeezed, but poorly exposed. Thus it is difficult to assess its real thickness. It contains some thin quartzite beds, and PRESTON (1959) reports also intercalations of quartz

conglomerate. In two localities I have found highly altered, fine-grained feldspar-rich shale, which may be of either arkosic or volcanic origin.

In the Dovrefjell-Wiechertfjellet exposures, mostly beds of fine- to medium-grained marble are found, some of which contain a fair amount of graphite which emphasizes the banded character of the rocks. Thin bands of schists are occasionally found, exhibiting greenschist- to biotite-grade facies. Dips are mostly steeply to the east in the upper part of the slope, flattening eastwards and even dipping W near the Vonbreen glacier. In the westernmost (upper) part of some of the exposures, strongly folded beds are observed, with an axis dipping S or SSE.

Devonian rocks

The Devonian rocks of the east side of the Hecla Hoek antiform are rather uniform grey-green sandstones and some rare, black shales. FØYN and HEINTZ (1943) suggested that they belong to the Red Bay series, whereas GEE and MOODY-STUART (1966) point to the strong lithologic similarity to the Siktefjellet group north of Liefdefjorden (considered to be a formation older than the Red Bay group).

PRESTON (1959) reports the presence of Devonian rocks on some of the mountain tops in the Dovrefjell-Wiechertfjellet range. We did not find them, however. The pre-Devonian erosion (and weathering) surface was observed, and Devonian rocks may exist under some of the highest located snowcaps.

The Devonian on the west side of the Hecla Hoek rocks are definitely of Red Bay affinity, starting with a weathering surface of an in situ limestone breccia, upwards grading into more heterogeneous, but still limestone-dominated breccia or conglomerate of typical Red Bay facies. This again grades into a quartz pebble conglomerate. Upwards follow alternating beds of sandstones and thin beds of rounded pebble conglomerate, succeeded by alternating beds of sandstone and shale (correlatable with the Andréebreen and the Frænkelyggen members north of Liefdefjorden).

Young volcanic rocks

In Sverrefjellet, close to the main post-Devonian fault at the head of Bockfjorden, a core of fresh alkali basalt is found (HOEL et HOLTEDAHL 1911; GJELSVIK 1963). It is considered to be of post-glacial age (HOEL et HOLTEDAHL loc.cit.; BUROV 1965; SEMEVSKIJ 1965).

From Sigurdfjellet, north of Frænkeltbreen, a similar rock is reported (HOEL et HOLTEDAHL loc.cit.). According to my recent observations, only pyroclastic rocks are present here, being deposited from a volcano, the neck of which is barely seen above the snow three kilometres to the west.

The top of the Eidsvollfjellet is capped by a rather flat-lying layer of unmetamorphic basalt, perhaps representing a flow. Mineralogically and structurally it is different from the Sverrefjellet basalt, and may be of Jurassic-Cretaceous or Tertiary age. A few dikes of it is found cutting the metamorphic rocks on the north slope of the mountain.

Tectonics

From the map and the profiles (Fig. 1) it is seen that the ridge is made up of a gentle antiform with NNW–SSE axis. Towards Liefdefjorden the antiform axis appears to plunge northwards. However, E–W faulting with downdrop to the N, has also taken place, and the plunge of the axis may not exceed 5–10°.

In the southern part, the elevation of the mountains is somewhat higher and greater thicknesses of the capping marble formation have been left by the erosion. Also here faulting occurred. The marble formation as a whole forms gentle structures; internally, however, it is often tightly folded. Both in the marbles and in the pelitic schists, folds of various amplitudes, from micro-size to some 100 m, are observed. Most of the larger ones are gentle and open, and dip to the SSE. Many small folds, however, are isoclinal. Sometimes they dip in directions opposite to the larger ones. Systematic and detailed tectonical observations are not sufficient to allow a proper tectonical analysis, but it is my impression that the small folding is largely due to plastic flow in the rocks, and does not generally affect the Hecla Hoek ridge on a larger scale. One possible exception should be mentioned. A limestone bed of a few tens of metres exposed thickness just N of Eidsvollfjellet has been subjected to a most intense folding and stretching. Numerous small isoclinal folds, whose axial planes dip steeply to the SW are observed, with axis plunging between 15 to 60°, averaging 30°, to the SSE. The lineation of this limestone is unusually strong. The profile G–H cuts across this locality, as well as across another, somewhat faulted and also intensely folded limestone with the same general directions. It may be the other limb of a large, overturned anticline. The extensive ice cover, as well as the complicated fault pattern, makes it difficult to ascertain this hypothesis.

Beside the two major faults which delimit the ridge, a number of smaller NS faults are observed, most of them located in the eastern border area (Fig. 3). Also faults in EW and other directions are observed, particularly in the areas just N of Eidsvollfjellet and near Liefdefjorden. In my opinion, EW faulting is a major element in the formation of this fiord. Some of them also intersect Devonian formations.

PRESTON (1959) indicates a NNW–SSE fault between Snøfjella and Dovrefjell with downdrop to the east. Unfortunately, the area is completely covered by a big glacier. However, some tectonically disturbed marble beds in the nearest nunataks suggest movement along a NNW–SSE zone halfway between Snøfjella and Dovrefjell. Similar features are observed at the saddle between Eidsvollfjellet and Falsenslottet (Fig. 3), following the same direction southwards. It is possible that some kind of a faulted flexure is located here.

In the western half of the ridge a system of reverse faults exists, which represents some kind of imbrication thrusting involving both the Hecla Hoek and the Red Bay formations. The profile C–D (Fig. 1) is most instructive to demonstrate this. Some of the thrust zones indicated on the map are inferred from abnormal structural trends and the occurrence of the basal Red Bay conglomerate in relation to them. The tectonic transport during imbrication

seems to have been towards NNE. The eastern border of this western thrust system, which is mostly covered by glaciers, is interpreted by me as part of the thrust system, since in most places where it is exposed, the contact is highly sheared and dips gently to the W. However, in the N, at Keisar Wilhelmhøgda, a steeply dipping shear zone (80°E) indicates more vertical movements. GEE and MOODY-STUART (1966) report that the Devonian formations in the eastern part of the ridge between Bockfjorden and Liefdefjorden in part are faulted, in part thrust against the Hecla Hoek rocks. Due to the extensive talus which covers the contacts, particularly those of the outlayers north of Børrebreen, I have not been able to verify this. The dip of the lower contact of the eastern Devonian formation is only about 10° steeper than that of the underlying marble formation (see Fig. 5 of GEE and MOODY-STUART (loc.cit.)).

Discussion

The outlined tectonic relationship within the ridge raises two interesting questions. The first one is of stratigraphical importance: are the two limestone formations which are so sharply contrasted in the northern half, two different formations? Or is the difference due only to the tectonic and metamorphic history? For a while I was inclined to believe in the first alternative, but a study of the metamorphic grade of the pelitic sediments, and of the grain size of the marbles in the southern part of the ridge, have brought me nearer to the second alternative. As mentioned above, there seems to be a gradual change southwards in metamorphic facies as well as in grain size and the difference in both respects between the eastern and western area to the south of Eidsvollfjellet is small indeed. Furthermore, the imbrication thrusts cannot be established south of the E-W faults on the north side of Eidsvollfjellet.

The other problem concerns the time of the formation of the antiformal structure. Some faults and thrusts affect also the Devonian formation present, and must therefore be of Svalbardian or later age. The trend of the antiformal, as well as most other folds, are the same as in the major Hecla Hoek block to the W. By interference it would mean a Caledonian time of folding. However, as seen from the northern profiles (Fig. 1), the doming may have involved also the outliers of Devonian (Siktefjellet group). In the southern part, the pre-Devonian peneplain also dips in conformance with the antiformal pattern: to the W in the western part, and to the E in the eastern part. This could mean that the antiformal structure is of a later age than the main Caledonian fold periods. The Soviet K/A age determination of two samples of the grey gneiss in the Bockfjorden area (KRASIL'SČIKOV et al. 1964) gave 385 million years for both samples. If this is the true age of the granite intrusion, or the granitization process, the antiformal could have been formed by this process after the deposition of the basal Devonian formations. If this is so, one would have expected that granite veins should have penetrated the Devonian somewhere. So far, it has not been observed. Alternatively, the antiformal structure could have been formed by a central uplift, longitudinal faulting and tilting systematically to the E and W. The structures in the southern part may fit this alternative, in

the north they seem better explained by folding. A more detailed tectonic analysis, as well as more and better geochronologic data, are needed to clarify this question.

As to the stratigraphical correlation of the Hecla Hoek rocks of the ridge, the thin slice of the stratigraphic column which is exposed, approximately 1,000 m, and the lack of fossiliferous or other marker beds make it possible only to give a rough guess. The similarity of the fine-grained limestone formation in Snøfjella and northwards to Liefdefjorden, and the marble formation of Blomstrandhalvøya in Kongsfjorden, makes it reasonable to suggest that the rocks exposed in the ridge belong to the Generalfjella Formation (GEE and HJELLE 1966). If it should include also the underlying Signehamna and Nissenfjella Formations, it would mean a most radical thinning to the east of these formations.

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Stratigraphy and tectonics of the Precambrian of Svalbard

By A. A. KRASIL'SČIKOV

Abstract

On the basis of a comparative historical-geological analysis, a general scheme of the stratigraphic division of the Precambrian of Svalbard is proposed. Two complexes of rocks have been recognized, corresponding to two periods of the development of the region in the Precambrian. The metamorphic successions of the lower complex have provisionally been assigned to the Lower-Middle Proterozoic; the mainly sedimentary formations of the upper complex which on the whole are comparable with the Riphean and Vendian of the folded framing of the East European Platform, are considered as Upper Proterozoic. According to set and succession of the upper Proterozoic formations, two types of section are clearly distinguished, characterizing different palaeotectonic zones, separated by a geoanticlinal uplift: the western-orthogeosynclinal and eastern-miogeosynclinal.

The Precambrian-Lower Paleozoic complex of metamorphic and sedimentary rocks, known in the literature as the "Hecla Hoek Formation", forms the Caledonides of Svalbard and is structurally clearly distinguishable from younger formations. The Precambrian-Early Paleozoic formations occur in four extensive but spatially separated areas in the northern part of the archipelago and along its west coast. These areas belong to different structural zones and have been studied by geologists from different countries and scientific schools. This accounts for the great variety of local stratigraphic standards and the occasional absence of reliable criteria for their correlation. Recent studies provide a possible basis for a single stratigraphic standard for the Precambrian of Svalbard.

A characteristic type section is recognized in each of the tectonic regions: Nordaustlandet, north-east (Ny Friesland), north-west, and western coast of Spitsbergen. The correlation of the four sections and their relation to the time-stratigraphic scale accepted in the USSR is shown in Fig. 1.

The comparative historical-geological method was used for the subdivision of the Precambrian of Svalbard because of the limited occurrence of fossils and the absence of reliable isotopic dates. Two contrasting complexes each corresponding to a major period in the Precambrian development of the archipelago have been recognized. The lower complex corresponds to the crystalline

North-East (Harland et al 1968, Krasil'shnikov 1973; Murašov 1973)

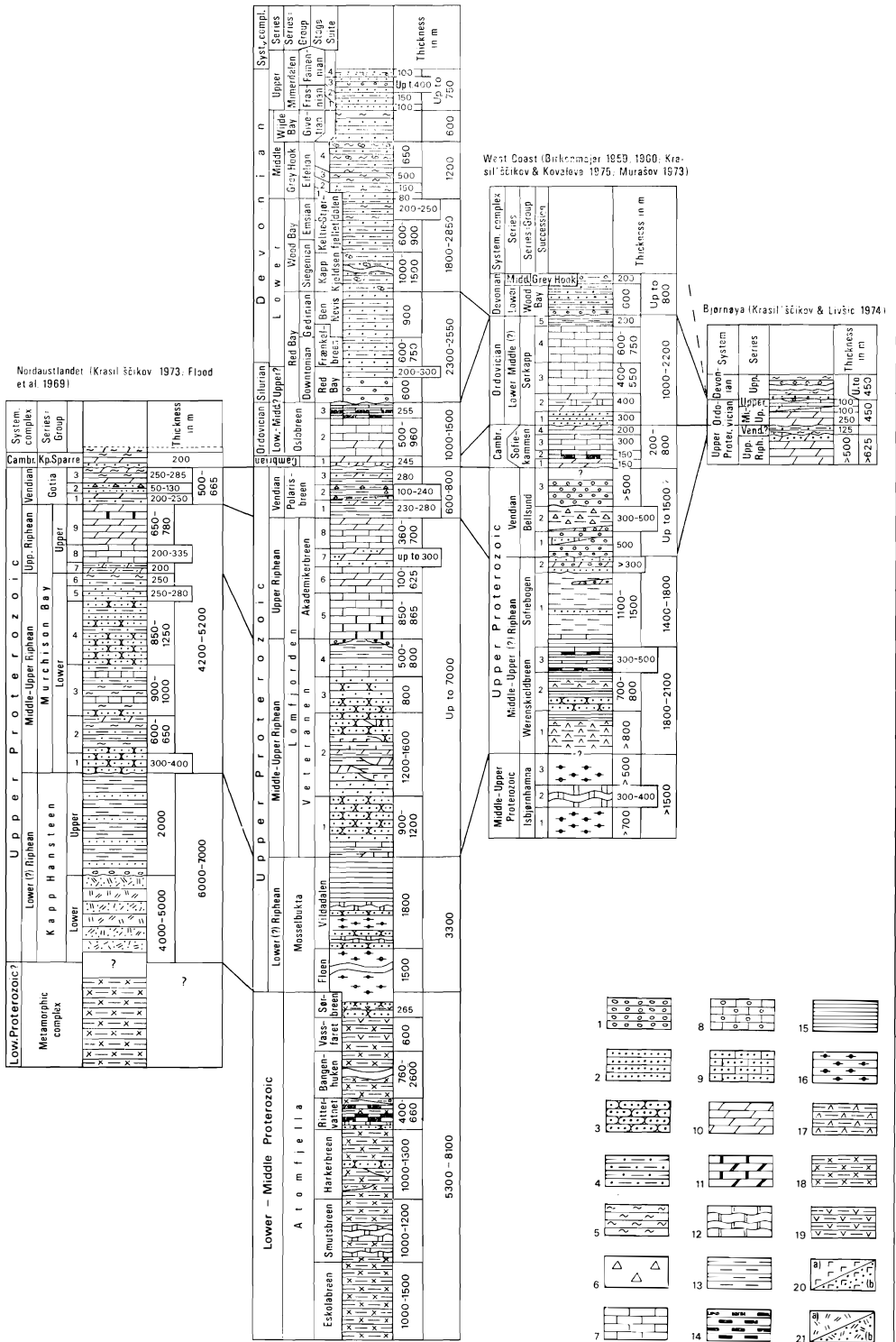


Fig. 1.

basement of the Caledonian geosynclinal belt of Europe and Greenland and is tentatively assigned to the Lower-Middle (?) Proterozoic (older than 1650 m.y.). The upper complex, composed mainly of sedimentary rocks, approximately correlates with the Riphean and Vendian of the miogeosynclinal flank of the East-European platform and is regarded as Upper Proterozoic (1650–570 m.y.).

The Lower-Middle (?) Proterozoic includes the relatively strong metamorphosed rocks of Spitsbergen (the Atomfjella Group and its granitized equivalents in the north-west) as well as gneiss-granites of eastern Nordaustlandet. The latter seem to be the result of ultrametamorphism of the oldest rocks of Svalbard. The base of the Lower-Middle (?) Proterozoic section is not known; the upper boundary usually coincides with the base of a thick phyllite sequence assigned to the Upper Proterozoic. In the type localities (Nordaustlandet, Ny Friesland) this boundary sharply separates two contrasting rock-units and is usually marked by an abrupt decrease in metamorphic grade. It is considered to be a structural unconformity modified and obscured by later tectonism.

Among the Lower-Middle (?) Proterozoic rocks, mica and garnet-mica plagiogneisses predominate. Carbonate rocks (marble and calciphyre) and quartzites are subordinate and are used to subdivide the rocks into formations. The north-west area is characterized by highly aluminous schists. Most of these rocks are undoubtedly metasediments, but there is no direct evidence for their age of deposition. A notable feature is the abundance of migmatites and granitized rocks as well as various amphibolite bodies derived, in the opinion of the majority of investigators (HARLAND 1959; BIRKENMAJER 1959), from the metamorphism of basic sills, lavas, and tuffs. The thickness of the complex is greater than 5000 m.

The Upper Proterozoic consists mainly of sedimentary rocks resting in deeply metamorphosed Lower-Middle Proterozoic rocks and overlain fossiliferous Cambrian deposits. Both lithologic and biostratigraphic criteria were used to subdivide the Upper Proterozoic into Lower Riphean (?), Middle-Upper Riphean, and Vendian complexes (Fig. 1).

The Lower Riphean (?) complex consists of distinctive shaly sequences (Kapp Hansteen and Mossel Groups and their possible equivalents on the western coast). These non-fossiliferous deposits are tentatively¹ assigned to the Lower Riphean. In some sections there are transitional beds between the shales and the overlying carbonate and terrigenous deposits. Microphytolites in these

¹ Considering that the Lower Riphean was tentatively assigned to the Upper Proterozoic (SALOP 1973) the age in stratigraphic columns that follow is shown as Middle Proterozoic, lower Upper Proterozoic inclusive.

←

Fig. 1. Correlation chart of the Precambrian and Lower Paleozoic (including Devonian) for Svalbard. 1. conglomerates; 2. sandstones; 3. quartzites; 4. siltstones; 5. mudstones; 6. boulders in tillite-like rocks; 7. limestones; 8. biogenic limestones; 9. clastic limestones; 10. dolomites; 11. bituminous carbonate rocks; 12. marbles and calciphyres; 13. shales; 14. "black shales"; 15. chlorite and sericite shales; 16. micaceous and garnet-micaceous schists (phyllites); 17. actinolite schists; 18. gneisses and plagiogneisses; 19. amphibolites; 20. basic lavas and tuffs; 21. acid and intermediate lavas and tuffs.

beds suggest a Middle Riphean age (RAABEN and ZABRODIN 1969; KRASIL'SČIKOV 1970). Argillaceous deposits containing occasional terrigenous and carbonate material are predominant in the Lower Riphean (?) deposits. In Nordaustlandet argillaceous deposits pass laterally into effusive and pyroclastic rocks within the Kapp Hansteen Group. Early Riphean volcanic deposits are unknown from other regions of the archipelago. A maximum thickness of more than 5000 m for the Lower Riphean was measured in Nordaustlandet (FLOOD et al. 1969; KRASIL'SČIKOV 1973). In Ny Friesland and the western coast of Spitsbergen it decreases to 3300 and 1500 m, respectively (HARLAND 1960).

The Middle-Upper Proterozoic complex is most complete in Ny Friesland (the Lomfjorden Supergroup) and western Nordaustlandet (the Murchison Bay Supergroup). Both sections are characterized by a transgressive sequence from quartz sandstones through silty-argillaceous and argillaceous-carbonate deposits to limestones and dolomites. Despite a general similarity, the total thickness of the complex varies from 4000–5000 m in Nordaustlandet (FLOOD et al. 1969; KRASIL'SČIKOV 1973) to 6000–7000 m in Ny Friesland (HARLAND et al. 1966). The complex is divided into two parts: lower, mainly terrigenous deposits (3700 to 4500 m), and upper carbonates (1300 to 2500 m). The boundary is drawn at the top of a variegated siltstone-mudstone sequence. The carbonates yield stromatolites and microphytolites of Upper Riphean age (KRASIL'SČIKOV, GOLOVANOV, and MILSTEIN 1965; RAABEN and ZABRODIN 1969). The lower part is Middle and lower Upper Riphean in age (RAABEN and ZABRODIN 1969; KRASIL'SČIKOV 1970).

On the West coast of Spitsbergen the Upper Proterozoic section has a completely different structure accounting for the tentative nature of earlier stratigraphic correlations (BIRKENMAJER 1960; HARLAND 1961; WINSNES 1965; KRASIL'SČIKOV 1973). A new preliminary scheme for the subdivision of the Upper Proterozoic for the west coast, based on tectonics (KRASIL'SČIKOV and KOVALEVA 1979), shows that three groups separated by unconformities of uncertain extent can be recognized. The upper Bellsund Group, carbonate-clastic, is correlated with "tillite-like" formations of the eastern regions of the archipelago, which are of Vendian age. The middle and lower groups, flysch-like Sofiebogen Group and carbonates and sedimentary-volcanics of Weren-skioldbreen Group are tentatively assigned to the Middle-Upper Riphean.

The lower sedimentary-volcanic Weren-skioldbreen Group, up to 2000 m thick, is subdivided into several formations. Basic volcanics at the base are overlain by a complex sequence of greenschists, possibly metamorphosed tuffs followed by a typical "black shale formation" consisting of homogeneous black limestones, silicified dolomites and shales rich in organic matter.

The middle flysch-like carbonate Sofiebogen Group is up to 1800 m thick and includes the Gåshamna and Höferpynten Formations (MAJOR and WINSNES 1955; BIRKENMAJER 1959, 1960). Assuming the beds in the Hornsund type locality to be the right way up (cf. BIRKENMAJER), the authors assume a break before the deposition of the upper Höferpynten Formation marked by a quartz gritstone horizon. A uniform microphytolite assemblage of Middle Riphean aspect (RAABEN and ZABRODIN 1969; MILSTEIN in press), presumably

reworked, has been found in the clastic dolomites of the Höferpynten Formation and in similar clastic-carbonate rocks of the Bellsund Group.

The Vendian complex includes the Sveanor Group of Nordaustlandet and the Polarisbreen Group of Ny Friesland as well as the carbonate-clastic Bellsund Group of the west coast of Spitsbergen. The Vendian age is suggested by its position directly below fossiliferous Cambrian rocks and is supported by the occurrence of microphytolites typical of the Vendian of the Urals and Siberia (KRASIL'SČIKOV et al. 1965; RAABEN and ZABRODIN 1969). The Vendian of eastern Svalbard includes fine-grained terrigenous rocks and characteristic horizons up to 50–150 m thick of coarse-clastic material in the middle part of the section (Sveanor and Wilsonbreen Formations). The total thickness of the Vendian is 500 to 700 m.

The upper carbonate-clastic Bellsund Group, up to 1500 m thick, includes the various conglomerate sequences of the west coast. Geologic-petrographic studies show no major differences between these sequences. Stratigraphic relationships between the Bellsund Group and previously defined stratigraphic units are not always clear. The Bellsund series consists of quartzite (rarely polymict) conglomerates of the Slyngefjellet type overlain by carbonate-clastic rocks and shales with varying amounts of boulders followed by a sequence of largely dolomitic conglomerates of the Kapp Lyell type often resting on eroded beds of the lower part of the succession.

Upper Proterozoic deposits are also known from southern Bjørnøya, where they are subdivided into two formations: the dolomitic Russehamna Formation and the sandstone-siltstone Sørhamna Formation (HOLTEDAHL 1920; KRASIL'SČIKOV and LIVŠIČ 1974). The Russehamna Formation contains microphytolite assemblages common in the top of the Upper Riphean and transitional between the Upper Riphean and the Vendian (RAABEN and ZABRODIN 1969; KRASIL'SČIKOV and MILSTEIN, in press). The Upper Proterozoic age given to the overlying non-fossiliferous Sørhamna Formation is quite tentative as there is a proven stratigraphic unconformity at its base. A Vendian age is suggested by the similar structural framework, the presence of underlying Upper Riphean dolomites and the lithological similarity of the formation to terrigenous Vendian rocks in other areas of the Barents Sea region.

Three tectonic complexes are recognized within the Caledonides of Spitsbergen: the pre-Upper Proterozoic crystalline basement, the major geosynclinal complex (Upper Proterozoic — Middle Ordovician), and the orogenic molasse complex (Lower-Upper Devonian). The intensive processes of the Caledonian metamorphism and rheomorphism gave rise to a new (Caledonian) infrastructure which obscured the contact between the crystalline basement and the major geosynclinal complex. The western and eastern zones of the Caledonian major geosynclinal complex are separated by a central graben filled by orogenic Devonian deposits.

Tectonic zonality was most distinct in Upper Proterozoic time (the Baikalian Stage). The eastern zone was an extensive depression (Hinlopen) formed at the site of the early-Riphean trough. The depression was asymmetrical as shown by sections in Nordaustlandet and Ny Friesland which, despite similar

lithology, differ greatly in thickness and internal structure of individual stratigraphic units. All the changes take place in a narrow zone related to Hinlopenstretet implying that during the Riphean, sedimentation was controlled by the Hinlopen fault. Both types of section consist of three units, each corresponding to a stage in the development of the basin. The lower terrigenous unit (transgressive) reflects the increasing downwarp of the basin floor. Intense downwarping, a higher gradient and a larger amplitude of oscillatory movements characterized the western margin of the depression. This is indicated by lateral changes in the lower terrigenous unit and the increase in thickness from 2600 m in Nordaustlandet to 4250 m in Ny Friesland.

The carbonate sediments of the second unit accumulated under stable conditions of long-term slow downwarping. The thickness also increases from east to west, from 1200 to 2700 m. Despite a high mobility of the western part of the basin inherited from the preceding stage, there is evidence that the regression (intraformational dolomite breccia beds) occurred almost contemporaneously across the entire Hinlopen depression. The change in regime resulted from tectonism at the close of the Riphean and was accompanied by shallow magmatism of trapp-type accounting for the slightly altered basalt and dolerite pebbles of the Vendian tillite-like rocks (KULLING 1934; KRASIL'SČIKOV 1967).

The upper regressive terrigenous unit corresponds to general emergence and filling of the Hinlopen depression. It consists of Vendian deposits (Gotia and Polarisbreen Groups) differing in lithology from the under- and overlying Upper Riphean and Cambrian carbonate sequences. The middle part of the upper terrigenous unit, mainly silty-argillaceous, contains an unsorted bed with scattered boulders and pebbles ("Tillite Formation" after KULLING 1934). The bed thickness and the amount and size of boulder-pebble clasts decrease northeastward. A predominance of well-rounded clasts of local carbonate rocks indicates the relatively short distance of transport from the source area, which presumably was in the south-eastern part of the archipelago. The clast composition suggests that the depth of erosion in the source area was unlikely to have been more than 500 m.

Reconstruction of the tectonic environment of the Baikalian Stage on the West coast of Spitsbergen (western zone) is complicated by the absence of reliable stratigraphic markers, and by extensive faulting and thrusting both of Caledonian and Tertiary age. In comparison with sections in the Hinlopen depression, sections in the west are considerably thinner (up to 6000 m). They are generally terrigenous, containing several conglomeratic horizons and characteristic volcanic sediments with basic intrusions.

The structure of the western zone is considered in more detail elsewhere (see KRASIL'SČIKOV and KOVALEVA in this volume). New data suggests the brief existence of a volcanic-orthogeosynclinal trough in the western part of the archipelago at the beginning of the Baikalian Stage. At the close of the Baikalian a compensatory trough developed in place of the volcanic trough as a result of adjacent geanticlinal uplifts. It was filled first by flysch-like and then by molasse-like carbonate-clastic deposits.

The stage of common inversion of geosynclinal troughs in Svalbard coincides approximately with the major phase of the Caledonian folding in North Europe (440–370 m.y.). The fault monoclinical structures of the Riphean of Bjørnøya, striking north-west (“Baikalian”) and overlain by subhorizontal Ordovician deposits, were most probably formed at the margin of an ancient land mass and do not belong to the Caledonian of Spitsbergen.

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Precambrian rock-stratigraphic units of the west coast of Spitsbergen

By A. A. KRASIL'SČIKOV and G. A. KOVALEVA

Abstract

On the basis of a formation analysis a correlation of the sections of different parts of the west coast of Spitsbergen is given, and a new scheme of division of the Precambrian for this region is proposed. Three groups separated by regional unconformity, have been recognized. The lower group (Lower Riphean?) includes the most ancient stratified successions of the crystalline schists. The middle group (Middle to Upper Riphean?), the sedimentary-volcanogenic one, encloses to a different degree changed basic intrusions. The upper group (Vendian), the carbonate-fragmentary one, is characterized by a considerable diversity of the section and by lateral variability; it is overlaid by polymict conglomerates, locally lying with a washout on more ancient horizons.

The Precambrian succession of the west coast of Spitsbergen is noted for its peculiarity and complex structure due largely to dislocations during the Alpine orogeny. Despite the relative accessibility of the western coast, its geology is not well known and data available does not provide a reliable basis for correlation. In this paper a Precambrian stratigraphic standard section is proposed for the western coast (south of Kongsfjorden), based on geological observation and tectonic analysis of the non-fossiliferous ("barren") sequences. A review of the data on the Precambrian of Spitsbergen published by the Norwegian Polar Research Institute (WINSNES 1965) and an excellent geological map of Southern Svalbard (1:500 000) (FLOOD, NAGY and WINSNES 1971) have contributed greatly to this work.

Field work was carried out by KRASIL'SČIKOV during the Spitsbergen Expedition of the Research Institute of the Geology of the Arctic in 1969–1972 in Brøggerhalvøya, north side of Aavatsmarkbreen, northern St. Jonsfjorden, north coast of Isfjorden, Nordenskiöld Land (a peninsula between Recherchefjorden and Dunderbukta) (the Wimsodden area), and the north and south coasts of Hornsund.

Prior to the present time, the stratigraphic standard compiled by Polish geologists for the Hornsund area was used (BIRKENMAJER 1959, 1960). However in many cases this does not reflect real stratigraphic relationships. First of all this concerns the widespread conglomerate sequences and metavolcanites.

A number of clear continuous and characteristic rock units were recognized through the analysis of geological data and detailed lithology and petrography.

The Precambrian rocks south of Kongsfjorden were divided into four groups each named after the type locality. They are from below: Isbjørnhamna Group (schists), Werenskioldbreen Group (sedimentary-volcanic), Sofiebogen Group (flysh-like-carbonates) and Bellsund Group (carbonate-clastic).

The Isbjørnhamna Group consists of a complex of crystalline schists, including marbles and quartzites, outcropping on the north coast of Hornsund (type locality) and on Brøggerhalvøya. In both areas the complex, 1500 to 2500 m thick, consists of two schist sequences separated by a sequence of carbonate rocks. Contacts with adjacent groups are tectonic. The whole complex was metamorphosed to amphibolite facies (almandine subfacies). This high grade of metamorphism provides the only grounds for its recognition as a single rock-stratigraphic unit. However, the petrography of these carbonates is comparable to that of the clastic dolomites and limestones of the Upper complex, and the Upper Isbjørnhamna schists (Revdalen Formation) are similar to distinctive graphitic schists of the Werenskioldbreen group.

The Werenskioldbreen Group, a sedimentary-volcanic complex, consists of volcanics and gabbros formed during a single episode and metamorphosed to greenschist facies and closely related sediments consisting mainly of homogenous argillaceous-carbonate deposits containing phosphates and organic matter.

The contact of the Werenskioldbreen Group with the underlying schists is tectonic. The predominantly volcanic lower part of the complex suggests the beginning of a new episode in the history of the region. The upper boundary is drawn at the abrupt lithologic change from a homogenous “black shale” sequence to almost unmetamorphosed flysh-like deposits. The Werenskioldbreen Group is best exposed north of Hornsund, between Hansbreen and Torellbreen. A new interpretation is proposed using the data available for the structure and stratigraphy of the different rock-units (BIRKENMAJER and NAREBSKI 1960; SMULIKOWSKI 1965, 1968), field data, rock collections and thin-sections (by KRASIL’ŠČIKOV 1969, 1972).

According to Polish geologists (BIRKENMAJER and NAREBSKI 1960) the older Isbjørnhamna Group is thrust over the Eimfjellet Group amphibolite complex with the Gulliksenfjellet Quartzite at the base. SMULIKOWSKI’s detailed structural maps (1965, 1968) show the Gulliksenfjellet Quartzite overlying the Eimfjellet Group in an asymmetric dome-like structure with an “amphibolite” core with which the so-called “Gangpasset granitization zone” is associated. This dome-like structure is complicated by intense folding on the flanks with alternations of quartzites and schists of different composition.

The Vimsodden Formation differs in composition, genesis, environment, and thermodynamic transformation, and occurs north of Werenskioldbreen, presumably obscuring a thrust fault system (BIRKENMAJER and NAREBSKI 1960). In BIRKENMAJER’s opinion the Vimsodden Formation is stratigraphically equivalent to the Eimfjellet Group while, according to SMULIKOWSKI (1968), it overlies the Eimfjellet Group as it is generally of lower metamorphic grade.

Most of the Eimfjellet rocks correspond in composition to actinolitized gabbro-diabases and albite-epidote-actinolite schists which cannot be termed "amphibolites". The amphibole of the schists is close in optical properties to glaucophane implying local fields of high pressure against a background of greenschist facies metamorphism.

The proposed subdivision of the Werenskioldbreen Group in the type locality is as follows:

1. Skålfjellet greenschist: (a) fine- to medium-grained actinolite schists with coarse-grained actinolite gabbro-diabase bodies, 550 m thick; (b) chlorite-muscovite schists with quartzite bands, 250 m thick.
2. Gulliksenfjellet quartzite: 250 to 300 m thick.
3. Kvislodden greenschists: chlorite and carbonate-mica schists with "rhyolitic conglomerate" beds, 100 m thick.
4. Pyttholmen arkose quartzites: 125 m thick.
5. Tonefjellet greenschists: carbonate-mica and chlorite schists with graphite-bearing schist bands and mica quartzites as well as lenticular and thick concordant bodies of actinolitized gabbro-diabases, 250 to 400 m thick.
6. Vimsodden "Black Shales": graphite-bearing schists with marble and quartzite beds, 200 to 250 m thick.

The Vimsodden quartzite-marble conglomerates and Tonedalen carbonate-clastic rocks are assigned to the younger Sofiebogen Group.

Further study of this region may prove that the Gulliksenfjellet quartzite, the Middle greenschists and the Skålfjellet greenschists are equivalent to the Pyttholmen quartzite, the Upper greenschist and the Tonefjellet Formation, respectively. The subdivision of the Werenskioldbreen Group would thus be simplified, and the total thickness would be less.

Volcanics of the Werenskioldbreen Group are also widespread south of Recherchefjorden where they consist of a thick (up to 1500 m) sequence of green and black shales with thin amygdaloidal basalt sheets and metagabbros (HJELLE 1969) sediments. The black phosphate-carbonaceous shales and limestones of the Werenskioldbreen Group are better exposed farther north (Nordenskiöld Land, northern Isfjorden, St. Jonsfjorden), but greenschist units are subordinate in the north. Gabbro intrusions on higher structural-stratigraphic levels are as a rule less altered. The thickness of the black shale-carbonate sequence in these regions is between 300 and 800 m.

Thus, the Werenskioldbreen Group, about 2000 m thick, consists of three formations; volcanic, sedimentary-volcanic, and homogenous argillaceous-carbonates ("black shale"). These proposed formations are named after the regions of their widespread occurrence:

1. Skålfjellet Formation (more than 800 m thick), the Tonefjellet sequence is its possible equivalent;
2. Vimsodden Formation (700 to 800 m thick) with the Gulliksen quartzites at the base, its equivalents are the Recherchefjorden and Nordenskiöld Land greenschists;

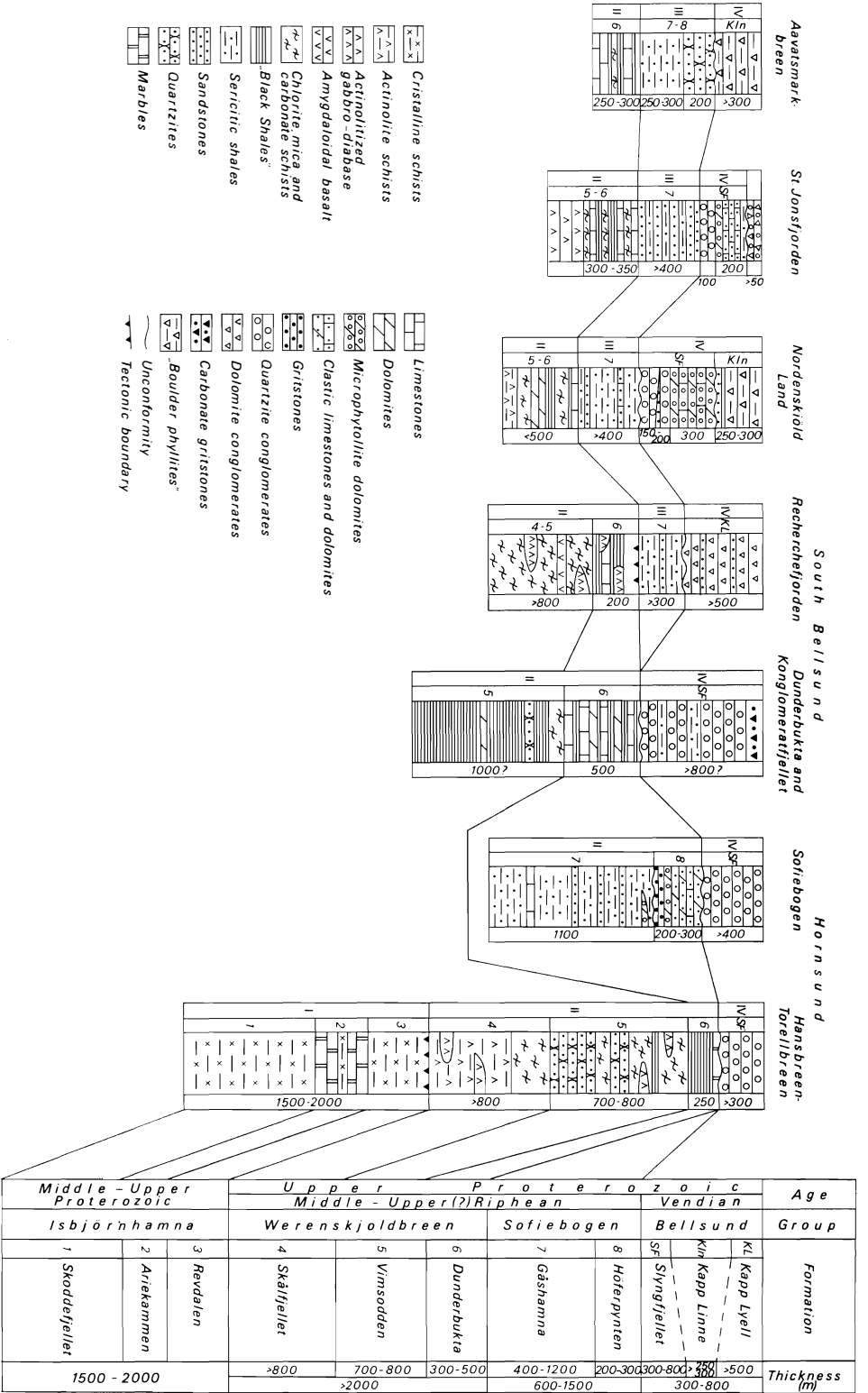


Fig. 1. Correlation chart of the Precambrian of the western coast of Spitsbergen.

3. Dunderbukta Formation (300 to 500 m thick), its equivalents are sequences of black shales and phosphate-carbonaceous limestones which can be traced from Vimsodden in the south to the Aavatsmarkbreen area to the north.

The Sofiebogen Group, a flysh-like carbonate complex, is widespread on either side of Hornsund where two lithologically different formations — the Gåshamna and the Höferpynten Formations can be recognized (MAJOR and WINSNES 1955, BIRKENMAJER 1960, KRASIL'ŠČIKOV 1970).

The Gåshamna Formation, up to 1200 m thick, is composed of sericitic shales (metamorphosed mudstones and siltstones) with sporadic bands of sandstone and carbonate. In the type section (Sofiebogen and Gåshamna), rocks of the formation form a steep west dipping monocline. In the east the monocline is thrust over a largely carbonate sequence assigned to the Cambrian; to the west the Gåshamna shales are overlain by a marker horizon of quartz gritstones with shale flakes. BIRKENMAJER (1959, 1960) considered this horizon to be the base of the Gåshamna Formation using the assumption that the beds in the Sofiebogen area are overturned. The equivalent of the Gåshamna Formation can be recognized in the Recherchefjorden area, on Nordenskiöld Land, and especially well on the northern coast of St. Jonsfjorden.

In the type locality the authors subdivide the Höferpynten Formation into three members: lower (40 to 60 m thick) — quartz gritstones, middle (50 to 150 m thick) — chiefly dolomites, and upper (about 100 m thick) — characterized by an alternation of metamorphosed limestones, dolomites, quartz-carbonate schists, and microquartzites.

North of Hornsund the lower member occurs in Nordbukta, in Nordenskiöld Land, and on the northern coast of Isfjorden. The middle member consisting of phytollite dolomites intercalated with limestones are widespread north of Kapp Martin.

Dolomites and rare limestones of the middle member contain numerous microphytollite nodules which help to distinguish this member as a good marker horizon. The microphytollites belong to a single *Osagia* group, *Osagia tenuilamellata*, characteristic of the Middle Riphean (RAABEN and ZABRODIN 1969, MILSTEIN 1971). However, the clastic nature of the enclosing beds indicates reworking of microphytollite nodules.

The Sofiebogen Group is transitional between the underlying group, and the overlying Bellsund Group. Within the Sofiebogen Group are rocks similar to the upper horizons of the Dunderbukta Formation. There is no reliable evidence about the contact between these sequences but spatially they are adjacent in many places suggesting a possible facies replacement of boundary beds. The upper mainly carbonate part of the Sofiebogen Group is similar to clastic dolomites and limestones occurring among the conglomerates of the overlying Bellsund complex. However the historical-geological distinction of the Sofiebogen Group is emphasized by its low grade metamorphism, the absence of basic magmatism and the occurrence of rocks of the Gåshamna and Höferpynten Formation type within the clastic material of the overlying conglomerates.

The Bellsund Group, a carbonate-clastic complex is characterized by strong lateral changes accounting for the recognition by earlier workers of a number of conglomerate horizons of different ages (BIRKENMAJER's three "pretillite" horizons). Analysis of geological setting and petrographic study proves that all the conglomerate sequences on the west coast of Spitsbergen belong to a single rock-stratigraphic complex formed during a single tectonic cycle.

In a number of scattered conglomerate outcrops the clastic material has a relatively constant composition and the matrix is similar in structure and mineralogy. The conglomerates largely consist of carbonate-mica-quartz shales which do not differ in composition from the numerous bands of calcareous-quartz sandstones and shales found within the conglomerate sequences. The psephitic material is comparable to the following rock-types:

- (1) various quartzite and quartzite-sandstones including the Gulliksenfjellet quartzites;
- (2) feldspar-quartz and calcareous-quartz sandstones and siltstones (including those of the Gåshamna Formation);
- (3) sericite, chlorite-sericite and chlorite (after metabasites?) schists (e.g. Vimsodden schists);
- (4) mica slates and mica-carbonate graphite-bearing shales (including those of the Vimsodden and Dunderbukta Formations);
- (5) black phosphate-bearing limestones (the Dunderbukta Formation type);
- (6) microphytolites and clastic dolomites (the Höferpynten Formation type);
- (7) granites, plagiogranites, quartz diorites.

The almost complete absence of pebbles of schists and metabasites is notable.

Visual differences of conglomerate sequences are mainly accounted for by large changes in the relative abundance of petrographically different pebbles. Three sequences were recognized, the lateral and stratigraphic relationships of which are not always clear.

The presumably lower sequence — the Slyngfjellet Formation — reported by BIRKENMAJER (1959) as mainly quartzite conglomerates, is 500 m thick. Besides predominant quartzite boulders the sequence is characterized by an abundance of boulder-pebble material (over 50%). The chiefly quartzite content of the conglomerates of Dunderbukta, Konglomeratfjellet, Kapp Martin, and Ankerbreen (north coast of St. Jonsfjorden) are equivalent to the Slyngfjellet sequence. In different areas quartzite conglomerates rest on different beds of the Dunderbukta, Höferpynten, and Gåshamna Formations.

The second sequence, the Kapp Linné Formation, was described by HJELLE (1962) as a 300 m thick sequence of "phyllites with boulder beds". Dolomite, quartzite, and altered quartz diorite were reported among the boulders. A non-persistent and as a rule small amount of boulder-pebble material without predominant rock, is a peculiar feature of the sequence. Shales with boulders usually occur within tectonic blocks, in association with clastic dolomites and

limestones. Thick schistose Comfortlessbreen conglomerates (HARLAND 1961) may be equivalent to the Kapp Linné “boulder phillites”.

The upper sequence, the Kapp Lyell Formation, is most complete on the south coast of Bellsund. It consists of schistose conglomerates more than 500 m thick with 5 to 40 per cent dolomitic psephite material. These conglomerates are thought to rest unconformably on the underlying rocks.

The clastic nature of the deposits, the overall unsorted character and the rapid lateral changes suggests that the Bellsund Group is a molasse deposit. The occurrence of conglomerates at different stratigraphic levels, and the composition of clastic material indicate tectonic activity prior to the deposition of the Bellsund Group. Most of the psephitic material cannot be identified with directly underlying rocks. This suggests that adjacent uplifted zones of early consolidation undergoing erosion provided the source for the coarse clasts. The tectonism most probably occurred in Prevendian times as the Bellsund Group is thought by many to correlate with the “tillite-like” Vendian formations of eastern areas of Svalbard. Therefore, the Sofiebogen Group is Middle-Upper Riphean in age (more probably Middle Riphean). The sedimentary-volcanic Werenskioldbreen Group presumably formed in a relatively narrow trough, the initiation of which may be tentatively assigned to the lower Upper Proterozoic.

Thus, the rocks of the West coast of Spitsbergen are thought to have formed in an Early Baikalian (Grenville) orthogeosyncline of embryonic development. At the close of the Baikalian stage a geosynclinal trough was buried under molasse-like deposits of Vendian age, then Early Paleozoic predominantly carbonate deposits accumulated under subplatform environment prior to the Caledonian orogeny. The complete inversion of sedimentary troughs, and the formation of the West Spitsbergen fold system, are related to the Caledonian orogeny.

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Alternative hypothesis for the pre-Carboniferous evolution of Svalbard

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Abstract

The stratigraphy of pre-Carboniferous Svalbard is reviewed in four major provinces. It is argued that at least three of these provinces exposed in Spitsbergen (Eastern, Central, and Western) have distinctive sequences whether their biostratigraphic or lithotectonic characters be compared. On this basis alternative palaeogeological models are compared.

One, a palaeofixistic model assumes for the time in question that the areas of Svalbard were in approximately their present relationship (having restored them to a position prior to late Phanerozoic continental drift).

The other, a palaeomobilistic model, separates these three provinces by late Devonian sinistral transcurrent faulting. The differences within Svalbard are then explained because the rocks were formed when far apart, and indeed when the constituent parts were nearer either East Greenland, or North Greenland and the Queen Elizabeth Islands. This hypothesis, suggested some years ago, is developed with more evidence than has previously been available.

I. Introduction

The Svalbard Archipelago contains critical evidence for the tectonic history of a very much larger part of the Earth's crust extending in the present Arctic to all the surrounding continental areas. There is an exceptionally rich record of events in late Precambrian and Phanerozoic strata. This record now makes it possible to reconstruct the pre-Carboniferous history of Svalbard, and a hypothesis with major horizontal displacements is argued.

1. OLDER ROCK COMPLEXES

The pre-Carboniferous rocks of Svalbard have traditionally been divided into the Devonian "Old Red Sandstone" and the pre-Devonian "Hecla Hoek" on the assumption that sequences throughout Svalbard can be best described and correlated in this way. HARLAND et al (1974) pointed out that this approach may be invalid because, if the possibility of large scale horizontal displacements along the Billefjorden Fault Zone be considered, there is no basis for assuming that the pre-Devonian sequences to the west should match the Hecla Hoek sequence of Ny Friesland.

We propose to make no such assumptions regarding pan-Svalbardian correlation here. Whether or not we interpret correctly, it is better at the outset of an investigation to distinguish diverse elements in a hypothesis, which can subsequently be equated should that be justified, rather than to assume a unity which, if not justified, could confuse the investigation.

The pre-Carboniferous rocks have been described in the literature as the result of surveys of geographical areas within which a unified stratigraphy can be discerned. We identify about twelve such areas, and use the term "complex" for the major rock bodies exposed. The term complex, larger than supergroup, implies that the sequence may not be fully understood, nor continuous, and it allows for the presence of extraneous elements.

2. PALAEOGEOLOGICAL PROVINCES

We suggest here that the twelve different complexes can be grouped in three or more provinces which correspond to major areas of present day Svalbard. The provinces are shown in Table 1 and each has distinct stratigraphic and

Table 1.

Provinces and complexes

Eastern Province (EP)

Ny Friesland (NF)
Nordaustlandet, western and central parts (NE)
More distant Nordaustlandet, Storøya and Kvitøya (EX)

Central Province (CP)

North Central with Andrée Land and related areas (NC)
South Central with Eastern Wedel Jarlsberg and Sørkapp Lands (SC)
North Western Spitsbergen from Raudfjorden to Kongsfjorden (NW)

Western Province (WP)

Brøggerhalvøya (WB)
Remainder of Oscar II Land (WO)
Prins Karls Forland (WK)
Western Nordenskiöld and Nathorst Land (WN)
Western Wedel Jarlsberg Land (WS)

Southern Province (SP)

Bjørnøya (BI)
Related part of South Barents Shelf (BS)

structural characteristics. Correlation between the complexes of one province is possible to a greater or lesser extent. We suggest, however, that detailed correlation between provinces may not be feasible, except by the long distance time-correlation characters reviewed in Chapter VI, and that this is because of the distant separation of the provinces at the time of formation of the rocks.

This division into three or more provinces (HARLAND 1972a) has recently become more plausible because of a growing understanding of the stratigraphic sequence in western Spitsbergen. Our group and others have now worked in this area for several years and, although many problems remain, the time is ripe to correlate the several complexes of the Western Province. The pre-Carboniferous sequence of this province contrasts strongly with the Central and Eastern Provinces both in facies and age, and the name Holtedahl Geosyncline has been proposed for these distinctive rocks (HARLAND, HORSFIELD, MANBY, and MORRIS, this volume).

3. PROVINCIAL BOUNDARIES AS FAULTS

Elsewhere (HARLAND et al. 1974) reasons have been given for supposing that substantial sinistral strike-slip motion took place along the Billefjorden Fault Zone (BFZ) during the Svalbardian (late Devonian) movements. Total movement of not less than 200 km and possibly 1000 km was suggested. That this major lineament separates regions of contrasting pre-Carboniferous strata has been assumed by others (e.g. KRASIL'SHCHIKOV 1973) and we take the BFZ to separate the Eastern and Central Provinces.

We also propose an analogous, but entirely speculative, fault or fault zone separating the Central and Western Provinces, with sinistral strike-slip motion

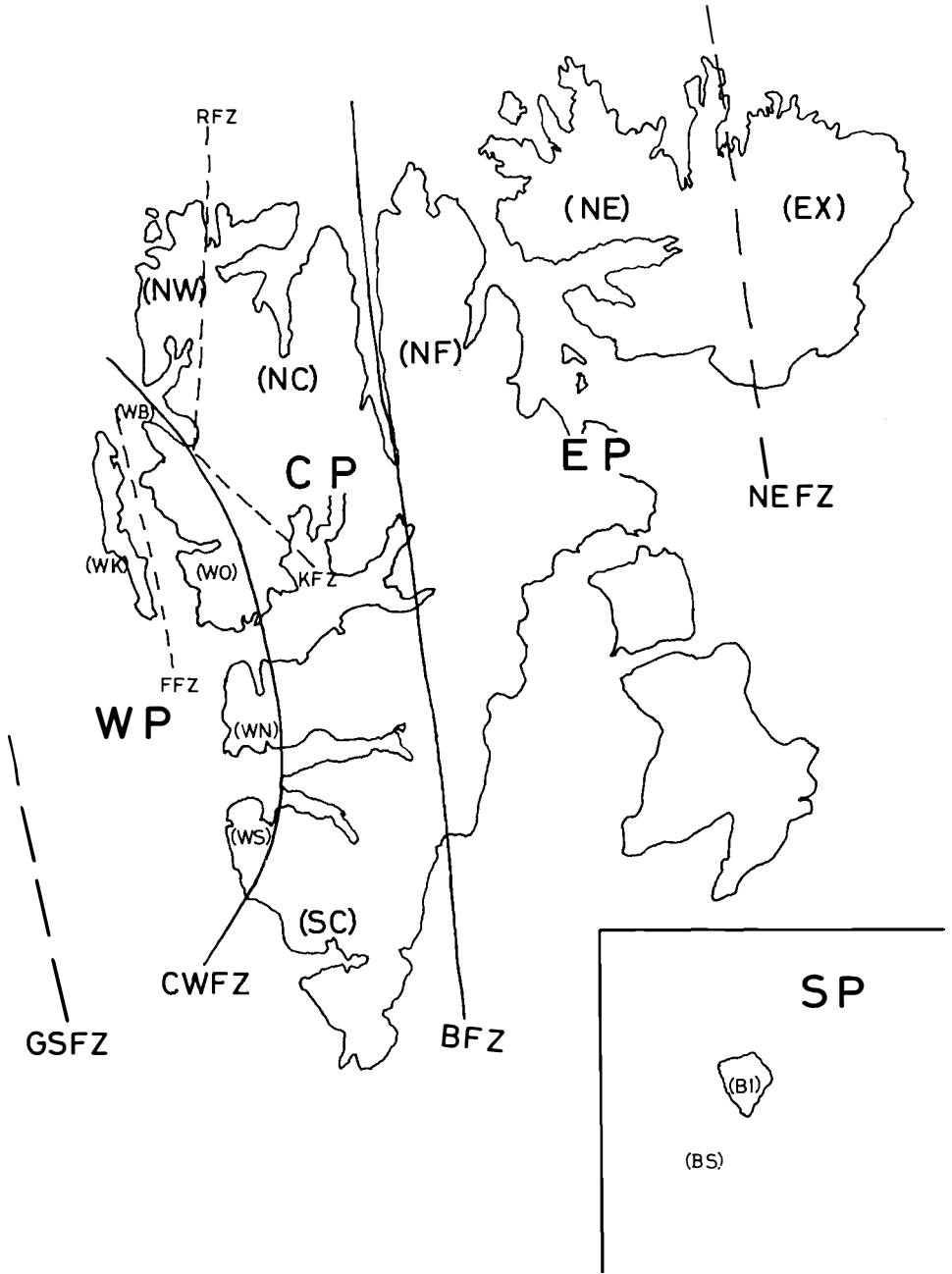


Fig. 1. Map of Svalbard showing location of pre-Carboniferous complexes, the suggested provinces and the boundary fault zones. Abbreviations of complexes and provinces are shown in Table 1 (in brackets).

- CWFZ - Central-West Fault Zone
- GSFZ - Greenland-Svalbard Fault Zone
- FFZ - Forlandsundet Fault Zone
- RFZ - Raudfjorden Fault Zone
- KFZ - Kongsvegen Fault Zone
- BFZ - Billefjorden Fault Zone
- NEFZ - North-East Fault Zone

of the same phase as the BFZ. We name this the Central West Fault Zone (CWFZ, see Fig. 1). We suggest that the fault zone is concealed beneath the front of the West Spitsbergen Orogen (it may even have provided some control over the location of the orogen) and thus no continuous surface expression of the fault zone is preserved. During sinistral transurrence in late Devonian time the CWFZ would have been straight; its present postulated curved shape would be the result of dextral transpression and overthrusting during the mid-Cenozoic West Spitsbergen Orogeny (LOWELL 1972; HARLAND and HORSFIELD 1974).

The CWFZ passes within Torellbreen, dividing the north-west from the rest of Wedel-Jarlsberg Land, and to the east of the older rock outcrops up the west coast to Kongsfjorden. While Kongsfjorden seems to be the most probable route, a splay of the fault might run directly to the north, determining the line of the Raudfjorden Fault Zone.

The Kongsfjorden–Kongsvegen line (KFZ) may be considered in two parts. The north-western part is equated to the northern extension of the CWFZ, while a possible extension to the south-east (as for example in KRASIL'SHCHIKOV'S map 1973) may be another splay.

The major tectono-stratigraphic provinces proposed here, along with their constituent complexes and the major fault boundaries suggested, are shown in Figure 1.

4. APPROACH TO THIS STUDY

The four provinces indicated above will be outlined, in terms of their constituent complexes, in Chapters II to V. In Chapter VI we consider those characters that make long-distance international time-correlation possible. We then compare the provinces by reference to those other characters of more regional or local significance, in order to assess alternative hypotheses of palaeopositions. Two such hypotheses are compared in the last chapter.

II. Eastern Province

1. NY FRIESLAND AND WESTERN NORDAUSTLANDET COMPLEX

The Eastern Province comprises the most completely elucidated pre-Silurian (Hecla Hoek) sequence in Svalbard, whose type development is in North Ny Friesland at Heclahuken and whose earliest investigators worked also in western Nordaustlandet. Lithofacial correlation across Hinlopenstretet is sufficiently detailed to justify including rocks on both sides of the strait in the same province. An apparently continuous sequence has been published and most formations have been divided into two or more members, many of which can be correlated, but only the formations and groups are correlated in Table 2. Convenient sources (taking only recent references in each case) are KULLING 1934; HARLAND and WILSON 1956; HARLAND, WALLIS, and GAYER 1966; FLOOD et al. 1969; KRASIL'SHCHIKOV 1973; FORTEY and BRUTON 1973; WINSNES 1965.

Table 2.

NY FRIESLAND		NORDAUSTLANDET	
<p><u>Osløbreen Group</u> (1.2km)</p> <p>Valhallfonna Fm (limestones)</p> <p>Kirtonryggen Fm (limestones and dolomites)</p> <p>Tokmannen Fm (dolomites and sandstones)</p> <p><u>Polarisbreen Group</u> (0.8km)</p> <p>Drakøisen Fm (shales)</p> <p>Wilsonbreen Fm (tillites)</p> <p>Eibobreen Fm (shales)</p> <p><u>Akadamikerbreen Group</u> (2km)</p> <p>Backlundtoppen Fm (dolomites and shales)</p> <p>Draken Conglomerate Fm</p> <p>Svanbergfjellet Fm (limestones and dolomites)</p> <p>Grusdievbreen Fm (limestones)</p> <p><u>Vesteranen Group</u> (3.8km)</p> <p>Oxfordbreen Fm (shales)</p> <p>Glasgowbreen Fm (greywackes and quartzites)</p> <p>Kingbreen Fm (quartzites and shales with greywackes and carbonates)</p> <p>Kortbreen Fm (quartzites and limestones)</p> <p><u>Flanefjella Group</u> (4.8km)</p> <p>Vildadalen Fm (semipelites, psammites and quartzites)</p> <p>Flåen Fm (semipelites, psammites and quartzites with acid pyroclastics)</p> <p><u>Harkerbreen Group</u> (4.1km)</p> <p>Sjøbreen Fm (quartzites and psammites)</p> <p>Vassfaret Fm (semipelites, psammites and amphibolites)</p> <p>Bangenhak Fm (feldspathites, psammites and amphibolites)</p> <p>Rittervatnet Fm (psammites, semipelites and amphibolites)</p> <p>Polhem Fm (quartzites and amphibolites)</p> <p><u>Finlandsvægen Group</u> (2.7km)</p> <p>Smutsbreen Fm (semipelites and marbles)</p> <p>Eskolabreen Fm (feldspathites, semipelites and amphibolites)</p>		<p>MURCHISONFJORDEN SUPERGROUP</p> <p>Kap Sparre Fm (limestones)</p> <p><u>Gófia Group</u> (0.6km)</p> <p>Klackberget Fm (shales with sandstones and marls)</p> <p>Sveanor Fm (tillites)</p> <p>Backaberget Fm (shales with sandstones and marls)</p> <p><u>Raoldtoppen Group</u> (1.3km)</p> <p>Ryssø Fm (dolomites with limestones, cherts and shales)</p> <p>Hunnberg Fm (limestones)</p> <p><u>Celsiusbreen Group</u> (2.1km)</p> <p>Raudstup-Skjold Fm (dolomites and sandstones)</p> <p>Norvik Fm (sandstones with shales)</p> <p>Flora Fm (sandstones)</p> <p><u>Franklinsundet Group</u> (1.8km)</p> <p>Kapp Lord Fm (sandstones with sandstones and limestones)</p> <p>Westmanbukta Fm (mudstones and siltstones)</p> <p>Persberget Fm (quartzites and shales)</p> <p><u>Botnialvalva Group</u> (>6km) UNCONFORMITY</p> <p>Kapp Platen Fm (quartzites and shales)</p> <p>Austfonna Fm (quartzites and limestones)</p> <p>Brennevinsfjorden Fm (quartzites, siltstones and shales)</p> <p>Kap Hansteen Fm (acid volcanics)</p>	
<p>HINLOPENSTRETET SUPERGROUP</p>		<p>LOMFJORDEN SUPERGROUP</p>	
<p>MOSSEL PHASE</p>		<p>ATOMFJELLA PHASE</p>	
<p>STUBENDORFFBREEN SUPERGROUP</p>			

In summary, this Ny Friesland Geosyncline totals at least 18 km in thickness. The two lower groups (Finnlandveggen and Harkerbreen), which are combined as the Atomfjella phase of development by KRASIL'SHCHIKOV (1973), are characterized by acid and basic volcanics but by only a few conglomerates and one tilloid sequence in the Rittervatnet Formation. The Planetfjella and equivalent Botniahalvøya Group (KRASIL'SHCHIKOV's Mossel phase) are characterized by mainly acid volcanics. Thereafter almost no igneous activity is evident and the sequence assumes miogeosynclinal characteristics by degrees, through greywackes to platform-type limestones and dolomites of late Riphean through Llanvirnian age, interrupted only by the clastic sequence correlated with the Varangian ice age. No significant break in this sequence has been established and there is very little evidence of tectonic mobility except for volcanic components becoming increasingly coarse to the east. The highest rocks in the sequence contain one of the richest known mid-Ordovician faunas.

2. EASTERNMOST COMPLEX (EX)

This comprises the most distant area of eastern Nordaustlandet and the islands further east such as Størøya and Kvitøya. The strata are entirely metamorphosed. The only radiometric ages are Caledonian (GAYER et al. 1966) and these are associated with extensive migmatites. The sequences could be coeval with the "Lower Hecla Hoek" but, as no precise correlation has been attempted, the possibility of an independent province should be considered. It was once proposed that this area was one of Archaean basement rocks and it seems that Krasil'shchikov separates this province on that account. If there was an ancient craton then parts of it have been remobilized. We keep an open mind on this question.

If a different easternmost province be considered then a zone of one or more faults may be implied, and we suggest the name North East Fault Zone (NEFZ) for this possibility. It may be useful not only as a key to the structures of Nordaustlandet but also for interpreting the structure of the Barents Shelf to the south.

III. Central Province

The province comprises three different terrains. The North Central area is occupied by Old Red Sandstone rocks, resting unconformably on a metamorphic complex that may be continuous with the north-west coast of Spitsbergen. There are faults dividing these older rocks and so, until correlation is secure throughout the region, a North West Complex is distinguished. Both these areas are separated from the South Central (Hornsund) Complex by outcrops of younger rocks of the Platform Sequence. We see no reason why all three complexes should not be combined into a single province.

1. NORTH CENTRAL COMPLEX (NC)

The Old Red Sandstone sequence began with the recently discovered Siktefjellet Group (GEE and MOODY-STUART 1966), of unknown (possibly late Silurian) age, that rests unconformably on metamorphic rocks with late

Table 3.
Old Red Sandstone sequences of Central Province.

<i>North Central Province</i>		<i>South Central Province</i>
<i>Andrée Land Group (5 km)</i>		<i>Marietoppen Group (1 km)</i>
Wijde Bay Fm	Mimer Valley Fm	Upper
Grey Hoek Fm		Middle
Wood Bay Fm	Wood Bay Fm	Lower
<i>Red Bay Group (2 km)</i>		
Ben Nevis Fm		
Fränkelryggen Fm		
Andréebreen Sandstone		
Red Bay Conglomerate		
UNCONFORMITY		
<i>Siktefjellet Group (<2 km)</i>		
Siktefjellet Sandstone		
Lilljeborgfjellet Conglomerate		

Caledonian radiometric ages. Then followed folding, faulting and possible thrusting (the Haakonian movements of GEE 1972), and a new unconformity developed. The clastic Red Bay Conglomerates are basal to the remainder of the relatively fine-grained, mostly continental, Old Red Sandstone sequence that persisted there without break or widespread conglomerates until Eifelian time (e.g. FRIEND and MOODY-STUART 1972).

The break which followed, before latest Devonian or earliest Carboniferous time, accommodated the Svalbardian movements with significant strike slip movement at least along the Billefjorden Fault Zone (HARLAND et al. 1974) and by analogy probably along other faults, e.g. CWFZ.

2. NORTH WESTERN COMPLEX (NW)

This large area comprises not only the mainland of northwestern Spitsbergen but also the off-lying islands (e.g. Amsterdamøya etc.), and extends from Raudfjorden to Kongsfjorden. It is a unified terrain with a coherent stratigraphy over large areas, with metasedimentary rocks of limited type passing downwards into migmatites, and cut by a late tectonic granite pluton.

The range of rock types led to correlation of these with the Lower Hecla Hoek Stubendorffbreen Supergroup of Ny Friesland (HARLAND 1960), and when the sequence was established with three formations (Table 4) GEE and HJELLE (1966) correlated them in detail (on the basis of marbles) with Ny Friesland. These arguments are good only insofar as the rocks might have formed at something like their present separation.

Most radiometric dates indicate that the main metamorphism and granite formations were Caledonian, but there remains the possibility of residual late Precambrian ages (GAYER et al. 1966). If late Precambrian diastrophism occurred this would distinguish the sequence from that in Ny Friesland where there is no such break, and this possibility was used to suggest closer proximity of the NW complex to NE Greenland where the Carolinian Orogen (HALLER

Table 4.

Sequence of older rocks of North West Complex.

Generalfjella Formation (2 km)

Marbles and interbanded marbles, pelites and quartzites in the lower part

Signehamna Formation (2–2.5 km)

Pelites with psammites and subordinate quartzites

Nissenfjella Formation (3 km)

Pelites with subordinate amphibolites and psammites. Feldspathic gneisses. Passes down into migmatites.

Migmatite

1971) appears similarly to contrast with the sequence in Central East Greenland (HARLAND 1969a).

Correlation of rocks north and south of Kongsfjorden is not at all obvious, and so in this paper we separate them by the CWFZ. Even if the rocks are not equivalent the NW complex could possibly underlie the western sequence stratigraphically or tectonically. Correlation with the South Central Complex is not obvious but the oldest rocks that might be of equivalent age are few and might not be coeval, so there is no contradiction in combining them into one province.

3. SOUTH CENTRAL COMPLEX (SC)

This comprises the older rocks outcropping in south Spitsbergen, and includes both pre-Devonian strata and the Devonian Marietoppen Group of Old Red Sandstone facies. These have been investigated from Hornsund, first to the south by Norsk Polarinstitut expeditions in Sørkapp Land (MAJOR and WINSNES 1955) and then to the north, in southern Wedel Jarlsberg Land, by Polish expeditions (e.g. BIRKENMAJER 1958). The Devonian strata have been correlated with the Old Red Sandstone sequence to the north, and are shown in Table 3. Correlation of the pre-Devonian strata north and south of Hornsund was effected in some detail (e.g. BIRKENMAJER 1960), so these are not distinguished in Table 5.

Seeking a provincial boundary to fulfil the function of the CWFZ proposed above, a hypothetical Birgebukta fault was proposed (HARLAND 1972a), but closer study shows that a N-S fault so far east would not fit the evidence. Therefore the CWFZ is now postulated to be west of the whole sequence shown in Table 5.

Correlation has been readily assumed between SC and NE sequences (e.g. HARLAND 1960 and BIRKENMAJER 1958) but a close connection between these areas is now called into question.

IV. Western Province (WP)

This province comprises the area from Kongsfjorden to Bellsund-Recherchefjorden. The constituent complexes outcrop in the following geographical areas: 1. Prins Karls Forland; 2. Brøggerhalvøya; 3. Oscar II Land (less

Table 5.
Sequence of older rocks of South Central Complex.

- Sørkapp Land Group* (1.7 km)
- h Arkfjellet Fm (slates and dolomites)
 - g Sjdano fjellet Fm (limestones and dolomites)
 - f Tsjebysovfjellet Limestone
 - e Rasstupet Limestone
 - d Nigerbreen Limestone
 - c Dusken Limestone
 - b Luciapynten Dolomite
 - a Wiederfjellet Quartzite

UNCONFORMITY

- Sofiekammen Group* (0.8 km)
- b Slaki Fm b2 Slaki Limestone
 b1 Olenellus Shale
 - a Blåstertoppen Dolomite

- Sofiebogen Group* (2.5 km)
- c Gåshamna Fm (phyllites with quartzites and limestones)
 - b Höferpynten Fm (limestones, dolomites and chert)
 - a Slyngfjellet Conglomerate

UNCONFORMITY

- Deilegga Group* (3.5 km)
- c Upper Fm (slates and phyllites with quartzites)
 - b Middle Fm (slates and phyllites with dolomites)
 - a Lower Fm (dolomites and conglomerate)

- Eimfjellet Group* (1.5 km)
- b2 Vimsodden Fm (slates and phyllites with amphibolites, limestones and schists)
 - b1 Skålfjellet Fm (amphibolites, schists and quartzites)
 - a Gulliksenfjellet Fm (quartzites)

- Isbjørnhamna Group* (1.5 km)
- c Revdalen Fm (garnetiferous mica schists)
 - b Arikammen Fm (garnetiferous mica schists and marbles)
 - a Skoddefjellet Fm (garnetiferous mica schists)

(After BIRKENMAJER 1960, with his "series" raised to the status of Formation etc.)

Brøggerhalvøya); 4. Western Nordenskiöld Land; 5. Western Nathorst Land (a very small outcrop); 6. Northern Wedel Jarlsberg Land. The northern boundary is a clear tectonic line. The boundary in the south may not be geologically obvious though a major fault line has been shown here by KRASIL'SHCHIKOV (1973 p. 103).

The northern part of the province, north of Isfjorden (areas 1 to 3), is the subject of another paper in this symposium (HARLAND, HORSFIELD, MANBY, and MORRIS) from which the details (in Table 6, columns 1 and 2) are entirely derived. Western Nordenskiöld Land and northern Wedel Jarlsberg Land are shown in column 3 and depend on work by HJELLE (1962 and 1969).

Table 6.

PRINS KARLS FORLAND	OSCAR II LAND	NORDENSKIÖLD LAND	NORTH WEDEL JARLSBERG LAND
<p><u>Graepian Group</u> (3.6 km)</p> <p>Geddesflya Fm (turbidites)</p> <p>Fugelbuk Fm (quartzites, slates)</p> <p>Barents Fm (turbidites)</p> <p>Conqueror Fm (quartzites)</p> <p>Utne Fm (slates, quartzites)</p> <p>Sofia Group (1.0 km)</p> <p>Roysha Fm (siltstones, slates)</p> <p>Keggen Fm (slates)</p> <p>Bakkla Fm (siltstones, slates, limestones)</p> <p><u>Penselfly Group</u> (1.3 km)</p> <p>Knyvødden Fm (phyllites)</p> <p>Hornnes Fm (phyllites, sandstones)</p> <p>Almasaihornet Fm (volcanics)</p> <p>Fisherlaguna Fm (phyllites)</p> <p><u>Geikie Group</u> (0.77 km)</p> <p>Rosbukta Fm (sandstones)</p> <p>Gordon Fm (limestones)</p> <p><u>Ferrier Group</u> (0.73 km)</p> <p>Neupiggen Fm (tillites)</p> <p>Peterbukta Fm (greywackes)</p> <p>Hardiefjellet Fm (tillites)</p> <p>Isachsen Fm (tillites, schists, volcanics)</p> <p>? Pinkie Fm (0.2 km) (metavolcanics)</p>	<p>BELLSUND</p> <p><u>Hullbreen Group</u> (0.7 km)</p> <p>Holmeleefjella Fm (calc slates, conglomerates)</p> <p>Kotalafjella Fm (limestones)</p> <p>Sarsflya Fm (0.5 km) (marbles, argillites)</p> <p><u>Comfortlessabreen Group</u> (2-4 km)</p> <p>Engelsbukta Fm (tillites)</p> <p>Annbreen Fm (quartzites)</p> <p>Hnaken Fm (tillites)</p> <p><u>St. Jørgen Group</u> (3.8 km)</p> <p>Alkhorn Fm (limestones)</p> <p>Lovlebbreen Fm (quartzites, volcanics)</p> <p>Meefjellet Fm (dolomites)</p> <p>Tromheimfjellet Fm (varied lithologies)</p> <p><u>Kongsvogen Group</u> (3.1 km)</p> <p>Nilsenfjellet Fm (Mullerneset Fm) (pelites, quartzites)</p> <p>Steenfjellet Fm (slates)</p> <p>Bogegg Fm (schistose pelites)</p> <p>? Vestgøtabreen Fm (blue schist metavolcanics)</p>	<p>BELLSUND</p> <p><u>Kapp Lindø Group</u> (1.3 km)**</p> <p>(NORDENSKIÖLD LAND)</p> <p>Laagneset Fm (tillites)</p> <p>Kapp Martin Fm (shales)</p> <p><u>Hullsaund Group</u> (1.4 km)**</p> <p>Laagnesflya Fm (dolomites, limestones)</p> <p>Konglomeratfjellet Fm (quartzites, dolomites, limestones, conglomerates)</p>	<p>BELLSUND</p> <p><u>Kapp Lyell Fm</u> (tillites, conglomerates)</p>

[** Hjelte (1962 and 1965)]

Visits by WBH to these areas south of Isfjorden have strengthened the view that correlation throughout the Western Province is feasible and correlation with Ny Friesland, for example, is not so obvious. Characteristic of the Western Province are thick flyschoid and conglomeratic sequences with volcanic horizons below, within, and above the tillite horizons. Also, the very rich sequence of turbiditic conglomerates associated with the Kapp Lyell tillite has no analogue a short distance away in the Southern Province; we satisfied ourselves from the literature that the Slyngfjellet conglomerate is not coeval.

HARLAND et al. (this symposium) postulate for the northern part of this Western Province a geosyncline of 18 to 20 km thick, of distinct facies, a substantial part of it consisting of Palaeozoic rocks up to Wenlock or Ludlow age. The sequence shows similarities with that of North Greenland or the Queen Elizabeth Islands and leaves place for orogeny of Ellesmerian rather than Ny Friesland-Caledonian affinity.

V. Southern Province (SP)

Bjørnøya is a distinct and isolated entity that completes the land areas in Svalbard exposing pre-Carboniferous rocks. KRASIL'SHCHIKOV and LIVSHITS (1974) recognised two pre-Devonian structural complexes, of late Precambrian and Ordovician age, and their sequence is related to the original descriptions of HOLTEDAHL (1920) and HORN and ORVIN (1928) in Table 7.

Table 7.
Sequence of older rocks in Bjørnøya.

KRASIL'SHCHIKOV and LIVSHITS 1974		HORN and ORVIØ 1928		
Devonian	Røedvika Formation	Ursa Sandstone		
Ordovician	L	Ymerdalen Fm.	Upper Member	Tetradium Limestone 240 m
	M-L		Middle Member	Younger Dolomite 400 m
			Lower Member	
Vendian	Sørhamna Formation	Slate-Quartzite 175 m		
Late Riphean	Russehamna Formation	Older Dolomite 400 m		

The Caledonian structures of Bjørnøya contrast strongly with the rest of Svalbard. A hiatus occurred during Cambrian time with a phase of minor folding (as has been suggested for south Spitsbergen) but the overlying Ordovician rocks are almost flat-bedded and show few signs of major tectonic activity.

VI. Time-correlation of sequences

Characters in the sequences of the three provinces that have value in international time correlation are tabulated, with the object of distinguishing those with long range correlation potential from those which would only be useful

over a short distance. Conclusions from the following are set out in Fig. 2. Age determinations are treated according to the nature of the characters used: 1 biostratigraphic; 2 palaeoclimatic; 3 radiometric; 4 lithotectonic.

1. BIOSTRATIGRAPHIC SEQUENCE

A. *Devonian*

The coal-bearing continental sandstones and shales of the Røedvika Formation in Bjørnøya are dated as Famennian to Tournaisian on the basis of macro- and microfloral assemblages (SCHWEITZER 1969; KAISER 1970, 1971). However, they are both lithologically and structurally part of the post-Svalbardian Billefjorden Group.

Middle and Lower Devonian fish of Spitsbergen are among the richest Old Red Sandstone faunas anywhere and compare closely with British continental faunas which, in southern England, interdigitate with the type marine sequence of Europe.

A relatively limited fauna is found in the Mimer Valley Formation, comprising several fish groups (especially Psammosteids and the Antiarch *Asterolepis*) which have long been recognised as being of late Middle Devonian age (Givetian) (for fuller list of Devonian faunas see FRIEND 1961). The formation also contains plant remains (and a cannel coal), and VIGRAN (1964) described two microfloral assemblages from near the top and near the base of the formation. These may be dated with some confidence as Frasnian and Late Givetian respectively.

The Wijde Bay Formation contains an assemblage of Heterostraci, Arthrodira, Antiarchi, Crossopterygii, molluscs and plants. FØYN and HEINTZ (1943) regarded the assemblage as Givetian.

The underlying Grey Hoek Formation contains Osteostraci, ?Heterostraci, Arthrodira, Petalichthyida, Crossopterygii, bivalves, gastropods, ostracods, charophytes and land plants. The fish genera *Lunaspis*, *Homostius*, *Heterostius* and *Porolepis* are significant, and together with the molluscs suggest an Emsian to early Eifelian age.

The Wood Bay Formation has yielded a very rich vertebrate fauna and has been subdivided into three faunal divisions (FØYN and HEINTZ 1943; FRIEND, HEINTZ and MOODY-STUART 1966), of which the lower two are used for international correlation. The "Lykta" fauna consists of Osteostraci, Heterostraci, Arthrodira, Crossopterygii and Charophyta; the significant genera are *Doryaspis*, *Arctulepis*, *Homostius*, *Actinolepis* and *Porolepis*, and suggest a mid-Siegenian to Emsian age (Breconian) (DINELEY 1960). The lowest faunal division is similar, but the Cephalaspids *Gigantaspis* and *Arctaspis* are present and suggest a Siegenian age.

The two higher formations of the Red Bay Group are fossiliferous, containing Osteostraci, Heterostraci, Acanthodii, Arthrodira, ostracods, bivalves, worms, Merostomata and plants, and, largely on the basis of close comparison of the Heterostracan and Cephalaspid genera with those of the Anglo-Welsh area, the assemblages have been dated as Dittonian (Ben Nevis Formation) and

Late Downtonian/Dittonian (Frænkelryggen Formation) (i.e. mid and late Gedinnian).

B. *Silurian*

The Bulltinden Member in the Bullbreen Group is a coral-bearing shelf carbonate limited to the Western Province (HARLAND et al. this Symposium). The fauna has been provisionally identified as of Wenlock or Ludlow age (SCRUTTON, HORSFIELD and HARLAND 1976).

C. *Mid-Early Ordovician faunas*

The "Tetradium Limestone" of Bjørnøya yields a fauna including *Tetradium syringoporoides* and the cephalopod genera *Gonioceras* and *Actinoceras*. HOLTEDAHL (1920) stated that the fauna indicates a Black River (Bolarian) age.

The underlying "Younger Dolomite" contains a fossiliferous horizon 250 m below the top which yields the genera *Calathium*, *Atchaeoscyphia* and *Piloceras*, indicating a Canadian age.

Better defined Canadian faunas occur in the south Central Province (Sørkapp Land Group), where gastropods, brachiopods, sponges and cephalopods occur. The fauna and facies resemble the Beekmantown Group of (eastern) North America (MAJOR and WINSNES 1955; BIRKENMAJER 1958, 1960).

The Eastern Province contains exceptionally rich trilobite and graptolite faunas in the Valhallfonna Formation. Several assemblages have been distinguished (FORTEY and BRUTON 1973); the youngest is dated as early Llanvirn, while the base of the formation contains an early Arenig graptolite fauna.

The underlying Kirtonryggen Formation contains abundant trilobites, brachiopods, nautiloids, gastropods and ostracods. Two assemblages, in the top and bottom members of the formation, are of stratigraphic importance and indicate Upper and Lower Canadian ages respectively when compared with western North America.

D. *Cambrian faunas*

No definite Middle or Late Cambrian faunas are known anywhere in the Western Arctic, although poorly preserved brachiopods (probably obolids)

←

Fig. 2. Correlation scheme of pre-Carboniferous strata in Svalbard. Height of columns represents thickness of strata preserved. Characters used in international tim. correlation and local correlation are shown in the key.

The names of each group, with brief lithology of formations, are given in Tables 2 to 7. Group names are abbreviated in the figure beside each column.

- (a) Bjørnøya (Table 7)
- (b) Nordenskiöld Land and N.W. Wedel Jarlsberg Land (Table 6)
- (c) Prins Karls Forland (Table 6)
- (d) Oscar II Land (Table 6)
- (e) North west Spitsbergen (Table 4)
- (f) North central Spitsbergen (Table 3)
- (g) South Spitsbergen (Tables 3 and 5)
- (h) Ny Friesland (Table 2)
- (i) Western Nordaustlandet (Table 2)
- (j) Eastern Nordaustlandet

occur in the top part of the Sofiekammen Formation of Hornsund which may be Late Cambrian. COWIE (1974) has suggested a hiatus as the explanation in the Ny Friesland area; alternatively tropical shallow marine conditions could have been chemically inimical to the occurrence or preservation of faunas within the lower Kirtonryggen dolomites.

In the south of the Central Province a rich trilobite fauna was described by MAJOR and WINSNES (1955) from the Slaki Formation (Sofiekammen Group), including *Serrodiscus*, *Callodiscus* and *Olenellus* species. They correlated this fauna with the late Early Cambrian (Georgian) of North America.

In the Eastern Province *Salterella* cf. *rugosa* has been known from the Tokamane Formation of Ny Friesland (HARLAND and WILSON 1956) with hyolithids and a single *Olenellus*. The supposed equivalent of these rocks in Nordaustlandet is the Kap Sparre Formation in which Hecla Hoek fossils were first found (inarticulate brachiopods). These may not be chronostratigraphically distinctive and a superficial re-examination of the Kap Sparre section by WBH suggested that correlation of these rocks with the Ordovician part of the Kirtonryggen Formation should be examined.

E. Vendian

Relatively abundant Vendian and Late Riphean floras of stromatolites, microphytolites, oncolites and katagraphites exist in the late Precambrian carbonates of Nordaustlandet, Ny Friesland, Hornsund and Bjørnøya. The stratigraphic value of some of these algal forms may be dubious but they have been used by Soviet workers to distinguish several Precambrian age-assemblages on a broad scale.

Vendian assemblages have been described from the Backaberget Formation of Nordaustlandet, where dolomite bands contain the oncolite species *Osagia svalbardica* and the katagraphite *Vermiculites irregularis* (KRASIL'SHCHIKOV, GOLOVANOV and MIL'SHTEIN 1965). In the Polarisbreen Group shales of Ny Friesland, MIL'SHTEIN reported microphytolite assemblages from the *Vesicularites*, *Nuberculerites*, *Radiosus* and *Volvatella* groups, correlated with groups in the Vendian of the Urals and Timan (MIL'SHTEIN in KRASIL'SHCHIKOV 1973 and this volume).

F. Late Riphean

Late Riphean stromatolite assemblages have been described from the Hunnberg and Ryssö Formations of Nordaustlandet by GOLOVANOV (1967), including species of the groups *Conophyton*, *Kusiella*, *Inseria*, *Gymnosolen* and *Tungussia*, which he compared with the Upper Riphean Karatau rocks of the Southern Urals. The Hunnberg Formation also contains the katagraphite *Vesicularites flexuosus* and the oncolite species *Osagia columnata*, again indicating Late Riphean age. Further down the succession, the Norvic Formation is said to contain oncolites of the *Astrospheroides* and *Radiosus* groups, of Riphean age, while *Osagia* of a mid-Riphean type is reported from the Kapp Lord Formation (GOLOVANOV and RAABEN 1967).

In Ny Friesland, stromatolite and microphytolite assemblages occur in the Backlundtoppen, Draken and Svanbergfjellet Formations of the Akademikerbreen Group. The highest formations contain oncolites of the *Osagia* group with both Upper Riphean and Vendian affinities, and stromatolites of the *Conophyton* and *Tungussia* groups. The Draken Formation contains microphytolites (*Vesicularites* and *Radiosus* groups) of possible Vendian age (RAABEN and ZABRODIN 1969). The underlying Svanbergfjellet Formation contains Upper Riphean stromatolites of the *Inseria*, *Conophyton*, *Tungussia* and *Gymnosolen* groups and many microphytolites of the *Vesicularites* and *Osagia* groups. A late Riphean microphytolite assemblage of *Radiosus*, *Asterospheroides* and *Volvatella* types occurs in the Oxfordbreen Formation, while the Kingbreen Formation contains stromatolites of the *Inseria* and *Collenia* types as well as microphytolites of the *Rachosis* and *Osagia* groups, also indicative of late Riphean age (RAABEN and ZABRODIN 1969).

In south Spitsbergen, stromatolites occur in the upper part of the Høferpynten Formation. They were described by BIRKENMAJER (1972) who doubted their stratigraphic value but correlated the horizon with the Draken Formation of Ny Friesland. MIL'SHTEIN (this Symposium) reports a microphytolite assemblage of possible Middle-Late Riphean age, suggesting perhaps a correlation with the Veteranen Group of Ny Friesland.

The Older Dolomite (Russehamna Formation) of Bjørnøya is reported by MIL'SHTEIN to contain microphytolites of Late Riphean type.

2. PALAEOCLIMATIC CORRELATION

A. Ordovician

The possibility that a tilloid in the Holmesletfjella beds of Oscar II Land was a tillite had been seriously considered (HARLAND 1972b), but a further visit in 1975 to the locality showed (by associated conglomerates) that this pebbly mudstone was probably not glacial. Indeed, almost simultaneously, its Silurian age was established. No other rocks have been suggested as Ordovician tillites.

B. Varangian

There is still some argument about the origin of the Svalbard tilloids but even those who are sceptical of a glacial origin agree that the Arctic tilloids provide good marker horizons for correlation. However, if a glacial origin be admitted, and the evidence for this is strong (e.g. CHUMAKOV 1968), then this distinct climatic episode has a wider international correlation potential. Low latitude glaciation in this part of the Arctic has been argued (e.g. HARLAND 1964; HARLAND and HEROD 1975) and so what has been referred to as the Varangian Ice Age provides a powerful correlation character, in association with which orthoconglomerates should not be ignored.

There can be little doubt of the correlation value, both local and international, of the Varangian tillites of the Eastern Province. The Sveanor tillite is well developed between the Klackberget and Backaberget Formations; similarly, in the Polarisbreen group, the Wilsonbreen tillite lies between shales

of the Drakoisen and Elbobreen Formation (see Table 2). The tillites often occur in two members as is characteristic of Varangian tillites generally.

Tillite horizons of the Western Province are different; they are more deformed and often slightly metamorphosed so that much primary evidence of sedimentation has been lost; they are in very much thicker flysch sequences so that they are associated with both slumped and turbiditic rocks and also with conglomerates. The Comfortlessbreen Group of Oscar II Land and the Ferrier Group of Prins Karls Forland are of the order of 3 to 4 km thick and each contains two tillite formations. In Nordenskiöld Land, if a similar thickness and varied facies be assumed, then most of the tilloids and conglomerates, from the Kapp Linné tillite in the north to the Kapp Martin tillite in the south, are of this age. Also in northwestern Wedel Jarlsberg Land the Kapp Lyell tillite first described by GARWOOD and GREGORY in 1898, is a remarkable, thick sequence consisting of a minority of beds that are tillitic and a great majority of coarse thick turbidite units, from 1 to 2 m, each with initial conglomerates (clasts up to 10 and 20 cm) and grading upwards to coarse or fine sandstone. The clast content has similar composition to that of the tillites.

No tillites have been recorded from the Central Province, with one possible exception (WILSON and HARLAND 1956, p. 216). From their stratigraphic position it seemed most likely to us and to Birkenmajer that the Gåshamna shales of Hornsund could be equivalent to the upper and/or lower shales of the Polarisbreen Group. The lack of obvious tillites in this sequence could be due to uplift of the Central Province at the time of lowest sea level. Conversely, the lack of conglomerates or obvious break between the pre-Varangian Høferpynten Formation and the Cambrian Sofiekammen Group would be surprising if the Kapp Lyell conglomerates and tillites were forming at anything like their present distance.

The upper and lower tillites of the type area in Varangerfjord (Norway) are separated by the Nyborg Formation which has yielded a radiometric age of 668 ± 13 Ma (PRINGLE 1973).

C. Pre-Varangian tillites

The only tilloids of this age reported from Svalbard are the Vimsodden tilloid of the South Central Province and the Rittervatnet tilloid of Ny Friesland. The latter occurs 12 km beneath the Varangian tillites, as does the Gnejssø tillite of East Greenland on which basis correlations have been attempted (e.g. HARLAND 1964). Similar correlation within Spitsbergen between the provinces is also feasible.

3. RADIOMETRIC AGES

Most radiometric ages obtained from the older rocks relate to the Central and Eastern Provinces and are Caledonian (*sensu stricto*), i.e. roughly 420 to 380 Ma. Nothing earlier than this would be expected in Ny Friesland where an unbroken sequence of older rocks is found, but there have been anomolous ages to the east and west that are difficult to interpret (GAYER et al. 1966).

In the Western Province not only is the earlier story overprinted by Mid-

Palaeozoic metamorphism (e.g. HORSFIELD 1972) but the region also suffered Mid-Cenozoic diastrophism.

In conclusion, little correlation can be done from present data (mostly K-Ar) but a further attempt at dating some of the old rocks is being made using Rb-Sr methods.

4. LITHOTECTONIC CORRELATION

The virtual absence of biostratigraphic, palaeoclimatic and radiometric characters in the oldest rocks of Svalbard presents a major problem for all but local time-correlation. The lithostratigraphic recognition of rock bodies and similar sequences of lithotypes is applicable only when the areas of comparison were physically close and can be shown to be part of a structurally continuous regime. If the four major provinces of Svalbard were once separated, attempts at lithological correlation between provinces (e.g. by volcanics or marbles) are suspect.

An attempt was made by HARLAND and GAYER (1972) to reach an independent estimate of the age of the Ny Friesland succession by comparing the observed thickness with known overall sedimentation rates from other Caledonian geosynclinal sequences. There is a clear distinction between the younger part of the succession (Akademikerbreen, Polarisbreen and Oslobreen Groups) which develops mainly shelf carbonate lithologies, and the older part (Finlandveggen, Harkerbreen, Planetfjella and Veteranen Groups) in which volcanics and clastics occur. Assuming sedimentation rates of 80–200 m/Ma for the older sequence by comparison with similar geosynclines, HARLAND and GAYER concluded an age of 1055–950 Ma (Middle–Late Riphean) for the base of the observed sequence. The method may be dubious, as it relies to some extent on a circular argument, but it gave an estimate when no other seemed possible. Because such an estimate is in effect based on a net-subsidence rate it is reasonable only to apply it in averaging large thicknesses of strata.

KRASIL'SHCHIKOV (1973) presented an overall scheme of time-correlation, with five separate large-scale chronostratigraphic divisions. His Cambro-Ordovician, Vendian and Late Riphean units are all based on independent time-correlation characters, but his older divisions (Early Riphean and Early to Mid Proterozoic) are supported by no independent evidence. We think that the correlation is based only on a lithological comparison of Svalbard sections with dated successions elsewhere and on the belief that the whole post-Archaeon sequence is present and exposed. Krasil'shchikov's correlation dated the oldest Ny Friesland rocks as 2600 Ma, which gives an average sedimentation for the "eugeosynclinal" sequence of 8m/Ma; this figure we consider to be too low by a factor of at least 10 when compared to similar dated sequences. Moreover we do not know of other geosynclines that have subsided without interruption for 2200 Ma.

VII. Characteristics of the different provinces

The object of this section is to consider those characteristics that provide evidence for the separation of the provinces.

1. BIOLOGICAL

A. *Devonian*

Outcrops of Devonian rocks in Spitsbergen are confined to the Central Province, and therefore no provincial comparisons can be made. However, a very wide correlation of the assemblages with those of the Old Red Sandstone in Central East Greenland, NE America and Britain suggests that the late orogenic "Old Red Land" was a unified faunal province with no major ecological barriers.

B. *Silurian*

Apart from the barren (possibly late Silurian) Siktefjellet Group in the Central Province the only established Silurian fauna is in the Western Province. No such fauna is known in East Greenland but Silurian faunas are known in North Greenland and in the Canadian Arctic Islands.

C. *Ordovician*

The Ordovician fauna of Ny Friesland is exceptionally rich, and is still in the process of being described. On the basis of the trilobites, FORTEY (1974) has commented on the close similarity of the fauna with that of the Cap Weber Formation in East Greenland. The youngest Ordovician fauna was compared by WHITTINGTON (1965) with the Table Head Formation of Newfoundland.

No direct comparison can be made with the Ordovician Sørkapp Land Group because, although fairly abundant gastropods, brachiopods, cephalopods and porifera occur, no trilobites have yet been found. Time-correlation has been based on comparison of the cephalopods with North American occurrences, but no detailed assemblage comparison has been noted.

The Ordovician fauna of Bjørnøya is similarly devoid of trilobites, and rather restricted in forms. Three of the genera found can be compared with genera from Sørkapp Land (MAJOR and WINSNES 1955), but the faunas are not closely related.

Thus the three provinces which yield Ordovician faunas do not as yet reveal any close similarities. Comparison of the Ny Friesland and South Spitsbergen faunas may prove to be closer than they appear at present when the cephalopods of Ny Friesland have been described. However, such a study may alternatively bring to light a real difference.

D. *Cambrian*

Cambrian faunas occur only in the Eastern and Central Provinces. In Ny Friesland the assemblage is restricted and consists mainly of inarticulate brachiopods and burrows (GOBBETT and WILSON 1960). In contrast, the Sofiekammen Formation of South Spitsbergen yields a wide trilobite assemblage of Pacific Province type, despite the development of very similar limestone/dolomite/clastic shelf facies in both provinces. This difference of faunas was noted by COWIE who concluded: "The contrast in Cambrian faunas between Ny Friesland and south-west Vestspitsbergen implies the existence at that time of a geographical-ecological barrier" (1974, p. 129).

E. *Precambrian*

Stromatolites, oncolites, and katagraphites are widespread in the shelf carbonate facies of the Eastern and Central Provinces but cannot be easily distinguished. The Western Province has so far yielded only oncolites except in the tillite clasts.

2. LITHOGENETIC

A. *Continental facies*

Old Red Sandstone facies of Early and Middle Devonian age are limited to the Central Province. The earliest Continental Sandstones in the Eastern, Western and Southern Provinces are of latest Devonian to early Carboniferous age. The Central East Greenland Old Red Sandstone is of Middle and Upper Devonian age. Old Red Sandstone facies of Devonian age are not familiar in North Greenland.

B. *Carbonate facies*

Carbonate shelf facies with relatively clean quartzites and shallow-water carbonates are developed to a greater or lesser extent in the Late Riphean to Ordovician strata of the Eastern, Central and Southern Provinces. In contrast, the Western Province is entirely lacking such “miogeosynclinal” sequences. Such facies are typical of the Cambro–Ordovician sequence of Central East Greenland but not so in Peary Land.

C. *Flyschoid facies*

Calcareous flyschoid turbidites are characteristic of much of the Western Province, especially during Varangian and Early Palaeozoic times, in contrast to stable shelf facies for these ages in the other three provinces.

D. *Conglomerates*

Apart from tilloids discussed already, the presence and abundance of conglomerates is indicative of tectonic mobility. In the Eastern Province conglomerates other than tillites are very rare, the principal exceptions being the Kapp Hansteen Formation volcanic agglomerates and the Draken intraformational conglomerates. In the Central Province there are remarkable conglomerates at the base of the Old Red Sandstone (both Siktefjellet and Red Bay) as well as the pre-Varangian Slyngfjellet Conglomerate. In the Southern Province there are no major conglomerates. In the Western Province, on the other hand, conglomerates occur at many horizons, mostly of Silurian Age (Sutor and Bulltinden Formations), as well as thick tillites and conglomerates of Varangian age.

E. *Volcanic facies*

No Palaeozoic or Varangian volcanics are known in the Eastern, Central or Southern Provinces. In the Western Province, volcanic rocks occur in three or more horizons above the Varangian tillites, probably of both Cambrian and

Ordovician age, and at least one major volcanic episode occurs within the tillite sequence.

Similarly, in late Riphean sequences no volcanics are known from the Eastern, Central or Southern Provinces but occur in the Western Province (Løvliebreen Formation).

Because of the difficulty of correlation of earlier rocks no precise meaningful comparison can be made, but it is clear that each province is different. It is probable that the Western Province does not have very old rocks of any facies – certainly the Kongsvegen Group are not volcanic. The Central Province has limited volcanics among the older rocks of the Nissenfjella Formation in the NW and the Eimfjellet Group in the South. On the other hand the whole of the lower 12 km of the Eastern Province geosyncline has a large volcanic component (acid in the Mossel phase and both acid and basic in the earlier Atomfjella phase).

3. TECTOGENETIC

The structural sequences of the provinces are distinct.

The Eastern Province evolved from a mobile eugeosynclinal sequence in late Riphean times to a stable shelf of miogeosynclinal character, and the Ny Friesland Orogeny probably spanned late Ordovician and Silurian times with batholithic emplacement. This is a typical orthotectonic Caledonian sequence, and only the local Svalbardian faulting followed in late Devonian time before a stable Carboniferous platform cover was deposited.

The Central Province is more fragmented. The North West complex may have suffered a late Precambrian orogeny (Carolinidian of HALLER 1971) before the main Caledonian sequence accumulated. The latter was associated with Old Red Sandstone continental detritus interrupted by the Haakonian diastrophism at about the end of Silurian time. In the South Central Complex the later Precambrian sequence is broken by unconformities, and was later affected by the Cenozoic West Spitsbergen Orogeny.

In both the Central and Eastern Provinces, the deposition of Billefjorden Group platform deposits began in earliest Tournaisian or late Famennian times (CUTBILL, HENDERSON, and WRIGHT, in press), and although minor faulting along the BFZ continued to affect sedimentation patterns, the two provinces were essentially united from that time.

The Western Province preserves a distinct sequence of 18–20 km (the Høltedahl Geosyncline) which is mobile and eugeosynclinal, and a substantial part of which is Varangian and Early Palaeozoic. The main deformation, involving metamorphism and overthrusting to the west, was late and/or post Silurian, and there is no Devonian record (cf. Ellesmerian rather than Caledonian). Stable Billefjorden Group platform deposition did not begin until Namurian times (Orustdalen Formation – see CUTBILL and CHALLINOR 1965), and it therefore seems probable that late orogenic movements continued in the Western Province after the Central and Eastern Provinces had stabilised.

In the Southern Province a major unconformity separates the older Precambrian strata from the Ordovician shelf carbonates, and some folding and

faulting occurred during this episode (KRASIL'SHCHIKOV and LIVSHITS 1974). However, the Ordovician strata are almost completely unaffected by Caledonian deformation; Silurian and Devonian "Old Red Sandstone" strata are absent, and Billefjorden Group sedimentation began in Famennian times.

VIII. Discussion and conclusion

When only a little information is available simplistic correlation encourages making a coherent hypothesis at an early stage. Up to now perhaps too few data have been available for alternative models to be challenging. Even now the evidence is insufficient to decide between major alternative hypotheses; but the time is ripe to consider some of them so as to focus on critical points of difference.

1. A PALAEOFIXISTIC MODEL

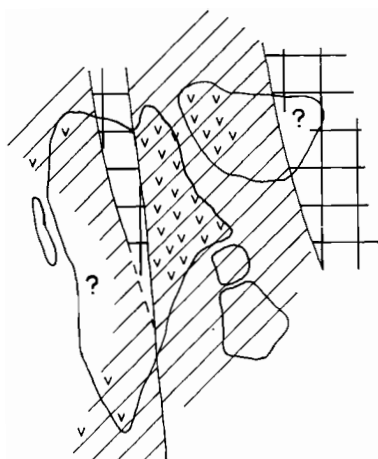
It is generally accepted that Mesozoic and Cenozoic opening of the Atlantic and Arctic Ocean basins moved Spitsbergen, along with the Barents Shelf, away from a position north of Greenland. However, many pre-Permian reconstructions have not adopted any major pre-Permian horizontal displacement. We present as one model for comparison a minimal displacement hypothesis (Fig. 3). KRASIL'SHCHIKOV (1973) postulated a sequence of early reconstructions of the Spitsbergen archipelago on a fixistic basis but we differ in postulating different boundaries to the provinces by emphasising different fault lines.

One implication of this model is that for latest Riphean through Ordovician time the thick, flyschoid sediments of the Western Province are not likely to have been transported across the Ny Friesland geosyncline which, although near sea level, was of finer sediment. The source could be in the Central Province to the north, or more probably the west.

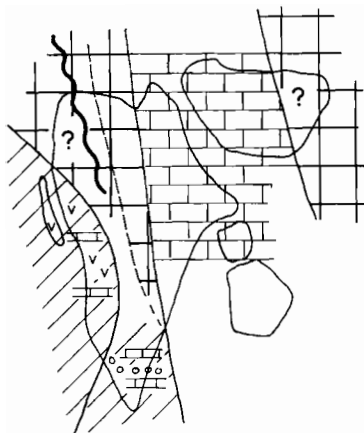
2. PALAEOMOBILISTIC HYPOTHESIS

As indicated at the outset the contrast between the three or four sequences suggests that the strata were formed at much greater distances apart than now. Comparisons between strata in Svalbard and elsewhere had already suggested an initial Svalbardian (late Devonian) strike slip so as to relate the Eastern pre-Devonian sequence to that of Central East Greenland and the post-Devonian sequence of Svalbard with that of the Canadian Arctic (HARLAND 1965). At that time the fault line was thought to be west of Spitsbergen. We now refer to this conceivable fault as the Greenland Svalbard Fault Zone (GSFZ). Subsequently (e.g. HARLAND 1969a) the possibility of faults both within and to the west of the archipelago, to achieve the same total displacement, was considered, and by 1974 (HARLAND et al.) the Billefjorden Fault Zone was established as a line of displacement (of between 200 and 1000 km).

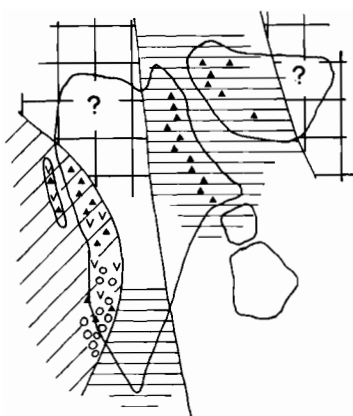
The revised hypothesis here accepts a minimum displacement of 200 km for the BFZ (say 500 ± 250 km) and postulates a further and greater displacement



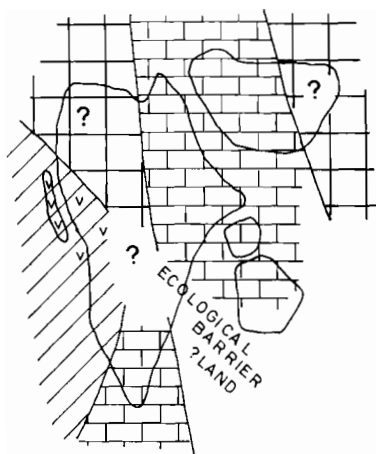
a) Middle -Late Riphean
1000(±) - 850



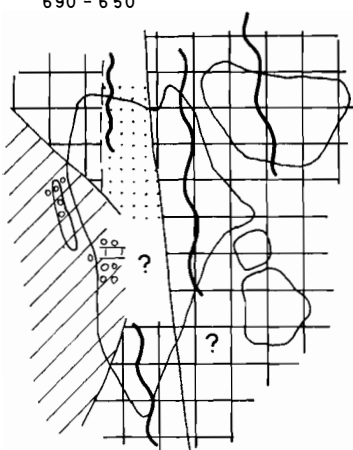
b) Latest Riphean
850 - 690



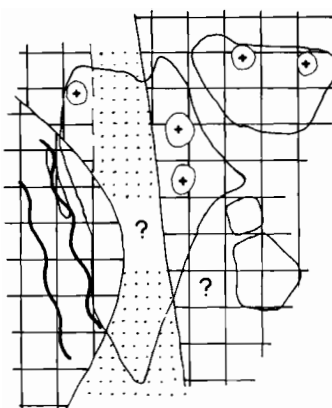
c) Varangian (Early Vendian)
690 - 650



d) Cambrian - Early Ordovician
570 - 450



e) Middle - Late Silurian
420 - 400



f) Early - Middle Devonian
400 - 360

Ages very approximate, in Ma

Fig. 4.

(say 1000 ± 500 km) along a nearly parallel WCFZ. This still leaves it open as to whether the GSFZ west of Spitsbergen operated with a similar displacement.

If we accept that within the western half of Svalbard the traces of two major faults occur (BFZ and CWFZ) of which the western line passes out to sea in the west, then it is not unreasonable to suppose that another fault still further west (GSFZ) took part in the same major sinistral displacement between Greenland and Europe. Moreover, such a fundamental fault would be the locus of Cenozoic separation. Accordingly, for this model we add another 750 ± 500 km sinistral displacement along the GSFZ.

Thus, three sinistral late Devonian transcurrent faults are proposed, of which two may be tested by the evidence given here, namely the BFZ and the CWFZ. A speculative reconstruction of palaeoposition prior to Early Palaeozoic movement is shown in Figure 4. This scheme, however, only refers to the latest pre-Svalbardian configuration (e.g. Middle Devonian) and does not accommodate any degree of orogenic compression or closure of any earlier ocean – a matter already discussed (HARLAND and GAYER 1972).

3. IMPLICATIONS OF THE PALAEOMOBILISTIC MODELS

A number of comparisons have been made in the previous chapter which suggest to us that the palaeomobilistic model is a serious contender. They will not be discussed further.

Further implications are that the two or three major Devonian sinistral faults proposed (BFZ, CWFZ and GSFZ) provided major fundamental lineaments that controlled the dextral transcurrence of the Cenozoic motions. The Cenozoic movements reactivated very slight motion on the BFZ; they were very largely concentrated on the CWFZ during the West Spitsbergen Orogeny for a limited mid-Eocene period of dextral transpression, and now operate along the Spitsbergen Fracture Zone. The line along which Spitsbergen and Greenland originally separated could have been the rejuvenated GSFZ.

It will be seen that these same ancient lineaments also controlled the events between the mid-Palaeozoic and mid-Cenozoic Orogenies. Epeirogeny during the platform sequence was limited and determined by these ancient fault

←

Fig. 3. *Sequence of facies in the provinces arranged according to palaeofixistic model. An unknown degree of E-W compression has not been restored so that the E-W distances are reduced here.*

The account generalizes facies as follows:

oblique ruling	= mobile sedimentary facies (often flyschoid);
brick ornament	= shelf type carbonate;
horizontal ruling	= shelf type shales;
stipple	= continental sandstones (ORS facies);
square ruling	= uplifted area;
circles	= conglomerates;
triangles	= tillites;
crosses	= acid plutons;
v	= volcanics;
sinuous curve	= tectogenesis;

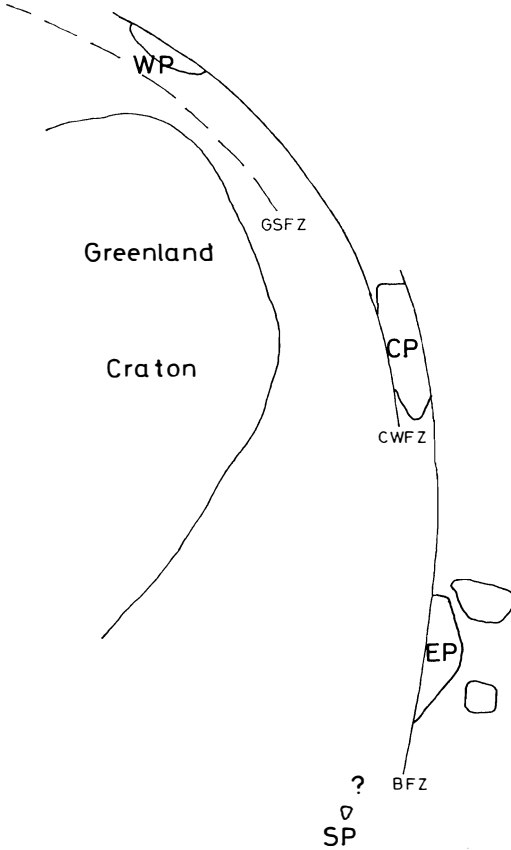


Fig. 4. Sketch to show the later Devonian transcurrent movement that juxtaposed the separate provinces. Earlier compression or possible oceanic closure is not restored, nor Cenozoic motions shown. The facies in each province in middle Devonian and earlier times may be taken from Figure 3.

blocks. This has been documented in detail for the BFZ (HARLAND et al. 1974) and a parallel study remains to be done for the CWFZ.

The postulated faults have further palaeogeological implications when the provinces are restored to their original positions. Studies of facies patterns, sediment sources, biological provinces and directions of tectonic transport will need to be considered.

The existence of Svalbard as a distinct land mass has at least two causes. The hot mantle that caused the separation of the Barents Shelf from Greenland also uplifted the north-west corner of the Barents Shelf that is the Svalbard archipelago (HARLAND 1969b). To this we may add the possibility that Spitsbergen contains at least two major Palaeozoic transcurrent zones that divided the Caledonian orogen and that this recurrent major tectonic zone of mobility accentuated the distinct structure of the archipelago.

4. CONCLUSION

We consider the palaeomobilistic model set out here to be reasonable and to provide a useful working assumption while the above implications are explored. It challenges renewed field work to seek out the many further critical points where it may be tested.

Acknowledgements

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An outline pre-Carboniferous stratigraphy of central western Spitsbergen

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Abstract

The stratigraphy of Prins Karls Forland and Oscar II Land is outlined by defining a sequence in each area with 19 formations in 5 groups and 15 formations in 4 groups respectively. Many of the units proposed and their names come from a long history of research but this paper for the first time relates these rocks into two nearly continuous successions that can be correlated by the common tillite formations (up to 4 km thick). Prins Karls Forland has 6.7 km above it and Oscar II Land has 6.9 km below it and an estimate of the total thickness is 18 to 20 km.

The only reliable fauna for dating is of Wenlock or Ludlow age in one of the upper formations. This sequence of flyschoid conglomerates and volcanic facies is distinct and is named the Holtedahl Geosyncline.

I. Introduction

This paper deals with the pre-Carboniferous stratigraphy of western Spitsbergen between Kongsfjorden and Isfjorden – that is Oscar II Land and Prins Karls Forland. These areas are two of the most accessible of all the pre-Carboniferous terrains of Svalbard and yet have recently been among the least understood.

The earliest substantial stratigraphic conclusions for Oscar II Land were drawn by HOLTEDAHL (1913) and ORVIN (1934), and later by WEISS (1953 and 1958) and WINSNES (1965), in the form of disconnected sequences for the north and south of the area. Our work began in 1958 with a reconnaissance survey by C. B. WILSON (who died in 1959). WBH visited St. Jonsfjorden and Engelskbukta briefly in 1959 in continuation of this work, and later A. CHALLINOR worked on the Tertiary Fold Belt to the east, dealing only incidentally with the older rocks. Geophysical investigations were carried out by K. HOWELLS and P. I. MATON between 1962 and 1968, but it was not until 1967 that geological work was resumed seriously – by WTH from 1967 to 1969, with visits by WBH in 1968, 1971 and later.

After many early visits Prins Karls Forland was first surveyed by the Scottish Spitsbergen Expedition (BRUCE 1910), and for long the only substantial account was that of TYRRELL (1924). Then, at the suggestion of one of us, D. J. ATKINSON and R. A. McDONALD in a tour de force reconnoitred the whole island in 1949, 1951 and 1952 (*Polar Record* 6 (44) 527 and 7 (49) 317). Thus a general idea of structure and stratigraphy was available but no single stratigraphic scheme emerged. This was the reason for our visits between 1968 and 1975, with two of us (GMM and APM) engaged in detailed work on Prins Karls Forland from 1973 to 1975; all these visits were organized by Cambridge Spitsbergen Expeditions whose support is acknowledged here by references to accounts in the *Polar Record* (1958, 9 (62) 464–465; 1959, 10 (64) 40–44; 1967, 14 (88) 43–45; 1968, 14 (91) 492–494; 1969, 15 (96) 331–332; 1971, 16 (100) 63–64; 1972, 16 (103) 579–580; 1973, 17 (106) 43; 1974, 17 (109) 383–384).

Our work is expected later to yield separate, more detailed, accounts of the geology of Prins Karls Forland and Oscar II Land. This present paper gathers together the principal stratigraphic conclusions to date. Modifications are likely as work proceeds but we believe the main outlines will stand, as they have for some time been tested.

Both areas have suffered both mid-Palaeozoic and mid-Cenozoic deformation and metamorphism. The rocks were overthrust in opposite senses (ATKINSON 1960, HARLAND and HORSFIELD 1974), and metamorphism produced rocks of biotite grade together with some mineralisation (SIGGERUD 1962, FLOOD 1969). This leads to a complex mutual dependence of stratigraphic and structural interpretation. Therefore, a fuller understanding of the stratigraphy will depend on a more detailed treatment of the structural units within which the many sequences are found. Until such time thicknesses given are provisional.

Both sides of Forlandsundet being familiar in some degree to all of us we have decided to set up two independent systems of nomenclature, since there is no

obvious correlation of many of the rocks. In the next two sections, therefore, we treat each area independently before attempting a synthesis in the last chapter.

The theme of this Symposium includes Devonian as well as earlier history. We have no knowledge of Devonian rocks occurring in our area (HEINTZ and SIGGERUD 1965). The youngest pre-Carboniferous rocks may be late Silurian. It once seemed that the first “Hecla Hoek” fossils in Oscar II Land had been discovered by the 1948 Birmingham expedition, but these proved to be of Carboniferous age (BAKER, FORBES and HOLLAND 1952). Similarly, LEE (1908) described fossils collected by BRUCE from Prins Karls Forland of Permian age; these are now presumed to have originated from glacial erratics. Tectonic slices of younger rocks are liable to be confused with the older rocks along the whole length of the area.

For the present we avoid referring to the rocks we are dealing with as Hecla Hoek because of the possibility that this sequence was formed at a great distance and in a different environment from the type Hecla Hoek rocks of Ny Friesland. For a general name we refer to the Western Complex or Sequence and the island part of it as the Forland Complex or Sequence (HARLAND et al. 1974). We describe the strata from the top downwards and begin with the Forland Complex where the preservation of the younger rocks appears to be more complete.

II. Prins Karls Forland

Although we have reconnoitred the whole island in outline, the area accessible from Scotiadalen has been surveyed in detail by two of us (GMM and APM) on a scale of 1:10,000. The stratigraphy found to apply throughout the island is named and described from the centre of the island where this detailed work has been done and where the relationship between the rocks north and south of Scotiadalen, long in doubt, has been established.

Our scheme is set out in Table 1 and the individual units are defined below, in sequence from top down. The nomenclature we have chosen follows so far as possible that of TYRRELL and ATKINSON, whose work we thereby acknowledge. This granting of historical priority is not only correct but has the advantage of yielding names that are shorter than we should now be obliged to adopt. The names we in turn have needed to introduce have been selected both for their topographical position and for their brevity (in so far as this is not seriously misleading).

As ATKINSON showed (1956 and 1960), the island consists of a number of thrust sheets each with a distinctive stratigraphy so that neither stratigraphy nor structure will be fully elucidated in isolation. It is often difficult to discover the way up for parts of the sequence and our interpretation of some units differs in this respect from that of Atkinson. In addition to sedimentary structures the clast content of the Sutorfjella Conglomerate Member has been critical for our interpretation.

One difficulty, that we hope has now been overcome, arises from the repetition of distinctive facies that were at first treated as one unit. For example,

Table 1.
Sequence of older strata in Prins Karls Forland

Group	Formation	Dominant lithology	Thickness in metres
Grampian [3.6 km]	Geddesflya	slaty turbidites	1800
	Fugelhuk	massive quartzites & slates	400
	Barents	slaty turbidites	500
	Conqueror	quartzites	850
	Utnes	grey slate & quartzite	80
Scotia [1 km]	Roysha	grey siltstones & black slates	400
	Kaggen	green & grey slates	300
	Baklia	grey siltstones, slates & limestones	200–300
Peachflya [1.3 km]	Knivodden	chloritoid phyllites	400
	Hornnes	phyllites & sandstones	350
	Alasdairhornet	volcanics	190
	Fisherlaguna	phyllites	350
Geikie [0.77 km]	Rossbukta	sandstones	300
	Gordon	limestone	470
Ferrier [0.73 km]	Neukpiggen	flyschoid mixtites	300
	Peterbukta	greywackes	160
	Hardiefjellet	flyschoid mixtites	120
	Isachsen	mixtites, schists & volcanics	150+
[0.2 km]	Pinkie	metavolcanites	200+

banded turbidite slates occur throughout the Grampian Group and characterise both the Geddesflya and Barents Formations. Hard black mudstones with white veins similarly occur throughout the Scotia Group and form an important part of both the Roysha and Baklia Formations. Green, grey and purple slates occur within both the Kaggen and the Knivodden Formations.

The Forland Complex shows a complex post-depositional history. The rocks have been subjected to diagenetic changes, followed by polyphase deformation primarily of Palaeozoic age but reactivated during Cenozoic time. Early folding was accompanied by metamorphism up to biotite grade; tightening of folds produced a slaty cleavage which was followed by thrusting and nappe emplacement in a south-westerly direction and later by more open refolding. The extent of deformation attributable to the Cenozoic West Spitsbergen Orogeny is difficult to disentangle, as trends are at a low angle to those of mid-Palaeozoic age (roughly NNW-SSE). However, later graben deformation can more easily be identified with some E-W fold axes.

1. GRAMPIAN GROUP

Rocks assigned to this group are typically flyschoid in character. This obvious stratigraphic grouping was first suggested by TYRRELL (1924, MAJOR et al. 1956) for his Northern Grampian Series. These rocks extend further north

than the Grampianfjella. The name Grampian was not used stratigraphically by ATKINSON (1956, 1960) and he described corresponding rocks as the Fugelhuk, Barents, Conqueror, Pinkie and Larsen Groups. We now define the Grampian Group as comprising the Geddesflya, Fugelhuk, Barents, Conqueror and Utnes Formations.

A. *Geddesflya Formation*

This unit comprises about 1800 m of dominantly quartzite lithology. It is the uppermost unit of the Grampian Group. It is found exposed in the type area of Geddesflya and from the tops of Tvihyrningen south to Kaldneset. It also occurs in the Grampianfjella northwards from Margaretfjellet. Similar facies have been noted west of Richardlaguna and to the north. In the type area the highest members seen consist of an alternation of quartzites with dolomite-banded siltstones, breccias and thin siltstones and slates. Lower down thinly bedded quartzites occur which give way to slate pebble breccias interbedded with banded siltstones. Beneath these are sequences dominated either by thinly bedded quartzites or by banded siltstones.

B. *Fugelhuk Formation*

The Fugelhuk Formation here refers to a sequence of quartzites, somewhat more restricted than those defined by ATKINSON (1956, 1960) who used this name for one of his groups comprising a thick sequence of massive bedded quartzites in Fugelhukfjellet at the northern end of the island. In central Prins Karls Forland they occur below the Geddesflya Formation, exposed in a series of folds from Tvihyrningen to Skarvnes. They also outcrop in the vicinity of Margaretfjellet, along the eastern slopes of the Grampianfjella to Djevletumen-Klöne-Neglene. This formation as defined here is most readily observed at Skarvnes and around the southern end of Hyrneknatten. It is dominated by thickly bedded quartzites often in units of more than one metre, interbedded with banded siltstones. The facies suggest a mobile environment. In the lower slopes of Tvihyrningen and north across Geddesflya they pass down into black – dark grey siliceous slates of the upper Barents Formation. Around Tvihyrningen this formation is approximately 400 m thick. The strata thicken northwards and reach a thickness of approximately 1000 m in Fugelhukfjellet (ATKINSON 1956, 1960).

C. *Barents Formation*

The Barents Formation, here as with ATKINSON's (1956) Barents Group, refers to a sequence dominated by siltstones which lies between the Fugelhuk and Conqueror Formations. In the Grampianfjella the rocks are uniform. However, to the west of Scotiafjellet the constituent facies are different. The Barents Formation lies beneath Fugelhuk strata and is exposed in the centre of the island, west of Scotiafjellet, on the lower slopes of Tvihyrningen north across Geddesflya to Normandalen, in a narrow band toward Ossianbekken, and widening out in an arc to the west coast. In this area the top of the formation is

marked by the appearance of black–dark grey, often siliceous, slates which lower down include more silty bands. These pass down into sandstones with frequent slate pebbly bands. Below are thin quartzites and black slate laminae. These overlie a characteristic grey metallic slate with 2–3 mm pyrite “blebs”. Below this is a sequence of banded siltstones which exhibits many minor folds and whose true thickness it is not easy to estimate, though it does not appear to exceed 100 m. The lowest part of this formation is marked by flaggy fine-grained sandstones which are frequently calcareous. The transition to the underlying Conqueror Formation is marked by a sequence of green pelitic quartzites containing a black limestone and impersistent pebbly quartzites. West of Scotiafjellet this sequence is poorly developed and does not contain the limestone. North of Conquerorfjellet it is much thicker and appears to be at its maximum development in the area west of Richardlaguna. Thickness is estimated at 500 m. Part of ATKINSON’s Larsen Group is a Klippe of Barents slates on Alfred Larsentoppen.

Sutorfjella Conglomerate Member

HOEL (1914) mentioned these conglomerates and commented on their similarity to the Red Bay Devonian conglomerate of north Spitsbergen. CRAIG (1916) similarly likened them to the lower Old Red Sandstone of Scotland. TYRRELL (1924), reviewing these observations, compared them rather with the Tertiary conglomerates of Thomsonfjella. ATKINSON (1956, 1960) concluded that these rocks are interbedded with strata of the Barents Formation and so belong to it. KRASIL’SCHNIKOV (personal communication and 1973) expressed uncertainty about their age but favoured TYRRELL’s view.

This conglomerate dips 60°SW and can be traced across the two Sutors. It strikes parallel to the shore at 165° for just over a kilometre northwards before disappearing out to sea. It is underlain by green pelitic bands interbedded with dolomitic quartzites very similar to those at the top of the Conqueror Formation.

The conglomerate itself contains clasts of a number of underlying formations, the dominant rock type being a brown-weathering pale grey quartzite, frequently cut by quartz veins which are not related to the jointing of the conglomerate itself. In a number of horizons clasts are of a green cleaved siltstone identical to that of the matrix. Subordinate clasts include quartzites not unlike those seen in the lower Barents Formation itself and the Grampian Group as a whole. A less common clast is a black mudstone with quartz dolomite veining of the type characteristic of two of the Scotia Formations, and this gives evidence for the order of events in the Forland Complex. The conglomerate has a slaty cleavage parallel to that of the adjacent slate belonging to the Grampian Group. The clasts themselves are also cleaved, also parallel to that of the rock as a whole. It is inconceivable, therefore, that this Member could be of any other than Barents age.

The boulders, especially the quartzites, are often well-rounded and have red oxidised skins suggesting subaerial erosion and transport. Some have a double skin in which the oxidised layer is superimposed by a green reduced layer and

this, together with the generally green matrix, confirms that the deposition environment was aqueous and reducing. The boulders show that the typical white quartz veins were formed before erosion so that some stresses possibly associated with uplift had already operated. This is consistent with a major tectonic fault scarp situation.

D. *Conqueror Formation*

This formation unites a distinctive sequence of quartzites and slates. The name is adopted from ATKINSON (1956, 1960). Although well exposed in the north-facing slopes of Conquerorfjellet, it is more fully developed further north along the strike. The passage from Barents to Conqueror Formations is taken as coinciding with the green pelite and quartzite sequence, which is well developed in Petuniadalen up onto the Ytteryggen system and can be traced as far north as the Sutorfjella. Below this quartzite follows a sequence of slates, dark grey when fresh, alternating with dark–light grey weathering quartzite. Below these the quartzites become more brown in colour when weathered and are lighter grey when fresh. A number of pebbly calcareous bands occur and in particular a 2 m conglomerate band. The latter is traceable for a distance from Normandalen, where it is thrust out north through Conquerorfjellet–Ytteryggen–Margaretfjellet and further north. Below this is a thick slate with thin quartzitic bands, towards the base marking the top of the Utnes Formation. The Conqueror Formation in the vicinity of the type area is about 500 m thick. Further from Ytteryggen, towards the northern end of Dyerlaguna, it thickens to about 850 m.

E. *Utnes Formation*

This formation represents the transition from the Roysha Formation of the Scotia Group to the Conqueror Formation of the Grampian Group. This is isolated here for the first time and although exposed on the western slopes of Conquerorfjellet is seen at Utnes. The passage down from the Conqueror Formation is marked by a thick grey slate with thin quartzite bands. Lower down the quartzites become more pyritic and the slate blacker with occasional calcareous bands. The passage downwards into the Roysha Formation is marked by the appearance of a soft black carbonaceous slate. This formation is no more than 80 m thick.

2. SCOTIA GROUP

As the name suggests this group occupies the region of Scotiafjellet–Scotiadalen. The rocks also outcrop in the strip north of the Thomsonfjella, on the eastern slopes of the Grampianfjella toward the region of Richardlaguna where they widen out considerably. TYRRELL (1924 and MAJOR et al. 1956) first isolated this group and referred to it as the Mt Scotia Series. It occurs both around Scotiafjellet and in Scotiadalen and we use ATKINSON's (1956, 1960) name Scotia Group. Previous workers have not applied any systematic scheme to divide it. It is redefined here as constituted by the Roysha, Kaggen and Baklia Formations.

A. *Roysha Formation*

The choice of this name for the upper part of the Scotia Group is arbitrary and this formation is best developed slightly to the north of the western slopes of Conquerorfjellet. It lies conformably below the Utnes Formation of the Grampian Group, the uppermost unit being a very soft black carbonaceous slate. The major part of the formation consists of an alternation of sequences dominated either by grey dolomitic siltstones interbedded with thin black slates or by black carbonaceous slates with occasional grey siltstones. The true thickness of this sequence is difficult to determine as tight-folding thickens to an unknown degree. A conservative estimate would not exceed 400 m.

B. *Kaggen Formation*

This formation lies below the Roysha Formation and consists of tight isoclinally folded slate-phyllonites. They outcrop in a widening band south of Scotiafjellet, over much of the lower slopes of the east Thomsonfjella, Krokodillen and Buchananryggen. They continue northwards, widening out west of Richardlaguna. The extent of folding and sliding hinders the elucidation of the complete sequence. The stratigraphy of the formation has been compiled from observations at a number of localities, none of which exhibits the full complement of rocks. Correlation between the various localities is purely lithological, based on distinctive green and purple striped slates, grey slates with quartzites, and chloritoid bearing slates. The name Kaggen is chosen from the ridge where one of the thickest sequences is exposed. Briefly, the formation consists of dark bottle-green slates and various shades of green–grey slates, often with a striped effect. The upper part of the sequence is best exposed on the ridges between Alanfjella and Scotiafjellet, from Normandalen across to Omondryggen; southwards are exposed light and dark grey variegated slates with abundant chloritoid in the darker layers. Below these are grey slates followed by a light green slate containing many large chloritoid porphyroblasts easily seen in hand specimens. Below these are a series of green and purple slates which outcrop particularly well in the south-eastern foothills of the Thomsonfjella–Scotiadalen area. Below these are grey–green to dark grey slates with an increasing incidence of quartzite bands towards the base. This formation is estimated to be 300 m thick.

C. *Baklia Formation*

The lower part of the Scotia Group is exposed in the region of the lake just east of Scotiadalen, called Baklia. The passage from black slates with quartzite to a black carbonaceous slate sequence marks the transition from the Kaggen into the Baklia Formation. The upper part of this formation is dominated by black slates. However, within these slates is a very variable sequence of grey–orange dolomite limestones. These limestones are characterised by the presence of intraformational breccias which appear to show a decrease in clast size eastwards. The lower part of the formation exhibits an alternation of grey, frequently cherty, dolomitic siltstones with black slates. The bottom of the formation is taken as the level of appearance of a quartzite sequence which is

frequently conglomeratic with green and black slaty laminae. Below this lies a sequence of brown-weathering grey slates and then a black slate with dolomitic cherty limestones. Below this the Knivodden Formation proper begins (part of the Peachflya Group). The whole of this sequence is conformable and is probably not more than 300 m thick.

3. PEACHFLYA GROUP

This group lies directly below the Scotia Group and is a subdivision of what TYRRELL named the Ferrier Peak series, more nearly ATKINSON's Kerr Group. Kerr is not appropriate because rocks of the Scotia and Barents groups outcrop in the vicinity of Kerrlaguna, whereas most of Peachflya (the strandflat of the west coast) is occupied by the lithologies described below. Each of the four formations is distinct: Knivodden (incompetent chloritoid phyllites); Hornnes (siliceous-phyllite, sandstone-quartzite, limestone alternation); Alasdairhornet (volcanic suite); Fisherlaguna (blue phyllites).

A. *Knivodden Formation*

Phyllites constitute the bulk of the Knivodden Formation which forms a wide band in the west of Peachflya. They are generally incompetent and structureless, although occasionally some bedding lamination and five broad compositional divisions can be distinguished. The top of the formation is marked by a dark grey to black siliceous phyllite, often with pyrite growing along the cleavage planes. Above this the rocks become much less siliceous and consist almost entirely of altered clay minerals with chloritoid laths growing in the matrix. Towards the base thin arenaceous layers occur, and the lowest phyllite unit is generally coarser than those above it.

Initially, these rocks were all finely laminated argillites but metamorphism and deformation have destroyed most of the sedimentary characteristics, imprinting a slaty cleavage produced by pressure solution and new mineral growth. Clay mineral crystallinity measurements indicate temperatures of 300°C during metamorphism.

B. *Hornnes Formation*

Lying below these phyllites is a 350 m sequence of sandstone-quartzites, limestones and phyllites which constitute the Hornnes Formation, named from the locality of its best developed coastal section. Essentially, this formation is made up of dark siliceous phyllites with thin (10 cm) layers and lenses of sandstone, containing a number of more massive sandstone-quartzite bands up to 4 m thick. In total there are about eight of these bands. In addition, there are three limestone horizons. The uppermost is very dark, almost black, and crystalline with thin convolute layers of dolomitic limestone. This member varies in thickness from 10 m to 40 m. The dark colour of the limestone is produced by a high carbon content and this, together with the fine banded nature of the dolomite, suggests a possible algal origin for the rock, although any organic structures have been destroyed by deformation. Near the base is a

2 m limestone layer, laminated, crystalline and dolomitic. The base itself is marked by a fairly substantial limestone which is usually pale in colour and contains some silici-clastic material, but is mainly composed of crystalline calcite-dolomite laminae. A basic sill intrudes this rock wherever it is exposed. Again, the calcite-dolomite banded nature of parts of this member contain significant amounts of carbon as thin discontinuous partings which could be of organic origin.

C. *Alasdairhornet Formation*

At the top of the Alasdairhornet Formation (190 m) is a transition rock consisting of alternating layers of carbonate material and volcanogenic material. Most of this formation consists of banded and welded tuffs with some basic flows, while the base is a reworked sediment of volcanogenic and silici-clastic material. The name Alasdairhornet is taken from the ridge along which good exposures occur. Most of the banding appears to be primary, but the mineralogy has been greatly altered. Retrograde metamorphism (retrograde from primary igneous) has chloritised and septonised most of the rock leaving occasional relict pyroxenes. Relict amygdales filled with quartz, feldspar and calcite occur in a thin flow near the base of the formation.

D. *Fisherlaguna Formation*

The lowest formation (350 m) in the Peachflya Group is the Fisherlaguna Formation, consisting of incompetent phyllites which have a characteristic blue sheen on cleavage surfaces, and contain very thin sandstone lenses and occasional pressure solution quartz pods. The formation is generally poorly exposed because of its incompetent nature. It has been named after Fisherlaguna for convenience since it outcrops there, but nowhere in the area mapped is there a well exposed section.

4. GEIKIE GROUP

The Geikie Group consists of two formations: Rossbukta (quartzite and phyllite) and Gordon (limestone). The boundary between the two is transitional but distinguishable. The group outcrops with limited exposure to the east of the Peachflya Group. Thrusting between the overlying Peachflya, the Geikie, and the underlying Ferrier groups has greatly reduced the outcrop width of this group to the west of the southern Grampians. A more complete section is seen east of the mountains and on parts of the Forlandsletta, though the exposures on the Forland are poor and widely separated. The name Geikie is employed for this group since it includes the extensive limestone and sandstone formations (Geikie and Gordon Groups) that ATKINSON (1960) thought to be older than the Ferrier Group. In the present study it has been found that these rocks form a stratigraphic sequence both with each other and with the Peachflya Group, but not with the Ferrier Group.

A. *Rossbukta Formation*

The younger, Rossbukta, Formation consists of 300 m of dark mainly siliceous phyllites becoming increasingly calcareous towards the base. The top

is marked by a change from the overlying blue phyllites to more cohesive brown phyllites. Within these phyllites are a number of impure, coarse, crystalline sandstones. Sorting is generally poorer near the top of the formation where the phyllites contain numerous thin lenses of sandstone and sand-sized particles with a matrix of assorted clay minerals and clay-silt-sized particles. A fairly well exposed section of these rocks occurs along part of the shore of Rossbukta and in patches inland, so this name was chosen for the formation. Where these rocks lie close to the thrust which separates them from the Ferrier Group they are partly mylonitised and massively quartz veined.

B. *Gordon Formation*

This formation is dominantly limestone and dolomite comprising 470 m but the base is thrust.

The topmost member is a calcareous phyllite with schistose partings in places. This is almost continuous with the lowest phyllite of the Rossbukta Formation, the difference being marked by the amount of calcareous material present. A 3–4 m layer of massive dolomite lies within these phyllites and the base is masked by a thin, laminated, crystalline limestone with carbonaceous laminae.

The base of the Gordon Formation is made up of a limestone series including thin silty beds, but consisting mainly of dolomite-limestone laminated horizons, massive dolomite bands, intraformational breccias and carbon-rich beds. In common with most of the limestones in the area, the possibility of organic origin arises with these rocks. In this case the evidence is somewhat stronger since the presence of pisolite-like structures indicates algal mat conditions during formation. Deformation and recrystallisation have, however, obliterated most of the finer structures.

5. FERRIER GROUP

This name was used by TYRRELL (1924 and MAJOR et al. 1956) for the oldest of his three series, and again by ATKINSON (1960). We redefine this group as comprising four formations: Neukpiggen, Peterbukta, Hardiefjellet and Isachsen. They are typically schistose mixtites, of biotite grade, that we interpret as distal flyschoid marine tillites. There are at least two tillite horizons – which is characteristic of the Varangian ice age. The rare occurrence of stromatolites and oncolites in the stones is also a Varangian characteristic. This group is a tectonic unit which is stratigraphically discontinuous at top and bottom. Neither the top of the upper mixtite nor the base of the lower is seen because the rocks occur as a nappe.

A. *Neukpiggen Formation*

Calcareous mixtites make up the highest formation (300 m) whose stratigraphic top is not seen. Calcareous mixtite schists contain fragments of dolomite and granite varying in size from 10 mm to 0.3–0.4 m across and 50–100 mm fragments of limestone. Near the top of the formation 1–4 m thick dolomite pebble beds occur in fairly rapid succession and for about 50 m below this there are occasional pebble beds of similar thickness. Towards the base of this upper

mixtite formation the schist contains numerous 50–500 mm bands of crystalline limestone and there is one apparently discontinuous dolomite-marble bed up to 3 m thick. This formation is also thrust to the north of Scotiadalen and was there included in ATKINSON's informal Larsen Group.

B. *Peterbukta Formation*

This formation (160 m) is fairly well sorted and strongly schistose meta-greywacke, distinguished by lack of large clasts. We name it from the bay where it is best exposed.

C. *Hardiefjellet Formation*

The mainly mixtite formation (120 m) consists of siliceous schists with occasional calcareous partings and 50–500 mm sandstone bands. In addition to the mixtites there are some pebble beds about 1 m thick. In general this formation is very similar to the upper mixtites but here they are darker in colour, more siliceous and higher grade. In places the schistosity is very strong, with mica flakes up to 5 mm across and small biotite crystals growing throughout the matrix.

D. *Isachsen Formation*

This formation of dark green biotite schist, with brown interlayers and numerous pressure solution-quartz segregations, is the lowest of the group. Thin layers of mixtite material do occur, about 1 m thick, but most of the rocks are fine-grained, thinly laminated (10–20 mm at the most) and fairly well-sorted. ATKINSON informally used this name for a group with metavolcanics which we cannot match exactly because of different structural interpretation. Nevertheless beds about 1.2 m thick of volcanic, possibly tuffaceous strata are dispersed throughout the formation, of which 150 m are exposed and the base is not seen.

6. FORMATIONS NOT YET FITTED INTO SEQUENCE

A. *Pinkie Formation*

The Pinkie Formation, one of ATKINSON's informal groups, is confined to the area between Bouréefjellet and Monacofjellet. On the southeastern spur of Bouréefjellet this formation structurally overlies a sequence of Geddesflya type siltstones and thin quartzites. To the west the Pinkie Formation is in turn overthrust by a sequence of Conqueror quartzites. The formation is readily distinguished by metavolcanics and a high grade of metamorphism; it includes quartz-biotite schists, feldspathic-magnetite-biotite schists, felsites and a calcareous brecciated slate with much biotite. The latter, described by Atkinson, is not part of the Scotia Group, being dynamically or cataclastically deformed. This formation may represent a more easterly facies of the upper Grampian Group thrust westwards, or because of its high metamorphic grade it could represent rocks older than the Ferrier Group.

III. Oscar II Land

This is a much larger area and our work has been more extensive, with mapping on a scale of 1:50,000 except for the eastern part where maps and sections to a scale of 1:25,000 were done by A. CHALLINOR. Not only is there a greater variety of strata here than in Prins Karls Forland but also more structural complexity, so that we are less certain of the relationships between the sequences worked out in the different areas. Nevertheless, the succession has survived some testing and the tillite formations, first noted by us in 1958 and 1959 (HARLAND 1960), and fossiliferous formations discovered by us in 1968 and 1969, give some stratigraphic control. We define the formations in sequence from the top (see Table 2).

We use HOLTEDAHL's (1913) observations and attempt to name formations with his observations in mind. We follow ORVIN (1934) exactly for the sequence of older rocks in Brøggerhalvøya and this account adds little to his. We have related our units to the main stratigraphic groups noted by the Birmingham expeditions of 1951 and 1958 (BAKER, GARRETT and other private communications; WEISS 1953, 1958; and *Polar Record* 5 (37/38) 340, 6 (44) 527-28, 9 (62) 463); and by BARBAROUX 1966 and CHALLINOR 1967.

Table 2.
Sequence of older strata in Oscar II Land

Group	Formation	Dominant lithology	Thickness in metres
Bullbreen [0.7 km]	Holmesletfjella (Bulltinden Mbr)	flyschoid calc slates conglomerates & slumped beds	500
	Motalafjella	limestone	200
[0.5 km]	Sarsøyra	limestones and slates	
Comfortlessbreen [2-4 km]	Engelsbukta	mixtite	500
	Annabreen	quartzite	2-4000
	Haaken	mixtite	
St. Jonsfjorden [3.8 km]	Alkhorn	limestone	1000
	Løvliebreen	quartzites & volcanics	1000
	Moefjellet	dolomites	800
	Trondheimfjella	various	1000
Kongsvegen [3.1 km]	Nielsenfjellet (& Müllerneset)	schistose psammities	1500-2000
	Steenfjellet	marbles	100
	Bogegga	schistose pelites	1500
Vestgötabreen Fm.			250

Our main debt, however, is to C. B. WILSON who in one season in 1958 covered a large part of the area and set up the first possible sequence for Oscar II Land. The intended joint visit in 1959 by CBW and WBH was prevented by his untimely death. His manuscript and reconnaissance map were used subsequently by our expeditions and we believe also by the Norsk Polarinstitut to whom the work was sent.

1. BULLBREEN GROUP

This unit, outcropping at the western end of St. Jonsfjorden, contains fossils of Palaeozoic age. Although strongly folded and cleaved, it shows only slight metamorphism, so that the fossils and sedimentary structures are well preserved. The group is a combination of the Holmesletfjella and the Motalafjella Formations. The whole group appears as overthrust sheets in a small Klippe at Ankerfjella (north of St. Johnsfjorden) and more extensively around Bullbreen to the south.

A. Holmesletfjella Formation

The uppermost unit is of calcareous siltstones, argillites and polymict conglomerates. It was so named by HARLAND (1960), but misplaced as older than the Comfortlessbreen Group. The siltstones weather to a pale buff colour, often with grey bands, and contain various sedimentary structures including graded bedding, ripple marks and trace fossils. Massive conglomerates occur but are very variable in thickness and composition. Two separate conglomerate beds occur on Ankerfjella but on Motalafjella there is only a single unit, several hundred metres in thickness with thin limestone intercalations. The conglomerate is designated the Bulltinden Conglomerate Member for convenience in reference because of its fauna. However, within the Holmesletfjella Formation, to the SE of that mountain, is a thin tilloid associated with conglomerates. This was first thought (WBH) to be evidence of Varangian age and when a Palaeozoic age was established an Ordovician tillite was suspected, but on a recent visit (WBH and APM) this interpretation could not be established.

There is apparent stratigraphic conformity between the Holmesletfjella and Motalafjella Formations. In the latter are poorly preserved fossils, but better specimens have been recovered from limestone clasts in the Bulltinden conglomerates.

Bulltinden Conglomerate Member

Thick conglomerates form a substantial but variable part of the Holmesletfjella Formation. This conglomerate member has long been known. HOLTEDAHL referred to it (1913) and his description clearly related it to Bulltinden, so we chose this name. He found oolitic boulders of the Alkhorn Formation on it. WILSON described the conglomerate and tentatively correlated it with the Haaken schists but with misgivings, noting that the clast content cannot be matched (WILSON and HARLAND 1964). WINSNES (1965) took a similar view, as did FLOOD, NAGY and WINSNES (1971) in mapping the whole group as Eo-Cambrian.

However, in 1968 WTH and WBH suspected fossils in the limestone clasts; in 1969 WGH collected fossils from the conglomerates in Motalafjella chiefly from a locality on northern Motalafjella. The assemblage was tentatively identified by C. L. FORBES of the Sedgwick Museum as late Ordovician or early Silurian. In 1971 WBH revisited the locality and made a further collection from a slightly different facies of penecontemporaneously slumped limestone. This later collection contains a similar but less diversified fauna which may not be much older than the member. These two collections have since been examined by C. T. SCRUTTON who considered that the fossils from the conglomerates are of Silurian aspect and probably of either Wenlock or Ludlow age. The collection from the slumped limestones lacks the more diagnostic forms found in the conglomerate but contains at least one coral in common and nothing that conflicts with a Silurian age for that horizon. Further details of these faunas will be published elsewhere (SCRUTTON, HORSFIELD and HARLAND 1976).

Other clasts in the Bulltinden Member conglomerates are of marble lithologies similar to the Alhorn Formation varieties, including some which were folded before inclusion in the conglomerate. Dolomite clasts are less common than in the Comfortlessbreen tillites and contain no stromatolites. There are no granite clasts but one or two pebbles have been found of retrogressed schists similar to those in the Vestgötabreen Formation. This absence of extra-basinal lithologies and the closely-packed and well-sorted distribution of clasts offer no support to interpretation as a tillite. It is considered that the conglomerates formed in a shallow marine environment during a phase of Silurian faulting, possibly with metamorphism and folding.

B. *Motalafjella* Formation

This consists of pale grey, massive limestones (200 m), well exposed in an overturned nappe structure on Motalafjella. Here it structurally overlies the Vestgötabreen Formation of coarse-grained glaucophane schists and metamorphosed basic intrusives (HORSFIELD 1972). An unfaulted lower boundary has not been found and mapping suggests that this horizon has acted as a zone of décollement, with the limestones becoming more crystalline and dolomitic towards this zone.

2. SARSOYRA FORMATION

The marbles and argillites of this formation were first described by HOLTEDAHL (1913). They were named by WILSON (Ms) the Sarsøyra Beds from the coastal plains on which they occur. It was later suggested that they were Cambro-Ordovician in contrast to all the other rocks (W9 of HARLAND 1960). The formation borders the Forlandsundet Fault Complex and its upper and lower relationships are probably tectonic; indeed, it is not clear whether or not the constituent units are an unbroken sequence. For the present one formation is assumed. It is at least 500 m thick and four members are recognized.

A. HOLTEDAHL's "Heller, massiger Kalk" (1913, p. 59) is strongly sheared and grades into dark grey calcareous shales. No dolomitization has been observed. Because of this we do not follow HOLTEDAHL nor WILSON in correlat-

ing this formation with the dolomites north of Engelsbukta (the Moefjellet Formation).

B. Black, purple and green cleaved argillites form another variable unit distinguished by HOLTEDAHL. They are not so metamorphosed as the pelites of the Comfortlessbreen Group not far to the east. HOLTEDAHL (1913, p. 58) described them as “. . . meistens grün bis fast schwarz gefärbte, gerade spaltbare Schiefer vor, denen einige sehr bitumenreich sind und einen fast schwarzen Strich haben”. Some have a soapy feel suggesting chlorite or talc.

C. Calcareous conglomerates and breccias form the next member. The clasts are small and show considerable flattening. Quartzites, dolomites and limestone clasts are found in a calcareous matrix. In two specimens collected from limestone clasts in the stream section north of Aavatsmarkbreen small fragments of fossils were found (by WTH) in 1969. One was part of a simple coral. This could be Ordovician or Silurian.

D. Semi-pelites and greenstones are also associated in the lowermost member.

The westernmost outcrops of this formation appear to be scattered within the plain of Sarsøyra. WILSON suggested that they belonged to a horst. WTH noted slumped blocks in the Tertiary sequence and considered they were all allochthonous. WBH accommodates both viewpoints by postulating a submerged system of Tertiary fault scarps with some blocks sliding a little way from their now submerged horst-like structure. This is typical of a step-faulted graben margin.

3. COMFORTLESSBREEN GROUP

This group is characterized by tillite formations. WILSON in 1958 described the Haaken schists which we thought might be tillites, and this view was confirmed in 1959 when the Comfortlessbreen Formation was named (W7 of HARLAND 1960). Subsequent correlation shows the group to be equivalent to parts of BAKER's Variable Group (private communication). WILSON also named the Annabreen Quartzites. We subsequently had difficulty in determining the sequence with a threefold division of two tillites and a quartzite formation. Where least tectonically disturbed the quartzite appeared in the middle. It is common in tillites of Varangian age for there to be two glacial episodes and this is not quite clear in Prins Karls Forland. We also distinguish here the units already recognised as follows and so define the group:

- A. Engelsbukta (tillite) Formation
- B. Annabreen (quartzite) Formation
- C. Haaken (tillite) Formation

The whole sequence appears to be 3 to 4 km thick but is in places thickened by folding and elsewhere reduced by faulting. Some of the characteristics of the upper and lower tillite are the same, namely: they are thick, calcareous, clastic possibly flyschoid turbidites with dispersed stones of varied composition and up to a metre or more in diameter but more commonly only a few centimetres across. They are usually of dolomite, limestone, quartzite or granite, in that order of decreasing frequency. The carbonate clasts are occasionally stroma-

tolitic and oncolitic. The clast composition is similar to that of Varangian tillites elsewhere in Spitsbergen and the Arctic. Varangian tillites generally follow a sequence of stromatolitic and pisolitic dolomites and limestones. The matrix is altogether different from that in Ny Friesland and Nordaustlandet and the sequence is thicker and suggests a mobile environment. Sedimentary structures, however, cannot easily be observed because of intense flattening and elongation with schistosity and subsequent chevron folding.

In these respects the tillite formations parallel those of Prins Karls Forland and Nordenskiöld Land rather than those further to the east. However, higher grades of metamorphism are evident as, for example, at Carlsfjella, Svartfjella and Eidembukta where biotite and garnet in a somewhat gneissose texture are found. The more usual grades are chlorite-sericite assemblages.

A. *Engelskbukta Formation*

This name is proposed here to distinguish this upper unit from the lower unit already named (Haaken). The formation is seen in situ very conveniently on the south cliff on Engelskbukta. Bands, up to a metre thick, of unsorted orthoconglomerate are restricted to this unit.

B. *Annabreen Formation*

So named by WILSON (W8 of HARLAND 1960), this unit was renamed by CHALLINOR (we think mistakenly) and classed as early Carboniferous (CUTBILL and CHALLINOR 1965). It consists of massive pink, brown and white weathering quartzites with no clasts. The granular texture typifies the kind of metamorphism associated with presumed Precambrian rocks. Quartzites and semi-pelites may show a fine compositional banding.

C. *Haaken Formation*

Dolomitic boulders, including stromatolitic and oncolitic structures, are common in this formation. Larger clasts, including the quartzites, are also more evident than in the Engelskbukta Formation.

4. ST. JONSFJORDEN GROUP

Two pairs of formations constitute this group. The upper pair (Alhorn and Løvliebreen) directly underlie the Comfortlessbreen Group with sedimentary contact. The lower pair (Moefjellet and Trondheimfjella) are isolated by faults but are inferred in the position suggested with breaks of unknown magnitude above and below. The group is unified because the two pairs have been confused, while the tillites above and the much more highly metamorphosed rocks of the Kongsvegen Group beneath are distinct.

A. *Alhorn (Limestone) Formation*

This is essentially a limestone formation that everywhere underlies the Haaken tillites of the Comfortlessbreen Group and overlies quartzites of the Løvliebreen Formation. The name (Alhornkalk) is taken from HOLTEDAHL

(1913 and MAJOR et al. 1956) who described the lower part of the sequence at Alkhornet (Type 3 a thick dolomitic limestone and Type 2 a phyllite and shaly marble, 80–100 m). From studies in the same region south of St. Jonsfjorden the Birmingham expedition surveyed this unit as thin “upper calcareous group, 765 m” (BAKER, private communication). C. B. WILSON independently described the Dahlbreen limestone from north of St. Jonsfjorden (W6 of HARLAND 1960) and we have referred to it by that name until now, when we acknowledge HOLTEDAHL’s prior contribution.

The formation contains a wide range of lithologies but very little dolomite (which distinguishes it clearly from the lower Moefjellet Formation). Banded pale and dark marbles of a distinctive variety are present and also fine grained tinted marbles, calc-argillites, grits, breccias, and conglomerates. No stromatolites have been observed but oolitic and pisolitic textures, which may be algal (oncolites), are common. These are often visible as dark spots where they have been replaced by chert (c.f. HORSFIELD 1973).

B. *Løvliebreen Formation (1000 m)*

The formation is named after the glacier on the southern side of St. Jonsfjorden. It corresponds to the dark quartzites of HOLTEDAHL (1913) at the bottom of his Alkhornet sequence and to the “massive quartzite bodies occurring in the eastern part of Holmesletfjella and in Gunnar Knudsenfjella” of WEISS (1953). In both localities there are interbedded pelites and volcanic rocks. The formation occurs in a recumbent syncline south of St. Jonsfjorden and the base is not seen, so that 1000 m may be an underestimate of its thickness. Two members are distinguished:

- (i) *Upper*: massive quartzites with intercalated pelites. The dark quartzites are cut by thin white quartz veins. In thin section they are equigranular and pure with rounded or sutured outlines. The pelites are dark, fine-grained, and fissile.
- (ii) *Lower*: the volcanic rocks (e.g. of Gunnar Knudsenfjella) weather dark brown, green and purple. They are fine-grained and contain amygdalae but have no glassy shards, crystallites or spherulites and no pillow structures. These metavolcanics are probably widespread in the inland areas of southern Oscar II Land, judging by their extensive occurrence in glacial moraines.

C. *Moefjellet Formation*

The formation was named by C. B. WILSON from one of the jagged dolomite peaks on Løvenskioldfonna and was the fifth in his system (W5 of HARLAND 1960). It is a massive dolomite, uniform, unfoliated and difficult to subdivide. In colour it is typically a cream-weathering grey rock, often a quartzite with gritty surface texture. Veining and internal small-scale brecciation are often seen on weathered surfaces. Small-scale banding may be due to alternating cream and grey dolomitic or cherty layers in dolomite. No organic structures have been seen.

The formation lies above the Trondheimfjella Formation at Trondheimfjella but its upper relationship is faulted. Its lower stratigraphic position is inferred because there is no break in the sequence Comfortlessbreen–Alkhorn–Løvliebreen. Interjacent strata could be missing. A minimum thickness of 800 m is suggested and the thrusting has produced an outcrop of variable width.

The formation outcrops north and south of Engelskbukta, but it is possible that marbles at Eidempynten and Daudmannsødden are correlative.

D. *Trondheimfjella Formation*

This is a mixed formation of interbedded marbles, quartzites, and pelites with calcareous conglomerates towards the base. It is well exposed along the northeast face of Trondheimfjella which runs east from Engelskbukta and was so named by WILSON (W4 of HARLAND 1960). Part of ORVIN’s (1934) “dolomites and limestones at Forlandsundet” is included.

The formation passes concordantly upwards into the Moefjellet Formation but its lower boundary is faulted. Its thickness is not less than 1000 m. Faulting brings it into close contact with rocks of Carboniferous age. CHALLINOR’s Bjørvigfjellet Formation on the northern side of Engelskbukta can therefore be correlated in some detail with this and the overlying Moefjellet Formation.

It is divided into three members:

- (i) Thin orange-weathering bands of calcareous conglomerates in a sequence of quartzites, psammities and massive dolomites c. 200 m
- (ii) Dark phyllitic semi-pelites and psammities with minor quartzites and calcareous beds c. 300 m
- (iii) Marble flags 500 m

Although it is suggested that this formation overlies the Kongsvegen Group the basal conglomerates do not contain typical Kongsvegen-type clasts.

5. KONGSVEGEN GROUP

This name (HARLAND, WALLIS and GAYER 1966) groups the formations that ORVIN (1934) first defined. He set up 11 units. 1–9 he named the “Quartzite and Mica Schist series”; 10 the “Steenfjell Dolomite”; and 11 the “Bogegg Mica Schist”, underlain by “Dolomites and limestones at Forlandsundet”. ORVIN considered the sequence inverted so that unit 1 would be the youngest. WILSON took the opposite point of view, (W1–3 of HARLAND 1960). HARLAND et al. (1966) and CHALLINOR (1967) confirmed ORVIN’s order because of the overturned Carboniferous unconformity, and introduced the name Nielsenfjellet (sic) Formation for units 1 to 9 and modified ORVIN’s two names to match the current place names. We accept ORVIN’s division of units 1–11 with the original names of ORVIN and CHALLINOR. But ORVIN’s fourth unit (CHALLINOR’s Bjørvigfjellet Formation) we here identify with the Trondheimfjella Formation described above.

We also include in this group the Müllerneset Formation from south of St.

Jonsfjorden which we correlate with the Nielsenfjellet Formation; it probably includes the outcrop at Kulmodden in western Brøggerhalvøya.

The group forms a distinct mountain range extending 30 km as the backbone of Brøggerhalvøya. It is overthrust from the south and a fundamental fault there could be concealed.

A. *Nielsenfjellet Formation*

This is ORVIN's (1934) "Quartz Mica Schist series" and was named by CHALLINOR (1967). Typically this is seen southeast of Ny-Ålesund in dark mountain ridges. It comprises monotonous dark phyllitic semi-pelites interspersed with paler, dolomitic quartzite bands. ORVIN suggested a thickness of 2500 m but that is probably taken from Brøggerhalvøya which is not a typical section and shows some repetition. The figure of 1500 m is derived from the more constant width to the southeast. On the other hand the Müllerneset Formation which we correlate with it is not less than 2000 m so we conclude 2 km to be a reasonable estimate.

B. *Steenfjellet Formation*

This formation (ORVIN 1934) consists of a prominent 100 m band of grey and creamy-orange marbles separating the Bogegg and Nielsenfjellet Formations. It is coarse-grained and slightly foliated, with a well-developed, internal, passive folding. The formation is locally cut by thrusting. It is the thickest and most prominent of the pale bands that occur in the overlying Nielsenfjellet Formation. It is cut out by thrusting at Veslebreen and Holtafjella. The foliation of the marble is accentuated by parallel muscovite and differential weathering of quartz-dolomite and calcite rich layers.

C. *Bogegg Formation*

CHALLINOR (1967) modified the spelling to Bogegga to conform to the official spelling of the place name but we use the name as first defined by ORVIN (1934 and MAJOR et al. 1956) – the "Bogegg mica-schist". It is a varied sequence of schistose and gneissose psammities, pelites, semi-pelites, pale dolomitic marbles, pink feldspathites and dark amphibolites. It is well exposed in the Edithbreen area and forms dark ridges with occasional thin paler bands. It is divisible into three members:

- (i) Pelites and semi-pelites with intercalated orange-weathering quartzites, marbles and psammities. The pelites and semi-pelites are biotite schists with quartzo-feldspathic bands and lenses, and occasional small garnets. The marbles are coarse-grained and gritty with slight muscovite schistosity and comprise half the bulk of the member 500 m
- (ii) Dark feldspathic and garnetiferous semi-pelite with quartzo-feldspathic bands. The unit is characterized by large quartzo-feldspathic segregations up to 2 m in diameter, with muscovite, and by pygmatic structures 400 m

- (iii) Gneissose porphyroblastic feldspathites and semi-pelites with schistose garnetiferous pelites, subordinate dolomites and impersistent concordant amphibolites. The feldspathites weather greenish-grey and typically develop an augen structure 600 m

The coarser schistosity and more prominent deformation of this formation relative to the structurally lower Nielsenfjellet Formation supports the view that this is the oldest of the three formations.

D. Müllerneset Formation

This formation consists of 2000 m of phyllitic and schistose pelites and semi-pelites with psammites and white quartzites. It is typically seen around Müllerneset, south of St. Jonsfjorden, and occupies the Svartfjellstranda coastal plain west of the thrust strip of Carboniferous and Permian rocks, and DINELEY (1958) suggested it was of this age. We correlate it, however, with the Nielsenfjellet Formation, and the tectonic setting is similar to that at Kulmodden west of Brøggerhalvøya.

6. VESTGÖTABREEN FORMATION

This is a suite of metamorphic rocks of blue schist facies, of which some lithologies at least are derived from basic igneous rocks. It occurs in a narrow zone about 10 km long by a few hundred metres wide which runs parallel to the regional NW–SE strike. Rocks of this suite were first discovered as loose blocks by C. B. WILSON in 1957. In 1962 D. G. GEE located a source on Motalfjella and in 1968/9 W. T. HORSFIELD located further outcrops on ridges to the north of Vestgötabreen (HORSFIELD 1972). Exposures are therefore situated only on steep craggy ridges about 10 km inland from Forlandsundet.

The zone along which these outcrops run is associated with considerable brecciation, iron and copper mineralisation, and runs parallel to several low angle faults which dip to the west. To the west lie sheared marbles and to the east massive metamorphosed greenstone. We cannot correlate this suite of rocks but they are probably older than the Bulltinden conglomerates, and probably suffered metamorphism at around 470 M.y.; however, since their occurrence is restricted to the area of Bullbreen Group rocks, they may well originally have been Palaeozoic rocks.

IV. Correlation and Conclusion

The main object of this paper is to outline the stratigraphic sequences as we understand them. There is no space to consider their palaeogeological implications. Our final task, to compare and correlate these sequences, is attempted in Table 3 and the notes below.

- (1) The fauna of the Bulltinden Conglomerate Member is of Wenlock or Ludlow age.

Table 3.

Correlation of older strata between Prins Karls Forland and Oscar II Land

The symbols (in brackets) used in this table serve as a key to the units on the geological map Fig. 1. Thickness in km of combined sequence = 16.5 to 18 km.

Prins Karls Forland		Oscar II Land	
Group	Formation	Formation	Group
Grampian 3.6	Geddesflya		
	(B) Fugelhuk		
	Barents	* Holmesletfjella (Bulltinden Mbr) Motalafjella	Bullbreen (B) 0.7
	(Sutorfjella Mbr)		
	Conqueror		
Utnes			
Scotia 1	Roysha C	M*	
	(S) Kaggen Vg	C Sarsøyra	0.5 (S)
	Baklia C	V	
Peachflya 1.3	Knivodden Vg		
	(P) Hornnes		
	Alasdairhornet V		
	Fisherlaguna		
Geikie 0.77	(G) Rossbukta		
	Gordon M		
Ferrier 0.73	Neukpiggen T	T Engelskbukta	} Comfortlessbreen 3 to 4 (T)
	(T) Peterbukta	Annabreen	
	Hardiefjellet T	T Haaken	
	Isachsen (Vg) T		
		M Alkhorn o	} St. Jonsfjorden 3.8 (J)
		V Løvliebreen	
		M Moefjellet Trondheimfjella	
		Müllerneset and Nielsenfjellet	} Kongsvegen 3.1 (K)
		M Steenfjellet	
		Bogegg	
	Pinkie V	V Vestgötåbreen	

* fossil horizon
T tillite horizon
— tectonic break

o oncolite horizon
C carbon or bitumen
M marble

V volcanic horizon
Vg volcanogenic

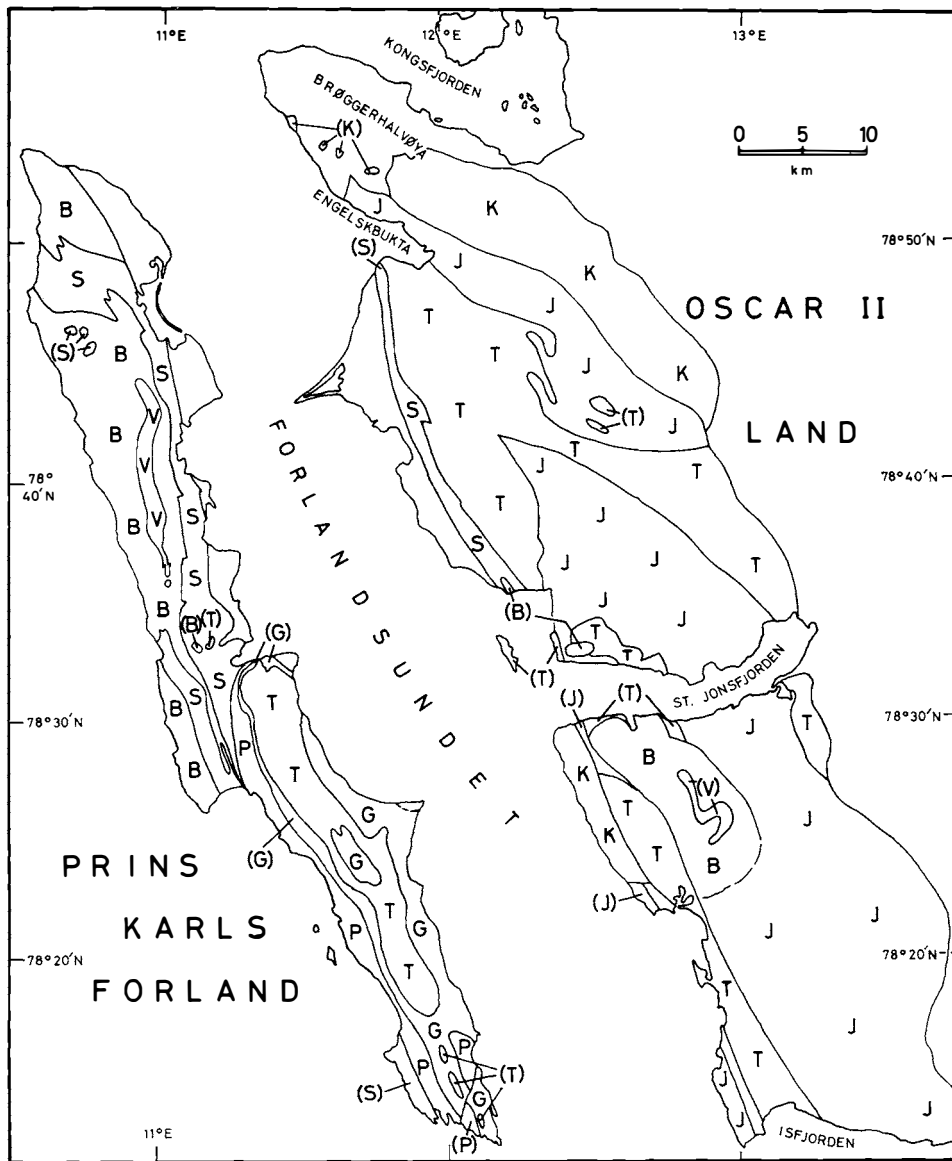


Fig. 1. Map of central western Spitsbergen to show the larger outcrop areas of the stratigraphic groups as described in this paper.

(2) The Bullbreen Group and the Grampian Group are analogous.

- (i) Both the Barents Formation and the Holmeslettfjella slates and siltstones are turbidites and their colouring is similar.
- (ii) Each has remarkably coarse and locally developed conglomerates, namely the Sutorfjella and the Bulltinden Conglomerate Members. Both are formed from the rapid erosion of uplifted areas, though the Sutorfjella conglomerate is limited in extent and thickness compared with the Bulltinden. The clast compositions are different also. The Sutorfjella clasts

all appear to derive from post-Varangian rocks, whereas the Bulltinden conglomerate derives from late Riphean and early Palaeozoic limestones up to at least Wenlock age. The latter would therefore appear to result from more widespread and deeper erosion. Nevertheless, both situations are highly mobile and could indeed be synorogenic.

(3) The varied strata of the Sarsøyra Formation, with limestones, multi-coloured slates and a rich carbon content, are reminiscent of the Scotia Group, i.e. the slates might compare with the Kaggen Formation and suggest a volcanic episode. The fossil reported suggests at earliest a mid-Ordovician age, and if the correlation suggested here is correct then these rocks would be older than the Bullbreen Group, i.e. pre-Devonian, probably pre-Ludlow.

(4) The tilloids of the Ferrier and Comfortlessbreen Groups correlate and probably represent the Varangian ice age, partly because each reveals two distinct tillite formations and each contains stones with stromatolites and oncolites.

(5) The underlying carbonate formation is not stromatolitic, but in both Oscar II Land and Nordenskiöld Land oolites and pisolites (oncolites) are found in limestones (HORSFIELD 1973) that, it has been suggested, correlate in time with the late Riphean Akademikerbreen Group. This is a reasonable time correlation, but the facies are very different.

(6) In the tectonically unstable area under consideration conglomerates are not infrequent but need not have been very extensive, so we do not value them for long distance time correlation. For example, we would not feel confident to match any part of our sequence with the Draken conglomerates of Ny Friesland.

(7) It may be significant that possibly the oldest rocks in each sequence, though tectonically isolated, are metavolcanics (Pinkie and Vestgötabreen).

V. Summary

We have established two sequences which are somewhat complementary. They are positively linked by the tillites and this enables us to add up the thickness for the whole. In doing so we have given throughout the most conservative thicknesses. Moreover, at several places the sequences are incomplete, so strata are probably tectonically removed. The sequence as a whole is no less than 18 km thick and probably more than 20 km. This ranges in age probably from Mid or Late Riphean through to Wenlock or Ludlow time. The facies indicates tectonic mobility almost throughout, with flyschoid and conglomeratic rocks and with several volcanic episodes. There is sufficient unity within this area and sufficient contrast from neighbouring sequences to justify naming a distinct geosyncline.

We have considered regional names, but they have been pre-empted (e.g. Forlandsundet, West Spitsbergen). Therefore, personal names have been considered; and in view of his outstanding contribution to the elucidation of the older rocks of western Spitsbergen we propose to commemorate OLAV HOLTEDAHL by referring to the Holtedahll Geosyncline.

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Hecla Hoek rocks of Oscar II Land and Prins Karls Forland, Svalbard

By AUDUN HJELLE, YOSHIHIDE OHTA and THORE S. WINSNES

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Abstract

The Hecla Hoek succession of the present area is strongly disturbed by the Tertiary diastrophism and most lithologic units are separated by faults. Five lithologic units have been distinguished in the Upper and Middle Hecla Hoek Supergroup, and three separated blocks of Lower Hecla Hoek Supergroup are also recognized. All correlations are based on lithologic characteristics and structural interpretation. The stratigraphic position of peculiar glaucophane schist-bearing formation is still a problem.

The metamorphic grade of the rocks generally increases with stratigraphic depth, from the green schist facies in the upper part to the upper amphibolite facies in the lower part, and the stilpnomelane-, chloritoid- and staurolite-bearing assemblages characterize the intermediate P/T type metamorphic facies series. This progressive metamorphism was achieved during the main Caledonian orogeny when the axial plane cleavages of isoclinal folds developed. A thermal matamorphism followed the regional metamorphism and a late deformation with

chevron folds superposed on the metamorphic rocks. The Tertiary movement with the stress from SW caused open gentle folds and thrust faults associating with local retrogressive metamorphism. The formation of the Forlandsundet graben is later than the thrust movement.

Introduction

Low grade Hecla Hoek meta-sedimentary rocks occur widely along the west coast of Oscar II Land and on Prins Karls Forland. These two areas are separated by the Forlandsundet Tertiary graben. The geologic structures of these rocks are very complicated as discussed by WEISS (1953) and ATKINSON (1960) and consequently many difficulties are encountered in attempts to restore the stratigraphy of these areas. This article presents a preliminary geological map, stratigraphic divisions and general structures, based on the field surveys carried out by many geologists of Norsk Polarinstitut since 1955.

Prins Karls Forland.

- T. GJELSVIK: 1968, 1970, 1971, 1972, 1973, 1974, mostly northern area, some in the middle part.
- A. HJELLE: 1972, middle part.
- Y. OHTA: 1972, northern part.
- T.S. WINSNES: 1972, southern part.

North of St. Jonsfjorden, Oscar II Land.

- E. TVETEN: 1968, from Uversbreen to Aavatsmarkbreen.
- T.S. WINSNES: 1967, west side of Kongsvegen to Engelskbukta.
1972, Haraldfjellet to Carlsfjella, and southern part of Prins Heinrichfjella.
1975, Løvenskioldfonna and Haaken Mathiesenfjella.
- A. HJELLE: 1972, southern part of Prins Heinrichfjella and Ankerfjella.
- Y. OHTA: 1972, from Konowfjellet to Gaffelbreen.

South of St. Jonsfjorden, Oscar II Land.

- H. MAJOR: 1955, Daudmannsøyra — Trygghamna — Venernbreen area.
- T. SIGGERUD: 1965, around Copper Camp, St. Jonsfjorden.
- Y. OHTA: 1973, from the west coast to Vegardfjella, and in the north of Eidembreen.
- T.S. WINSNES: 1975, Sparrefjellet.

Besides these data, the works of TYRRELL (1924), WEISS (1953), and ATKINSON (1956, 1960, and 1963) are referred to.

It is hoped that this article can open a discussion on this problematic Hecla Hoek area of Svalbard.

Stratigraphy

The area of Hecla Hoek rocks is divided into two by the Forlandsundet Tertiary graben: Prins Karls Forland in the west and Oscar II Land in the east (Fig. 1). Each of these areas is divided into three, and six lithostatigraphic

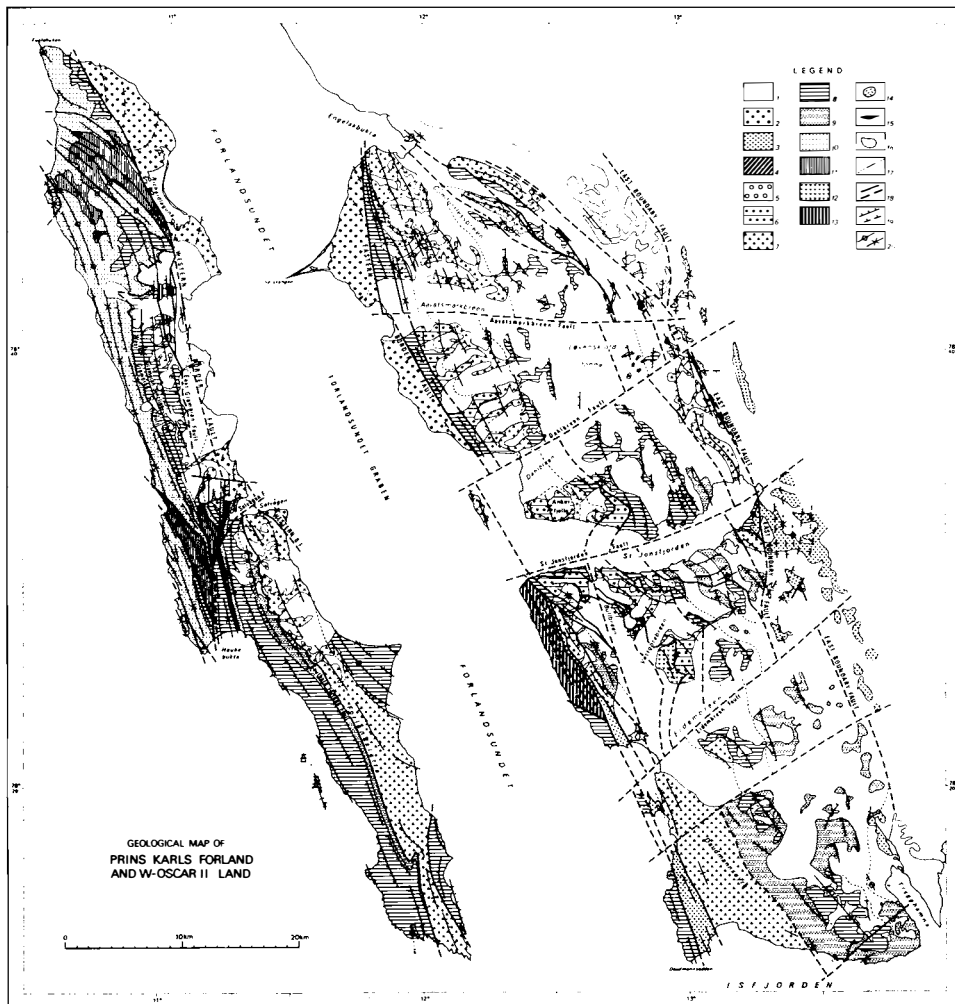


Fig. 1. Geological map of Prins Karls Forland and the western part of Oscar II Land.

Legend: 1: glacier and moraine, 2: Tertiary rocks, 3: Permo-Carboniferous rocks, 4: Devonian, 5: Sutor Conglomerate, 6: Bulltinden Formation, 7: Tillitic Conglomerate, 8: Calc-argillo-volcanic Formation, 9: Quartzite-shale Formation, 10: Quartzite-sandstone Formation, 11: Black shale Formation, 12: Vestgötabreen Formation, 13: Lower Hecla Hoek rocks, 14: basic intrusive rocks, 15: dolomite and serpentinite along faults, 16: exposure border, 17: lithologic border, 18: fault, 19: structural trend, 20: anticline and syncline.

sections have been established (Fig. 2). The stratigraphic columns from the northern part of Prins Karls Forland and the southern side of St. Jonsfjorden are summarized from many geologic profiles, while the rest were synthesized from a few sections.

The formation names already proposed by previous authors are followed with some modifications. No new names are proposed but the rock units are distinguished by their dominant lithology. Eight lithologic units, hereinafter termed formations, have been distinguished.

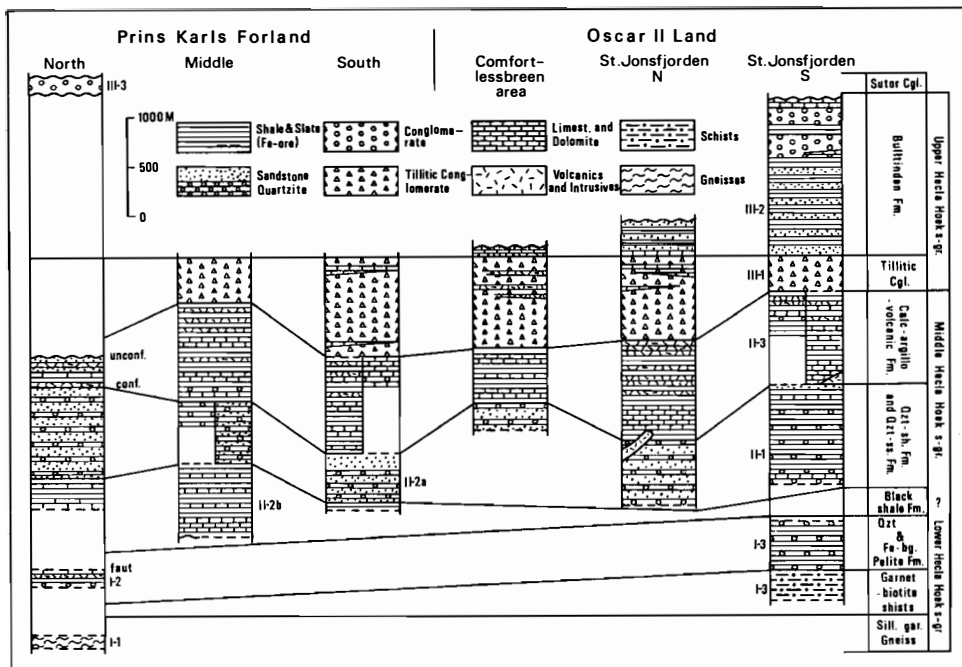


Fig. 2. Lithostratigraphic columns of the Hecla Hoek successions in Prins Karls Forland and Oscar II Land. The numbers: I-1, II-1a etc., refer to the description in the text.

In spite of the short distance across the Forlandsundet graben, the correlation of these formations between the west and east is not straightforward. The most reliable correlation is the Tillitic Conglomerate, which we consider to be the glacial deposit of Vendian age, widely occurring all over Oscar II Land and in the southern part of Prins Karls Forland. However, many other formations are separated by faults and their stratigraphic positions are mainly based on structural interpretations and on the comparisons of their lithology with those of other areas of western Spitsbergen.

The succession above the Tillitic Conglomerate of the Upper Hecla Hoek Supergroup was grouped into one formation. Three formations are considered to be of the Middle Hecla Hoek Supergroup and three isolated successions are assumed to be of the Lower Hecla Hoek Supergroup. The observed thicknesses are: Upper Hecla Hoek 2,500 m and Middle Hecla Hoek 3,500 m. The lower Hecla Hoek rocks, - 1,000 m, occur as small fault blocks.

LITHOSTRATIGRAPHIC DESCRIPTION

Lithological characteristics of each formation are described, based mainly on field observations, supported by some microscopic work (Figs. 1 and 2).

I. Lower Hecla Hoek rocks

These rocks are distinguished by their notably higher degree of metamorphism. The occurrences are always bounded by faults.

I-1 *Sillimanite-garnet-biotite gneisses at the Bouréefjellet, middle Prins Karls Forland*

These rocks occur along the eastern foot of Bouréefjellet with a thickness of about 150 m. The rocks are dark, dense and schistose, having alternations of calcareous and pelitic layers of cm to dm thicknesses. Mylonitic texture is evident in the matrix. Sillimanite occurs as relics in small prismatic grains enclosed in the bundles of sericite. Garnet grains are idiomorphic and often helicitic with numerous inclusions of plagioclase, potash feldspar, biotite and quartz. Pale brown large flakes of biotite occur on the mylonitic matrix of quartz and feldspars. Epidote, clinozoicite, sphene, tourmaline and calcite are common accessories. These rocks form a wedge-like block, bounded by a steep fault on both the east and the west side, extending parallel to the Western Border Fault of the Forlandsundet Tertiary graben (Fig. 1).

I-2. *Iron-ore-bearing pelites*

On the southern faces of Bouréefjellet, a slate rich, 150 m thick, succession with very gentle structure occurs. Many quartzite layers are interbedded and a brecciated limestone of green and white tints occurs in the upper part. Amphibolite lenses are intercalated in the middle part. Between the limestone and amphibolites, an iron-ore bed of 3 m thickness occurs, being composed of dense layered magnetite and hematite.

The slate-rich succession is cut by steep faults both in the west and the east, and extends to the south at least for 2–3 km as a narrow zone of about 1 km width. This succession can be correlated to the iron-ore-bearing rocks east of Recherchebreen, south of Bellsund.

I-3. *Garnet-biotite schists and quartzite-shale alternation*

These rocks occur along Svartfjellstranda and are separated from the main Hecla Hoek region in the east by a narrow fault zone of Late Paleozoic rocks. The rocks are divided into two members; the Garnet-biotite schist of 300 m thickness in the east, and the Quartzite-shale alternation with a thickness of 500 m in the west along the shore. The schistosity and bedding of both members are nearly vertical, and the border of these two shows rapid but gradational change. Due to the high grade metamorphic assemblages, the Garnet-biotite schist Member is considered to be in the lower position. The quartzite of the upper member is pure and banded with pink, grey, and white beds. This colour distinguishes it from that of the Quartzite-shale Formation (see II-1). These rocks form a NNW-SSE trending zone parallel to the eastern border of the Forlandsundet Tertiary graben.

I-4. *Vestgötabreen Formation: the glaucophane schist and associated rocks*

This rock unit was named the Vestgötabreen suite by HORSFIELD (1972), and occurs from Bulltinden to Motalafjella for 10 km as a narrow thrust schuppen a few hundred meter thick. The original definition of this suite by HORSFIELD was coarse-grained glaucophane bearing garnet-muscovite-schists, but this is modified here to include all rocks closely associating in the thrust zone, such as: epidote-actinolite greenstone, serpentinite, glaucophane-bearing

schists, eclogite, muscovite-quartz-schist, calcareous schist and dolomite, and the name is changed into the Vestgötabreen Formation.

This formation is sandwiched between rocks of the Bulltinden Formation (see III-2). The rock association of the Vestgötabreen Formation: frequent alternation of quartz schist and impure limestone with a large amount of meta-basites, represent typical eugeosynclinal shallow sea deposits. The stratigraphic position of this formation is problematic, correlative successions are the Calc-argillo-volcanic Formation (II-3) of the present area in the Middle Hecla Hoek succession and the Eimfjella Formation of Hornsund in the Lower Hecla Hoek succession.

II. *Middle Hecla Hoek rocks*

The stratigraphic positions of two formations presented below are unknown by direct evidences, but are supposed to be of Middle Hecla Hoek Supergroup from lithologic comparisons.

II-1. *The Quartzite-shale Formation*

This formation occurs in two areas: 1) from Wollertoppen, north of St. Jonsfjorden to the south and west of Lexfjellet, north of Isfjorden, and 2) along the ridges of the southern part of Prins Karls Forland.

The lower border of this formation is probably a fault contact with the Tillitic Conglomerate at the southeastern edge of Gunnar-Knudsentjella and at the northwestern edge of Vegardfjella, and at the western tip of Carlsfjella, south and north of St. Jonsfjorden, respectively.

This formation is composed of an alternation of quartzite and shaly sandstone. The quartzite is pure and dominant in the lower part of the succession, with beds of 0.5 to 10 m in thickness, occupying about 1/3 of the succession. Some quartzite layers on the ridge west of Charlesbreen show pink and grey coloured bands similar to the Quartzite-shale alternation Member of the I-3. Quartzite-rich successions also occur further to the south, in Stortrollet and the foothills of the mountains from Kinnefjellet to Protektorfjellet in the northern side of Isfjorden.

The shaly rocks of the Quartzite-shale Formation consist of individual beds thicker than the quartzite, up to 30 m thick. The dark shale increases in the upper part. Here three or four layers of black oolitic limestone; less than 5 m in thickness, occur on the ridge between Anna-Sofiebreen and Gunnarbreen.

In the southern Prins Karls Forland, the upper part of the formation is exposed, being composed of homogeneous flaggy siliceous shale and some quartzite beds.

A minor gabbro stock cuts this formation around Donpynten on the northern shore of Isfjorden.

II-2. *The Quartzite-sandstone Formation and the Black shale Formation in Prins Karls Forland*

Two formations, the Quartzite-sandstone apparently underlying the Black shale Formation, occur above the Calc-argillo-volcanic Formation west and

north of Prins Karls Forland. The conformable relation of the Quartzite–sandstone Formation to the underlying one was observed at three localities and this formation can apparently be considered younger than the Calc–argillo–volcanic Formation and older than the Tillitic Conglomerate.

However, the rocks conformably underlying the Tillitic Conglomerate are phyllites and limestones similar to the rocks of the Calc–argillo–volcanic Formation. From the structural interpretation of the Grampianfjella, the Calc–argillo–volcanic Formation occurring in the northern Prins Karls Forland is the upper limb of an overturned syncline. Thus, there is a possibility that these two formations may be inverted and older than the Calc–argillo–volcanic Formation. This problem is still an open question and may be solved by a detail study of the sedimentary structures of these two formations.

II–2a. The Quartzite–sandstone Formation

This is a more than 900 m thick alternation of quartzite and sandstone and occurs along the west coast from Kaldneset to Utnes, and over the whole northern part of Prins Karls Forland. In the Grampian area, the lower boundary with the Calc–argillo–volcanic Formation is cut by the steep west dipping West Grampian Fault (Fig. 1). The quartzite is less than a half of the whole succession and always sandy and impure. Dark alternating layers of fine– to medium–grained sandstone, occasionally conglomeratic, and silty stones often show turbidite structures. Graded beddings are occasionally observed, but the up–down relations observed are not sufficient to determine whether this formation is overturned or not. Some thin layers of conglomeratic quartzite with angular limestone pebbles are characteristic in the lower part of this formation.

II–2b. The Black shale Formation

In the northern part of Prins Karls Forland, a distinct black formation apparently overlies conformably the Quartzite–sandstone Formation. The rocks comprise calcareous black shale, partly slate, and impure limestone. Characteristic siliceous concretions of irregular rod shape, 1–5 cm across, often occur in the shale. Some of them show weak concentric pattern around the margins, and may be relics of organic structures.

This formation extends to the south to Haukedalen, central part of Prins Karls Forland, together with the Quartzite–sandstone Formation. The thickness is 550 m in the north and 750 m in the central area.

This is apparently the uppermost formation of any regional significance in Prins Karls Forland. However, if the structural assumption of an overturned syncline in Grampianfjella is accepted, this becomes the lowest succession below the Quartzite–sandstone Formation which may be correlated to the Quartzite–shale Formation (II–1) in Oscar II Land.

II–3. *The Calc–argillo–volcanic Formation*

This formation is commonly composed of grey banded limestone in the lower part, black shale or slate in the middle and green phyllite with meta–volcanics

and –intrusives in the upper part in general. The successions observed in five separated localities show large variations as outlined below:

1) The largest outcrop of this formation is a NNW–SSE zone from Engelsbukta to Trygghamna, north of Isfjorden, 5–9 km wide and 80 km long. In the northern part of this zone, Trondheimsfjella, several limestone layers alternate with slate in the lower part, the middle part is shale, and a few relatively thick limestones (up to 100 m) and shale occur in the upper part of the succession. Observed thickness is 550 m.

Along the northern side of St. Jonsfjorden, this formation was observed with a slate of 80 m thickness and a limestone–rich succession of more than 100 m thickness in the lower part, about 100 m thick black slate in the middle, and grey–green phyllites of about 300 m thickness in the upper part. The lower limestone extends to the north around Løvenskioldfonna. Two small basic intrusive bodies, hornblende gabbro and porphyrite, occur in the lower slate and they may be feeder of the eruptive rocks which are now represented by the green phyllites in the upper part of this formation.

To the south of St. Jonsfjorden, the limestone–rich character of this formation continues to Isfjorden. The limestone is prominent in the lower and upper parts and shale dominates in the middle part of the succession. A thick light coloured, banded dolomitic limestone on the ridges of Kinnefjellet and Lexfjellet to Protektorfjellet, conformably overlying the Quartzite–shale Formation (II–1), is the lower part of this formation.

A remarkable volcanic complex occurs on the top of Gunnar–Knudsenfjella. It is composed of basalt lava, varicoloured viscular agglomerates and green phyllites, and the total thickness is up to 150 m. The thickness decreases rapidly north– and southwards and the complex grades laterally into thin layers of green phyllites. The green phyllite horizons below the limestone of the mountains west of Trygghamna and the hornblende gabbro stock around Donpynten are the southern extremities of this basic rock unit.

The basic rocks show coarse–grained gabbroic textures when they occur in the lower part of the present formation and in the Quartzite–shale Formation below, while they are made up of extrusive rocks and green phyllites in the upper part of this formation. This volcanic member is the only comparable rocks with the epidote–actinolite greenstones of the Vestgötabreen Formation (I–4).

2) The second occurrence of this formation is in the western part of Oscar II Land; along the Forlandsundet Tertiary graben in the north of St. Jonsfjorden and along the fault zone of Permo–Carboniferous rocks south of St. Jonsfjorden.

In the northern localities, the exposure is very poor and the observed rocks are mostly grey dolomitic limestone with black slate, and the estimated thickness is 550 m north of Dahltoppen to the north of Aavatsmarkbreen. At the shore south of Andreasbreen, from the Tertiary graben fault and 1.5 km eastwards, green phyllite alternates with grey and black phyllites and quartzite and calcareous rocks are subordinate. The total thickness here is estimated to maximum 450 m. A serpentinized green phyllite occurs close to the Eastern Border Fault (Fig. 1).

To the south of St. Jonsfjorden, the Calc–argillo–volcanic Formation occurs along the shore between Thorkelsenfjellet and Bulltinden and is composed of thin alternation of grey limestone, quartzitic sandstone and green phyllite. A narrow fault zone of green phyllite follows along the eastern side of Thorkelsenfjellet and Svartfjella. The thickness is 200 m.

The limestone–rich succession along the western shore of Daudmannsøyra has some dark and green phyllites, and can be correlated with this formation.

3) Southern part of Prins Karls Forland

Green phyllite and shaly rocks are dominant in Scotiadalen and Haukedalen, with a few layers of white limestone, while the limestone intercalations become very frequent to the south on Forlandsletta and in Vestflya in the lower part of the formation. These limestone beds form the isolated mountain peaks around the southern tip of Prins Karls Forland. The middle part of the succession is shaly and the upper part is rich in green and dark chocolate coloured phyllites. The width of the occurrence is more than 5.5 km from the middle of Forlandsletta to Forlandsøyane, however, the steep dipping structures certainly include many structural imbrications. The thickness is estimated to be about 1,000 m here. The transition to the lower formation is observed to be gradational along the ridges of the southern Prins Karls Forland and the upper boundary is also gradational to the overlying Tillitic Conglomerate.

A similar phyllite–rich succession occurs as a wedge–shaped zone in the middle of Omondryggen in the west, and is considered the upper part of this formation. The phyllite grades into the apparently overlying Quartzite–sandstone Formation, while the lower calcareous part is cut by steep faults in the east. The main rocks here are black phyllite and various calcareous rocks, occasionally with glossy, coal–rich surfaces, both with some green phyllite intercalations.

4) Grampianfjella in the central Prins Karls Forland

The Calc–argillo–volcanic Formation is folded into a large overturned syncline along the main ridge of Grampianfjella and outcrops along the eastern steep slopes (Fig 3, B and C sections). The succession is rich in dark slate with many limestone beds in the lower part, but the detail is unknown because of poor accessibility.

5) The Laurantzonfjellet area and the eastern foothills of Barentsfjellet and Fuglehukfjellet

The overturned syncline of the central Prins Forland plunges gently north and the overturned upper limb of the present formation extends down below the surface. An incomplete tectonic window occurs north of Laurantzonfjellet, where this formation is composed of green phyllite, shale and sandstone with a thin layer of white limestone. The exposed thickness is 250–300 m.

A green phyllite and shale–sandstone succession occurs along the Western Border Fault in the northern–most part of Prins Karls Forland, apparently underlying conformably the Quartzite–sandstone Formation. These rocks are

considered to be the upper part of the present formation and the exposed thickness is about 150 m.

II-4. *The successions just below the Tillitic Conglomerate*

Concordantly underlying successions below the Tillitic Conglomerate have been observed in several localities, having somewhat different lithologies. All these are thought to belong to some part of the Calc-argillo-volcanic Formation.

II-4a. The limestone succession in southeastern Prins Karls Forland

A 300 m thick limestone occurs along the eastern foothills of Methuenfjellet as the core of a local anticline. The contact to the overlying Tillitic Conglomerate is observed in the west to be concordant. The limestone is shaly and phyllitic with some occasional quartzite beds. The general nature of the succession is unknown, because of poor exposure on the strand-flat, but gives the impression of being similar to the lower part of the Calc-argillo-volcanic Formation.

II-4b. The phyllite succession of Vegardfjella, southeast of St. Jonsfjorden

The rocks here are mostly black and green phyllites, shaly sediments of volcanogenic origin, with a few thin beds of partly oolitic limestone. These phyllites occur concordantly below the Tillitic Conglomerate and are overlain with an unconformity by a varicolored conglomerate of Late Paleozoic age. The exposed thickness is 420 m.

II-4c. The phyllite succession in Svartfjella, southwest of St. Jonsfjorden

Similar phyllites as those of the Vegardfjella area occur here under the Tillitic Conglomerate, where the shaly phyllite is associated with three thin beds of grey limestone. Some leucocratic injection veins, less than 10 cm thick, occur in the phyllite with pinch-and-swell structures. The observed thickness of the succession is 250 m here.

III. *Upper Hecla Hoek rocks*

The rock succession from the Tillitic Conglomerate and above are grouped into two formations, and the Sutor Conglomerate of unknown stratigraphic position is also described here.

III-1. *The Tillitic Conglomerate*

This formation occurs in the western part of Oscar II Land as a 6–7 km wide zone from Engelskbukta to Dahlbreen and a narrow wedge-shaped zone along eastern side of the Late Paleozoic rock zone in Svartfjella, widening to the south in Daudmannsøyra, north of Isfjorden. This formation also occurs further to the east along the western side of Osbornebreen, inner-most St. Jonsfjorden. The same rocks are seen in the southern part of Prins Karls Forland along the eastern slopes of the ridge from Doddsfjellet to Methuenfjellet. The thickness is more than 1,000 m in both Oscar II Land and Prins Karls Forland.

The upper boundary is conformable with the overlying Bulltinden Formation in Prins Heinrichfjella, north of St. Jonsfjorden. This formation overlies different rocks presumably of the Calc-argillo-volcanic Formation (II-4a-c). Although the contacts are obliterated by strong schistosity, these evidences indicate that the base of this formation may show an angular unconformity.

This formation is mainly composed of characteristic schistose conglomerates with numerous grey dolomite pebbles. Some thin dolomitic limestone beds occur in the upper part. Many thin phyllite have been recorded in the upper part in Haaken Mathiesenfjella, west of Comfortlessbreen.

Besides the boulders of dolomite, grey limestone and quartzite are common among the pebbles and boulders and the limestone occasionally contains stromatolite and oolite. Pebbles of granitic rocks are rarely found. The pebbles and boulders are strongly flattened and elongated, some are more than 1 m long. They are scattered and show very poor sorting, and a few to about 10 pieces per 1 m² are observed in the schistosity plane. The matrix of the rocks is a calcareous and siliceous shale with large amounts of dolomite, calcite and sericite. Quartz, albitic plagioclase, graphite and opaques are subordinate.

From the scattered and unsorted occurrences of pebbles and boulders in the fine-grained matrix, the rocks of this formation are considered as glacial sediments of Vendian age.

III-2. *The Bulltinden Formation*

This formation was defined by HORSFIELD (1972). The type locality is Bulltinden, western Holmesletfjella and Motalafjella in the south of St. Jonsfjorden. The lower boundary is a low angle fault and no depositional contact has been found in this area. The formation is composed of three members; a frequent alternation of coarse-grained, often conglomeratic, sandstone and shale in the lower part, very coarse conglomerate and shale in the middle and a limestone in the upper part.

In the lower member, the sandstone is relatively well sorted and occupies more than half of the succession. The conglomeratic nature of this sandstone distinguishes it from those of the other formations below. The alternating shale often shows well developed laminar structures. The thickness of this member in Holmesletfjella is about 1,000 m.

The conglomerate of the middle member is moderately sorted, the thickness of the sorted units is several metres, from coarse-grained sandstone to polymict boulder conglomerate. The pebbles and boulders are sub-rounded to round, up to 1.5 m across. Boulders of grey limestone and brown dolomite are abundant with subordinate quartzite, sandstone, shale, black slate, chlorite phyllite, quartz schist, schistose amphibolite, meta-diabase, garnet-mica schist and skarn. These pebbles and boulders are scarcely deformed. Two beds of conglomerate are present in the southern part of Motalafjella, the lower bed is more than 200 m thick and the upper one 150 m. Between these two beds a 100 m thick dark shale occurs. This shale has well developed laminations of shale and silt, and numerous thin sandstone layers.

A grey limestone of 10 m thickness occurs in the middle member on the

northeastern edge of Motalafjella. This unit is composed of conglomeratic limestone with angular breccias and dense grey limestone, and the latter includes many fossils of Lower Paleozoic age. The rocks are only slightly recrystallized, and brachiopods, cephalopods, gastropods, stromatolites, crinoids, table and horn corals have been collected. An upper Ordovician to lower Silurian age is suggested by a brief preliminary examination.

The conglomerate is very well developed in Bulltinden. A conglomerate with such large boulders suggests that this is an intermontane deposit at the foots of a rapid raising hinterland.

A distinct grey limestone of 50 m thickness occurs above the middle member in the area from Bulltinden to Motalafjella. HORSFIELD (1972) excluded this from the Bulltinden Formation. However, doubtful fragments of brachiopods were found in a scree block directly lying on this limestone in Motalafjella and, therefore, this limestone is included in the Bulltinden Formation as its upper member. The limestone sandwiches the Vestgötabreen Formation and is seen on the ridge of southwestern Holmesletfjella with the associated black shale above.

The Bulltinden Formation occurs across St. Jonsfjorden to the north, in Ankerfjella and Prins Heinrichfjella, on the northern side of Aavatsmarkbreen where it forms part of a syncline whose western limb is cut by a steep fault (Fig. 3, section F). The conglomeratic rocks here are mostly schistose and are often accompanied by sandstone and quartzite beds. Intercalated limestone beds are occasionally observed in the lower part on the slopes of Ankerfjella and the northern side of Aavatsmarkbreen. Poorly preserved fossil-like fragments were found at a locality in Ankerfjella. Although most rocks are obliterated by strong schistosity, the lower boundary with the Tillitic Conglomerate is considered to be conformable in Prins Heinrichfjella.

III-3. *The Sutor Conglomerate*

A remarkable conglomerate occurs at Sutorfjella on the west coast of Prins Karls Forland. This was firstly described by ATKINSON (1960). The conglomerate is very coarse with sub-angular boulders up to a few metres across, very poorly sorted and have a green or dark brown coloured matrix of coarse-grained sandstone. White quartzite boulders with dark reddish weathered crust are notable. The quartzitic conglomerate with angular limestone breccias from the Quartzite-sandstone Formation (II-2) below are also included as pebbles and cobbles. This formation overlies the Quartzite-sandstone Formation by a low angle fault, and its stratigraphic position is unknown.

SUMMARY OF STRATIGRAPHY

Due to a strong deformation and faulting during Tertiary, most rock units in the present area are separated by tectonic breaks which make stratigraphic correlations difficult.

TYRRELL (1924) made the first classification of the rock successions in Prins Karls Forland and his divisions generally agree with those of the present paper. ATKINSON (1956 and 1960) made further sub-divisions, but his descriptions are

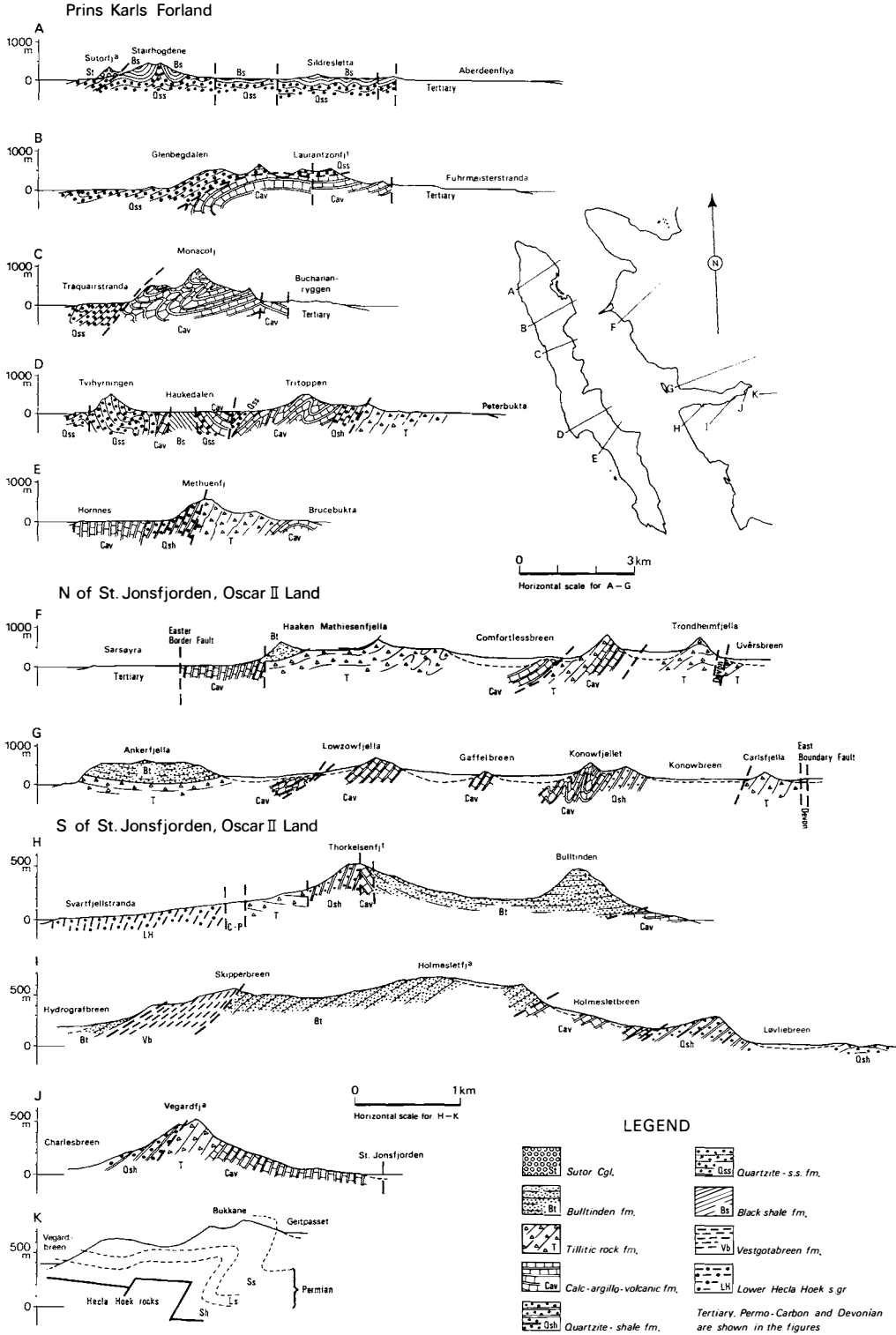


Fig. 3. Geologic profiles from Prins Karls Forland (A-E) and Oscar II Land (F-K).

mainly concerned with the structural units and not based on the lithology. Therefore, it is difficult to make correlations with his divisions, but an attempt is shown in Table 1.

The Tillitic Conglomerate can be correlated well over the whole western Spitsbergen, and the correlation of the Bulltinden Formation with the Sørkapp Land Group is certain. However, the correlations of the other older formations are still tentative and based on general lithologic similarities and the grade of metamorphism. The correlations of the Quartzite-sandstone Formation and the Black shale Formation of Prins Karls Forland are the most problematic since these are based on the structural interpretations.

The important points concerning the stratigraphy are summarized below:

1) The Bulltinden Formation, presumably Ordovician to Silurian, includes the youngest succession of Upper Hecla Hoek Supergroup in Svalbard and is an intermontane molasse of the Caledonian orogeny.

2) The Tillitic Conglomerate, a Varangian glacial deposit, is the best key for regional correlation. A clino-unconformity is estimated at the base of this formation.

3) Two volcanic horizons have been distinguished in the studied area: in the upper part of the Tillitic Conglomerate and in the upper part of the Calc-argillo-volcanic Formation. The holo-crystalline basic rocks in the lower part of the Calc-argillo-volcanic Formation and the Quartzite-shale Formation are considered to be the feeders and intrusives related to the volcanism in the upper Calc-argillo-volcanic Formation of the Middle Hecla Hoek succession.

The basic rocks of the Eimfjellet Group in Hornsund are the third basic igneous activity in the Hecla Hoek succession of western Spitsbergen.

The Vestgötåbreen Formation may be correlated to one of the older two.

4) The Quartzite-shale Formation of Oscar II Land and the Quartzite-sandstone Formation of Prins Karls Forland are correlated to the Slyngfjellet Conglomerate of southwestern Spitsbergen. If this is the case, the conglomeratic nature of these two formations are far less than the Slyngfjellet Conglomerate and their lithologies are similar to that of the Veteranen Group of Lomfjorden Supergroup in Ny Friesland, although their thickness is about 1/4 of the latter.

The Black shale Formation below the Quartzite-sandstone Formation may belong to the Lower Hecla Hoek succession.

5) The high grade metamorphic rocks including biotite, garnet, sillimanite are certainly of the Lower Hecla Hoek succession. The iron-ore-bearing slate in central Prins Karls Forland is correlated to the similar succession below Slyngfjellet Conglomerate in the east of Recherchebreen, south of Bellsund.

Table 1.

Upper Hecla Hoek	Hornsund (Birkenmajer 1974)	Bellsund (Hjelle 1969)		Oscar II Land (present paper)	Prins Karls Forland		Ny Friesland (Harland et al. 1966)
		south	north		present paper	Tyrrell 1924	
Upper Hecla Hoek	Sørkapp Land Gr.			Bulltinden Fm.			
	Sofiekammen Gr.						Oslobreen Gr.
Upper Hecla Hoek	Gáshamna phy.	Bellsund-Dunderdalen tillite	Kapp Linné Tillite Lågneset Tillite	Tillitic Cgl.	Tillitic Cgl.	Ferrier Gr. Geikée Gr. Gordon Gr.	Polarisbreen Gr.
	Höferpynten dol.	Konglomeratfj. shale-quartzite volcanics calc beds	Lågneset-Kapp Martin green-grey shale Lågnesrabbanne calc beds	Calc-argillo-volcanic Fm.	Calc-argillo-volcanic Fm.	Ferrier Peak Series	Akademikerbreen Gr.
M. Hecla Hoek	Slyngfjellet Cgl	Konglomeratfjellet Conglomerate	Kapp Martin Cgl.	Quartzite-shale Fm.	Quartzite-sandstone Fm.(conglomeratic)	Fuglehuken Gr.	Veteranen Gr.
	Deilegga Gr.	East of Recherchebreen calc. + phy. + iron-bed calc. qtz.			Black shale Fm. Pelite with Fe-ore	North Grampian Series	Planetfjella Gr.
Lower Hecla Hoek	Eimfjellet Gr.			(Vestgötabreen Fm.?)			Harkerbreen Gr.
	Isbjørnhamna Gr.			Quartzite-shale alternation Gar-bi-schist Sill-gar-bi-gneiss			Finlandveggen Gr.

Hornsund Superg. Torellbreen S-gr. Stubendorffbr. S-gr.

6) The Sutor Conglomerate resembles the basal Devonian conglomerate elsewhere in the northern Spitsbergen as well as the Tertiary basal conglomerate on the eastern ridges of Grampianfjella, and no definite correlation of this conglomerate can be made at the present stage.

Geologic structures

Complicated nappe structures were proposed by ATKINSON (1960) in Prins Karls Forland, and the mesoscopic structural elements were studied by WEISS (1953) in the south of St. Jonsfjorden. A new tectonic interpretation has been proposed by HARLAND and HORSFIELD (1974) based on the studies mentioned above and the work on Brøggerhalvøya by CHALLINOR (1967) and others. In the present paper, the simplest possible interpretations are adopted. The observed structures are summarized below in retroactive order (Figs. 5 and 4).

THE STEEPLY DIPPING TRANSVERSE FAULTS OF ENE–WSW STRIKE

These faults cross cut the general strike of the regional structures of both Tertiary and Caledonian origin.

a) *The Selvågen Fault*: The Western Border Fault of NNW–SSE strike is dislocated and rotated slightly between the north and the south side of Selvågen, and a local fault of SW–NE strike is assumed here. This fault may converge into the fault in Scotiadalen, which is considered to be a branch of the Western Border Fault.

A E–W striking minor fault was observed from Magdabreen to Alfredbreen in the southeast of Grampiafjella.

b) *The Dahlbreen Fault*: The Eastern Border Fault goes into the sea at Snippen, northwestern entrance of Dahlbreen, while Hermansenøya to the west of the southern extension of the fault is composed of the Tillitic Conglomerate. Besides, the syncline axis of the Ankerfjella is displaced to the west in Prins Heinrichfjella. Thus, a transverse fault is assumed along the northern part of Dahlbreen.

These local displacements show that the ENE–WSW striking steep faults are later than the border faults of the Forlandsundet Tertiary graben. Another four steeply dipping faults of similar strike are supposed in the southwestern part of Oscar II Land (Fig. 3).

Faults with similar trends are known elsewhere along the west Spitsbergen Tertiary folded zone south of Isfjorden. This young trend is one of the most distinct fracture trends all over Svalbard and is manifested by the direction of shore lines and major fjords.

THE TERTIARY GRABEN

Faults of NNW–SSE trend are important structures in western Spitsbergen and bound the graben structures along the west coast.

The Western and Eastern Border Fault are typical in that they are almost

vertical and diverge only locally. Three steep faults branch off from the Western Border Fault west of Richardlaguna in the northern Prins Karls Forland. They curve gently to a NW–SE trend in the north, which suggests that a horizontal SW–NE component of stress was present during the early time of graben formation.

The N–S striking steep faults along Scotiadalen and Haukedalen are also branches of the Western Border Fault. These faults produced step-wise movements of the wedge-shaped blocks enclosed by them. Some steep faults west of Scotiafjellet and Omondryggen are of the same origin.

The Eastern Border Fault also diverges in the south of St. Jonsfjorden and encloses a narrow zone of Late Paleozoic rocks from Svartfjella to the east of Daudmannsodden. Here the Eastern Border Fault is located some distance offshore to the west.

It is noteworthy that the rocks of high metamorphic grades occur within the fault zones of both the Western and the Eastern Border Fault: the sillimanite–garnet–biotite gneisses (I–1) and the iron–ore–bearing pelite (I–2) along the Western Border Fault and the garnet–biotite schist (I–3) along the Eastern Border Fault. Dolomite and serpentized greenrocks occur along these faults as discontinuous, squeezed-out thin bodies along the Eastern Border Fault from Kaffiøyra to Sarsøyra and along a splay fault from Richardlaguna to Balfourfjellet in northern Prins Karls Forland.

A steep fault of similar strike is supposed to occur along Bullbreen.

The fault bounding the Hecla Hoek rocks to the east from Kongsvegen to Trygghamna, the East Boundary Fault, is steep and has similar strike as the graben faults. Although initiated earlier, this fault was certainly reactivated during the period of graben formation. The occurrence of Devonian rock along this fault is noteworthy.

The Forlandsundet Tertiary graben is considered to be formed in Eocene–Miocene time (HARLAND and HORSFIELD, 1974 and BIRKENMAJER, 1972) and extends offshore western Spitsbergen. Another graben occurs around Øylandsodden and Sørkapp Land, and all these structures are parallel to the Spitsbergen Fracture Zone some 150 km offshore to the west along the continental margin. These graben structures are younger than the main folding phase of the Tertiary orogeny.

MAIN TERTIARY OROGENY

Low to moderately westward dipping faults and regional folds with NNW–SSE axial trend were formed during the Tertiary orogeny and are the main elements of tectonic architecture of western Spitsbergen.

The youngest structures formed during this period is open folds of NNW–SSE strike, which are observed in the northern part of Prins Karls Forland. Most folds are gentle and shallow with 1–2 km wave length and a few hundred metres in amplitude, and almost concentric style with vertical axial plane. Some of the folds are asymmetric conical folds with axial planes dipping steeply to the west. Axial surface cleavages are weakly developed locally in the eastern part of Barentsfjellet. The anticlines tend to have acute crests which turn into

thrust faults with steep westward dip. This suggests that the development of these folds preceded the formation of the graben faults. The N-S and NNW-SSE striking minor fold axes (Fig. 4-A) correspond to these folds in the middle and northern Prins Karls Forland. These open folds refolded the overturned syncline of Grampianfjella in the northern part of Prins Karls Forland and are therefore younger than the main phase of strong folding and thrusting. An open syncline in Ankerfjella belongs to this young deformation.

Typical fold structures of the main Tertiary orogeny are represented by the overturned syncline of Grampianfjella, middle Prins Karls Forland (Fig. 3 section C). Although the details are not known because of the inaccessible steep cliffs, the syncline core was observed from a distance and can be reconstructed from associated minor folds of dm wave length with moderately westward dipping axial planes. The main axis of the syncline plunges very gently to the north and the overturned upper limb is roughly horizontal in northern Prins Karls Forland, and the axial plane has very gentle dip to the west.

In the western part of middle Prins Karls Forland, a steep westward dipping fault, the West Grampian Fault, is present in the overturned upper limb, and separates different lithologic units. This fault disappears somewhere around Laurantzofjellet in the north. The eastern limb of the syncline is cut by the moderately westward dipping East Grampian Fault along the eastern foothills of Grampianfjella. From this fault splay a few faults around Bouréefjellet enclosing the highly metamorphosed deeper rocks. The Southern Forland Fault is the extension of the East Grampian Fault and the Calc-argillo-volcanic Formation to the west of the fault is supposed to include some imbrications of tight folds which are southern extensions of the overturned syncline of Grampianfjella.

In Oscar II Land, all lithologic units occur in zones with a NNW-SSE trend, bounded by low to moderately westward dipping faults. These faults are reverse, mostly less than 45° , and brought the older rocks up upon the younger ones. They were later cut by the graben border fault in the western part. Local tight folds accompanying weak cleavages produced by the thrust movement were observed around the faults: in the west of Comfortlessbreen, in Konowfjellet, eastern ridge of Ankerfjella and west of Piriepynten.

The large scale thrust structures of Oscar II Land are certainly comparable to the major structures of Prins Karls Forland, judging from the common axial trend and thrusting from the same SW direction. Structural imbrications are of a larger scale in Oscar II Land than in Prins Karls Forland.

The Carboniferous to Triassic rocks east of the Hecla Hoek region have been folded in step-style folds with the NNW-SSE axial trend with small wavelengths and partly overturned to the east in the western part like in the Osbornebreen-Charlesbreen-Trollheimen area (Fig. 3, Section K), and large step folds with a long flat western limb in the eastern part like in the Borebreen-Wahlenbergbreen-Kongsvegen area. These folds of the sedimentary blankets are considered to be the reflections of the thrust block movements of the underlying Hecla Hoek rocks as seen in western Oscar II Land. Mesozoic dolerites are involved in these folds elsewhere.

The arc-shaped zone of crystalline schists and phyllites of Brøggerhalvøya and Holtafjella has harmonious thrust structures (CHALLINOR, 1967) with the present Hecla Hoek rocks, and also involves the coal-bearing Tertiary strata at Ny-Ålesund.

From the structural evidence in the younger rocks, some gentle folds and thrusts of the present area are considered to be caused by Tertiary deformation preceeding the formation of the Forlandsundet graben. Although some of them may be of Caledonian origin, they were reactivated during the Tertiary orogeny. This main phase of Tertiary orogeny occurred sometime in the Paleogene Tertiary period and is probably related to the relative motion between the Greenland and the European plate during the early opening of the Norwegian–Greenland Sea. The tectonic transport from SW to NE is opposite to what ATKINSON (1960) discussed, and the dextral strike slip movement is represented by a gentle swing of westward dipping faults from a NNW–SSE to a NW–SE trend as seen in the northern part of Prins Karls Forland, around Comfortlessbreen and in the southern side of St. Jonsfjorden.

THE MESOZOIC DOLERITES

A few small dykes of dolerite cut the Tillitic Conglomerate northwest and southeast of Løvenskioldfonna, Oscar II Land. Sheets of dolerite with various thicknesses occur in the Permo–Carboniferous rocks east of the Hecla Hoek region. No systematic trend can be distinguished.

This igneous activity marks the end of the stable platform period from Late Carboniferous to Middle Mesozoic, and the beginning of Tertiary movements which are connected with the opening of Northern Atlantic.

THE LATE PALEOZOIC ROCKS

A narrow downfaulted zone of Permo–Carboniferous rocks, a few hundred metres wide, occurs from Svartfjella to Kapp Scania along the western shore of Oscar II Land. Two more small occurrences were found along the fault from Trondheimfjella to Kregnestoppen.

Some hundreds of pieces of angular blocks of doubtless Permo–Carboniferous rocks occur on a small hill to the northeast of Svartfjella, on the amphibolites of the Vestgötabreen Formation. These blocks cannot be regarded as a glacial drift, and the occurrence is problematic.

These Permo–Carboniferous rocks indicate that the Late Paleozoic shallow sea sediments, probably together with the Mesozoic ones, once covered the whole area of Oscar II Land.

DEVONIAN ROCKS

A conglomerate with grey and red limestone pebbles and red sandstone has been seen at the eastern tip of Haraldfjellet, west side of Osbornebreen. Another fault slice of Devonian rocks with fossils was found at Kregnestoppen, northeast of Løvenskioldfonna. These are in fault contact with the Tillitic Conglomerate and have a lithology which resembles that of the Devonian rocks

in northern Spitsbergen. The eastern occurrence is along the N-S striking East Boundary Fault of the Hecla Hoek region, and the northern extension joins the Devonian rocks of Lovénøyane in Kongsfjorden (GJELSVIK 1974). This suggests that one of the faults bounding the Devonian graben in the northern Spitsbergen might pass along the East Boundary Fault of Hecla Hoek region in the middle Spitsbergen.

TWO PHASES OF CALEDONIAN DEFORMATION DISTINGUISHED FROM THE MESOSCOPIC STRUCTURAL ELEMENTS (FIG. 4)

All Hecla Hoek rocks, even those of the early Paleozoic Bulltinden Formation, show strong cleavages, developed parallel to the axial surfaces of small tight folds. The regional fold structures are represented by the lithologic units of some hundreds of metres in thickness, and the cleavages also show the same folded structures. This means that the cleavages were involved in the Tertiary regional folds and are consequently older than the Tertiary deformations.

The mesoscopic structural elements from Oscar II Land are summarized in Figs. 4-B and 4-C. The axial crests of the folded beddings were lost by strong cleavages in most cases, and the measurable beddings are mostly sub-parallel to the cleavages. Therefore, these two planar elements were counted together in the diagrams.

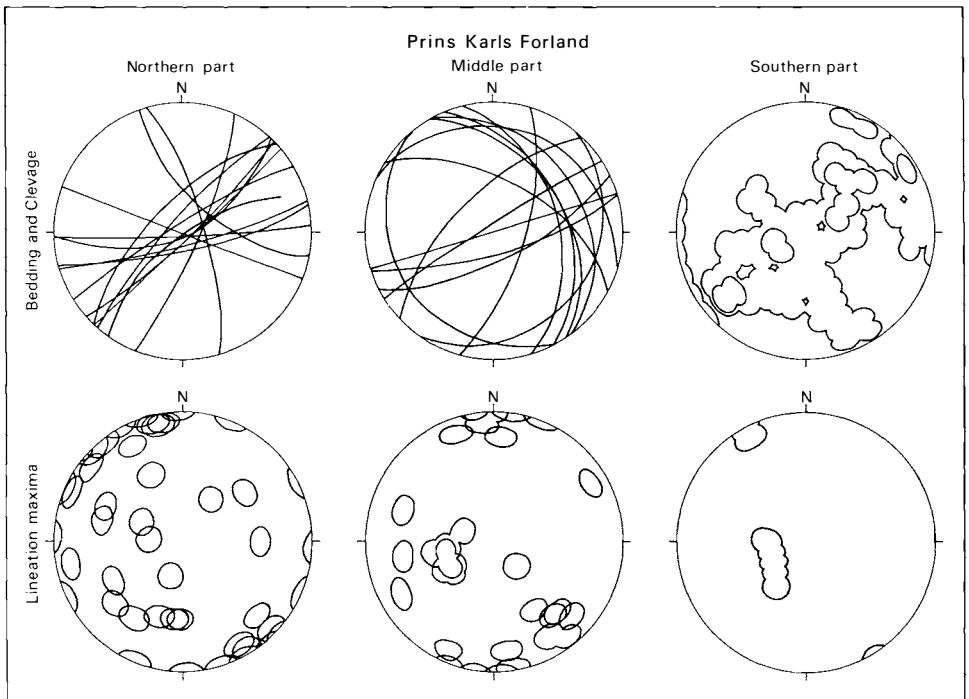


Fig. 4a. Fabric diagrams of the mesoscopic structural elements.

A. Summaries of the planar and linear structural elements from Prins Karls Forland. The girdles are drawn from the maximum of the bedding-cleavage diagrams of the subareas from the northern and middle parts of Prins Karls Forland, while the southern part is shown by a common contoured diagram. The maxima and sub-maxima of lineations from the subareas are projected in the lineation maxima diagrams.

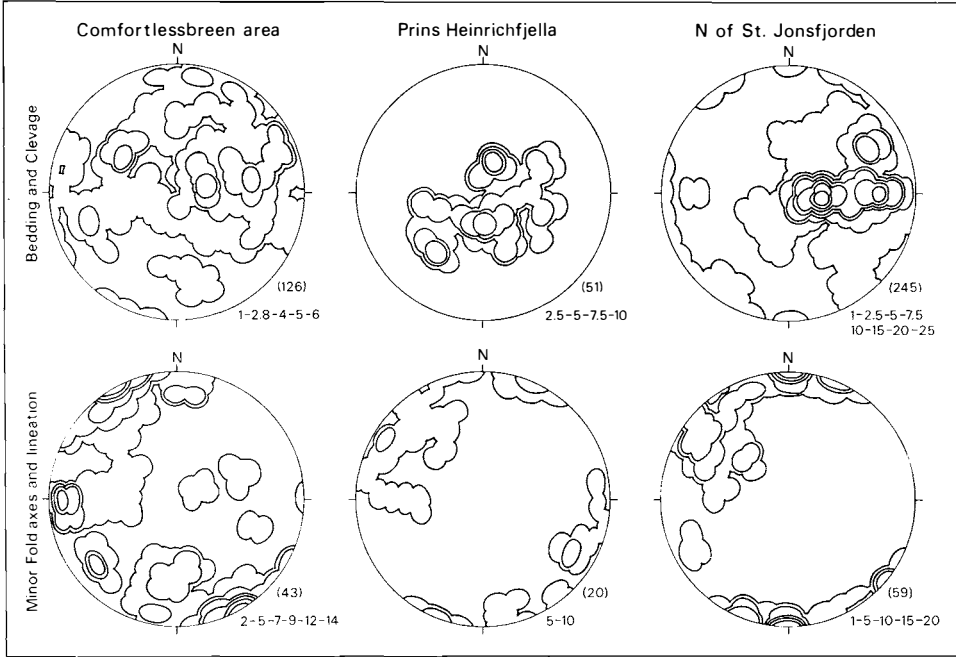


Fig. 4b.

B. Fabric diagrams from northern Oscar II Land.

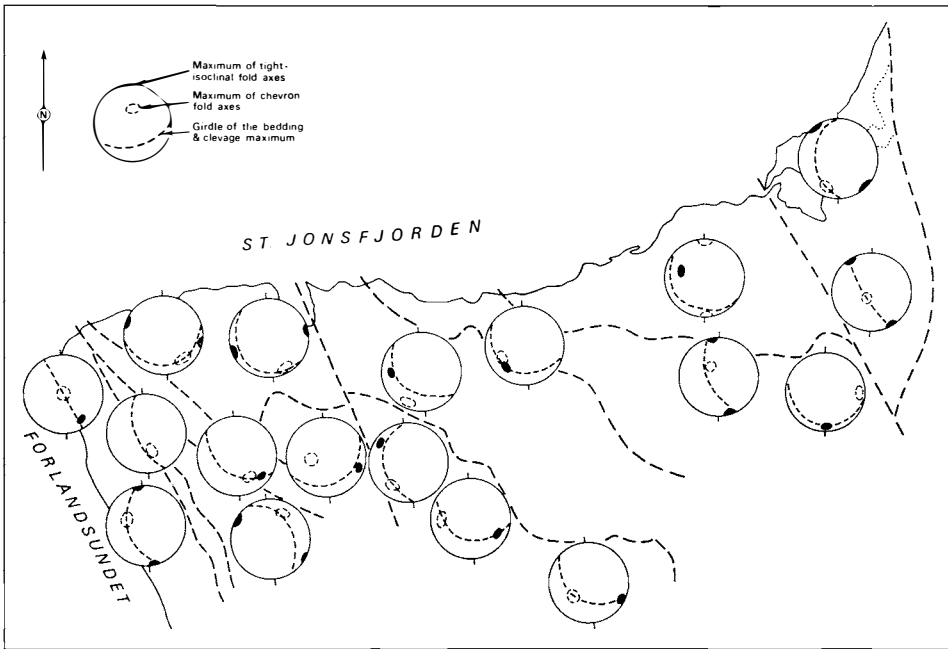


Fig. 4c.

C. Simplified fabric diagrams from the southern side of St. Jonsfjorden.

Two types of minor fold axes were distinguished in the area south of St. Jonsfjorden: 1) younger chevron folds and 2) older tight-isoclinal folds. In most diagrams of Fig. 4-C, the maxima of observed fold axes of both types are projected on or near the girdle, while they deviate distinctly from the girdle in a few diagrams. This discrepancy indicates that the present bedding-cleavage structures do not have their original trend, but are rotated by the later Tertiary movements. The rotation of both types of minor fold axes by the Tertiary deformations is evident since the trend of fold axes maxima changes from one sub-area to another in a thrust block. WEISS (1953) concluded that the N-S striking lineations are younger than the E-W striking ones and are of Tertiary origin, but our detailed analysis shows that both must be older than the Tertiary movements.

In the southern part of Prins Karls Forland, ATKINSON (1956) associated the younger lineations of NE-SW strike, the chevron- and kink-folds, with the a-lineation of his nappe movement. But, chevron- and kink-type folds are not likely to be a-lineations, and the scattering and rotation of statistical maxima of these fold axes (Fig. 4-A) strongly suggest that these are older than the Tertiary deformation which is represented by N-S to NW-SE striking lineations.

In St. Jonsfjorden no comparable lineation corresponding to the two types of minor folds in the Hecla Hoek rocks was seen in the Permo-Carboniferous rocks involved in the Tertiary deformation, which means that both chevron- and tight-isoclinal folds were formed prior to Lower Carboniferous.

The tight-isoclinal folds always associate the cleavages and all metamorphic minerals occur under the control of cleavages and fold axes. Therefore, the folds represent the main recrystallization phase in the regional metamorphism. The axes of the tight-isoclinal folds have a more consistent trend than the chevron fold axes. This gives them an impression of being the youngest of the two types. However, the chevron folds evidently deformed the axes and axial planes of the tight-isoclinal folds and are definitely younger. The small scattering of the tight-isoclinal fold axes is due to the similarity in axial trend between these and that of main Tertiary deformation.

The N-S to NW-SE striking minor folds in the northern half of Prins Karls Forland (Fig. 4-B) are caused by the Tertiary deformation. To distinguish Tertiary structural elements from the more intense Caledonian ones is very difficult in the Hecla Hoek rocks of western Spitsbergen.

PRE-CALEDONIAN EVENTS

Lack of a major calcareous Cambro-Ordovician succession in this area may be due to either a structural missing or tectonic events during early Paleozoic as proposed in Hornsund by BIRKENMAJER (1972).

An unconformity at the base of the Tillitic Conglomerate is suggested by the different underlying rocks. The occurrence of a few granitic rock pebbles in the conglomerate indicates the existence of crystalline basement in the vicinity somewhere. However, the grade of metamorphism is progressive downwards in the stratigraphic sequence and the biotite isograd was reached in parts of the Tillitic Conglomerate.

The Quartzite–shale Formation and the Quartzite–sandstone Formation, which can be correlated to the Slyngfjellet Conglomerate at the base of Middle Hecla Hoek succession, have no distinct conglomerate as in the southern area of western Spitsbergen.

Mainly due to the strong disturbances from the Tertiary deformation, the analysis of pre–Caledonian events is very difficult in this part of Spitspergen.

Metamorphism

The phyllitic cleavages, formed as axial surface cleavage of the tight-isoclinal folds, are associated with progressive metamorphic mineral assemblages, while the younger structures as the chevron folds and the shear cleavages along the Tertiary disturbance zones are accompanied by retrogressive metamorphism only. Therefore, the metamorphic recrystallization of the Hecla Hoek rocks of this area is mainly of Caledonian origin.

The grade of metamorphism increases with stratigraphic depth. In the uppermost unit, the Bulltinden Formation, detrital muscovite and biotite are well preserved in the sandstone, while small flakes of sericite represent strongly crenulated microfolds and secondary cleavages in the shaly rocks. Detrital muscovite is preserved in the stratigraphic sequence down to the level of the Tillitic Conglomerate. Recrystallized chlorite and sericite occur in large amounts in the grey and green phyllites of the Calc–argillo–volcanic Formation. Chloritoid occurs in the phyllites of this formation in the middle and southern part of Prins Karls Forland (ATKINSON 1956).

The holocrystalline basic rocks in the Quartzite–shale Formation and the Calc–argillo–volcanic Formation are affected by post– or late–magmatic alteration and most mafic constituents have been converted into (biotite)–actinolite–epidote–chlorite–rutile–opaque aggregates. Quartz occupies irregular interstitial spaces and the plagioclase is totally dusty. The metamorphic assemblages of these rocks are characterized by (biotite), epidote, actinolite, chlorite and stilpnomelane of brown and green varieties.

The rocks of the Tillitic Conglomerate in St. Jonsfjorden have chlorite–sericite–bearing assemblages, while those of the southern Prins Karls Forland, east of Doddsfjellet, have biotite. Thus, the biotite isograd is not exactly parallel to the boundary of the lithologic units. Anyway, the biotite isograd was attained and typical nemato– and grano–blastic textures were achieved by recrystallization around the base of the Upper Hecla Hoek succession in the present area.

The rocks just below the Tillitic Conglomerate are essentially chlorite–sericite and graphite phyllites on the northern shore of Vegardfjella, eastern St. Jonsfjorden. The green phyllites at this locality are partly of spotted texture with albite poikiloblasts, some of which show S-shaped outline which represents syntectonic growth. The green schists in the same stratigraphic position southeast of Svartfjella show banded structures with boudinaged leucocratic layers of albitic rock. Chessboard plagioclases of albitic composition and quartz are the main constituents in the leuco-layers and the matrix is the same as

surrounding fine-grained schistose rocks. Although the occurrence is limited to a small area, one may suggest an introduction of quartz-dioritic materials into this structural level. This resembles the granitization described from Hornsund by BIRKENMAJER and NAREBSKI (1960), although it is there of the Lower Hecla Hoek succession.

Highly metamorphosed rocks are correlated to Lower Hecla Hoek. The garnet-biotite schists of Svartfjellstranda show two stages of recrystallization: the older with helicitic garnet and lepidoblastic biotite and muscovite, and the younger represented by large flakes of pale brown biotite growing regardless of the cleavages. The chevron and kink type minor folds made very weak local cleavages at the crests and have no progressive recrystallization. Therefore, these two phases of metamorphic recrystallization probably belong to the Caledonian main deformation phase, the older is syntectonic and the later is late- or post-kinematic thermal recrystallization. Later growth of biotite is also evident in the sillimanite-garnet-biotite gneiss of Bouréefjellet, Prins Karls Forland. Sillimanite and garnet are granulated relics enclosed in chlorite and sericite aggregates and many mylonitic fractures cut the granoblastic matrix. The large biotite flakes occur on the mylonitic matrix and were crushed and altered by the fractures formed during the Tertiary movements.

ATKINSON (1956) reported staurolite from the east of Monacofjellet, Gramplanfjella. It is likely to conclude, from the occurrences of stilpnomelane and chloritoid in the lower grade rocks and staurolite, almandine-garnet and sillimanite in the higher grade ones, that the Hecla Hoek rocks of this area were metamorphosed under the conditions which correspond to the intermediate temperature/pressure facies series, in a range from greenschist facies to upper amphibolite facies.

The glaucophane-bearing rocks and eclogite of the Vestgötabreen Formation have different characteristics from the other Hecla Hoek rocks and will be discussed in a separate article of this issue (OHTA, this volume).

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Blue schists from Motalafjella, Western Spitsbergen

By YOSHIHIDE OHTA

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Abstract

The occurrence of Caledonian glaucophane schists and eclogitic rocks in the Vestgötabreen area, south of St. Jonsfjorden, western part of Spitsbergen, has been studied in detail, and the metamorphic conditions are discussed based on new chemical analyses of 30 bulk rocks and 25 constituent minerals. The rocks occur as a thin thrust schuppen, lying between the lower grade metamorphic and meta-sedimentary rocks of lower Paleozoic age. The original rocks of the glaucophane schists are considered to be argillaceous quartzite and mixtures of argillaceous sediments and basic volcanic rocks. The metamorphic conditions of the schists are similar to those of the Type IV glaucophane schists of California, and the formation conditions of the eclogitic rocks have been calculated to 9.7 kb and 540–570°C from the Fe/Mg partitioning between clinopyroxene and garnet. The metamorphism of these rocks is a transitional facies series between the blue schist type and the intermediate P/T type of metamorphic facies series. The metamorphism of Hecla Hoek rocks of western Spitsbergen is the intermediate P/T facies series. These metamorphic facies series are considered to be developed from the eugeo-

synclinal Torellian basin of pre-Cambrian time. The possibility of two sets of paired metamorphic zones during the Caledonian period and of a pre-Cambrian metamorphism of the intermediate P/T facies series in Svalbard are discussed.

I. Introduction

An occurrence of glaucophane schists and eclogitic rocks in the Vestgötabreen and Eidembreen areas, south of St. Jonsfjorden, has been known since 1957 among British geologists studying Svalbard. The extension of the occurrence, the outline of mineral assemblages, and bulk rock chemistry were reported by HORSFIELD in 1972, with five K–A ages. Although the area including this locality has been affected by the Tertiary deformation, these particular rocks are considered to be formed during the Caledonian Orogeny, as indicated by the radiometric ages more than 410 m.y. old. Thus, this is one of the oldest in the world among the glaucophane schists hitherto reported, only revealable is the Ballantrae complex in Ayrshire, Scotland.

This area was mapped in detail by the present author in the summers of 1973 and 1975, and all rock varieties of this particular rock formation have been collected. Details of the field occurrence and chemical properties of the rocks and constituent minerals are presented in this paper, and some comparative considerations are made between these rocks and other metamorphic rocks in Svalbard.

II. Geological setting

The distribution of blue schists and their associated rocks is shown in Fig. 1. HORSFIELD (1972) distinguished the following rocks in his geological map: (1) Bulltinden Formation: conglomerate, sandstone and shale; (2) marbles; (3) mica schists, mylonites and breccias; (4) epidote-actinolite greenstones; and (5) glaucophane-bearing rocks. Coarse-grained glaucophane, garnet and muscovite schists were referred by him to the Vestgötabreen suite. Rocks with abundant chloritoid, serpentinite, and others having sodic clinopyroxene, i.e. eclogite type rocks and pyroxenite pods, are also described by him from this rock suite.

HORSFIELD's map has been modified a little north of Skipperbreen and in the southern half of Motalafjella, based on new mapping by the present author. He found a fragment of a badly preserved brachiopod from a scree block of limestone in the middle-north of Motalafjella. This block came no doubt from the limestone lying above the conglomerate of the Bulltinden Formation which has a fossiliferous limestone lens of lower Paleozoic age. For this reason, this limestone (the (2) marbles of HORSFIELD) is included in Bulltinden Formation in this paper.

The rocks of (3), (4), and (5) of HORSFIELD, occur as a narrow zone less than a few hundred metres thick and closely associated for more than 10 km along the strike. The development of cleavage and the degree of recrystallization of these rocks are strikingly different from those of the Bulltinden Formation. Therefore, all these rocks are tentatively grouped as the Vestgötabreen Formation in the present paper.

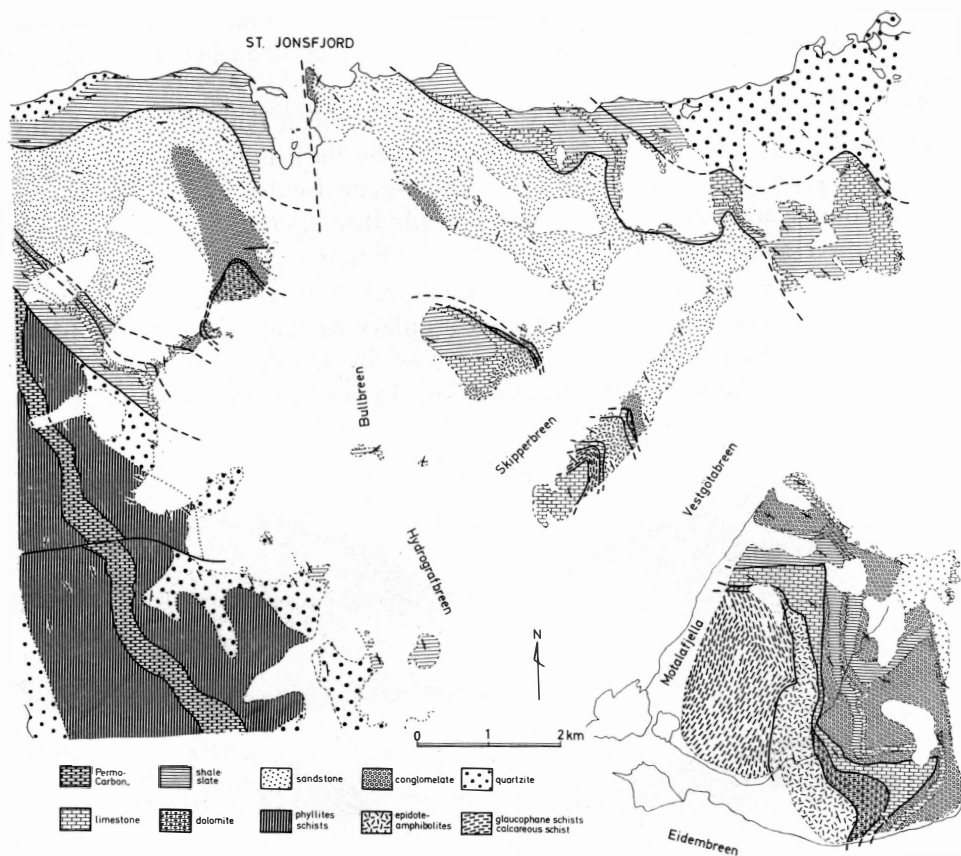


Fig. 1. Geological map of the area from St. Jonsfjorden to Motalafjella.

The Bulltinden Formation is composed of (1) coarse-grained sandstone (often conglomeratic) and shale alternations more than 1,000 m thick, (2) boulder conglomerate 500 m thick, (3) fine-grained sandstone-shale alternations 300 m thick, and (4) banded grey limestone less than 50 m thick, in ascending order. A conglomeratic limestone bed 20 m thick occurs in the

boulder conglomerate and includes many fossils which suggest upper Ordovician — lower Silurian age (the fossils will be described in a separate paper).

Both the lower and upper sides of the Vestgötabreen Formation are in low angle fault contact with the grey limestone of the Bulltinden Formation.

The rocks of the Vestgötabreen Formation can be divided into two members: (1) the epidote actinolite greenstones, black phyllite, dolomite and serpentinite in the lower part and (2) the glaucophane-bearing rocks, eclogitic rocks, calcareous schists and dolomite in the upper part. The distribution and thickness of each rock type differ very much from place to place (Figs. 1 and 2) and the lower and upper members are separated by a subordinate low angle, probably reverse fault, locally intercalating the grey limestone layers of the

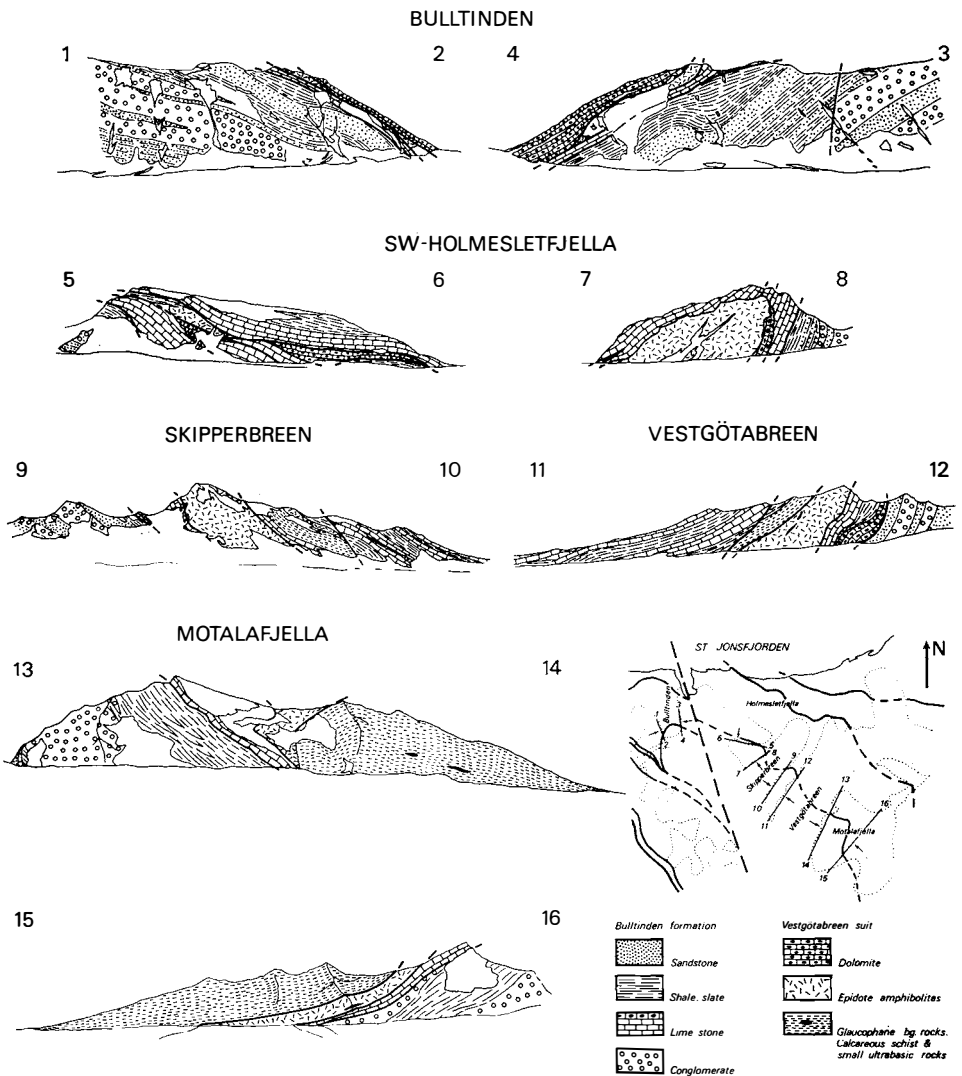


Fig. 2. Occurrences of the Vestgötabreen Formation.

Bulltinden Formation as a schuppen. The upper member is lacking north of Skipperbreen and Bulltinden.

Short notes on the lithology of the rocks of the Vestgötabreen Formation are given below.

A distinct yellow dolomite follows along the basal thrust fault of the Vestgötabreen Formation. It is a dense, massive rock, partly gneissose and marked by fresh green chlorite and occurs as large lenses hundreds of metres long with a thickness of up to 150 m. The epidote-actinolite greenstone occurs in sharp concordant contact with the dolomite and limestone, and often shows strong schistosity. A large stock-like body is exposed on the southern face of the ridge between Vestgötabreen and Skipperbreen (Fig. 2, profile 7-8). In Motalafjella, this rock is well banded greenschist, having small garnet grains in some layers, and is interlayered with black phyllite. Small lenses of schistose serpentinite occur concordantly in the greenstone. No gradational change from the greenstones to the overlying glaucophane-bearing rocks has been observed and the border is a subordinate thrust fault.

The upper member of the Vestgötabreen Formation shows frequent alternation of muscovite-quartz schist, glaucophane schist and calcareous schist, and the last-named rock makes up nearly half the volume. The calcareous schist has thin layers of muscovite, chlorite, and garnet with a small amount of glaucophane in medium-grained banded marble. The muscovite-quartz schists show strong diaphtholitic cleavages undulated by large garnet porphyroblasts and scattered prisms of chloritoid. The compositional banding, several cm thick, is defined by the layers of different ratios of the constituent minerals. Glaucophane occurs in these rocks as idiomorphic prisms without linear arrangement on the cleavage surfaces. The glaucophane schist shows dark blue shiny cleavage and has many large garnet grains up to 3 cm across. The groundmass of the rock is often composed of pure glaucophane aggregate.

The eclogitic rocks occur as concordant lenses, from several dm to ten metres thick, in the muscovite-quartz schists and calcareous schists, and have idiomorphic glaucophane around the margins and cracks.

A thick grey limestone and slate occur above the Vestgötabreen Formation on the ridge north of Skipperbreen (Fig. 2, profile 5-6). These rocks can be correlated with the grey limestone and shale of the Bulltinden Formation.

III. Bulk chemical composition of the rocks

In order to evaluate the nature of the original rocks and to study metamorphic mineral parageneses, 30 rocks were analysed for major elements by the following methods: Si, Al, Ti, total Fe, Mn, Mg, Ca, and P by X-ray fluorescence analysis, FeO by titration, Na and K by the atomic absorption method, and gas+water by the penfield-tube ignition method.

The analysed rocks are listed in Table 1. Three analyses of HORSFIELD (1972) from the Vestgötabreen area and four metabasites of HJELLE (1962, and unpublished) from Nordenskiöldkysten are discussed together. The amphibolites from Hornsund (BIRKENMAJER and NAREBSKI 1960, and

Table 1
Chemical analyses of the rocks

	1	2	3	4	5	6	7	8
SiO ₂	48.49	48.80	49.66	46.77	47.00	47.31	47.91	47.75
TiO ₂	1.43	1.58	1.01	2.19	1.30	1.29	0.51	1.41
Al ₂ O ₃	15.66	14.30	13.51	14.27	14.78	15.84	13.41	11.93
Fe ₂ O ₃	3.30	8.20	4.39	5.95	9.23	3.50		6.86
FeO	9.81	4.02	8.69	10.74	3.74	11.28	14.64	6.47
MnO	0.25	0.14	0.18	0.15	0.15	0.21	0.23	0.28
MgO	10.54	9.71	11.30	9.02	10.58	9.48	7.73	8.02
CaO	2.22	3.07	2.11	2.11	1.99	2.75	5.86	10.31
K ₂ O	1.73	1.88	0.78	1.69	3.02	1.66	1.40	0.19
Na ₂ O	3.10	4.44	4.40	3.17	3.85	2.03	0.85	4.34
P ₂ O ₅	0.05	0.16	0.04	0.06	0.19	0.10	0.88	0.01
H ₂ O _±	3.60	3.82	3.60	3.39	3.62	4.77	5.64	2.71
Total	100.18	100.12	99.67	99.51	99.45	100.22	99.06	100.28

	9	10	11	12	13	14	15	16
SiO ₂	53.78	46.06	44.39	42.52	48.33	45.85	48.71	47.62
TiO ₂	0.81	1.79	1.84	2.38	1.02	1.80	1.52	1.36
Al ₂ O ₃	9.88	13.90	13.78	13.33	14.32	14.58	15.84	14.25
Fe ₂ O ₃	5.13	6.98	4.05	4.35		5.49	2.28	4.71
FeO	3.37	6.47	9.66	11.35	12.08	8.87	8.84	6.00
MnO	0.12	0.22	0.22	0.21	0.19	0.29	0.21	0.18
MgO	8.54	7.19	7.76	8.23	7.65	9.16	7.38	7.95
CaO	10.65	11.31	11.00	7.30	10.28	4.58	6.73	12.54
K ₂ O	0.06	0.76	0.78	0.01	0.40	1.41	1.50	0.23
Na ₂ O	6.75	1.70	1.92	2.23	2.50	2.08	3.37	1.89
P ₂ O ₅	0.03	0.14	0.15	0.21	0.01	0.16	0.18	0.13
H ₂ O _±	0.62	3.56	3.86	7.55	5.01	5.15	3.69	2.87
Total	99.74	100.08	99.41	99.72	101.79	99.42	100.25	99.73

	17	18	19	20	21	22	23	24
SiO ₂	47.95	47.43	51.32	46.98	53.89	41.42	46.91	43.80
TiO ₂	1.55	1.98	1.94	2.84	1.78	2.66	1.50	3.17
Al ₂ O ₃	15.26	11.51	15.22	15.51	16.95	7.28	14.46	13.17
Fe ₂ O ₃	2.13	5.19	1.94	2.88	2.16	2.13	3.25	0.92
FeO	9.05	7.69	6.79	6.97	6.18	12.04	9.05	9.50
MnO	0.20	0.26	0.14	0.13	0.15	0.19	0.19	0.23
MgO	6.34	7.33	6.22	6.98	3.13	18.84	8.08	8.78
CaO	8.58	9.85	6.13	9.17	3.37	8.93	9.54	9.89
K ₂ O	1.27	0.82	1.66	0.83	3.46	0.68	0.27	1.56
Na ₂ O	3.48	3.97	4.55	3.43	6.40	0.35	3.25	2.65
P ₂ O ₅	0.18	0.20	0.52	0.45	0.47	0.41	0.11	0.35
H ₂ O _±	3.78	3.29	3.28	3.42	2.18	5.22	3.55	6.17
Total	99.77	99.52	99.71	99.59	100.12	100.15	100.16	100.19

Table 1 cont.

	25	26	27	28	29	30	31	32
SiO ₂	38.47	46.72	50.50	56.01	51.19	67.94	71.79	49.25
TiO ₂	1.63	3.28	1.87	1.80	1.60	0.39	0.18	1.14
Al ₂ O ₃	6.54	11.98	14.16	13.23	21.21	12.13	13.32	24.15
Fe ₂ O ₃	4.53	1.85	1.52	3.30	4.11	1.89	1.51	16.80
FeO	7.60	12.46	10.88	12.43	7.51	2.80	1.19	
MnO	0.20	0.24	0.13	0.10	0.12	0.04	0.05	0.23
MgO	26.25	7.67	6.79	6.46	3.66	4.13	2.58	1.98
CaO	6.06	9.41	6.93	1.41	0.55	1.65	0.62	0.00
K ₂ O	0.57	0.05	0.02	1.85	4.34	4.29	2.87	0.70
Na ₂ O	0.24	2.74	4.35	0.40	2.08	1.76	3.79	2.60
P ₂ O ₅	0.15	0.22	0.10	0.15	0.23	0.05	0.02	0.05
H ₂ O _±	7.96	3.45	2.89	2.64	3.21	2.70	1.60	3.04
Total	100.20	100.07	100.14	99.78	99.81	99.77	99.52	99.94

	33	34	35	36	AK0232
SiO ₂	58.80	74.40	70.96	58.70	45.08
TiO ₂	0.82	0.60	0.28	0.96	3.23
Al ₂ O ₃	16.56	12.41	13.80	20.98	12.24
Fe ₂ O ₃	3.07	1.00	0.86	2.17	4.08
FeO	4.60	3.63	1.98	5.14	11.96
MnO	0.20	0.28	0.03	0.17	0.31
MgO	3.65	2.52	2.98	3.13	5.99
CaO	1.32	0.62	0.21	0.36	8.80
K ₂ O	1.70	1.36	3.30	2.86	1.13
Na ₂ O	5.95	1.05	3.12	0.77	2.65
P ₂ O ₅	0.08	0.13	0.04	0.19	1.06
H ₂ O _±	3.18	1.99	1.98	4.18	3.07
Total	99.93	99.99	99.54	99.61	99.60

Explanation to Table 1

Analysed samples (including four analyses by HJELLE 1962 and three by HORSFIELD 1972).

Glaucofane schists from the Vestgötabreen Formation (○):

1. Garnet-glaucofane schist with muscovite and chlorite.
2. Garnet-muscovite-glaucofane schist.
3. Banded garnet-muscovite-glaucofane schist.
4. Garnet-glaucofane schist.
5. Garnet-muscovite-glaucofane schist.
6. Garnet-muscovite-glaucofane schist.
7. Garnet-epidote-muscovite-glaucofane schist (Analyst: HORSFIELD (E1022)).

Eclogitic rocks from the Vestgötabreen Formation (⊙):

8. Schistose glaucofane-garnet-omphacite eclogite.
9. Glaucofane-garnet-omphacite eclogite with more than 90% omphacite.

Epidote amphibolites from the Vestgötabreen Formation (●):

10. Actinolite-epidote-sericite-chlorite-plagioclase-quartz schist.
11. Actinolite-epidote-chlorite-plagioclase schist with calcite veins.
12. Banded epidote-actinolite-chlorite schist.
13. Epidote-calcite-chlorite-actinolite-plagioclase greenstone (Analyst: HORSFIELD (B3023)).

Garnet-bearing epidote amphibolite from the Vestgötabreen Formation (x):

14. Garnet-epidote-actinolite-chlorite meta-diabase.
15. Garnet-epidote-sericite-actinolite schist.
16. Dense actinolite-epidote-chlorite-sericite-plagioclase schist.
17. Garnet-actinolite-epidote-chlorite-sericite schist.
18. Garnet-chlorite-sericite-plagioclase-quartz meta-gabbro.

Meta-basites from Holmsletfjella, south of St. Jonsfjorden (●):

19. Chlorite-plagioclase meta-diabase.
20. Hornblende-actinolite-epidote-chlorite-plagioclase meta-hornblende gabbro.

Meta-basites from the northern side of St. Jonsfjorden (●):

21. Coarse-grained meta-hornblende gabbro.
22. Hornblende porphyrite.
23. Meta-diabase.

Meta-basites from the Bellsund area, all by HJELLE (●):

24. Meta-basite. Recherchebreen, S Bellsund, unpublished.
25. Ultrabasic rock. Recherchebreen, unpublished.
26. Amphibolite. Diabasodden, N Bellsund (HJELLE 1962, Table 3.3).
27. Amphibolite. 4 km SE of Orustosen, N Bellsund (HJELLE 1962, Table 3.3).

Glaucofane-muscovite-quartz schists from the Vestgötabreen Formation (▲):

28. Glaucofane-garnet-chlorite-muscovite-quartz schist.
29. Glaucofane-garnet-chlorite-muscovite-quartz schist.
30. Glaucofane-garnet-muscovite-quartz feather schist.
31. Glaucofane-hematite-muscovite-quartz schist.

Muscovite-quartz schists from the Vestgötabreen Formation (▲):

32. Garnet-chloritoid-muscovite-quartz schist. (Analyst: HORSFIELD (E1020))
33. Cataclastic chlorite-plagioclase-quartz schist.
34. Chloritoid-garnet-muscovite-quartz schist.
35. Chloritoid-muscovite-quartz schist.
36. Chloritoid-chlorite-muscovite-quartz schist.

A meta-gabbro from northern Hornsund, collected by ORVIN (AK0232), has been analysed and is included in the data of the Hornsund amphibolites in the figures.

Symbols in parentheses refer to Figs. 3a, 4a, 5, 6, 7, 8, 16, 17 and 18.

SMULIKOWSKI 1968) are summarized separately and compared with the present data. Hereafter, the epidote-actinolite greenstones and metabasites will be called the epidote amphibolites.

In the AFM diagram (Fig. 3a), the glaucofane schists cannot be distinguished from the other amphibolites. Most of the rocks project in the field of calc-alkaline and alkali basalt. Four plots are slightly off from the border curve of this field, but the border curve itself is rather arbitrary in this part of the

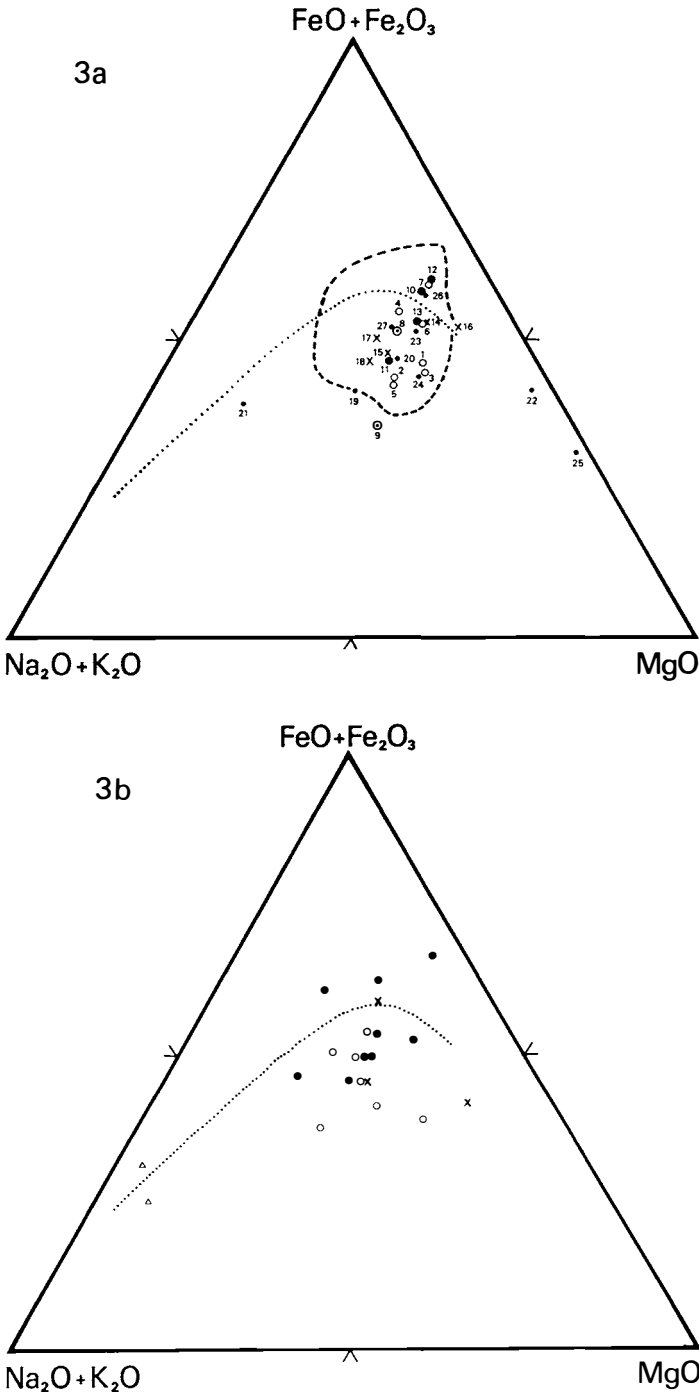


Fig. 3. Bulk rock compositions on the MgO—FeO+Fe₂O₃—Na₂O+K₂O diagram. (The symbols and suffixed numbers refer to Table 1.)

a. The rocks of the Vestgötabreen Formation, with some greenrocks from St. Jonsfjorden and the Bellsund area. Dotted curve: border of the tholeiitic and alkalic rock series, Hawaii (KUNO et al. 1957); broken curve: glaucophane schists fo California (COLEMAN and LEE 1963).

b. The rocks of the Hornsund area (SMULIKOWSKI 1968). Dots: fine-grained amphibolites; open circles: coarse-grained amphibolites; crosses: schistose amphibolites; triangles: acidic volcanic rocks; dotted curve: same as in Fig. 3a.

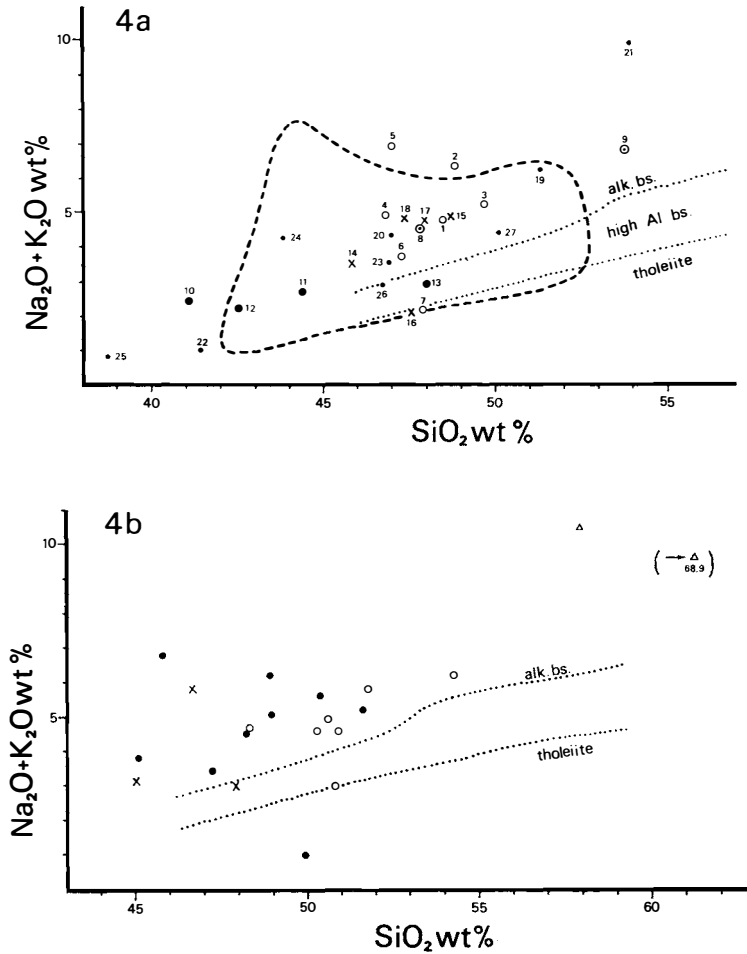


Fig. 4. Alkali- SiO_2 diagram.

- a. The rocks of the Vestgötabreen Formation with some amphibolites from western Spitsbergen, excluding those from Hornsund. The symbols are the same as in Fig. 3a, refer to Table 1.
- b. The rocks from Hornsund. The symbols are the same as in Fig. 3b.

diagram (KUNO 1957). In the SiO_2 -alkalis diagram (Fig. 4a), there are no points in the field of tholeiitic basalt. The basic rocks of the Hornsund area show similar characteristics in the same diagrams (Figs. 3b and 4b). No definite calc-alkaline rock has been found in the Vestgötabreen Formation.

The relatively high alkali contents in the glaucophane schists compared with the epidote amphibolites are due not to Na_2O but to K_2O , as seen in the K_2O - Na_2O diagram (Fig. 5), in which the following zones can be distinguished with increasing K_2O : epidote amphibolites (excluding Nos. 19 and 21) — garnet amphibolites — glaucophane schists — muscovite-quartz schists. Two epidote amphibolites, Nos. 19 and 21, are rich in SiO_2 and have higher Na_2O than the others, somewhat like keratophyre. All other rocks have less than 4.5 wt.% Na_2O , accordingly they are not typically spilitic in com-

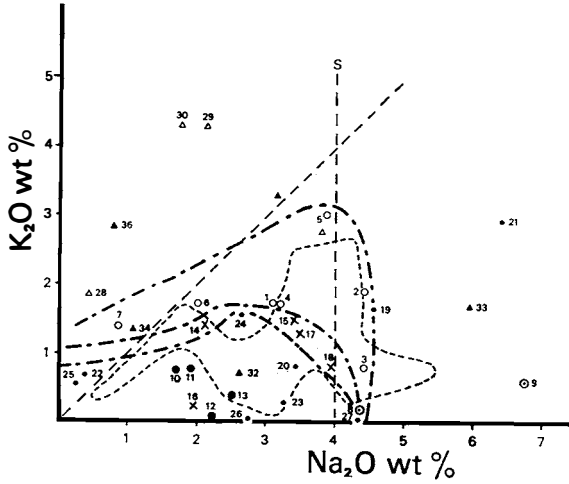


Fig. 5. K_2O — Na_2O diagram. Thin broken curve: the field of the rocks from Hornsund; thick dash and point curve: see text. The symbols and numbers are the same as in Fig. 3a, refer to Table 1.

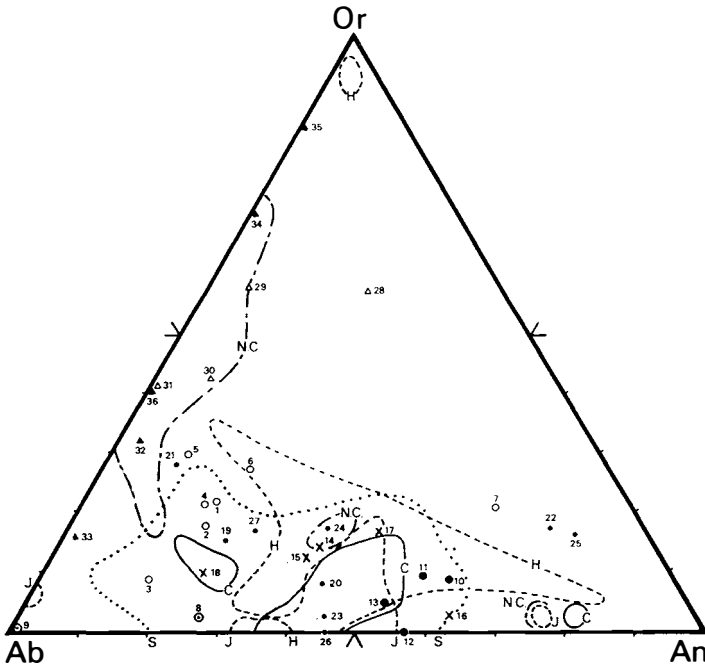


Fig. 6. Normative Or-Ab-An diagram of the Vestgötabreen Formation rocks. C: Californian glaucophane schists; H: Hornsund rocks; J: Japanese glaucophane schists; NC: New Caledonian glaucophane schists; S: Spilites (YODER 1967).

The symbols and numbers are the same as in Fig. 3a, refer to Table 1.

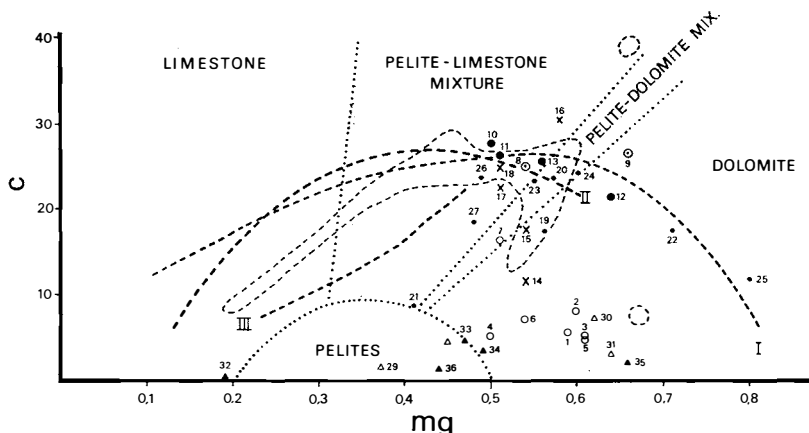


Fig. 7. NIGGLI's *mg-c* diagram for estimating the nature of original rock types, after LEAKE (1963). I: differentiation trend of Karroo dolerite; II: differentiation trend of Na-rich alkali rock series of Etiopia (MOHR 1960); III: DALY's average calc alkaline rocks, thin broken curve: field of the rocks from Hornsund.

The symbols and numbers are the same as in Fig. 3a, refer to Table 1.

position (YODER 1967 and AMSTUTZ 1974). The glaucophane schists project in the field between the epidote amphibolites and the muscovite-quartz schists in this diagram and have relatively high SiO_2 contents, as shown in Fig. 4a. This suggests that the glaucophane schists could be a mixture of basic volcanic rocks and argillaceous quartzitic sediments.

On the Or-Ab-An diagram (Fig. 6), the glaucophane schists, except for No. 7 (HORSFIELD's analysis), project in the field between the epidote amphibolites and the muscovite-quartz schists field. This supports the idea mentioned above. Two keratophyre-like epidote amphibolites, Nos. 19 and 21, plot in the same field as the glaucophane schists; thus they might also be of similar origin.

Figs. 7 and 8 are prepared after LEAKE (1963) and SIMONEN (1953), respectively. Fig. 7 suggests that most epidote amphibolites are originally intermediate differentiates of basaltic magma, and the muscovite-quartz schists without glaucophane are impure argillaceous quartzite. Most glaucophane schists and some muscovite-quartz schists can be considered as mixtures of argillaceous quartzite and early differentiates of basaltic magma. The epidote amphibolites with keratophyre-like composition are suggested to be a mixture of argillaceous quartzite and intermediate differentiates of basaltic magma.

In Fig. 8, some early differentiates of basaltic magma are plotted in the upper left field and this means that the field of volcanogenous sediments extends further up to the left in this diagram. The glaucophane schists project in the field between argillo-siliceous sediments and the early differentiates of basaltic magma in this diagram. This conforms with the conclusions drawn from Figs. 5, 6, and 7.

Excluding probable mixed rocks, classification in terms of bulk chemical composition, was tried on selected basic rocks of basaltic composition with

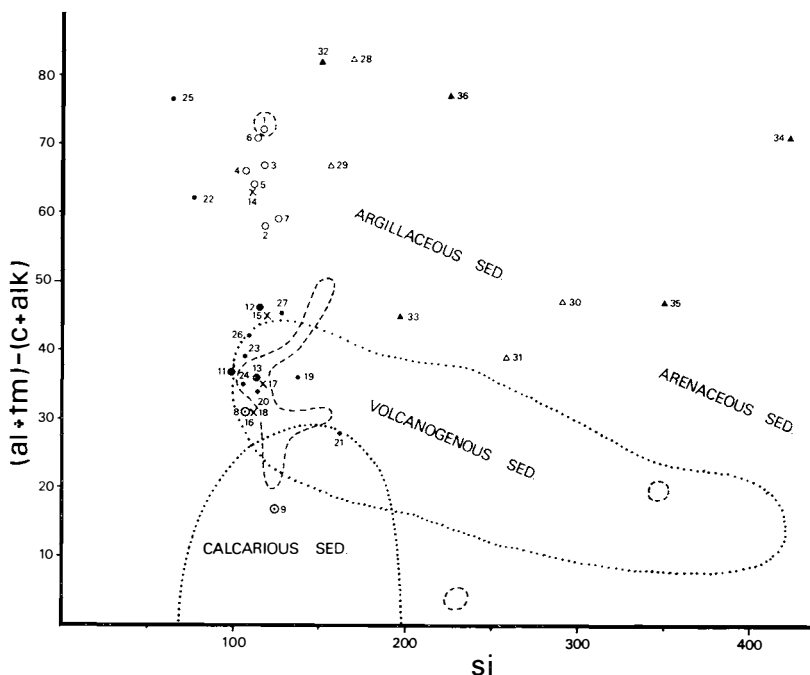


Fig. 8. NIGGLI's $si - (al+fm) - (c+alk)$ diagram to estimate the original rocks, after SIMONEN (1953).
Thin broken curve: field for the rocks of Hornsund.
The symbols and numbers are the same as in Fig. 3a, refer to Table 1.

44% < SiO₂ < 53.5%, 10 from the present analyses and 17 from Hornsund. Almost all rocks were totally recrystallized and mineralogical and textural information on original rocks has not been preserved at all. Table 2 shows the classification based on the existence of the norm hypersthene and olivine, referring the weight percentages of Al₂O₃ and TiO₂, and was prepared from Figs. 9a, 9b, 9c, and 9d, after the classification suggested by MIDDLEMOST (1975). The sub-alkalic rocks appear as tholeiitic rocks by this classification.

These tables show that the rocks do not show any dominant rock type and have a large variation in their chemical characteristics of major elements, from tholeiite to alkali olivine basalt. The most probable reason for this variation is the fact that the rocks still contain some mixture of terrestrial sedimentary material and modifications of metamorphic differentiation.

Fig. 9. Classification of the selected basic rocks after MIDDLEMOST (1975), recalculated without H₂O.

a. The classification of alkalic-subalkalic-low K₂O alkalic rock series.

b. Division of the subalkalic and transitional rocks.

c. and d. Divisions of the alkalic and transitional rocks. Open circles: basic rocks from the Vestgötabreen Formation, north of St. Jonsfjorden and the Bellsund area, suffixed numbers refer to Table 1. Other symbols are the rocks from the Hornsund area and the suffixed numbers refer to SMULIKOWSKI 1968; dots: fine- to medium-grained amphibolites; triangles: coarse-grained amphibolites; crosses: schistose amphibolites. Circled numbers: alkalic rocks; squared numbers: subalkalic rocks; underlined numbers: low K₂O transitional rocks; overlined numbers: low K₂O subalkalic rocks; other numbers: transitional rocks. The alkaline and transitional rocks are plotted together in Fig. 9c, and the subalkaline and transitional rocks are plotted together in Fig. 9d.

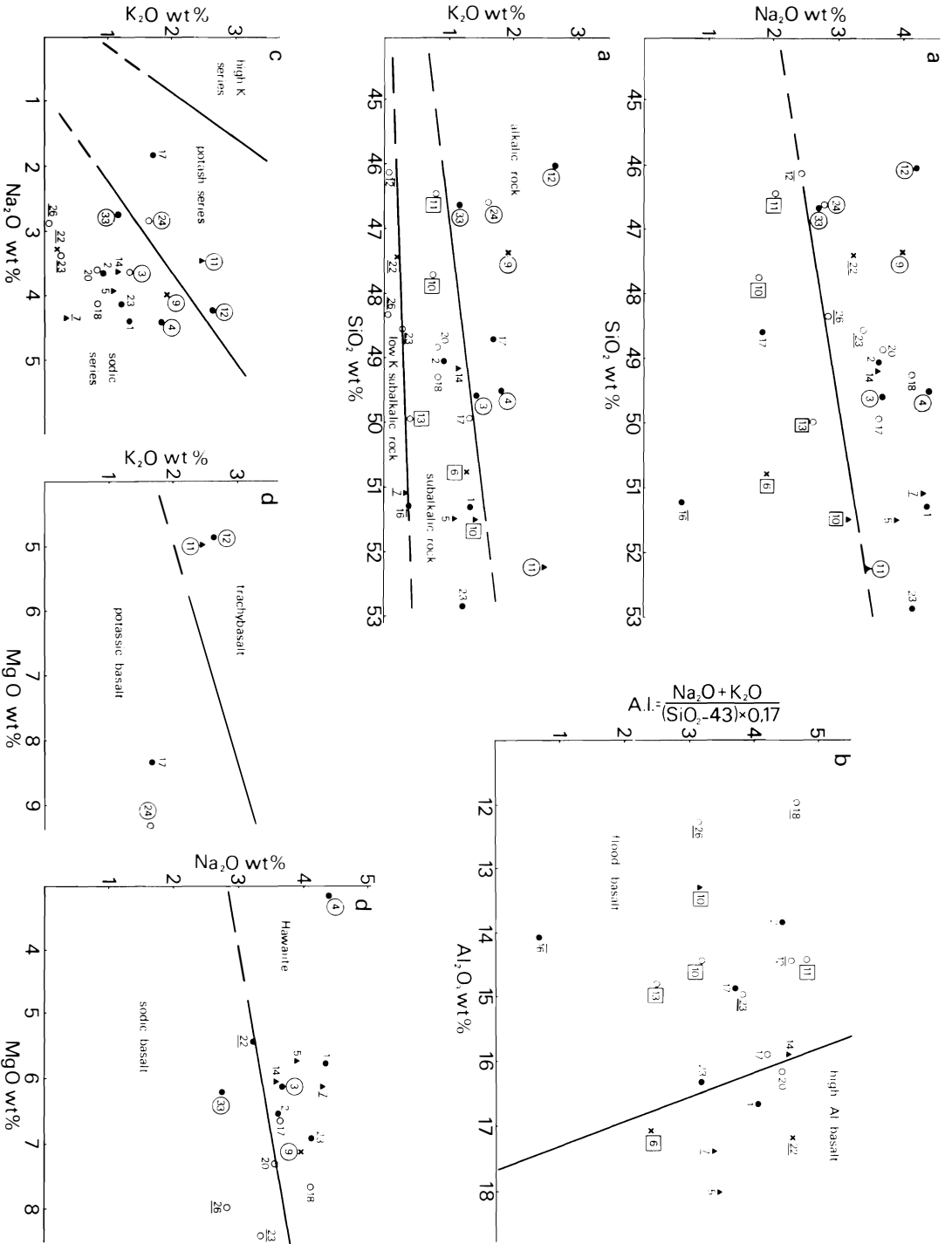


Fig. 9.

Table 2
Classification of the selected basic rocks.

Locality	Alkalic basalt			Subalkalic basalt			Transitional basalt			Total
	High-K basalt	K-alkalic basalt	Na-alkalic basalt	Low K-subalk. basalt	Flood basalt	High-Al basalt	Low-K transitional basalt	K-rich transitional basalt	Na-rich transitional basalt	
St. Jonsfj. etc.	0	1	0	1	3	0	2	0	3	10
Hornsund	0	2	4	1	1	1	2	1	5	17
Total	0	3	4	2	4	1	4	1	8	27

Locality	Alkalic and transitional basalt				Subalkalic and transitional basalt			
	High-K series	Potash series		Sodic series		Flood basalt	High-Al basalt	
		Trachybasalt	Potassic basalt	Hawaiite	Sodic basalt			
St. Jonsfj. etc.	0	(1)	0	1	2	(5)	8	1
Hornsund	0	(3)	2	1	9	(11)	6	5
Total	0	2	2	2	11	5	14	6

When terrestrial sediments are mixed, rocks tend to be enriched by SiO_2 and K_2O ; therefore, some rocks classified as belonging to the potash-alkaline group are suspicious. The coarse-grained rocks from Hornsund might be modified by the granitization (BIRKENMAJER and NAREBSKI 1960; SMULIKOWSKI 1968). The second reason may be found in their occurrence; some of them should not be included in the basaltic rock group. For example, the low potash tholeiite of No. 26 actually occurs in shallow sea sediments as an intrusive vent and may be considered a lamprophyre; accordingly this is not representative of ocean floor basalts. The third possibility implies that these rocks may be from different stratigraphic horizons of the Hecla Hoek succession of western Spitsbergen. This is apparent from the literature (BIRKENMAJER 1975; HJELLE 1962). Thus, more systematic sampling to avoid mixture of foreign material and different stratigraphic horizons and the studies of minor elements are important in the future study of basic rocks in western Spitsbergen.

Summary of the original nature of the Vestgötabreen Formation

1. Most epidote amphibolites were derived from intermediate differentiates of basaltic magma.
2. Some relatively acidic varieties are not later differentiates of basaltic magma, but mixtures of intermediate differentiates and argillo-siliceous sediments.
3. The muscovite-quartz schists with or without glaucophane were formed from impure argillaceous quartzite.
4. The glaucophane schists are mixtures of early differentiates of the basaltic magma and argillo-siliceous sediments.
5. Na enrichment is not prominent in the glaucophane schists and the original rocks were not typically spilitic.
6. The original volcanic rocks of these basic rocks are unknown, and they have large variation presumably by tectonic stirring.

Essential association of volcanic rocks in the Vestgötabreen Formation is alkaline and possibly tholeiitic rocks. This fits well in with what MIYASHIRO (1975) pointed out from the Sanbagawa and Franciscan blue schist zone. The tectonic setting of this volcanic association is problematic.

IV. Mineral chemistry

The main constituent minerals of the epidote amphibolites and glaucophane schists were analysed by an electron probe microanalyser by the courtesy of Dr. M. KOMATSU of Niigata University, Japan, and Prof. H. RAMBERG, Uppsala University, Sweden (Table 3). Natural and synthetic minerals were used as standards and each analysis is the average of five measurements. Physical properties of these minerals have not been fully examined yet, and their chemical nature will be presented here with brief comments on their occurrences.

Table 3

Chemical composition of the constituent minerals (analyses Nos. 1 and 2 of garnet and Nos. 4 and 5 of alkali amphibole by H. RAMBERG, Uppsala Univ.; all others by M. KOMATSU, Niigata Univ. Japan).

<i>Garnets</i>					
	1	2	3	4	5
SiO ₂	38.15	37.0	37.93	38.91	38.30
TiO ₂	0.8	0.2	0.6	0.2	—
Al ₂ O ₃	20.75	20.8	20.88	21.70	21.85
FeO	30.64	28.1	32.38	25.5	28.12
MnO	2.49	0.7	3.88	0.9	0.88
MgO	2.67	4.8	2.86	7.2	5.52
CaO	6.41	6.7	3.16	6.1	5.59
Total	101.19	98.3	101.14	100.51	101.26

<i>Calcic and sub-alkalic amphiboles</i>					
	1	2	3	4	5
SiO ₂	54.23	53.20	48.94	50.05	37.84
TiO ₂	0.02	0.02	—	—	—
Al ₂ O ₃	1.16	1.17	6.20	7.60	15.49
FeO	13.38	15.60	21.08	16.40	27.96
MnO	0.07	0.20	—	—	—
MgO	15.93	13.13	12.67	13.08	3.58
CaO	11.58	11.31	8.90	8.63	11.54
K ₂ O	0.03	0.07	—	—	—
Na ₂ O	1.14	1.25	1.24	2.98	2.01
Total	97.54	96.54	99.03	98.99	99.74

<i>Alkali amphiboles</i>							
	1	2	3	4	5	6	7
SiO ₂	54.69	55.9	55.62	55.72	55.1	56.73	56.71
TiO ₂	—	0.1	—	0.3	0.2	—	0.04
Al ₂ O ₃	9.88	8.8	7.19	7.68	8.4	10.42	9.97
FeO	15.54	15.1	18.57	14.52	15.9	15.20	13.00
MnO	—	0.1	—	0.03	0.1	—	0.08
MgO	8.45	9.5	7.81	10.42	8.8	8.25	9.94
CaO	1.39	1.6	0.54	0.46	1.1	0.35	0.39
K ₂ O	0.05	0.00	0.00	—	0.00	0.08	0.01
Na ₂ O	6.31	5.5	7.03	7.15	5.8	6.28	7.03
Total	96.58	96.6	96.76	95.99	95.4	97.30	97.16

(cont. next page.)

(Table 3 cont.)

<i>Chloritoids</i>			<i>White micas</i>			
	1	2		1	2	3
SiO ₂	24.27	25.25	SiO ₂	49.27	49.34	50.41
TiO ₂	0.15	0.16	Al ₂ O ₃	26.90	27.62	25.58
Al ₂ O ₃	39.58	40.78	FeO	3.48	3.30	4.61
MnO	0.12	0.53	MgO	2.75	3.01	3.36
MgO	2.71	4.72	K ₂ O	10.05	9.60	10.16
CaO	0.08	0.08	Na ₂ O	0.53	1.02	0.65
K ₂ O	0.05	0.04	Total	92.97	93.89	94.77
Na ₂ O	0.07	0.05				
Total	93.90	94.28				

<i>Epidotes</i>			<i>Chlorites</i>		<i>Clinopyroxene</i>		
	1	2		1	2	1	
SiO ₂	37.71	38.05	SiO ₂	28.55	25.84	SiO ₂	56.20
Al ₂ O ₃	23.97	24.47	Al ₂ O ₃	18.63	20.49	TiO ₂	0.09
FeO	10.76	10.91	FeO	25.84	30.34	Al ₂ O ₃	8.98
MnO	0.33	0.29	MgO	17.62	12.70	FeO	6.05
CaO	22.64	22.45	Total	90.64	89.37	MnO	0.01
Total	95.42	96.16				MgO	8.48
						CaO	11.56
						Na ₂ O	7.38
						Total	98.75

Garnet

Garnet of the epidote amphibolites occurs as small scattered grains in certain compositional layers, while those of the glaucophane schists and muscovite-quartz schists are remarkably large idioblastic crystals up to 3 cm across. Such large garnets are unique among the glaucophane schists from the world. Most garnet grains in the present glaucophane-bearing rocks are weakly fractured and the cracks are filled with chlorite (Plate 2-4). The core parts of the garnets are often homogeneous with a few inclusions of chloritoid and opaque minerals (Plate 2-3). The margins of the garnets are always poikiloblastic and skeletal, including many elongated quartz and glaucophane grains. Chloritoid makes a thin rim around those garnet grains which are almost free from inclusions.

The garnet of eclogitic rocks is scattered in the dense omphacite groundmass and has always many cracks filled with chlorite.

Five garnets from the glaucophane-bearing rocks were analysed and the results were calculated as shown in Table 4 and Figs. 10a and 10b. Their pyralspite components are more than 80%. The first component is always almandine, followed by pyrope, and the third is grossularite, except for No. 1 the second component of which is grossularite and the third pyrope. Comparing with TRÖGER's statistical work on garnet (TRÖGER 1959), the present

Table 4

Molecular contents of the garnet

	1	2	3	4	5
Host rock	Glaucophane schist	Mus-qt schist	Mus-qt schist	Eclogitic rock	Eclogitic rock
Pyrope	10.50	18.93	11.30	27.48	21.04
Almandine	65.83	60.55	71.04	53.77	58.99
Spessartine	5.57	1.58	8.70	1.99	1.91
Grossularite	16.53	16.82	8.13	15.52	16.95
Andradite	1.57	2.11	0.83	1.23	1.12
Pyralspite	81.90	81.07	91.04	83.25	81.93
Ugandite	18.10	18.93	8.96	16.75	18.04
Bulk rock analysis No. (Table 1)	28	30	34	8	9

garnets have compositions similar to those from charnockites (Nos. 3, 4, and 5), pelitic metamorphics (No. 2) and mica schists (No. 1).

Cation percentages in the garnets relative to those of the host rock are shown in Table 5. Samples 1, 2, and 3 show higher CaO and MnO concentrations than the others and the values are comparable to those of the type III glaucophane schists of California (LEE et al. 1963). The high pyrope and low spessartine contents of all present garnets are characteristic for the garnets from the Type IV rocks of California. In general, the present garnets are chemically similar to those of the type IV glaucophane schists of California, and project in the composition field between the garnets from the type III glaucophane schists of California and true eclogites in Figs. 10a and 10b.

Calcic and sub-alkaline amphiboles

A progressive development of colourless actinolite from chlorite is well observed in the epidote amphibolites (Plate 2-2). Nematoblastic actinolite needles occur at the crests of tight crenulations in chlorite and occasionally show distinct bluish green pleochroism. Similar, but more bluish amphibole was found in an amphibolite from Nordenskiöldkysten (its bulk analysis is No. 26, Table 1), collected by A. HJELLE. Another type of deep bluish green amphibole has been found in the meta-gabbroic rocks from the northern side of St. Jonsfjorden (bulk analyses are Nos. 21 and 23, Table 1) and in some amphibolites from the northern Hornsund area, collected by ORVIN (No. 37, Table 1). The rocks of the latter locality were studied by SMULIKOWSKI (1968) by optic means and the amphibole was found to be hastingsite. These deep bluish green amphiboles show a different mode of occurrence from the nematoblastic actinolite and occur as rims or irregular patches in large relict brown-green hornblendes which are certainly of primary igneous origin.

Four nematoblastic amphiboles (Nos. 1 to 4) from the epidote amphibolites of the Vestgötabreen Formation, and one deep bluish green amphibole (No. 5)

Table 5
Cation percentages in the garnets relative to those of the host rock.

	California III 27-CZ-59 59-CZ-59 58RGC-58C			California IV 50-CZ-60 62-GC-58		Zernatt PUB878 PUB891 PUB941		Western Spitsbergen 1 2 3 4 5					Knocknoral A B		
	Rock MgO G/R	0.43 0.70 1.63	0.16 0.08 0.50	3.10 0.44 0.14	5.15 2.10 0.41	5.40 1.70 0.32	4.5 3.3 0.73	4.6 4.68 1.00	3.5 5.54 1.57	6.46 2.67 0.41	4.13 4.7 1.14	2.52 2.86 1.14	8.02 7.2 0.90	8.54 5.52 0.65	6.45 — —
Rock FeO G/R	2.00 15.60 7.80	3.50 13.8 3.94	10.80 24.6 2.28	6.64 25.3 3.81	9.10 27.6 3.03	6.8 26.8 3.94	5.9 25.2 4.27	4.9 26.1 5.33	12.43 29.92 2.41	2.80 28.08 10.03	3.63 32.01 8.82	6.47 24.64 3.81	3.37 27.60 8.20	10.12 21.00 2.08	7.04 13.45 1.91
Rock CaO G/R	3.2 7.1 2.22	0.22 6.3 28.64	2.1 6.4 3.05	16.14 11.1 0.69	11.0 9.5 0.86	6.4 8.2 1.28	9.1 7.6 0.84	8.3 7.2 0.87	1.41 6.41 4.55	1.65 6.70 4.06	0.62 3.16 5.10	10.31 6.10 0.59	10.65 6.59 0.62	8.61 — —	13.01 9.11 0.70
Rock MnO G/R	2.3 17.6 7.65	0.65 21.3 32.77	2.8 10.8 3.86	0.33 1.6 4.85	0.23 1.3 5.65	0.2 1.6 8.00	0.2 0.6 3.00	0.2 1.0 5.00	0.10 2.49 24.9	0.04 0.70 17.5	0.28 3.88 13.86	0.28 0.90 3.21	0.12 0.88 7.33	0.25 4.04 16.16	0.18 0.64 3.56
	LEE, COLEMAN and ERD (1963)						BEARTH (1973)						BLOXAM and ALLEN (1959)		
	present paper														

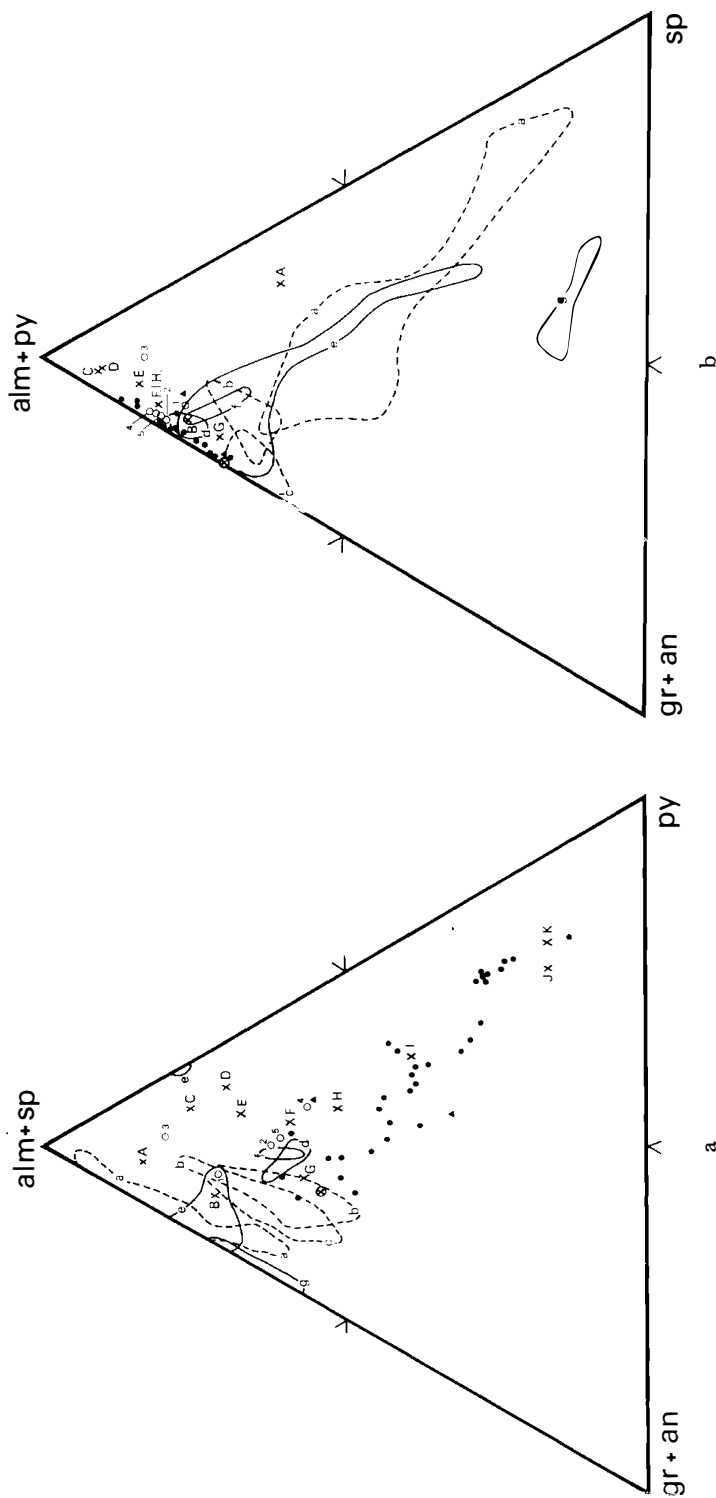


Fig. 10. Garnets from the glaucophane-bearing rocks.

a. *Pyrope-Almandine + Spessartine-Andradite + Grossularite* diagram. Open circles 1-5; garnets from the rocks of this area, see Table 3; open circle with cross inside: garnet of the eclogite, from Biskayerhuken, NW Spitsbergen; solid triangles: garnet from the glaucophane schists of Knockermal, Scotland; dots: garnets from eclogites of the world; crosses: TROGER's averages of garnets from pelitic metamorphic rocks (average of 31); B: from mica schists (3); C: mica schists and hornfels (33); D: paragneisses (19); E: granulites and pegmatites (10); F: charnockites (15); G: amphibolites (21); H: meta-gabbros (8); I: eclogites (14); J: kimberlites; K: ultrabases. Broken curves: a: garnets from the Type III glaucophane schists of California; b: from the Type IV glaucophane schists of California; c: from the eclogites in the glaucophane schists of Zermatt; e: from the glaucophane schists of Japan; f: from New Caledonia; g: from the green schists of New Zealand.

b. *Almandine + Pyrope-Almandine-Andradite + Grossularite* diagram. The symbols are the same as in Fig. 10a.

Table 6
Calcic and sub-alkaline amphiboles

	Al ^{IV}	Na+K	100 Mg	100 (Na+K)	Chemical formula
			Mg+total Fe+Mn	Ca+Na+K	
1 Na actinolite	0.159	0.327	67.7	15.4	(NaK) _{0.327} (Fe''Mn) _{1.627} Mg _{3.433} (TiAl) _{0.141} (AlSi) ₈ O ₂₂
2 Na actinolite	0.143	0.369	59.7	17.1	(NaK) _{0.369} Ca _{1.790} (Fe''Mn) _{1.953} Mg _{2.891} (TiAl) _{0.168} (AlSi) ₈ O ₂₂
3 Common hornblende	0.799	0.355	51.0	20.2	(NaK) _{0.355} Ca _{1.405} Fe'' _{2.598} Mg _{2.783} Al _{0.277} (AlSi) ₈ O ₂₂
4 Edenite	0.762	0.884	58.0	37.8	(NaK) _{0.884} Ca _{1.337} Fe'' _{1.983} Mg _{2.819} Al _{0.533} (AlSi) ₈ O ₂₂
5 Ferro hastingsite	2.076	0.876	15.1	31.2	(NaK) _{0.876} Ca _{1.935} Fe'' _{3.442} Mg _{0.834} (Fe''Al) _{1.0} (AlSi) ₈ O ₂₂

1, 2 and 3 are from the same rock: No. 11 of Table 1. Total FeO = FeO, except for No. 5.

from a meta-gabbro of the Hornsund area (from ORVIN's collection), have been analysed (Table 3), and their chemical formulae and classification are shown in Table 6 and Fig. 11.

Two nematoblastic bluish green amphiboles are Na-actinolite and the other two are common hornblende and edenite. They seem to have a composition range from common hornblende to sub-alkaline amphibole, with a little amount of tschermakite component. These amphibole compositions are common in green schists from the blue schist metamorphic zones of the world.

The deep bluish green amphibole from Hornsund (No. 5) is pargasite or ferro-hastingsite and is a typical sub-alkaline variety. This agrees with the optic studies of SMULIKOWSKI (1968), and the mode of occurrence of these amphiboles indicates a type of metasomatic change associated with the introduction of quartz and biotite into various kinds of gabbroic rocks. Thus, the metamorphism of the Hornsund amphibolites can be distinguished from the amphibolites of the Vestgötabreen Formation.

Alkali amphiboles

Blue amphiboles show various modes of occurrence; the dense granoblastic groundmass of the glaucophane schists, scattered large idiomorphs (Plates 3-1 and 3-2) and vein-like monomineralic mass in the eclogitic rocks, and thin long crystals showing feather-amphibolite texture in the muscovite-quartz schists (Plates 1-1 and 1-2). The idiomorphic and vein-like varieties may have crystals up to 4-5 cm long. Typical glaucophane schist is composed of more than 90% blue amphibole, with subordinate garnet, epidote, muscovite and some secondary chlorite. Idioblastic blue amphiboles are often poikiloblastic

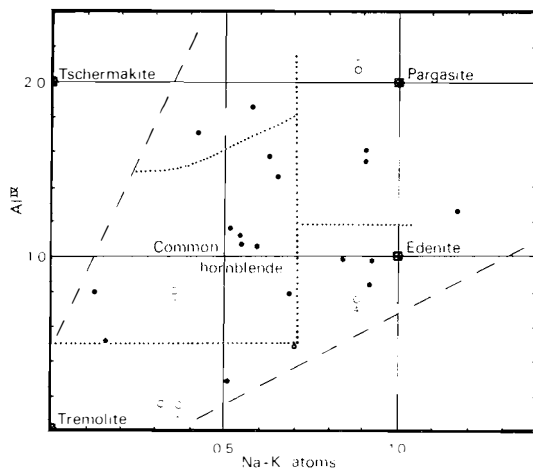


Fig. 11. *Calcic and sub-alkaline amphiboles. Open circles: amphiboles from the present area, see Table 4; dots: amphiboles from Japanese glaucophane schists.*

with inclusions of quartz in the muscovite-quartz schists and of clinopyroxene in the eclogitic rocks (Plates 3-2 and 3-4). They often show distinct zonal structure with a zone of strong pleochroism around the margins (Plates 3-1 and 3-2) and sometimes along cracks (Plate 3-3). The blue amphibole in the vein-like mass occurs as long prismatic crystals, perpendicular to the wall of the vein, and sometimes represents remarkable radial aggregates, and the core of vein is often occupied by quartz. Some blue amphibole grains are included in the marginal poikiloblastic parts of the garnet.

Table 7
Alkali amphiboles

	Al ^{IV}	$\frac{100\text{Fe}''}{\text{Fe}'' + \text{Al}^{\text{VI}} + \text{Ti}}$	$\frac{100\text{Fe}'' + \text{Mn}}{\text{Fe}'' + \text{Mg} + \text{Mn}}$	$\frac{100(\text{Na} + \text{K})}{\text{Ca} + \text{Na} + \text{K}}$	Host rock (Nos refer to Table 1)
1 Glaucophane	0.130	23.1	43.7	89.2	fine grained, banded ga-gl-mus-chl schist
2 Glaucophane	—	27.0	42.3	89.4	gl-ga-mus-qt schist (No. 30)
3 Crossite	—	38.5	46.8	75.8	ga-gl schist with mus crots (No. 2)
4 Crossite	—	35.0	31.8	96.6	mus-ga-gl schist (No. 28)
5 Glaucophane	0.001	26.1	38.7	84.7	gl-ga-cpx eclogite (No. 8)
6 Glaucophane	0.014	14.3	46.5	97.1	gl-ga-cpx eclogite
7 Glaucophane	0.004	20.2	34.9	97.0	gl-ga-cpx eclogite (No. 9)

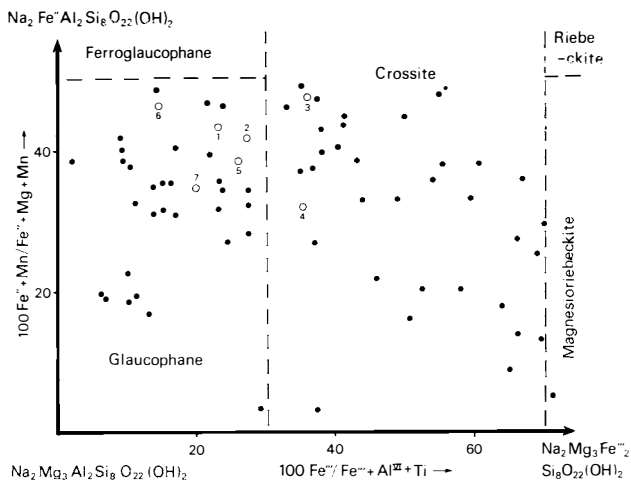


Fig. 12. Alkaline amphiboles. Open circles: from the present area, see Table 5; dots: glaucophane and crossite from the world (DEER et al 1962, KOCTROK 1970).

Seven blue amphiboles were analysed (Table 7, Fig. 12). Five are glaucophane and two are crossite. No systematic difference of composition has been found among those from different host rocks. The crossite makes up the deep pleochroic marginal zone and narrow zones along cracks of the idiomorphic glaucophane crystals. The $Fe^{III}/Fe^{III} + Al^{VI} + Ti$ ratios vary in the range of 14–38% from glaucophane to crossite, and the transitions are gradational in most zoned crystals, but some show sharp borders (Plate 3–2).

The radial occurrence of glaucophane suggests they might be formed along open cracks in the host eclogitic rock. The idioblastic texture of this mineral is suggestive of an origin later than the main recrystallization phase of the host rock and some segregation or circulation of materials may be expected. For example, addition of Fe and subtraction of Ca from omphacite to make glaucophane in the eclogitic rocks, and concentration of Fe, Mg and Na and subtraction of Al and K in the muscovite-quartz schists might have occurred even though there is no positive evidence to suggest Na-metasomatism in the study of the bulk chemical compositions.

Chloritoid

This mineral occurs in the muscovite-quartz schist as prismatic crystals up to 3.5 cm long. Most chloritoid grains are in the quartz-rich groundmass and some are concordantly enclosed by undulated muscovite. Chloritoid is a dominant inclusion in the core parts of large garnets. All chloritoid grains show polysynthetic twin lamellae and abnormal birefringence colours. They show no secondary alteration.

Two chemical analyses of chloritoid are shown in Fig. 13: (1). from garnet-muscovite-quartz schist, and (2). from inclusion in the garnet of typical glaucophane schist. The data have been given on the anhydrous base of $O = 12$ and total iron as FeO; therefore, the projections may move towards the $Al + Fe^{III} + Ti$ direction on the figure when Fe^{III} is detected.

The chloritoid (2) included in garnet has less Mg and projects in the composition field similar to those from the rocks of green schist facies, i.e., the Mn-rich low grade schists from the NW Rheine schist area (KRAMM 1973), but the latter have a large amount of Mn instead of Fe^{''}. The chloritoid (1) from the garnet-muscovite-quartz schist has a little higher Mg than (2), but still less Mg than those from the glaucophane eclogite of Zermatt (BEARTH 1973).

White micas

White mica is a main constituent mineral of the muscovite-quartz schists, up to about 25% in volume, and is always found in all other schistose rocks of the Vestgötabreen Formation. It occurs as large flakes and defines the strong diaphlotitic cleavage. The flakes show no trace of alteration and have no inclusions of other minerals.

Three chemical analyses of white mica show that all of them are phengite with less than 44% trioctahedral Fe^{''} (Table 8). They project in the same field as the phengites from other glaucophane schists of the world (Fig. 14). The glaucophane schists from Bessi, Japan (BANNO 1964) and the glaucophane eclogites from Zermatt (BEARTH 1973) have white micas of paragonite composition, however, the paragonite contents of the present ones are less than 14%.

Epidotes

Epidote is granular and is abundant in the epidote amphibolites. It is also common in the glaucophane schists, eclogitic rocks, calcareous schists and muscovite-quartz schists, and is in secondary appearance to some extent, dusty granular alignments along schistosity. Clinozoisite is often associated with epidote.

Two epidotes, from the epidote amphibolite (1) and typical glaucophane schist (2), have very small piedmontite contents (2.58% and 2.99%, respectively) and their zoisite contents are 26.2% and 31.3%, respectively. Thus, they are classified as Al-epidote (Fig. 15).

Table 8

White micas

	Paragonite ratio 100 Na <hr/> Na + K	Di-octahedral degree 100 Al ^{VI} <hr/> Al ^{VI} + Fe + Mg	Ferri-phengite comp. 100 Fe ^{''} <hr/> Al ^{VI} + Fe ^{''}	Ferro ratio in trioct. 100 Fe ['] <hr/> Fe ^{''} + Mg	Host rock (Nos. refer to Table 1)
1 Phengite	7.36	12.14	---	41.54	chd-ga-mus-qt schist (No. 34)
2 Phengite	13.83	18.24	—	38.03	gl-ga schist + mus.crots (No. 2)
3 Phengite	8.90	—	2.00	43.52	gl-ga-mus-qt schist (No. 28)

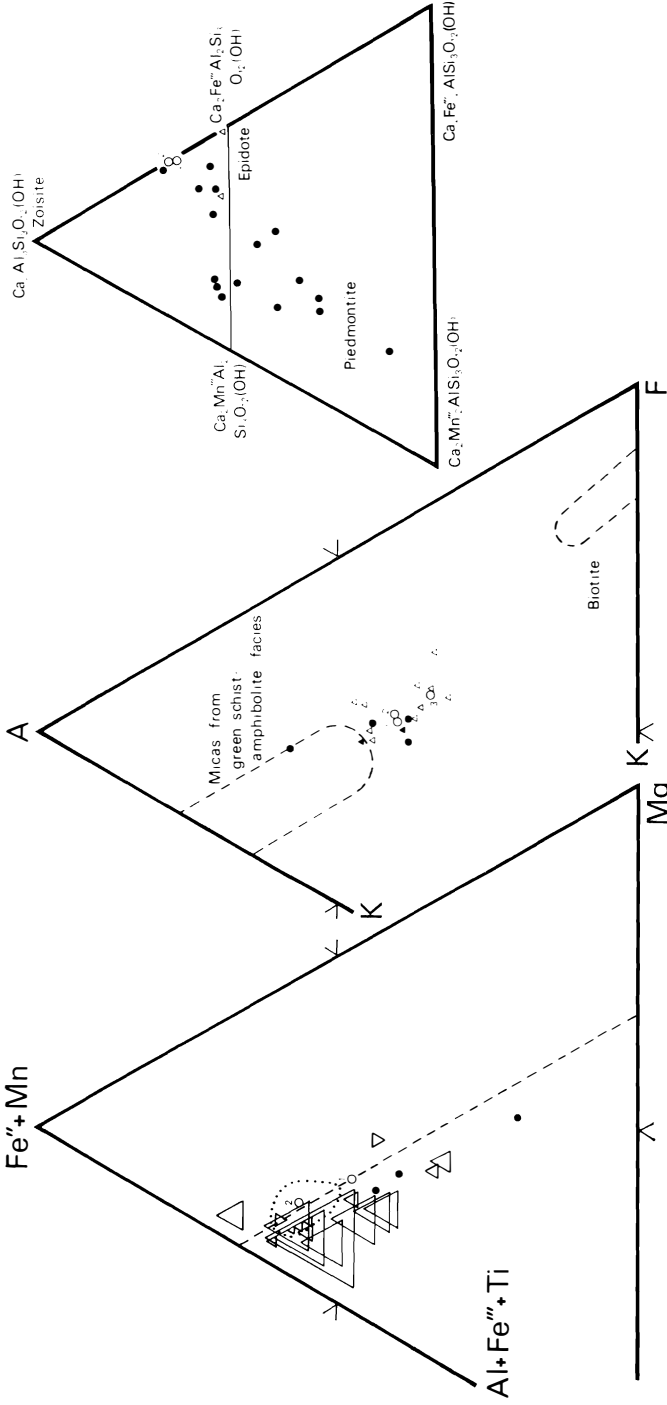


Fig. 13.

Fig. 14.

Fig. 15.

Fig. 13. Chloritoids. Open circles: from the present area, see text; dots: from the glaucophane eclogites of Zermatt; triangles: from various metamorphic rocks in the world (DEER et al. 1962). Dotted curve: from the schists of Venn Staelet, north-west Rheinischesfergebiet (KRAMM 1973).

Fig. 14. White micas. Open circles: from the present area, see Table 6; dots: from Japanese glaucophane schists; open triangles: from California; solid triangles: from New Caledonia.

Fig. 15. Epidotes. Open circles: from the present area, see text; open triangles: from Japanese glaucophane schists; dots: from various metamorphic rocks of the world (DEER et al. 1962).

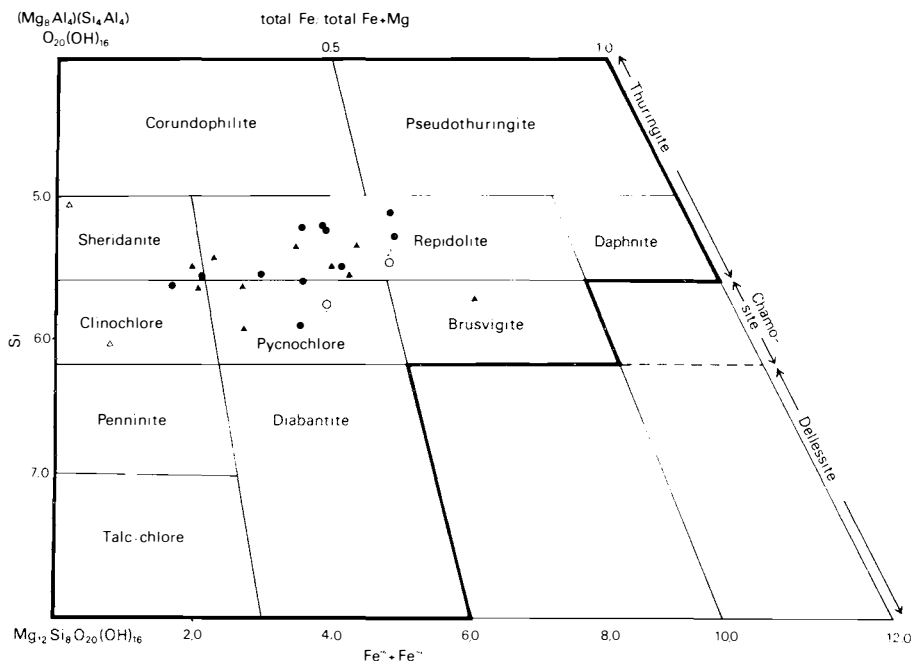


Fig. 16. Chlorites. Open circles: from the present area, see text; dots: from Japanese glaucophane schists; solid triangles: from chlorite schists of the world; open triangles: from serpentine schists of the world. The diagram after HEY (1954).

Chlorites

Chlorite is one of the main constituents of the epidote amphibolites and is always a secondary alteration product filling cracks in the garnets of the glaucophane-bearing rocks.

Two chlorites, from the glaucophane-epidote-actinolite-chlorite schist (1) and garnet-glaucophane schist (2), were analysed. Both have similar Si/Al ratios, but their Fe/Mg ratios are somewhat different. The (1) chlorite is pycnochlorite and the (2) is repidolite (Fig. 16); the latter occurs in the cracks of garnet. All chlorites filling the cracks of garnet in various rocks show almost the same pleochroism and birefringence as the analysed repidolite, while the chlorites from the epidote amphibolites may have a certain compositional variation judging from the difference of the birefringence.

Clinopyroxene

Clinopyroxene is an essential constituent of the eclogitic rocks, most of which have some amount of glaucophane. The host rock of the analysed clinopyroxene is a small lens, 10–15 cm thick, in the glaucophane-garnet-muscovite-quartz schist, composed of more than 95% of slightly dusty, dense clinopyroxene aggregate and less than 5% medium grained garnet which is often cracked and chloritized.

The molecular components of the analysed clinopyroxene are: Jd = 38.4%, Ac = 13.6%, Di = 45.0%, and (Fe,Mg)Si₂O₆ = 3.06%. This composition

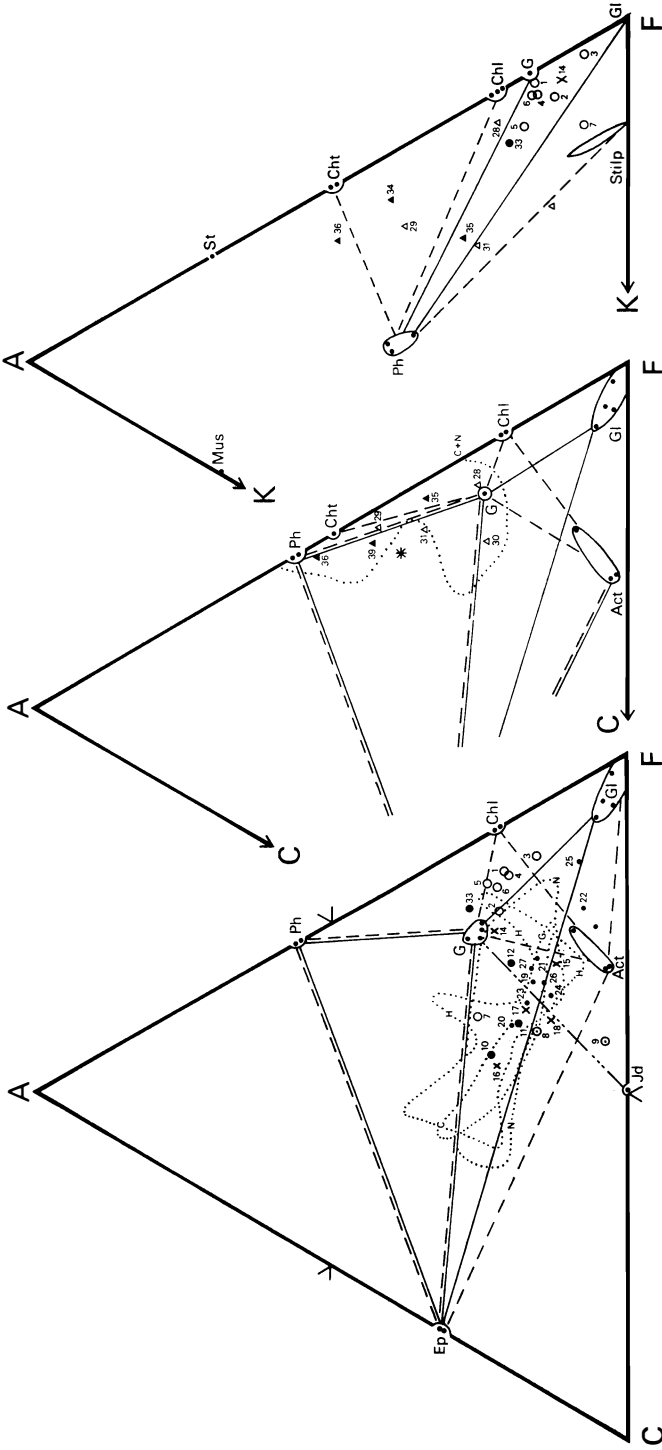


Fig. 17a.

Fig. 17b.

Fig. 18.

Fig. 17. A-C-F diagrams of the rocks from the Vestgötabreen Formation.

a. *Meta-basic rocks.* Broken tie lines: assemblages from the rocks without glaucophane; solid tie lines: assemblages with glaucophane; dash and point tie line: assemblage of the eclogitic rocks. Dotted curves: C, the field of Californian glaucophane schists; H, the field of Hornsund amphibolites; N, the field of New Caledonian glaucophane schists; small dots without No.: analysed minerals. Other symbols and numbers are the same as in Fig. 3a and Table 1.

b. *Meta-sediments.* Legends are the same as in Fig. 17a.

Fig. 18. A-K-F diagram. The tie lines are the same as in Fig. 17 and the symbols and numbers are the same as in Figs. 3b and 17, refer to Table 1.

is in the middle of the omphacite field of the pyroxene diagram (ESSENE and FYFE 1967), and is similar to those found in the metamorphic rocks of blue schist facies and in eclogite from different parts of the world.

V. Metamorphism

To map successive metamorphic zones in the Vestgötabreen Formation in the field is almost impossible because most lithologic units are separated by faults. However, petrographic observations suggest some sequences of metamorphic reactions in these rocks.

(A). Medium grained hornblende-plagioclase rocks are rarely observed as domains of relictic texture (not original igneous texture) in the epidote amphibolites with weak schistosity. In the schistose rocks, all mafic constituents were converted to a chlorite-epidote-sphene-opaque assemblage and plagioclase decomposed into sericite-epidote-carbonate-albite-quartz assemblages. Actinolite occurs at the crests of tightly crenulated chlorite (Plate 2-2), and small garnet grains are scattered. Thus, the reactions chlorite→actinolite and chlorite→garnet are characteristic in these rocks, and indicate the metamorphic condition of the epidote-actinolite subfacies of greenschist facies. No definite indicator of high pressure metamorphic facies series, such as pumpellyite, lawsonite and jadite-quartz assemblage, has been found, except for stilpnomelane (Plate 2-1).

The most dominant rock in the Vestgötabreen Formation is the calcareous schists with muscovite, epidote, clinozoisite, and garnet. There is no mineral indicating higher metamorphic grade than the epidote amphibolite facies. Glaucophane occurs as idiomorphic prisms or flaky aggregates in the muscovite-rich layers of these rocks.

(B). The muscovite-quartz schist, with or without garnet and chloritoid, is one of the dominant rocks in this formation. Staurolite has never been observed. The mineral assemblages observed in these rocks, including (A), are shown by the broken tie lines in Figs. 17a, 17b, and 18. Judging from the zonal distribution of inclusions in the garnet, the chloritoid-bearing assemblages may be earlier than the formation of glaucophane.

The co-existence of chloritoid and Al-silicate polymorphs is suggested to be stable under the condition of 500–570°C and 3.5–5.5 Kb. from experimental studies (NEWTON 1966; RICHARDSON, BELL and GILBERT 1969; and GANGULY 1969). Since no Al-silicates have co-existed with chloritoid even in strongly micaceous layers, for example No. 36, Fig. 17, in the present rocks, it is suggested that they were formed outside the above mentioned condition. Occurrence of stilpnomelane and a slightly higher Mg content in the chloritoid, may indicate relatively high pressure (CHINNER and DIXON 1973). This condition corresponds to the lower amphibolite facies of the intermediate P/T facies series.

(C). The eclogitic rocks occur as isolated bodies and are concordantly enclosed by the chloritoid-garnet-muscovite-quartz schist. Strong diaphrolitic

cleavages of the enclosing schist never penetrate into the eclogite. Direct contact between the eclogite and the calcareous schists has not been found and the eclogite does not occur in the epidote amphibolites. Thus, primary occurrence and the time relation of the eclogite in relation to various schists and amphibolites are unknown.

However, from petrographic observations, the primary bimineralec eclogite constituents are evidently older than the formation of glaucophane. This assemblage is shown by the dash-point tie line in Fig. 17a.

Based on the Fe''/Mg partitioning between co-existing omphacite and garnet (Table 3, No. 5) in the eclogite (an almost glaucophane-free part was selected), the formation conditions of the eclogite was estimated as suggested by BANNO and MATSUI (1965). Since total iron was given as FeO by the electron microprobe, Fe'' was calculated by the formulae proposed by RYBURN et al. (1976).

$$K_G^{Fe''-Mg} = 2.706, K_{cpx}^{Fe''-Mg} = 0.205, K_{G-cpx}^{Fe''-Mg} = 13.21$$

$$Jd \text{ content of cpx} = 38.4\%$$

The P-T condition was estimated referring the diagrams of GREEN and RINGWOOD (1967 and 1972) and RÅHEIM and GREEN (1975), K_D and Jd content as functions of pressure and temperature:

$$\text{minimum pressure} = 9.7 \text{ kb (without co-existence of albite)}$$

$$\text{temperature} = 540\text{--}570^\circ\text{C}$$

These data give a first approximation to the formation conditions of the eclogite from the present area.

The same estimation was attempted for the eclogite from Biskayerhuken, northwest Spitsbergen, reported by GEE (1966).

$$K_G^{Fe''-Mg} = 3.196, K_{cpx}^{Fe''-Mg} = 0.285 \text{ (Fe'' by chemical analyses),}$$

$$K_{G-cpx}^{Fe''-Mg} = 11.18, \quad Jd \text{ content of cpx} = 20,$$

$$\text{minimum pressure} = 4 \text{ kb, } \text{temperature} = 550\text{--}620^\circ\text{C}$$

The low pressure of the Biskayerhuken eclogite can be explained by secondary modification under the conditions to form extensive amphibole-plagioclase symplectite which has not been observed in the eclogite of the present area.

(D). The glaucophane-bearing assemblages are younger than all others and are similar to the type IV glaucophane schist of California (the solid tie lines in Figs. 17a, 17b, and 18). The formation conditions of the latter have been studied by the Oxygen isotope method by TAYLOR and COLEMAN (1968) and are given as 400–500°C. The phengites (Fig. 14) from the glaucophane-bearing rocks have the Si^{+4} value = 3.39 in average, based on the assumption

of O_{10} per unit formula. Taking the estimated maximum temperature from the Californian type IV rocks, the maximum pressure for the present glaucophane schists can be assumed to be about 8 kb from the diagram of VELDE (1967) for the estimate of P–T conditions from the Si^{+4} value of natural phengites. These conditions may be not far from the present case.

All estimated physical conditions for different types of rocks from the Vestgötabreen Formation are shown on the phase diagram (Fig. 19) as A, B, C, and D with the same divisions as discussed above. The P–T conditions of B, C, and D are in the similar temperature range, while the pressure differs to some extent. If the estimates made above are near the truth, the rocks of the Vestgötabreen Formation represent a transitional facies series between the blueschist type and the intermediate P/T type metamorphic facies series. This corresponds to the high-temperature glaucophane schist facies of WINKLER (1967) and TAYLOR and COLEMAN (1968). This conclusion agrees with the lack of typical high pressure minerals in the lower grade rocks and also with the mode of occurrence of glaucophane in vein-like and radial aggregates.

VI. Metamorphic facies series of the western Spitsbergen Hecla Hoek rocks

Epidote amphibolites similar to those of the present area are known from several places along the west coast of Spitsbergen: the north and south sides of St. Jonsfjorden, the southern half of Nordenskiöldkysten, west of Recherchebreen, and the northern entrance of Hornsund. All these rocks have mineral assemblages of the epidote-actinolite subfacies of greenschist facies in general, but an important characteristic is the association of stilpnomelane and bluish green sub-alkaline amphiboles.

Many thin, green phyllite horizons occur in Prins Karls Forland and everywhere along the west coast of Spitsbergen, and their essential mineral assemblages are chlorite-sericite-epidote with occasional chloritoid, for example in Prins Karls Forland (ATKINSON 1956).

All these basic rocks are stratigraphically lower than the possibly Vendian tilloid formation; the Hornsund amphibolites are of the Eimfjellet Group of the Torellbreen Supergroup (BIRKENMAJER 1975), while those of the Bellsund area cut the lower calcareous successions of the Sofiebogen Group and are found as pebbles of tillitic conglomerate (HJELLE 1962). The amphibolites from St. Jonsfjorden and Prins Karls Forland are not correlated with certainty yet, but may be lower than the tilloid formation.

A staurolite-biotite assemblage was reported by ATKINSON (1956) from Prins Karls Forland, and the garnet-sillimanite and staurolite-garnet assemblages have been found by the present author in the pelitic metamorphic rocks from Prins Karls Forland and Sørkapp Land, respectively. Garnet-biotite gneisses occur north of Hornsund, and similar garnet-biotite schists are distributed in Svartfjellstranda on the east coast of Forlandsundet. All these high grade rocks

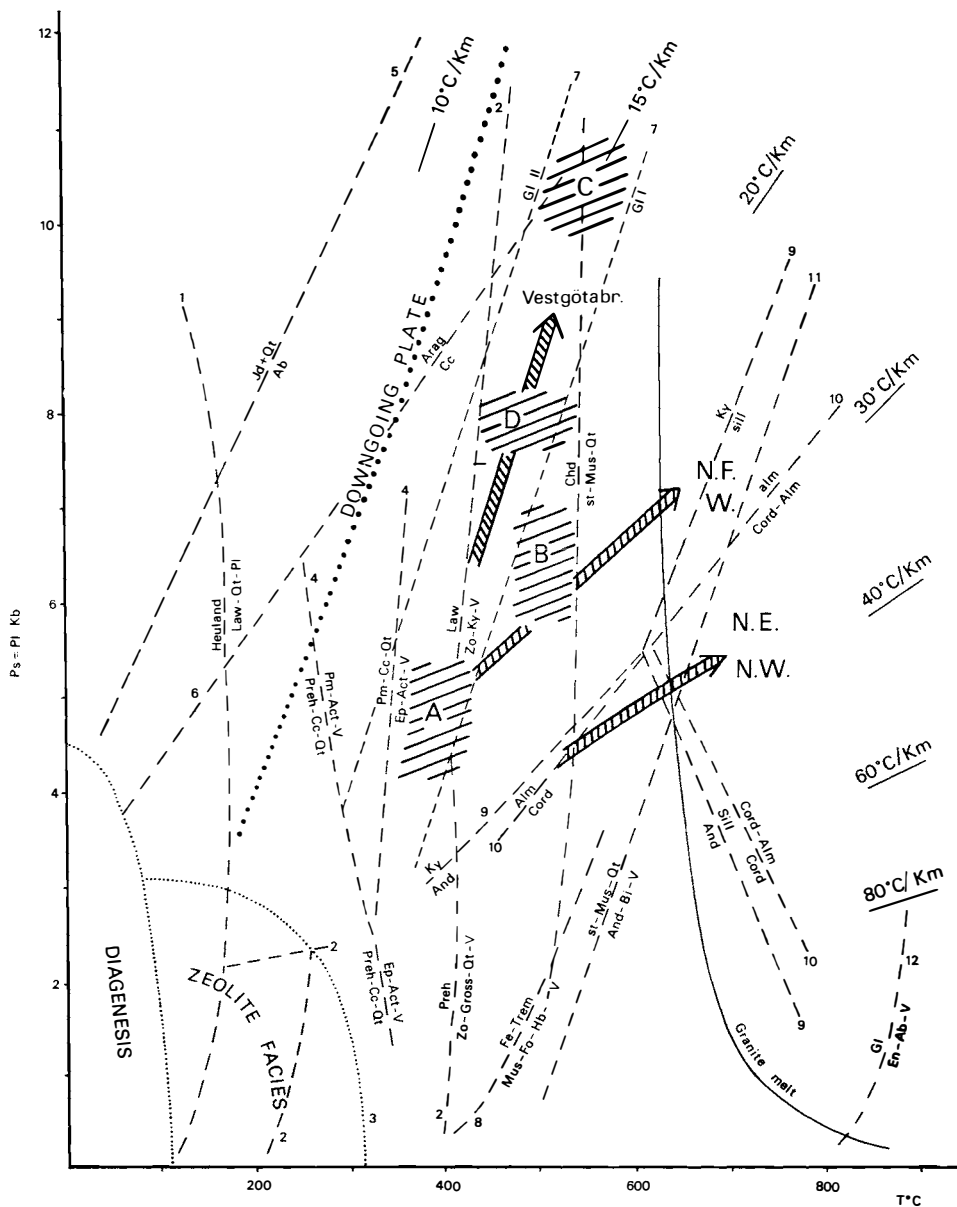


Fig. 19. Metamorphic facies series.

A: epidote-actinolite-greenstones; B: chloritoid-muscovite-quartz schists; C: eclogitic rocks; D: glaucophane schists. Thick arrows: the metamorphic facies series of: N.A., Nordaustlandet zone; N.F., Ny Friesland zone; N.W., Northwestern Spitsbergen; W, West Spitsbergen metamorphic zone. (1) NITSCH 1968, (2) LIUS 1971, (3) SEKI 1973, (4) NITSCH 1971, (5) NEUTON and SMITH 1967, (6) CLARK 1957, (7) ERNST 1963, (8) ERNST 1966, (9) RICHARDSON 1969, (10) HIRSCHBERG 1968, (11) HOCHAK 1969, (12) ERNST 1961.

can be grouped as Bjørnhamna Group of the Torellbreen Supergroup and are the deepest rocks so far observed in western Spitsbergen.

The mineral parageneses of these rocks indicate that the metamorphism of this area belongs to the intermediate P/T type metamorphic facies series.

It is said that the type of metamorphism in a certain metamorphic zone is restricted to one metamorphic facies series, and may extend into types intermediate to the neighbouring facies series (MIYASHIRO 1961; ZWART et al. 1967). The occurrence of high-temperature glaucophane schist facies in the intermediate P/T type western Spitsbergen zone does not contradict to this statement. A similar example has been reported from the Betic orogenic zone, southern Spain (KAMPSCHUUR 1975).

The occurrence of idioblastic large crystals of glaucophane in the Vestgötabreen Formation indicates that this mineral was formed later than the muscovite-quartz schists and eclogitic rocks. This suggests a polyphasial metamorphism.

A peculiar chemical composition with a large amount of Na-rich pyroxene in the eclogitic rocks, might have played an important role in the formation of the glaucophane schists of the Vestgötabreen Formation. Recent studies of eclogite amphiboles (MOTTANE and EDGAR 1970) show that much glaucophane, Ca-glaucophane and barroisite were formed in the symplectite amphiboles after the omphacite of eclogites, and this suggests a retrogressive formation of the glaucophane schists from eclogites.

VII. A Caledonian subduction zone?

HORSFIELD (1972) mentioned that the lithologic characteristics of the Vestgötabreen rock suite (in his definition) suggest a possible subduction zone, involving oceanic crust, during the Caledonian period: Indeed, some young blueschist metamorphic zones elsewhere in the world can be explained by a subduction-zone model (ERNST 1972 and 1975; MIYASHIRO 1972), but there are still many problems to be solved when this idea is going to be applied far back into geologic time (MIYASHIRO 1975).

Although the Vestgötabreen Formation is an isolated thrust schuppen, it probably had not travelled a long distance into the present area. From what we know of the metamorphic grade, these rocks are comparable to those of the upper Isbjørnhamna Group or younger, the most possible correlation to the epidote amphibolites being to the Sofiebogen Group of the Bellsund area.

BIRKENMAJER (1975) proposed two-fold geosynclinal cycles in the pre-Cambrian of western Spitsbergen based on the stratigraphic breaks in the Hecla Hoek successions: the older Torellian eugeosynclinal cycle and the younger Jarlsbergian miogeosynclinal cycle.

The observed thickness of the Torellian eugeosynclinal succession is about 11 km. The metamorphic condition of the chloritoid-staurolite-bearing rocks of the Isbjørnhamna Group is supposed to be 5–7 kb and 500–700°C (Fig. 19); this is a depth of about 15–20 km. This condition is not impossible by large

lateral shortening in a folded belt and some introduction of heat from the depth brought up by granitic intrusion, such as in the northern Hornsund area. The general eugeosynclinal characteristics of the Torellbreen Supergroup (BIRKENMAJER 1975) are in harmony with the development of an intermediate P/T type metamorphic facies series, and there is no need to introduce the idea of a subduction zone.

The K–A ages of the glaucophane schists are mostly in the range of 402 ± 14 – $475 \pm$ m.y. (muscovite and whole rock), indicating a main Caledonian phase, and one is 621 ± 12 m.y. (whole rock) (HORSFIELD 1972). The older one may be correlated to the 556 ± 24 and 584 ± 25 m.y. biotite ages of the garnet-biotite schist from Hornsund (GAYER et al. 1966), which suggest the middle Cambrian event similar to the west-Finnmark event in North Norway (STUART and MILLER 1967) and the Grampian event of Scotland (DEWEY and PANKHURST 1969).

The rocks involved in the proposed Jarlsbergian and Hornsundian event (BIRKENMAJER 1975) are of miogeosynclinal character and are not likely to produce an intermediate P/T facies series metamorphism. It is reasonable to consider the original rocks of the Vestgötabreen Formation to be of the Torellian eugeosynclinal deposits, probably metamorphosed in the Torellian event and modified during the Caledonian period to renew the K–A ages.

An important piece of geological evidence for the conclusion of a subduction of oceanic crust would be to find out to what extent the rock succession of such metamorphic zones, belongs to the oceanic ophiolite suite. The rock assemblage of the Vestgötabreen Formation is very similar to the classic “ophiolite trinity” of STEINMANN (1927). The original succession of these rocks, phyllarenite — basic volcanics — limestone, suggests a sedimentary condition around the margins of an eugeosynclinal basin, if there is any dominant ophiolite suite associated in the same geologic regime. The percentage of basic rocks in the observed type succession of the Torellian eugeosyncline is about 12%, and is far smaller in the areal distribution. The association of alkalic and tholeiitic volcanic rocks in the Vestgötabreen Formation fits well in with that of the Sanbagawa and Franciscan blueschist zone as summarized by MIYASHIRO (1975). A large amount of sodic alkaline rocks is characteristic for ocean islands of probable hot-spot origin; however, if these rocks were detached from descending ocean plate and implicated in a subduction zone, a large amount of oceanic tholeiite might be associated. The evidences — association of shallow sea sediments and very small amounts of tholeiitic rocks in the Vestgötabreen Formation — are not in favour of a subduction zone. Moreover, some potassic alkaline rocks in the Vestgötabreen Formation and occurrence of granite and potassic rhyorite ($K_2O = 9.40$ and 10.21 wt.%) pebbles in the conglomerate of the Vimsodden Subgroup of Hornsund, suggest the existence of a continental crust during the development of the Torellian eugeosyncline in western Spitsbergen. Thus, the idea of a Caledonian subduction zone is still an open question at the present stage of our knowledge. The idea of a large transcurrent displacement proposed by HARLAND (1969) and others, is beyond the scope of the present paper.

VIII. A discussion on the metamorphic facies series in Svalbard

It is evident that the majority of Hecla Hoek rocks in Svalbard were involved in the Caledonian metamorphism. The radiometric ages already available support this (GAYER et al. 1966). These metamorphic rocks occur around the north and west sides of Svalbard, and three zones of N-S trends are distinguished on the northern coast: the Nordaustlandet (NE) zone, the Ny Friesland (NF) zone, and the Northwestern (NW) zone. Another zone (W zone) occurs along the west coast of Spitsbergen (Fig. 1).

The characteristic mineral parageneses and metamorphic facies series of these zones are shown in Table 9. Although the Vestgötabreen Formation is not a typical high-pressure type metamorphic facies series, the Caledonian metamorphism of Svalbard covers nearly all three varieties of a metamorphic facies series within the width of about 300 km (Figs. 1 and 19). The zones of the high T/P facies series, the NE and the NW zones, are characterized by extensive development of migmatite and granite. The zones of the intermediate P/T facies series, the NF and the W zone, have a little granite intrusions and have some eugeosynclinal sediments in the lower parts of their geosynclinal successions.

These four zones can be considered two sets of paired metamorphic zones (MIYASHIRO 1961 and 1973) if all these zones were formed during the Caledonian period.

Different metamorphic facies series are closely related to different histories of the geosynclinal development. Therefore, the idea of paired metamorphic zones should be justified by the difference in the geosynclinal successions of the Hecla Hoek from these zones. Some differences have already been noticed, but their characteristics are not so contrasting as in the younger paired zones which are typically made up by a pair of a high P type and a high T/P type metamorphic zones.

The present author assumes that these two sets of paired zones, the NE-NF pair and the NW-W pair, are essential geologic units in the pre-Devonian

Table 9

Mineral parageneses and metamorphic facies series of Svalbard (ref. Fig. 1)

Zone	Diagnostic mineral assemblages	Metamorphic facies series
Nordaustlandet zone	Cordierite, almandine, sillimanite (staurolite in the contact zone)	High T/P type facies series (the older unknown)
Ny Friesland zone	Staurolite, almandine, kyanite, sillimanite	Intermediate P/T type
NW zone	Cordierite, almandine, sillimanite (kyanite and staurolite relic)	High T/P type (the older: intermediate P/T type)
W zone	Stilpnomelane, chloritoid, staurolite, sillimanite, sub-alkaline amphiboles, almandine (younger; alkaline amphiboles)	Intermediate P/T type

structures of Svalbard, and they should be considered fundamental tectonic units when any large scale transposition is assumed on the plate tectonics (HARLAND 1969; HARLAND et al. 1974; CHURKIN 1973; BIRKENMAJER 1972).

An alternative interpretation of the difference of pre-Devonian rocks in Svalbard is superposed metamorphism.

Two occurrences of kyanite have been found in the NW zone: Biskayerhuken (GEE 1964) with staurolite, and Danskøya–Amsterdamøya. The kyanite of the latter locality was recently conformed by HJELLE and the present author from HJELLE's collection. The kyanite occurs as small fragmental relics, sericitized along cracks, in the cordierite-sillimanite gneiss.

Pseudomorphs of spinel-corundum clots after presumably staurolite or chloritoid, were reported by the present author (Plates 3 and 4 of HJELLE and OHTA 1974) from the Smeerenburgfjorden area.

These evidences suggest that there was a regional metamorphism prior to the migmatization, having the nature of the intermediate P/T facies series in the NW zone.

Staurolite has been known from Nordaustlandet (FLOOD et al. 1969), but this occurs around a granitic intrusion and does not show any older metamorphic phase. Although Nordaustlandet is completely soaked by migmatization products and it is very difficult to distinguish older phases, a deformation of late pre-Cambrian age, prior to the deposition of Murchisonfjorden Supergroup has been suggested (GEE in FLOOD et al. 1969).

In the W zone, the metamorphism of the intermediate P/T facies series can be considered to be older than the main Caledonian phase as discussed before.

It is worth mentioning that the older phases of both the NW and W zones are the intermediate P/T facies series as in the NF zone (GAYER and WALLIS 1966). One possible conclusion based on this evidence is that the metamorphism of the earlier phase of Caledonian or Torellian (BIRKENMAJER 1975), was of the intermediate P/T facies series over all three zones, and the later main Caledonian high T/P facies series was superimposed intensely in the NE and NW zone.

Existence of continental crust in the basement of the Hecla Hoek geosyncline is suggested by the occurrences of thick acidic volcanic rocks in the Kapp Hansteen Formation of Nordaustlandet and in the Harkerbreen Group of Ny Friesland, potassic rhyorite in the Vimsodden Subgroup of Hornsund, and the granite pebbles from the Vimsodden Subgroup, Rittervatnet Formation of the Harkerbreen Group, and from the Vendian tillitic rocks elsewhere in Svalbard. Without the study of lower Hecla Hoek volcanic rocks, the problem of a proto-Iapetus ocean (HARLAND and GAYER 1972) can not be discussed with any certainty.

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PLATES

PLATE I

Glaucophane-bearing rocks.

- 1-1. Feather amphibolite texture of glaucophane-muscovite schist.
- 1-2. Garnet-muscovite-quartz schist with glaucophane needles.
- 1-3. Garnet-glaucophane schist, the groundmass is pure glaucophane aggregate.
- 1-4. Garnet-glaucophane schist, idiomorphic large garnet and pure glaucophane groundmass with secondary calcite veins.

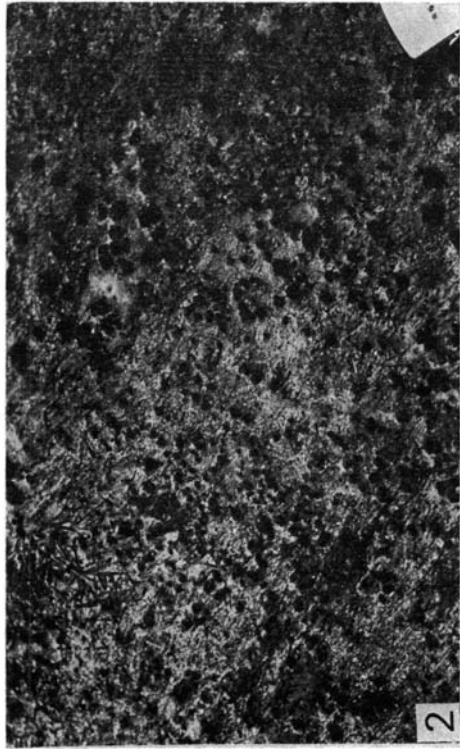


PLATE II

Microphotographs.

- 2-1. Stilpnomelane in the epidote-chlorite schist. The short side of picture is 0.7 mm.
- 2-2. Actinolite needles along the border between the chlorite-rich and quartz-rich band of the epidote-chlorite schist. The short side of picture is 0.7 mm.
- 2-3. Garnet with chloritoid inclusions in the inner part and poikiloblastic texture along the margins. The short side of picture is 2.0 mm.
- 2-4. Cracked garnet in dense glaucophane groundmass. The short side of picture is 2.0 mm.

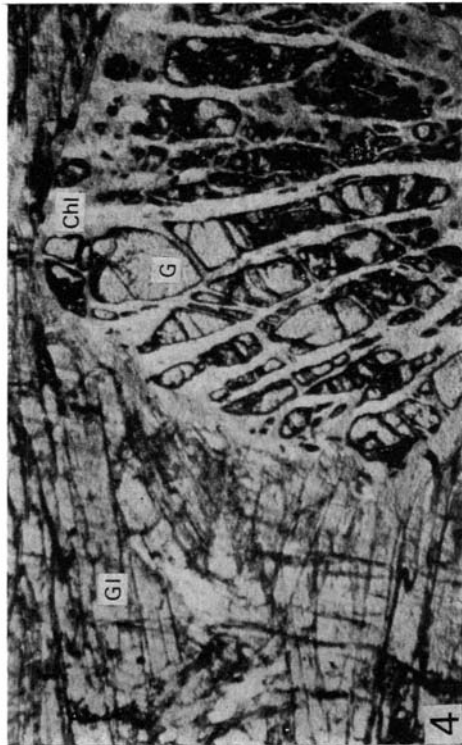
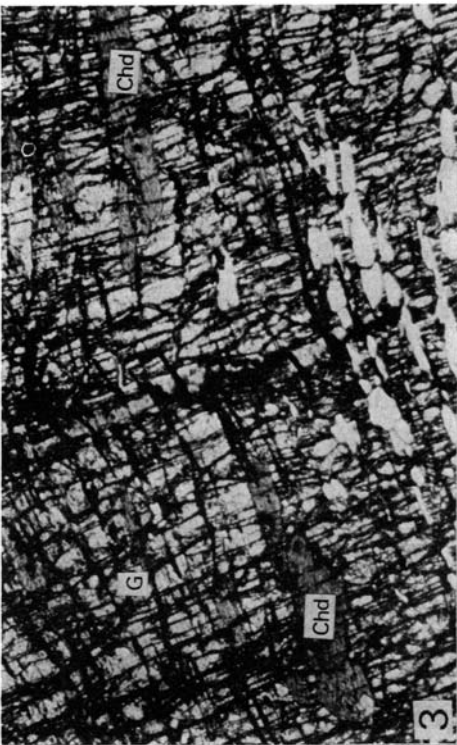
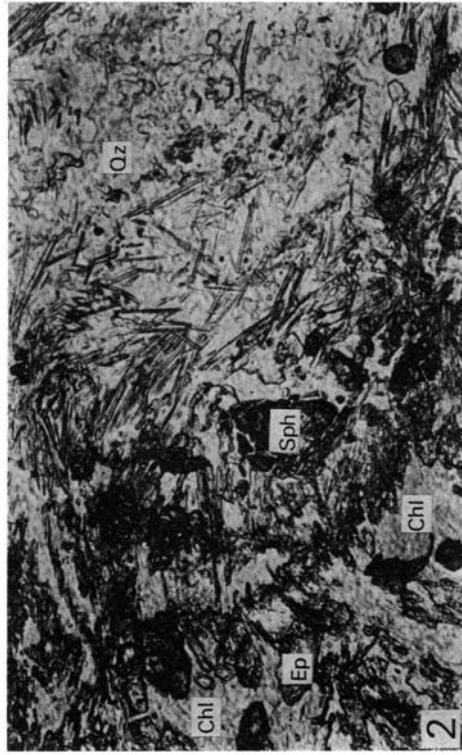
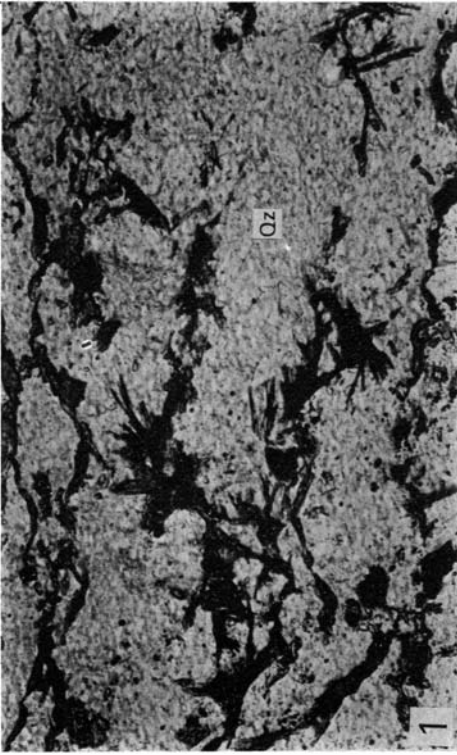
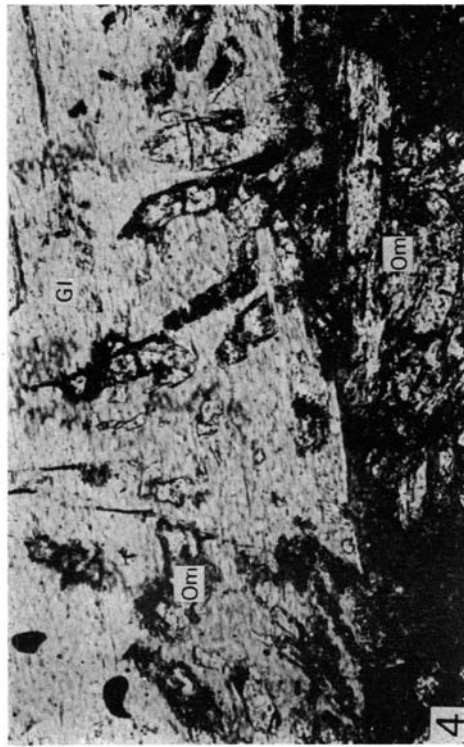
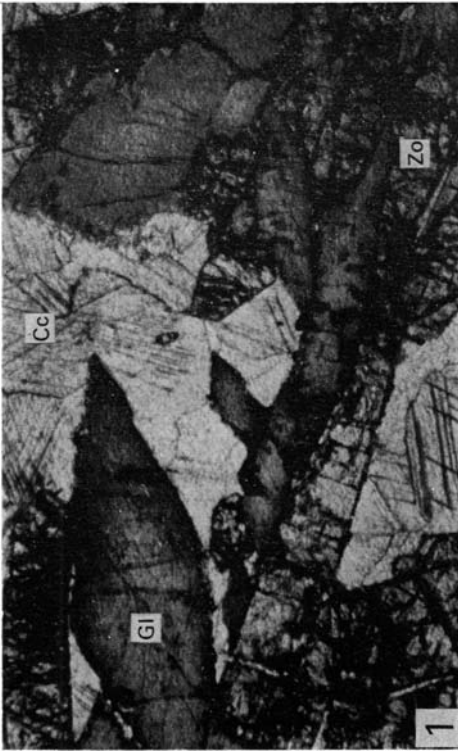


PLATE III

Microphotographs.

- 3-1. Zoned idiomorphic glaucophane in the calcareous schist. The short side of picture is 2.0 mm.
- 3-2. Zoned idiomorphic glaucophane in the dusty groundmass of omphacite, eclogitic rock. The short side of picture is 2.0 mm.
- 3-3. Idiomorphic glaucophane, the dark parts along cracks and cleavages are crossite. The short side of picture is 0.7 mm.
- 3-4. Omphacite inclusions at the margin of idiomorphic glaucophane in the eclogitic rock. The short side of picture is 0.7 mm.



Upper Precambrian microphytolites and stromatolites from Svalbard

By V. E. MILSTEIN and N. P. GOLOVANOV

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Abstract

Result of the areal and vertical distribution of Riphean and Vendian communities of microphytolites of Spitsbergen and Bjørnøya have been dealt with in the paper, and data on distribution of upper Riphean stromatolites of Nordaustlandet are presented. Stromatolites from other regions of the archipelago have been described in detail by RAABEN (1975).

Microphytolites and stromatolites or phytolites were formed by carbonate secreting blue-green algae and possibly bacteria by processes of homogenic and mechanic sedimentation. Stromatolites are undulate-layered structures attached to a substrate; microphytolites are nodular or pelletoidal carbonate structures not attached to a substrate.

A. Microphytolites of Spitsbergen and Bjørnøya

Study of Upper Precambrian microphytolites collected by KRASIL'ŠČIKOV from Spitsbergen and Bjørnøya began in the early 1960's and was later continued by MILSTEIN. A main contributor to the study of the formations in the regions discussed was ZABRODIN.

Microphytolites and stromatolites are widespread in the Upper Precambrian deposits of Spitsbergen and Bjørnøya. They have been found in sections of Nordaustlandet (western part), north-eastern Spitsbergen (Ny Friesland, Olav V Land), southern Spitsbergen (Hornsund area), and Bjørnøya. A number of microphytolite assemblages were recognized through analyses of geographical and stratigraphical distribution of microphytolites, based on data obtained by RAABEN, ZABRODIN, and GOLOVANOV. The assemblages were named after the formation in which they occur and index fossils.

Assemblage “Höferpynten”

Osagia tenuilamellata — *Vesicularites raabena*

Distribution: Regional. The sequence at Höferpynten including the Sofiebogen Formation, Höferpynten Formation and the base of the Gåshamna Formation, also presumably the lower member of the Flora¹ Formation, Nordaustlandet.

Composition: After MILSTEIN and ZABRODIN: *Osagia* of the Middle Riphean with *Osagia tenuilamellata* REIT., *O. cf. tenuilamellata* REIT., *O. minuscula* MILST. (in collection), *Vesicularites* aff. *igaricus* MILST., *V. raabena* ZABR.

Age: Middle — Upper Riphean after MILSTEIN and ZABRODIN (?).

Assemblage “Bogen”

Osagia kingbreensis — *Radiosus sculeatus* — *Radiosus decipiens*

Distribution: Local. The Bogen Member of the Kingbreen Formation.

Composition: After ZABRODIN: *Osagia kingbreensis* ZABR., *Radiosus decipiens* ZABR., *R. aculeatus* ZHUR., *R. elongatus* ZHUR., *R. pachyradius* ZABR.

Age: Upper Riphean Bir’yan division after RAABEN and ZABRODIN.

Assemblage “Norvik”

Asterosphaeroides — *Radiosus*

Distribution: Local. The Norvik Formation.

Composition: After MILSTEIN: *Asterosphaeroides* REITL., *Radiosus* ZHUR.

Age: Riphean after MILSTEIN.

Assemblage “Enpiggen”

Asterosphaeroides tubulosus — *Radiosus limpidus* — *Radiosus fasciculatus*

Distribution: Local. The Enpiggen Member of the Oxfordbreen Formation.

Composition: After ZABRODIN: *Radiosus aculeatus* ZHUR., *R. limpidus* ZHUR., *R. manjaricus* ZABR., *R. Fasciculatus* ZABR., *Asterosphaeroides tubulosus* ZABR., *Volvatella Svalbardica* ZABR.

Age: Upper Riphean, Bir’yan division after RAABEN and ZABRODIN.

Assemblage “Grusdievbreen–Hunnberg”

Osagia pullata — *Vesicularites elongatus* — *Vesicularites raabena*

Distribution: Regional. The upper member of the Grusdievbreen Formation, Lower Hunnberg Formation.

Composition: After MILSTEIN: *Vesicularites* REITL., *Vesicularites flexuosus* REITL., after ZABRODIN: *Vesicularities vapolensis* ZABR., *V. raabena* ZABR., *V. elongatus* ZABR., *V. enigmatus* ZABR., *V. parvus* ZABR., *Osagia maculata* ZABR., *O. pullata* ZABR., *O. milsteinae* ZABR., *Astrosphaeroides* (?) *ruminatus* ZABR.

Age: Middle-Upper Riphean after MILSTEIN and GOLOVANOV. Upper Riphean, Min’yar division, B’yank subdivision after RAABEN and ZABRODIN.

¹ The terms “regional distribution” and “local distribution” refer to the distribution of microphytolites within Spitsbergen and Bjørnøya only.

Assemblage “Svanbergfjellet–Hunnberg”

Osagia milsteinae — *Vesicularites enigmatus* — *Vesicularites vapolensis*

Distribution: Regional. The Svanbergfjellet Formation, Upper Hunnberg Formation.

Composition: After MILSTEIN: *Osagia columna* REITL., *Asterosphaeroides* REITL. (was found in an isolated outcrop of limestone). After ZABRODIN: *Vesicularites vapolensis* ZABR., *V. raabena* ZABR., *V. enigmatus* ZABR., *Osagia maculata* ZABR., *O. milsteinae* ZABR., *Radiosus licudus* ZABR., *R. aculeatus* ZHUR., *Asterosphaeroides* (?) *ruminatus* ZABR.

Age: Middle — Upper Riphean after MILSTEIN and GOLOVANOV. Upper Riphean Min’yar division, B’yank subdivision after RAABEN and ZABRODIN.

Assemblage “Draken”

Vesicularites bothrydioformis — *Vesicularites concretus* — *Nubecularites abustus*

Distribution: Local. The Draken Formation.

Composition: After ZABRODIN: *Vesicularites concretus* ZHUR. ZHUR., *V. bothrydioformis* (KRASNOP.), *V. lobatus* REITL., *Radiosus polaris* ZABR., *Nubecularites abustus* ZHUR.

Age: The top of the Upper Riphean, Vendian Yuomian division after RAABEN and ZABRODIN.

Assemblage “Russehamna”

Distribution: Local. The Russehamna Formation.

Composition: After ZABRODIN at the base of the section: *Osagia crispa* ZHUR., *Radiosus* cf. *aculeatus* ZHUR., somewhat higher: *Vesicularites vapolensis* ZABR., *V. elongatus* ZABR., *V. enigmatus* ZABR., *V. parvus* ZABR. *Osagia* aff. *maculata* ZABR., after MILSTEIN in ascending order four local assemblages are recognized:

- Assemblage I: *Vesicularites lobatus* REITL., *Nubecularites* MASL.
Assemblage II: *Osagia crispa* ZHUR., *O. medwezhziella* n.sp., *Vesicularites alexandrovi* n.sp., *Radiosus aculeatus* ZHUR.
Assemblage III“A”: *Osagia maculata* ZABR., *O. pullata* ZABR., *O. milsteinae* ZABR. (samples from separate blocks), *Vesicularites elongatus* ZABR., *V. raabena* ZABR.
Assemblage III“B”: *Vesicularites compositus* ZHUR., *V. enigmatus* ZABR., *V. vapolensis* ZABR., *V. lobatus* REITL., *V. aff. bothrydioformis* (KRASNOP), *V. elongatus* ZABR., *V. cf. elongatus* ZABR., *V. parvus* ZABR., *Asterosphaeroides* (?) *ruminatus* ZABR.
Assemblage IV: microphytolites, oncolites, *Vesicularites* REITL.

Age: After RAABEN and ZABRODIN: Upper Riphean Min’yar division, B’yank subdivision. After MILSTEIN: assemblage II Upper Riphean, assemblage III “B” Upper Riphean, Vendian (Yudomian) complex.

Local assemblages together with the assemblages “Russehamna” are correlated with “Grusdievbreen–Hunnberg”, “Svanbergfjellet–Hunnberg” and “Draken” assemblages.

Assemblage “Backlundtoppen–Ryssø”

Osagia maculata — *Osagia porrecta* — *Osagia milsteinae*

Distribution: Regional. The lower and upper members of the Backlundtoppen Formation, the Ryssø Formation.

Composition: After MILSTEIN in the lower part of the section (in alluvial disintegrated rocks): *Osagia frislandica* MILST., *O. aff. torta* MILST., *Asterophaeroides* REITL., *Catagraphia* MASL., *Vesicularites eniseicus* MILST. In the upper part of the section *Catagraphia* MASL. After ZABRODIN in the lower part of the section *Osagia musculata* ZABR., *O. pullata* ZABR., *O. milsteinae* ZABR., *O. porrecta* ZABR., in the upper part of the section: *Radiosus polaris* ZABR., *Nubecularites abustus* ZHUR., *Vesicularites lobatus* REITL., *V. concretus* ZHUR.

Age: After MILSTEIN and GOLOVANOV: Riphean, Upper Riphean. After RAABEN and ZABRODIN: the top of the Upper Riphean, Vendian (Yudomian) division.

Assemblage “Elbobreen–Backaberg”

Osagia svalbardica — *orbiculatus* — *Vermiculites irregularis*

Distribution: Regional. The Elbobreen Formation, the upper unit of the Backaberg Formation.

Composition: After MILSTEIN: *Osagia svalbardica* MILST. *Vermiculites irregularis* (REITL.). After ZABRODIN: *Vesicularites lobatus* REITL., *V. bothrydiiformis* (KRASNOP.), *V. concretus* ZHUR., *V. orbiculatus* ZABR., *Radiosus polaris* ZABR., *Nubecularites abustus* ZHUR., *Volvatella* sp.

Age: After MILSTEIN: Vendian (Yudomian) complex. After RAABEN and ZABRODIN: the top of the Upper Riphean, Vendian (Yudomian) division.

Assemblage “Drakoisen”

Vesicularites lobatus — *Nubecularites abustus*

Distribution: Local. Drakoisen Formation.

Composition: After ZABRODIN: *Vesicularites lobatus* REITL., *Nubecularites abustus* ZHUR.

Age: The top of the Upper Riphean, Vendian (Yudomian) division.

The composition of species and genera change up the section. Each assemblage is named after a group of index fossils, in order to mark its appearance.

The oldest “Höferpynten” assemblage (*Osagia*) is similar in composition to the second assemblage by ZHURAVLEVA but it cannot be assigned to the Middle Riphean because it contains *Vesicularites raabena*.

The assemblages “Bogen”, “Norvik”, “Enpiggen” include the genus *Radiosus*. this and the presence of forms from the third assemblage of ZHURAVLEVA confirms RAABEN and ZABRODIN’s date of Upper Riphean in age.

The “Grusdievbreen–Hunnberg” and “Svanbergfjellet–Hunnberg” assemblages include *Osagia–Vesicularites*. These assemblages have nothing in common with the third assemblage of ZHURAVLEVA and hence with the Upper Riphean deposits. The above assemblages contain the Middle Riphean *Osagia columnata*

(from an isolated outcrop) and a number of forms known to the author from the Kolosov Formation of Taimyr which were previously assigned to the Middle Riphean. So, at present we cannot postulate with certainty an Upper Riphean age for the “Bogen”, “Svanbergfjellet–Hunnberg”, “Enpiggen”, “Grusdievbreen–Hunnberg” assemblages.

These assemblages and the divisions established by RAABEN within the Upper Riphean are dated as Middle–Upper Riphean. The “Draken” assemblage (*Vesicularites*) is equivalent to the fourth, Vendian (Yudomian) assemblage of ZHURAVLEVA. The “Backlundtoppen–Ryssø” (*Vesicularites–Osagia*) assemblage and the fourth (*Vesicularites*) assemblage of ZHURAVLEVA contains *Osagia maculata*, *O. pullata*, *O. milsteinae* which belong to the older *Osagia–Vesicularites* assemblages such as “Grusdievbreen–Hunnberg” and “Svanbergfjellet–Hunnberg”. The Middle–Upper Riphean *Vesicularites eniseicus* was also found there (unfortunately, in disintergrated alluvial blocks). Therefore the authors regard the “Backlundtoppen–Ryssø” and the “Draken” assemblages as transitional between the Upper Riphean and Vendian, and not as pure Vendian (Yudomian).

The “Elbobreen–Backaberg” and “Drakoisen” assemblages may be correlated with confidence with the fourth Vendian (Yudomian) assemblage of ZHURAVLEVA.

On the basis of the microphytolite study it is possible to come to the following general conclusions about the formation ages. The formations of the Lomfjorden Supergroup of Ny Friesland, the Murchisonfjorden Supergroup of Nordaustlandet and probably the Höferpynten Formation of the western coast of Spitsbergen belong to the Middle–Upper Riphean. The sequence of “old dolomites” of Bjørnøya (the Russehamna Formation) is Upper Riphean in age. The Polarisbreen Group of Ny Friesland belongs to the Vendian.

Many forms of the microphytolites of Spitsbergen and Bjørnøya were reported by ZHURAVLEVA and ZABRODIN from the Upper Riphean rocks of the Urals and Timan. Microphytolite assemblages recognized from these regions correlate with those from Spitsbergen and Bjørnøya. In the present author’s opinion it is impossible to draw a boundary between Upper Riphean and Vendian (Yudomian) assemblages. The older assemblages are considered to be Middle–Upper Riphean. In this connection earlier views of RAZNITSYN concerning the Middle Riphean age of the Bystra Formation must be considered.

B. Riphean stromatolites of Nordaustlandet

The first information about stromatolites from Nordaustlandet was recorded by KULLING (1934). Later they were studied by GOLOVANOV and RAABEN (1967).

The Riphean deposits of Nordaustlandet are represented mainly by the Murchisonfjorden Supergroup, subdivided on the basis of recent data into the following formations (above the Franklinsundet Group): Flora, Norvik, Raudstup–Sälodd, Hunnberg, and Ryssø (FLOOD et al. 1969, KRASIL’SČIKOV 1973).

The stromatolites were collected by KRASIL'SČIKOV in 1963 and 1964. They were found only in the Ryssø and Hunnberg formations, where they consist of columnar forms (GOLOVANOV 1967). Both branching and non-branching columnar stromatolites occur in the Hunnberg Formation. The lower part of the formation contains actively branching forms of *Gymnosolen murchisonicus* GOLOVANOV. The upper horizons yield actively, passively, branching, and non-branching columnar stromatolites. *Inseria blingica* GOLOVANOV, *I. chunbergica* GOLOVANOV, *Yungulsia* sp. are among the actively branching forms. Passively branching stromatolites include *Kussiella* (?) sp., and non-branching columnar forms consist of *Jacutophyton spitsbergensis* GOLOVANOV and *Conophyton* sp.

Strongly recrystallized columnar stromatolites occur in dolomites of the lower part of the Ryssø Formation. The upper parts yield columnar actively branching stromatolites *Gymnosolen* cf. *ramsayi*.

The stromatolites studied from the Hunnberg and Ryssø Formations are very similar to those of the Upper Riphean assemblage of the Karatau Series of the south Urals and its equivalent from the Polyudov Range, Timan, Kanin Peninsula, and Kildin Island (GOLOVANOV and RAABEN 1967). Thus, on the basis of stromatolites studies, the Hunnberg and Ryssø Formations are assigned to the Upper Riphean. The overlying deposits of the Gorja Series, on the basis of microphytolites are dated as Vendian (MILSTEIN 1967), while the underlying deposits of the Franklinsundet Group, the Flora, Norvik, Raudstup-Sälodd Formations on the basis of macrophytolites and their position in the section are assigned to the Middle- Upper Riphean.

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Variation in a species of «worm» from the Ordovician of Spitsbergen

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A sample from the Valhallfonna Formation (Arenigian-Llanvirnian Age) has produced a large number of enigmatic steinkerns in the 1-mm size range; rare cephalopods, ostracodes, and ?tentaculitids also occur. The steinkerns are smooth and may have filled tubes of calcium carbonate. The hundreds of tubes, presumably an unsorted population, show an incredible degree of variation, ranging from straight through curved and irregularly bent to coiled. Coiling may be irregular, helical, or bilaterally symmetrical. Both bilaterally and helically coiled forms cannot be distinguished from material that other authors have assigned to the Gastropoda, but the continuous variation from coiled to straight tubes demonstrates that these tubes cannot be gastropods.

An extremely difficult problem in paleontology is assignment of fossils to the phylum Mollusca. When one deals with an internal shell filling (steinkern) of small size, it is particularly difficult to find criteria that distinguish mollusks from some other organisms. It is too little appreciated that some groups within the “worms” can construct tubes that mimic the shells of small mollusks.

We thank authorities at the Paleontological Museum, Oslo, for making facilities available and for providing the samples studied. YOCHELSON acknowledges the permission granted him by the Director of the U.S. Geological Survey, for a temporary transfer of headquarters, which allowed him to participate in this research. We are both in debt to TOR MELLEME, Electron Microscopical Unit for Biological Sciences, University of Oslo; his assistance with photographs allowed us to document the great variation in the fossils examined. All photographs were taken with a JEOL JSM-S1, and specimens were uniformly coated with carbon followed by gold-palladium; the apparent shadowing seen on a few photographs is the result of background charging of electrons.

Sample locality, age, and preparation

The material discussed below was collected by DAVID BRUTON, Paleontological museum, Oslo, and RICHARD FORTEY, British Museum of Natural

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History, during the 1971 Paleontological Museum of Oslo Expedition to Northern Ny Friesland, Spitsbergen. A summary of the stratigraphy has been given by FORTEY and BRUTON (1973), the sample that concerns us having been obtained from melt stream “A” on their Figure 1, of the Olenidsletta, Hinlopenstredet. It was collected 17 m above the base of the Profilbekken Member, the upper member of the Valhallfonna Formation; this member is 110 m thick. The sample was fine-grained limestone, medium dark grey, with a tint of brownish grey (N4–5YR 4/1).¹

FORTEY and BRUTON consider the Valhallfonna Formation to be of late Canadian–Whiterockian Age (latest Early Ordovician to earliest Middle Ordovician) and close to the Arenigian–Llanvirnian boundary. 50 m below our sample, conodonts obtained from the middle part of the Olenidsletta Member have been tentatively identified as a form preceeding *Prionodus evae* (LINDSTRÖM). *P. evae* is considered by BERGSTRÖM and COOPER (1973, p. 330) to span the *Didymograptus bifidus* zone, which they judge to extend across the upper Canadian and lower Whiterockian of the American stage sequence; thus the age of our sample probably corresponds to that of the *Didymograptus bifidus* zone.

Two slabs less than 2 cm thick and weighing 314 grams were treated with dilute acetic acid for about one week until all reaction was completed; the slabs were cubed before acid digestion. After the sample was digested in acid, it was washed in distilled water and sieved while wet; no additional chemical preparation was performed. The principal objective of preparation was to investigate ultramicrofossils, but all the insoluble residue was retained.

The insoluble residue amounts to 30.8 grams, indicating a limestone composed of about 90% calcium carbonate. Twenty-two grams of this residue consist of chunks, 2 to 4 cm long, composed of clays and silica particles cemented with silica; it need not be discussed further. The finer residue was fractionated with a saturated solution of zinc chloride. Isolated clay particles smaller than 45 μm were lost during washing.

Coarse fraction

Except for the chunks of matrix noted above, the largest pieces in this fraction are silicified shells of cephalopods as much as 1 cm in size. The silicified cephalopods are light gray, are curved or coiled, and have a rugose surface. The attached larval shell expands abruptly and also bears rugosities; a few of the juvenile forms have been found in the 1–2-mm size range and were identified by the rate of expansion and the rugosities. One example is illustrated in Fig. 6C. The few isolated larval shells are steinkerns and are not silicified.

Echinoderm columnals and fragments of plates are present and form about 25% of this fraction. The cavities in the skeletons are filled with white silica. Smaller fragments of echinoderms are easily identified by this color and by the spongy texture of the infillings. A few large dark-gray balls of silica and small irregular pieces of very light gray drusy silica compose the remainder of this fraction.

¹ Rock-Color Chart. Prepared by the Rock-Color Chart Committee. Distributed by the Geological Society of America. Boulder, Colorado 1970.

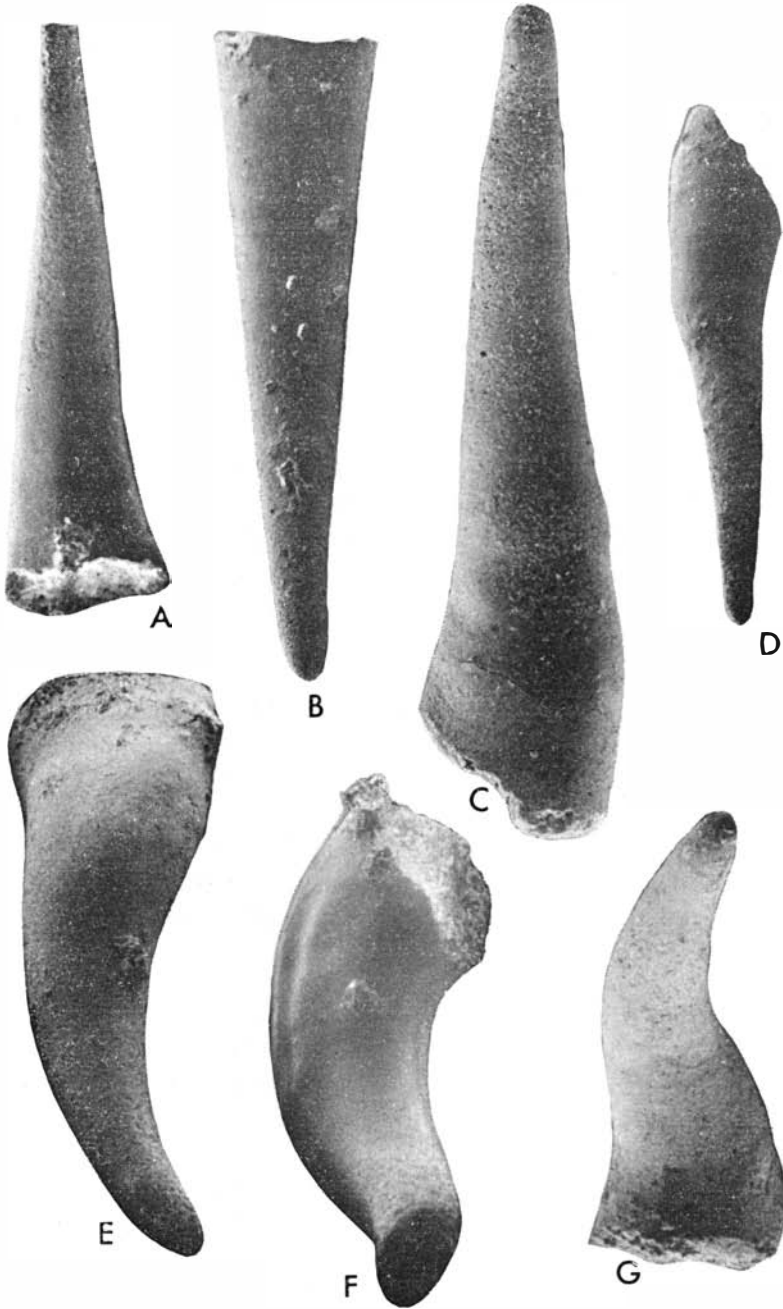


Fig. 1. Steinkerns of straight to curved "worm" tubes from 17 m above the base of the Profilbekken Member. A. PMO NF 3201/1, $\times 50$. B. PMO NF 3201/2, $\times 75$. C. PMO NF 3201/3, $\times 100$. D. PMO NF 3204/9, $\times 75$. E. PMO NF 3202/2, $\times 100$. F. PMO NF 3201/4, $\times 75$. G. PMO 3201/9, $\times 75$.

Fine fraction

Approximately 5.4 grams of fine-sized residue was obtained, most of it in the 1–2-mm size. More than 10% of this size range is organic. Dark-grey silicified ostracodes are present but rare. Trilobite fragments, also silicified, occur even more sporadically. Commonly, thorax fragments are gray, whereas glabella fragments are a rich dark brown. The preponderant forms in this fraction are steinkerns of tubes. Some of the nondescript material of small size could be partial filling of tubes, the smooth outer surface of the steinkern having been broken away.

Only four conodonts of the distacodid type were obtained, although samples from 15 m higher and above have produced a variety of conodonts in abundance. Conodonts have so far not been found in other samples from the lower part of the Profilbekken Member at this locality.

The texture of the steinkerns of the tubes varies from coarse to smooth; commonly both textures are found on the same specimen. The apical part is often coarser and more pitted than that near the aperture (Fig. 3C). Probably the finer texture is a result of easier infilling of matrix toward the aperture after death of the organism. Some of the steinkerns are so smooth that they appear polished (Fig. 1F). Still others, in addition to being smooth, have a bluish sheen on the surface, suggestive of a phosphate mineral.

In the size range below 1 mm, organic-walled fossils including acritarchs, chitinozoans, graptolites, and scolecodonts are absent, although these have been found in samples both above and below the 17 m level. Much of this finer material in our sample seems to consist of silicified echinoderm debris. The color and texture easily distinguish this debris from inorganic silica in sizes as small as 80 μm . Echinoderm debris of such fine size is quite rare in other samples from the Profilbekken Member, although larger pieces of silicified echinoderms do occur.

Steinkern A

We recognize two forms of tubelike steinkerns. The form arbitrarily designated A is characterized by annulations irregularly spaced. Two representative specimens are illustrated in Fig. 6B, E. Tubes are straight to slightly irregular, but are not curved or bent. The annulations are suggestive of those that occur in *Tentaculites* and *Cornulites*; some annulations have a slightly sinuous course around the tube. Commonly, the tip is not preserved in this form of Steinkern. We are unable to assign this material to any genus because it lacks all surface features, but we are reasonably confident of this generalized assignment. About 5% of the steinkerns recovered show annulations.

Steinkern B

We assign the bulk of the fossils obtained to another taxon. These consist of simple elongated tubes that seem to expand at a continuous low rate; no “flaring” of the apical area has been observed.

The extreme tip of the steinkern varies slightly from subpointed in a few specimens (Fig. 1D) to most commonly rounded (Fig. 1B, E). The maximum

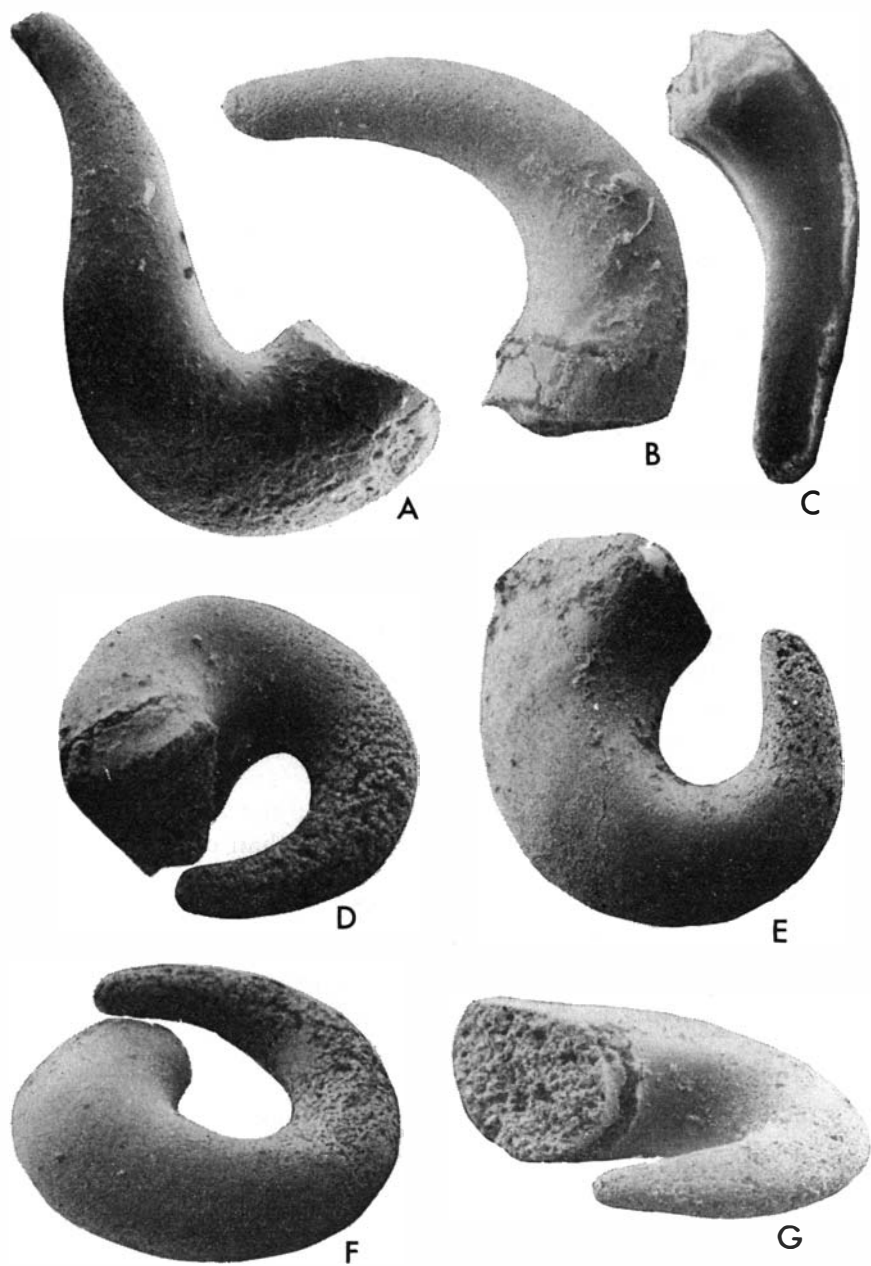


Fig. 2. Steinkerns of curved "worm" tubes from 17 m above the base of the Profilbekken Member.
A. PMO NF 3201/5, $\times 75$. B. PMO NF 3204/8, $\times 100$. C. PMO NF 3201/11, $\times 50$. D. PMO
NF 3202/2, $\times 100$. E. PMO NF 3202/3, $\times 75$. F. PMO NF 3203/5, $\times 100$.
G. PMO NF 3202/6, $\times 100$.

variation in this feature is trivial, and we do not consider it significant. The extreme apical area of a shell or tube is exceedingly thin. Some present-day animals strengthen this area by depositing calcium carbonate on the interior of the shell. Such a deposit may account for the minor variation observed. An equally plausible alternative is that the apex on the tube interior was relatively pointed, but the physics of infilling such a restricted area with tiny particulate matter commonly caused a rounded surface to form.

Beyond the apex, however, variation is remarkable in that a variety of bizarre shapes may be seen in the slightly curved shapes (Figs. 1C-E, 2B, C); some appear irregularly bent (Figs. 1F, 2A). In three-dimensional coiling, the coils may be irregular (Figs. 4E, 5I); one coil may be on top of another so that a cylinder shape is formed (Fig. 5C, D, F-H); or there may be helical coiling (Fig. 5A, E). Oriented with the apex upward, all the examples of helically coiled tubes have the aperture on the right.

Tubes also vary in shape from hooked-specimens (Fig. 2D, E) and those that are coiled in an extremely low helix, so that the “upper surface” is nearly in one plane (Fig. 2F, G), to those that are nearly bilaterally symmetrical, showing only a slight asymmetry of the whorl profile (Fig. 2A, B), and those that appear to be truly bilaterally symmetrical (Fig. 4D). This last form differs very slightly from the others in that a suggestion of a bulbous expansion appears at the tip (Fig. 4F). We have seen this in only a few examples and suspect that it is related to the tight bending required to complete the initial coil in a small restricted area.

Variation in the tubes is continuous from straight to coiled. There appears to be a few more straight to slightly irregular tubes than other “end member” forms illustrated. However, the vast majority of the tubes that deviate markedly from being straight, seem to have no preference for any particular shape. Specimens that appear to be perfectly bilaterally symmetrical are less common than specimens that deviate a small degree from these ideal shapes. Helically coiled specimens having an ideal logarithmic spiral in three dimensions are rare.

Paleoecological speculations

A prime consideration in any discussion of presumed life habits of former living animals is the question of whether material studied has been transported. In this sample, it is safe to assume that little movement has occurred. We base this interpretation on the occurrence of larger specimens along with small tubes. Supplemental support is provided by the delicate nature of larger pieces of silicified echinodermal debris. Perhaps the single most compelling evidence of significance of lack of bottom transport is the large numbers of steinkern B that have the tip preserved.

Because all specimens of steinkern B fall within a limited size range, a meaningful histogram of sizes cannot be made to see whether a survivorship curve can be constructed. We cannot determine whether we are dealing with a true sample of a living population or with an accumulation of organisms that lived and died through the interval of time necessary for the total thickness of the slabs to accumulate. We suggest that the latter situation is closest to the

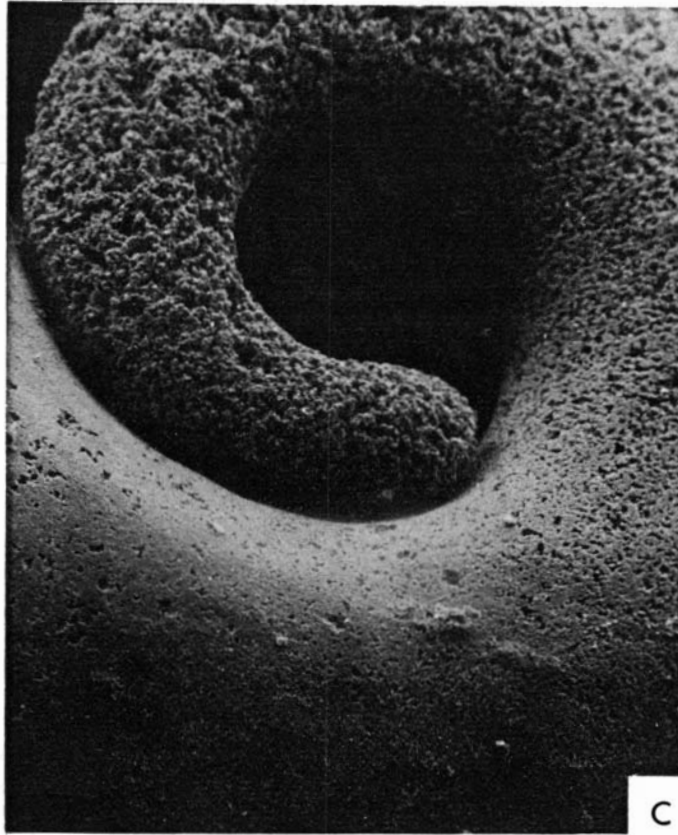
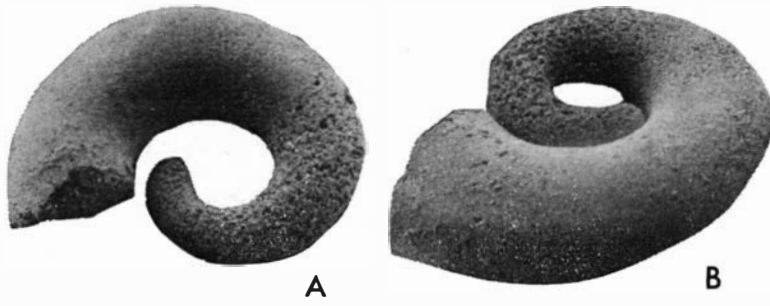


Fig. 3. Steinkerns of curved to coiled "worm" tubes from 17 m above the base of the Profilbekken Member.
A. PMO NF 3203/4, $\times 100$. B. PMO NF 3203/3, $\times 75$. C. Detail of shell of Fig. 3B, $\times 300$.

truth, for even when one deals with bedding-plane accumulations rather than a three-dimensional sample, there is no assurance that specimens dead prior to a sedimentological event have not been included. Nevertheless, even if we assume that the assemblage accumulated for a few years, an exceptionally large number of specimens is present, and it is appropriate to use the term “gregarious” for this accumulation.

None of the specimens show any evidence of attachment, which might be indicated by a flattening on one side; specimens that are essentially straight are nearly radially symmetrical. Likewise, the apex comes to a sharp tip in the better preserved specimens; had there been attachment in this area, the tip might have been flattened or bulbous.

The variety of shapes seen in the steinkerns suggests that there must have been some support during change in shape. Indeed, it is hard to see how some of the specimens that coil in three dimensions could have grown had they not been apex downward and coiling upward. We suggest that these specimens could have lived in or within an algal mat, their shape in part, perhaps, controlled by the reaction of a potentially deforming substrate to the increasing weight of the animals during growth.

There is no evidence from the limestone of algal mat deposition. Our speculation cannot be confirmed, but we find it difficult to arrive at a more satisfactory explanation for the great variety of shapes that occur together. This sample shows differences in faunal content from other Profilbekken Member samples, but we cannot arrive at any firm conclusions as to why these differences occur. Much of the Valhallfonna Formation is interpreted as a slope deposit; perhaps the thin layer containing abundant worm tubes represents a slightly shallower, nonpersistent shelf deposit.

Discussion

Steinkern A, presumably a tentaculitid, provides no important biologic data. It is also of limited stratigraphic utility, as tentaculitids significantly older in the Ordovician have been reported (FISHER and YOUNG 1955).

Steinkern B, by contrast, is remarkable for both its diversity and its abundance. Of the 70 samples from the Valhallfonna Formation studied to date, only this one has produced the wealth of tiny tubes. Although a few other residues have produced similar steinkerns, in all these samples the individuals are much less abundant. The steinkern is judged by us to be the filling of a calcareous tube, soluble in acetic acid. We find it interesting that in this sample, which contains the illustrated “worm” steinkerns in protusion, echinoderms and arthropods are silicified, as well as the larger cephalopods; the isolated larval cephalopod shells were not silicified. It is not uncommon for silicification to be differential among various groups in the same sample, but we think it significant that at least some of the undoubted mollusks have been silicified and these tubes have not. This is at least suggestive of some difference in the organic matrix, mineralogy, or structure which distinguishes the tube from a molluscan shell.

In another sample from this locality, 35.3 m above the base of the Profil-

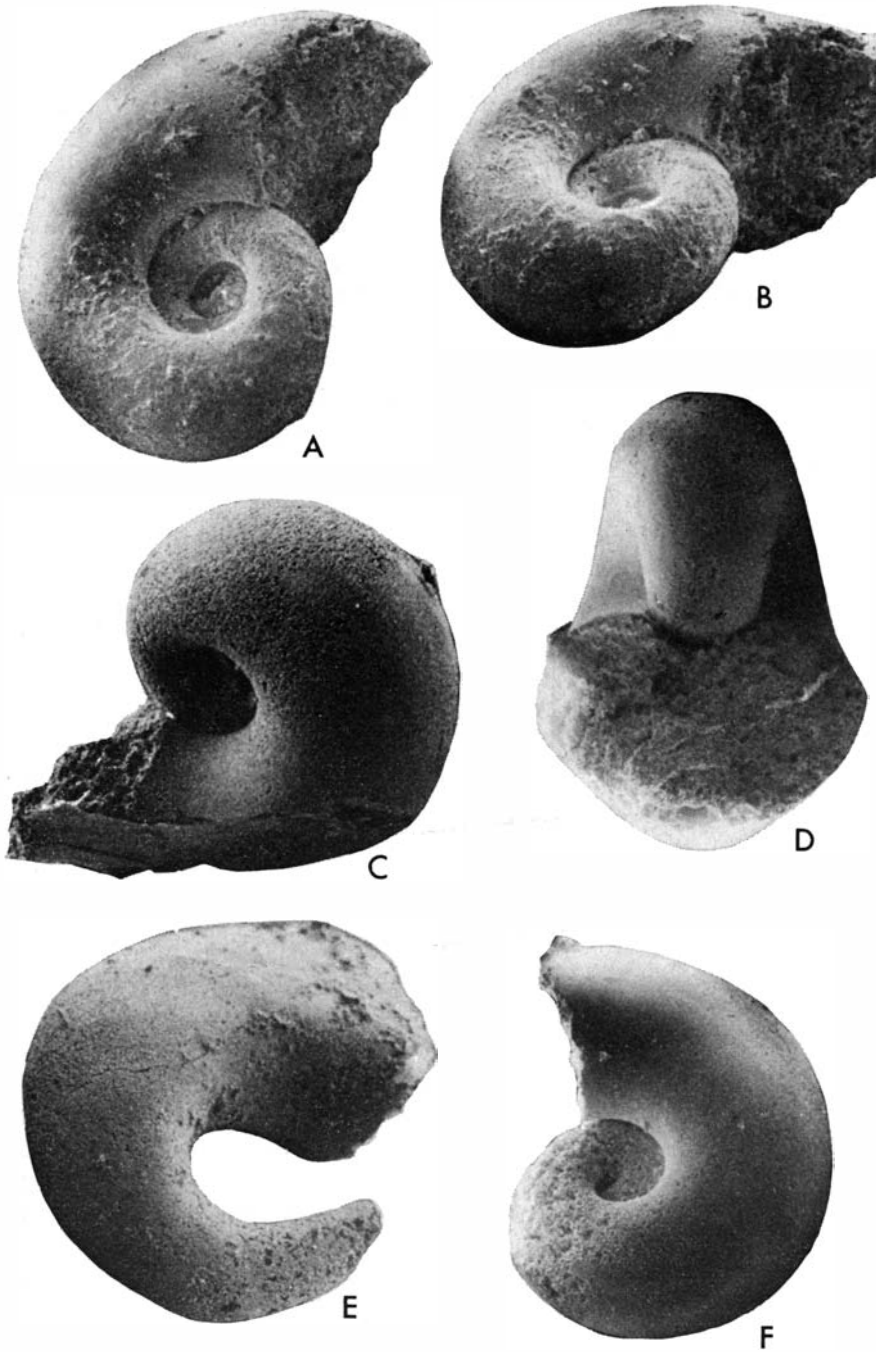


Fig. 4. Steinkerns of curved to bilaterally symmetrical coiled "worm" tubes from 17 m above the base of the Profilbekken Member. A. PMO NF 3203/6, $\times 75$. B. Specimen shown in 4A in an oblique view. C. PMO NF 3202/5, $\times 100$. D. PMO NF 3203/7, $\times 75$. E. Specimen shown in Fig. 2E from slightly different angle, $\times 100$. F. PMO NF 3202/4, $\times 75$.

bekken Member, one hollow specimen has been recovered. It is coiled much like a bellerophortid gastropod (Fig. 6D), though the whorls do not touch, suggesting that it may be a steinkern; the shell is thin and shows no obvious layering (Fig. 6F). It does not appear to be silicified but, rather, has close similarity in texture and color to calcium phosphate shells, rich in organic matter, of inarticulate brachiopods, which also occur in that sample. The specimen may have a true shell, may be silicified, or, less likely, may be composed of a thin phosphatic layer between a shell and more conventional steinkern material. The rarity precludes any chemical test to determine mineralogy. Some steinkerns of tubes are composed of glauconite. Unlike the lower sample, other faunal elements are both diverse and abundant. Ostracodes occur both as glauconitic steinkerns and silicified valves.

A third sample, 38.8 m above the base of the section, has produced a single straight tube (Fig. 6A), which is hollow, and a few steinkerns of tubes. Glauconite is not so common in this sample as in that immediately underlying, but the fauna of the insoluble residue is similar. Except for these two other samples, “worm” tubes are exceedingly rare in insoluble residues from the Valhallfonna Formation.

We are unable to provide any high-level classification of steinkern B. Although it might be a coleolid (FISHER 1966, p. W133), it might equally well fall within the fossil sedentary Annelida (HOWELL 1966, W155), or might belong in yet a third higher taxon. The Paleozoic worms are poorly known. We also choose not to assign any generic or specific name to the material, for an additional name not rigorously defined will simply add further confusion to the Paleozoic “worms”.

The steinkerns provide no information about the tubes exterior, although we suspect that the exterior may have been relatively smooth because of the absence of any elongated ridges impressed on the steinkerns. Several steinkerns do show the impression of a broad ridge on one side of the tube (Figs. 1F, 2C), but we are unable to separate them otherwise from the majority of the tubes. We prefer to regard this as an irregularity in outline of aperture of a few individuals rather than as a specific character on an indication of sexual dimorphism.

The incredible degree of variation in shape suggests that extreme caution be used when one is faced with a sample of a single or a few “worm” tubes, especially when the tube surface is smooth and principal characters are to be derived from the shape.

Perhaps, equally important is the homeomorphy of these tubes to tiny, helically coiled and bilaterally symmetrical gastropods. Were it not for the fact that all intermediate shapes leading to the straight tubes occur, we would have assigned those shapes to the gastropods without question. It is one of the advantages of mollusks that all growth stages are preserved during life. This study indicates that it is a more cautious procedure to observe the early growth stages of larger specimens, which are undoubted gastropods, than to simply examine isolated, tiny coiled specimens, for the assumption that all such coiled shells are larval gastropods is not correct.

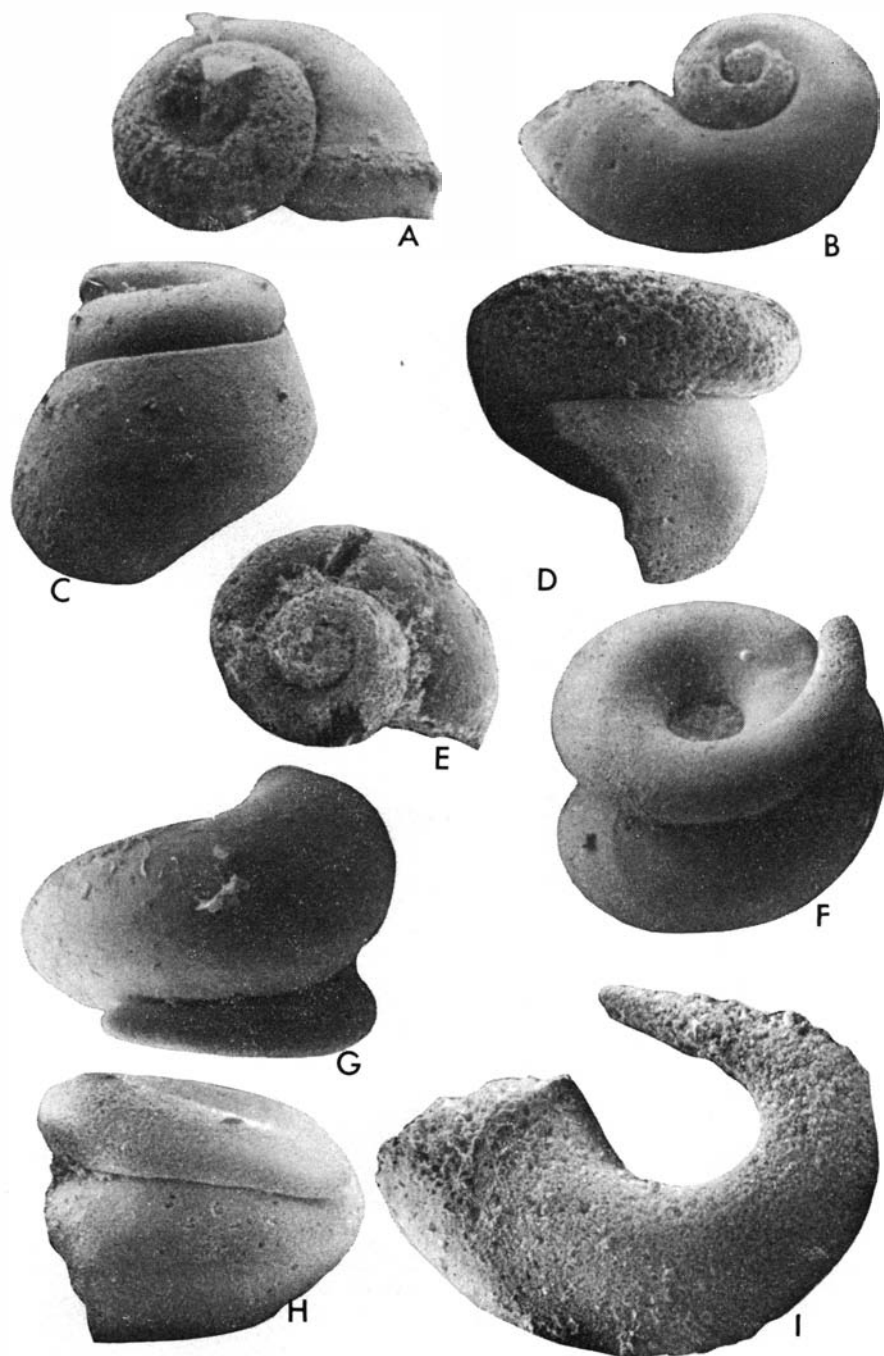


Fig. 5. Steinkerns of curved to helical coiled "worm" tubes from 17 m above the base of the Profilbekken Member. A. PMO NF 3204/1, $\times 100$. B. PMO NF 3204/2, $\times 100$. C. PMO NF 3204/3, $\times 100$. D. PMO NF 3203/1, $\times 100$. E. PMO NF 3201/8, $\times 100$. F. PMO NF 3204/5, $\times 75$. G. PMO NF 3204/4, $\times 75$. H. Specimen shown in 5D rotated about 90° , $\times 100$. I. PMO NF 3202/1, $\times 75$.

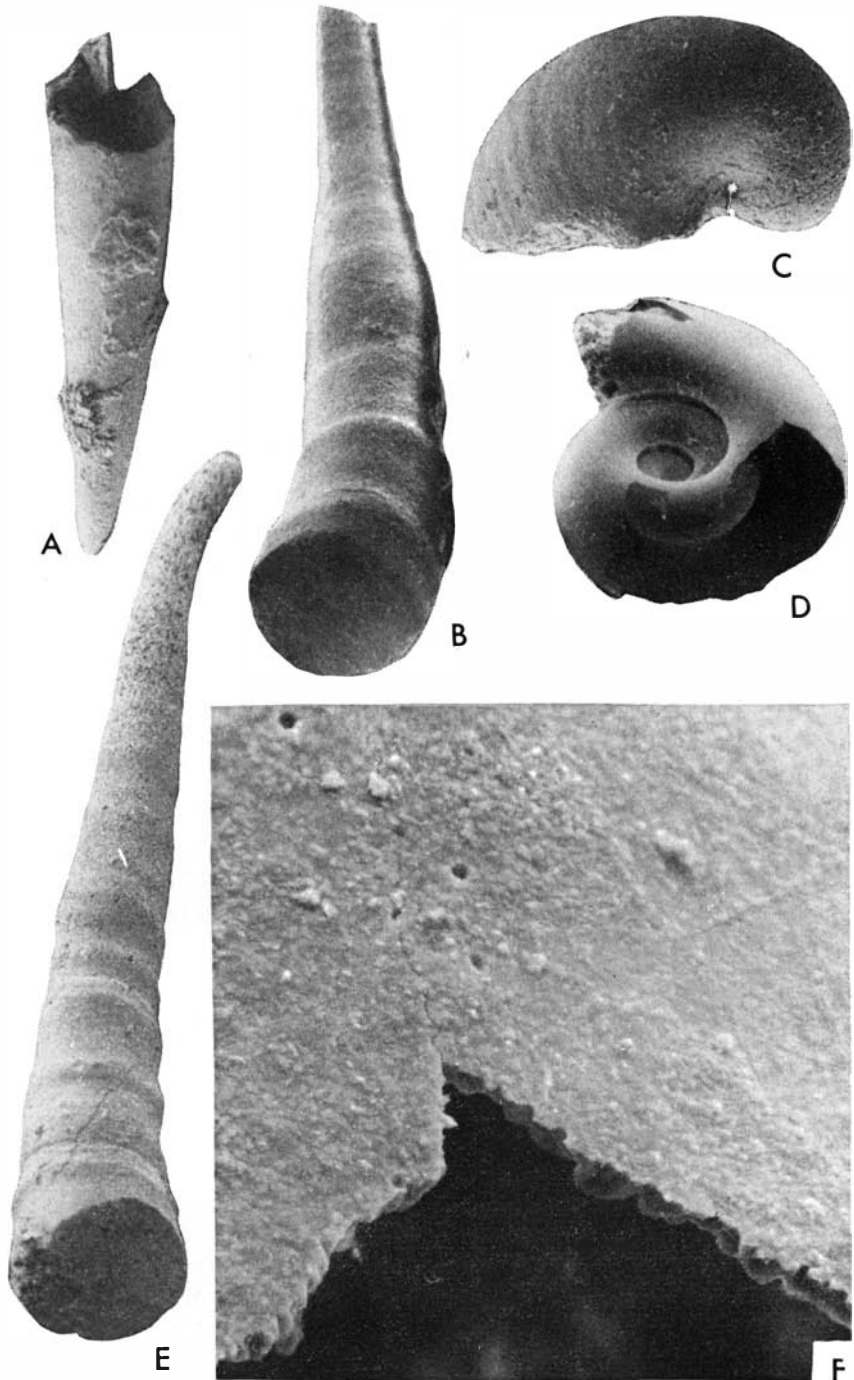


Fig. 6. A. Straight, hollow "worm" tube from 38.8 m above the base of the Profilbekken Member. PMO NF 3204/6, $\times 150$. B. E. Annulated steinkerns (B) from 17 m above the base of the Profilbekken Member. B. PMO NF 3201/6, $\times 75$. E. PMO NF 3204/7, $\times 75$. C. Cephalopod steinkern from 17 m above the base of the Profilbekken Member. PMO NF 3201/10, $\times 100$. D. F. Coiled hollow "worm" tube from 35.3 m above the base of the Profilbekken Member. D. PMO NF 3201/7, $\times 100$. F. Detail of shell of Fig. 6D, $\times 3000$.

The homeomorphy between worm tubes and small gastropods has to be kept in mind. It is especially important when dealing with steinkerns, where even more of the limited information available to the paleontologist for assignment of fossils is lost. To cite one example bearing on bilateral symmetrical forms, EISENACK (1966, Pl. 20, Fig. 4) illustrated one from a Silurian phosphatic residue as a gastropod. It is comparable with our Fig. 4D. We judge that this form illustrated by EISENACK is as plausible a “worm” tube as it is a gastropod.

To cite another example, involving helical forms, MISSARZHEVSKIJ (1969) described the genus *Aldanella* as a gastropod from the lowermost Cambrian (Tommotian) of Siberia. The type specimen of the type species is a phosphatic steinkern. RUNNEGAR and POJETA (1974, p. 313–314) accepted this assignment to the gastropods and used it as one of their points in reconstructing the early phylogenetic history of the mollusks. Yet in size and shape, *Aldanella* and similar enigmatic, small coiled forms that are known only from steinkerns should not be assigned to the mollusks, for there is no way to prove that they are mollusks, and our evidence from Fig. 5 shows that they could fit easily into the “worms”.

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The tectonic structure of the Devonian Graben (Spitsbergen)

By YU. P. BUROV and D.V. SEMEVSKIJ

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Abstract

A description is given of the tectonic structure of the Devonian Graben (island of Spitsbergen) and the history of its development. It is stated that the sedimentation within its confines has been controlled by fractures of deep location, and the field of development of Devonian deposits at the period of sedimentation was essentially larger than its present dimensions. Main plicative and disjunctive structures have been recognized, the latter being united into differently aged erathems (groups). The present look of the graben has to a considerable extent been conditioned by vertical movements.

Description

Study of the Devonian Graben of Spitsbergen is all important because of its key position as the extreme north-westernmost projection of Eurasian continental structures, and the variety of its geological formations concentrated within a relatively small and accessible area.

Devonian rocks occur in northern Spitsbergen as a thick variegated molasse sequence with a basaltic conglomerate at the base. The Devonian sequence rests on the Caledonian basement with a sharp unconformity and is of great interest both for correlation of not well known coeval sequences in the Arctic and their implication for the geological history of Spitsbergen before its platform development, starting early in the Carboniferous.

The study of Devonian deposits of Spitsbergen was initiated by Swedish expeditions under the leadership of NORDENSKIÖLD, NATHORST, and DE GEER between 1868 and 1899 and has continued up to the present. Studies included the construction of a stratigraphical standard and environmental reconstructions of the Devonian deposits. A synthesis of results may be found in NATHORST (1910), FØYN and HEINTZ (1943), and FRIEND (1961).

Since 1964 Spitsbergen expeditions of the Institute for the Geology of the Arctic (Leningrad) have studied the Devonian deposits and carried out geological mapping (generally, at a scale of 1:200 000) in the Liefdefjorden, Bockfjorden, Woodfjorden, Wijdefjorden, and Austfjorden areas, as well as in the regions of Raudfjorden, Mimerdalen, and Hornsund. Almost all of Andrée Land and most of Dickson Land have been mapped.

Fold and fault structures revealed by mapping within the Devonian graben aid the reconstruction of the general history of the area in the postgeosynclinal period and subsequently during the development of its present day form.

The Devonian Graben is exposed in central Spitsbergen to the north of Isfjorden. Its history is closely related to fundamental faulting, particularly with faults bordering the graben to the west and the east. The major displacements seem to have been initiated before or during the early stages of the Caledonian orogeny. Movement along existing fault planes took place later, particularly during the Devonian, influencing the environment of deposition. This reflects in non-depositions, distribution of facies, thicknesses, and lithologies of Devonian deposits. In Upper Devonian and post-Devonian time block movements affected the mode of structure formation. In the Jurassic–Triassic the displacements along major fault planes were accompanied by trappean magmatic activity.

Thus, the history of the Devonian Graben is in fact that of the block movements.

The modern graben is distinctly divided by a north-south trending submeridional fault (Breibogen to Bockfjorden) into two parts: the western inner horst and the eastern deep depression.

The western boundary of the inner horst coincides with the western boundary of the Devonian Graben, probably representing the major fault zone, and occurs along the whole western coast of Spitsbergen. The lowest beds of the Devonian, the Lower Devonian Siktetjellet and Red Bay Groups with a total thickness of the order of 3000 m, rest unconformably on the eroded surface of the Caledonian basement within the inner horst. The deposits consist mainly of rhythmically alternating red sandstones and siltstones of lagoonal and non-marine origin together with a 600 m sequence of coarse conglomerates containing some beds of coarse-grained sandstones at the base of the section. The highest beds of the succession are found in the northern end of the horst in the Raudfjorden area; to the south successively lower horizons of the succession are exposed. Some relics of Red Bay conglomerates reach the latitude of Kongsfjorden. Faults, generally trending north or northwest, divide the inner horst into the blocks of varying size and with vertical uplift from some hundreds of metres to at least one kilometre. Within these blocks the Devonian deposits dip steeply to the west ($60\text{--}80^\circ$) while in the southernmost inner horst (south of 79°N) they dip mainly to the south. In the vicinity of normal and strike-slip faults the Devonian rocks are folded to form local syn- and anti-forms which are steep-limbed, usually with small amplitudes, and near-fault in character. Their axes are generally subparallel with the fault trends. These structures do not usually exceed 300–500 m with amplitudes amounting to

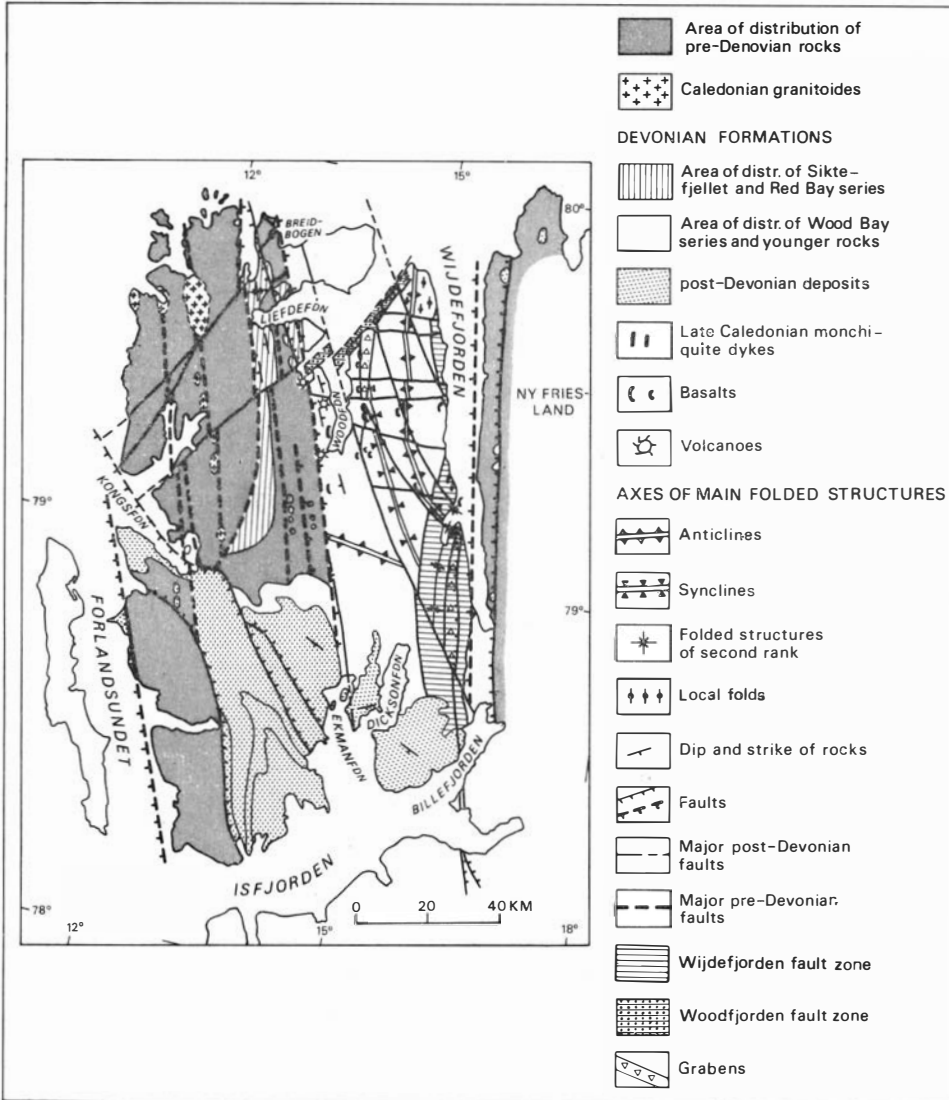


Fig. 1. Tectonic map.

some dozens of metres. The most significant one is the synform in the mountain Ben Nevis area between Raudfjorden and Liefdefjorden. This is an asymmetrical fold with its axis trending north-west. The structure attains 4 km in width, and the maximum depth of subsidence is probably not more than 500 m. The eastern limb dips at angles of 25° to 30° while the western limb dips at 15° to 20° although the dip increases westward as the western limb approaches the major submeridional faults bordering the structure. The synform is complicated by a fault pattern with a throw of the order of 400–500 m.

The eastern part of the inner horst, formed by northward radiating major faults (the Rivieratoppen-Bockfjorden line to the west and the Breibøgen-Ekmanfjorden line to the east), represents the most elevated terrain of the

inner horst and has a structurally complex interior. Mainly pre-Devonian rocks are exposed there, but Lower Devonian deposits belonging to the lowest Red Bay Group occur and generally dip to the west at angles of 10° to 35° .

The eastern part of the graben (to the east from the Breibogen–Ekmanfjorden fault) contains younger Devonian groups. These include the Lower Devonian deposits of the Wood Bay Group, with the Kapp Kjeldsen Formation in the Woodfjorden area and the “Austfjorden Sandstone” at the base in eastern Dickson Land. Devonian deposits in the eastern part of the graben are represented by all three groups including variegated and grey-coloured lagoonal — non-marine, coastal — marine and partly non-marine sediments. Upper Devonian deposits of the Mimerdalen Group (Esteriahaugen Formation, Fiskekløfta Formation and Planteryggen Formation) outcrop only in a narrow zone in easternmost Dickson Land. This zone was affected by intensive block faulting in post-Devonian time. The great majority of investigators suggest that the Upper Devonian in the Mimerdalen area rests on the eroded surface of the Lower-Middle Devonian, probably with an angular, unconformity. The total thickness of the Lower-Middle Devonian deposits is evaluated at 4.5 km in the eastern part of the graben, whereas the thickness of the Upper Devonian deposits of Andrée Land and Dickson Land is evaluated at 650 m. Devonian deposits are characterized by cyclic sedimentation on a large formational scale and also by smaller scale rhythms grade as a rule from gravelites or interformational conglomerates, interbedded with coarse-grained sandstones, to sandy siltstones and claystones. In the Lower-Middle Devonian sequence as a whole the grain size as well as the order of cycles, increases south-eastwards from the head of Woodfjorden to the head of Vestfjorden.

The eastern deep depression is sharply divided into two areas by a saddle-shaped elevation not far to the south from the latitude of Vestfjorden. The northern area contains a complex monoclinial structure dipping to the north-east, in the southern area the rocks dip to the south and south-east. The major meridional fault zone associated with Wijdefjorden is a natural boundary for these two areas at the present erosion level. Within this zone the Devonian rocks are strongly deformed and folded by many near-fault diastrophic movements.

The Andrée Land anticline and adjacent syncline to the west are the largest known fold features complicating the northern part of the monoclinial structure. The fold axis of the Andrée Land anticline trends to the north-north-west. The structure can be traced for 70 km from Gråhukken in the north through the head of Purpurdalen to Kapp Petermann to the south, where it “fades” in the zone of strong folding and faulting near Wijdefjorden. The lowest parts of the Wood Bay Group, Kapp Kjeldsen Formation, and Keltiefjellet Formation outcrop near the axis of the structure, whereas the younger Wood Bay Group and Grey Hoek Group are exposed on its limbs. To the north and south of Kartdalen the anticline plunges in north-western and south-eastern directions, respectively. The limb span amounts to not more than 20 km and the vertical uplift of the crest is at least 1000–1200 m. The structure has an asymmetrical form, the eastern limb dipping at angles of 12° to 15° increasing toward Wijdefjorden to 20° and 25° . Angles of dip of the western limb usually

do not exceed 15° . A succession of axis offsets to the west due to a series of east-west trending faults may be observed. A number of north-eastward trending faults extending for tens of kilometres and controlling the basalts in Andrée Land are subparallel with the axis of this structure.

This major anticlinal feature is adjacent to the Andrée Land syncline to the west. This syncline has been traced from Jakobsenbukta to the latitude of Høegdalen, over 60 km, and is 7 to 11 km in width. Its axis is subparallel with the axis of the anticline, and similar offsets westward due to sublatitudinal faults are developed. The syncline is possibly closed at the head of the Vestfjorddalen, south of which the Devonian rocks plunge stably and gently southwestward. On the western limb of the syncline the dip angles are $8\text{--}10^\circ$ gradually decreasing towards Woodfjorden to $5\text{--}6^\circ$. Further to the west a monoclinical plunge of Devonian rocks at the angles of 4° to 5° may be demonstrated throughout the Woodfjorden area. The eastern limb dips with angles of 10° to 15° . In the core of the syncline the grey-coloured deposits of Tavlefjellet Formation and Forkdalen Formation outcrop, whereas on the limbs red beds of Keltiefjellet Formation and Stjørdalen Formation occur. The depth of subsidence reaches 1000 m in the Burfjellet area increasing northward as the axis plunges to the north.

The major structures above are in turn complicated by a number of small-scale folds and near-faults. The most important of them are the Prinstoppen graben-anticline (north-western Andrée Land), the Vestfjorden syncline and Gråkammen graben (south-eastern Andrée Land).

The Prinstoppen graben-anticline is formed by two northward radiating faults with a throw of 300–450 metres. The structure extends about 12 kilometres. This graben contains grey-coloured deposits of the Forkdalen Formation which compose an anticlinal feature complicated by small, steep folds and faults with offsets of some tens of metres. The folds are distinct near-fault in character. Their axes trend submeridionally, the limb span varying from 100 to 500 m, and ascending and descending displacements of the limbs being 30–50 m. The folds seem to be limited to the graben. The faults bordering the graben dip at the angles up to 65° .

The Vestfjorden syncline strikes almost meridionally from Krosspynten to the latitude of Høegdalen, for more than 20 km. Morphologically, the syncline is expressed by low relief: Vestfjorden, a bay, and Vestfjorddalen, its south tributary. The structure is 4 to 5 km in width. Its axis coincides with a fault striking meridionally along Vestfjorden. The elongation of the structure (1:4 to 1:5) indicates its genetic association with the faults. Sublatitudinal and northwestern faults successively displace its axis to the west. The beds of the east limb dip more steeply with angles up to 20° and 30° . In the core the “Dicksonfjorden Sandstone” outcrops, whereas on the limbs the “Austfjorden Sandstone” and the lower part of the “Dicksonfjorden Sandstone” are exposed. The depth of subsidence reaches 500 to 600 m.

To the east of the Vestfjorden syncline Gråkammen graben, extending north-south for about 20 km, occurs, its northern end (Kapp Petermann area) and southern end (Høegdalen area) being 1.5 and 4 km in width, respectively.

It is bordered by the major Wijdefjorden fault zone to the east and by the Vestfjorden syncline to the west. It contains red beds and grey-coloured deposits of Wood Bay Group and Grey Hoek Group folded and complicated by faults. The folds are distinctively near-fault in character. Their axes usually have an east-west strike. The limb span ranges from 100 to 500 m and the amplitudes vary from 50 to 100 m. Dips of 10–15° occur on the limbs, increasing up to 25° and 30° in the vicinity of faults.

Many small ups and downs structural features have been observed throughout the region. They are closely associated with major faults. These features are as a rule small steep folds ranging from a few hundred metres to some kilometres in length. Their axes are subparallel with faults situated near the folds. The strike of the axial plane is north-south or north-east-south-west, parallel to the main fault directions in the area. The folding is most intense in the Wijdefjorden fault zone (many folds with limb span of 100 to 300 m and dip angles from 30° to 50°). The intensity of folding decreases westward from this zone.

Devonian deposits outcropping between the “saddle” and the heads of Dickson- and Ekmanfjorden are in general slightly deformed. A constant (monoclinial) plunge to the south and south-west at the angles of 3–4° is typical in this area. Only a few important faults striking north or north-west with displacements of 500 to 700 m have been recognized in this area. These complicate the monocline, and folds are developed in nearby Devonian rocks. Axial planes of the folds are subparallel with the strike of the faults; the dip angles do not exceed 15° to 25° on the limbs. Approaching the Wijdefjorden fault zone, the intensity of folding increases gradually with a simultaneous increase in faulting.

Classification

The Devonian Graben is multiple faulted with a predominance of steep, normal faults. Most investigators suggest that they are connected with the terminal episodes of the Caledonian orogeny. Generally, the fold structures and flexures are related to the faulting. Tectonism also occur in post-Devonian time. The absence of post-Devonian rocks in most of the region hampers a chronological classification of the known faults. However, observations on relationship between faults, data on their orientation, morphological features, position within the graben and facies and thickness analysis may be combined to divide faults into pre- and post-Devonian.

The pre-Devonian faults are represented by the major faults; movements along these planes determined all the general features of the present day block structure of the graben. These faults strike north-south throughout northern Spitsbergen and probably further south where they are buried under younger sequences. During their prolonged evolution the major faults were reactivated both during the Devonian time and in the later epochs. Among these, two major faults, complicating the western part of the graben, may be recognized; the fault along the western coast of Raudfjorden with a vertical amplitude of not less than 1000 m in the Liefdefjorden area and the Breibogen-Ekmanfjorden fault with the amplitude of about 1500 m in the Bockfjorden area. Both

faults are approximately north-south trending and have an axial plane dipping to the east at an angle of 70° to 80° . Shear and mylonization zones associated with these faults are up to 300 m in thickness. Some eastern normal faults and strike-slip faults having a north-eastern strike and belonging to the Wijdefjorden fault zone, may also have originated before the Devonian. The most intense movements associated with these faults and the Breibogen–Ekmanfjorden fault, occurred in the Lower Devonian after the formation of the Red Bay Group within the inner horst.

The steeply dipping faults of post-Devonian origin, widespread within the Devonian Graben, generally have small vertical amplitudes (100 to 500 m) and represent a single set of tectonic deformations related to block movements of the Earth's crust. The main Devonian field is cut up by these faults into many block fragments. The great majority of them formed during the initial stages of platform evolution in this region. The most intense block movements took place during Lower Carboniferous, although along most major faults the movements were reactivated in the later epochs too.

Among the faults of post-Devonian origin, four groups may be recognized (from older to relatively younger): faults trending north-east, submeridional (north-south), sublatitudinal (east-west), and north-west.

A relatively small group of steeply dipping, almost vertical normal faults trending between 30° and 60° NE occur throughout the graben, are 10 to 15 km long, and have variable vertical displacements of 100 to 800 m. They are difficult to see, and only the most important coincide with morphological features such as valley directions. The lack of hydrothermal alteration of the brecciated material is characteristic and the zone of brecciated or intensively fissured rocks hardly exceeds 20 to 30 m. The entire absence of fold structures near the fault zone is also typical.

Normal and strike-slip faults trending from 350° to 20° are grouped as submeridional faults. These displace the north-eastward trending faults, the most significant movements along their planes probably taking place during the Middle Carboniferous (?Bashkirian age). In Andrée Land and Dickson Land the submeridional faults form two major zones related to the eastern coast of Woodfjorden and the western coast of Wijdefjorden and Austfjorden, respectively, the latter being the wider.

Most of the faults belonging to the Woodfjorden zone are normal faults. The exposed fault planes are either vertical or they dip westward at an angle of 70° to 80° . The vertical displacement ranges from tens to hundreds of metres, the eastern block as a rule being downthrown. The thickness of the zone of intensively deformed and fissured rocks is 2 to 5 km.

The fault zone related to Wijdefjorden and Austfjorden consists of a basal strike-slip fault subzone 2 to 5 km wide (to the east) and a shearing-linear fold subzone 12 to 15 km wide (to the west). They are separated by a major strike-slip fault which extends along the eastern coast of Andrée Land, through Petermannfjellet and further southwards up to the head of Mimerdalen. To the south of Kapp Petermann the vertical displacement is about 1200 m, the eastern limb being upthrown.

The strike-slip fault subzone is characterized by curved fault planes dipping westward at angles of 50 to 60° and by vertical displacements of one or more kilometres. Zones of quartzitized and calcitized tectonic breccia 30 to 100 m thick are related to these faults. Within this subzone Devonian rocks are strongly deformed and folded. Middle Carboniferous monzonite dykes are associated with these major submeridional faults.

The shearing-linear fold subzone contains many normal faults with vertical displacements of 100 to 300 m. Narrow elongated folds, fault-adjacent in character, are associated with them. The subzone extends from the head of Mimerdalen northward up to Gråhukuken, the most significant normal fault extending for 15 to 30 km, the eastern limb usually being downthrown. The shear zones contain quartz-calcite veins 0.5 to 3.0 m thick. On the peninsula between Vestfjorden and Austfjorden the subzones are separated by a meridional graben, the Gråkammen Graben, extending for 20 km and being between 1.5 and 4.0 km wide.

The group of sublatitudinal faults consists of a normal fault series which intersect the Devonian Graben from west to east and displace the other tectonic features except the north-west faults. Many of them coincide with the directions of the widest valleys. Fault planes dip northward at an angle of 65° to 70°, the northern block usually being downthrown. The vertical displacements generally do not exceed 300 to 400 m. The absence of low-temperature mineralization in the fault zone is typical for this group. The most significant faults are the latitudinal Jakobsenbukta fault and those coinciding with the valleys of Verdalen, Stjørdalen, Purpurdalen, and some others.

The north-western fault group contains tectonic displacement trends of 320° to 340° and intersects all the other faults. Most of them are tens of kilometres long and have relatively small vertical displacements (up to 400 m). Faults of this group are known throughout the graben, and are especially widespread in its eastern deep subsidence structure. They complicate the near-crest part of the Andrée Land anticline, and displace the Devonian rocks, forming another major structure of the graben, the Andrée Land syncline. Shear and mylonitization zones, folds, and sharp deviation in dip angles and directions of the rocks are associated with these faults. The zones of brecciated rocks do not usually exceed 20 to 30 m in thickness and as a rule show only slight hydrothermal alteration. The major faults of this group control the distribution of basalt rocks in Andrée Land. The most significant representative of this group is the normal fault extending from the mouth of Junkerdalen in Woodfjorden through the head of Purpurdalen and Kartdalen up to the western coast of Austfjorden. It extends for about 50 km. The fault trends 330° and has a vertical displacement of about 300 m, the north-eastern block being upthrown. Most of the known basalts in Andrée Land run parallel to this fault direction.

Analysis of the data shows that the initial size of a graben-like subsidence may have been significantly larger than the present Devonian Graben. This is indicated by the presence of the same facies of Devonian rocks both in the Kongsfjorden–Hornsund area and in the present-day Devonian Graben. The

western boundary of the graben is probably a northern extension of the western bordering fault zone originating early in the Caledonian orogeny. The position of the eastern boundary of the early graben is not clear. In our opinion this boundary was located to the east of the modern graben boundary, in the Ny Friesland area, or maybe in the Edgeøya and Barentsøya region. This is based on the presence of marine deposits in the Wijde Bay Group in the north-easternmost part of the graben, the zone of their distribution being open eastward. Significant ancient tectonism in the Ny Friesland area is indicated by relics of Lower Carboniferous rocks which occur in the western part of the peninsula. Generally, the presence of Devonian deposits in the area is not to be ruled out.

Geological History

A geological history of the Devonian Graben may be reconstructed from the data. Block movements along faults and orogenesis gave rise to a grabenlike subsidence structure trending north-east, which was superimposed on the Caledonian structures. This subsidence structure was situated between the Spitsbergen west coast high and the Nordaustlandet high (or anticlinoria) and may have extended southwards to Barentsøya and Edgeøya. Its original extent is unknown, but it is possible that it was bordered by the Bjørnøya–Hopen high to the south.

The initial subsidence structure differs from its present form. The western region (to the west of the Breibogen–Ekmanfjorden fault) was deepest, and during Gedinnian, the Siktefjellet and Red Bay Groups were deposited there. Block faulting with slight folding occurred later in this area. The eastern part of the structure was not prominent during the Gedinnian, and sedimentation was slow or absent. Late in the Gedinnian block faulting along the Breibogen–Ekmanfjorden fault and partly along the Wijdefjorden fault zone within the initial graben gave rise to the formation of a vast, deep trough which may have extended eastward to Nordaustlandet. The southern part of Andrée Land and Ny Friesland may represent an area of lower subsidence in the newly formed trough. This is reflected in progressive psammitization of the Wood Bay Group in a south-eastern direction. Steady subsidence with a centre in the eastern part of Wijdefjorden, may have continued to Lower Carboniferous. During that time the trough was filled up by a thick sequence including Lower, Middle, and Upper Devonian formations of lagoonal-non-marine, coastal-marine, and non-marine facies. During the final stage of orogenesis (“Svalbard phase of folding”) movement along fault planes was renewed, giving rise to most of the existing faults and producing the block structure of the graben in its present form. The eastern termination was caused by later uplift of the Ny Friesland horst where Devonian deposits were eroded. The Andrée Land anticline formed, together with an adjacent syncline in the west and possibly in the east, along the graben axis and the bordering fault planes.

The platform development of the area was characterized by predominantly continental sedimentation. In the Middle Carboniferous (?Bashkirian) intense block faulting occurred along the ancient Wijdefjorden–Kvalvågen

tectonic lineament together with the formation of most of the submeridional faults. Monchikite dykes dated at 309 ± 5 m.y. (K-Ar age) are associated with these faults.

General uplift and block movements continued to and into the Early Cretaceous, but short-term subsidence may have occurred intermittently. Around Early Cretaceous, the sublatitudinal fault system formed, and the region of Andrée Land may represent a mountainous area with rough topography. In the Late Cretaceous, tectonic movements steadied, and the territory was intensively eroded, eventually giving rise to a vast (?Late Cretaceous) peneplain.

At the Late Cretaceous–Paleogene, boundary block movements along the graben border faults were renewed, which formed a north-eastern fault system parallel to the former and controlling the basalt distribution in Andrée Land. The basalts flooded over a smooth surface dipping gently to the north and mostly submerged below sea level.

It is probable that during Paleogene and early Neogene the region was peneplained. Only in the Pliocene were the intensive tectonic movements initiated in Spitsbergen. North Spitsbergen may have been lifted uniformly considering the hypsometric uniformity of the raised peneplain surface. Reactivation of the bordering faults especially conspicuous in Central Spitsbergen, occurred simultaneously.

The final tectonic episode was characterized by uplift complicated by displacements along rejuvenated faults, with associated volcanic activity. Short-term subsidence intervened.

Glacial isostatic rebound caused an uplift of 45 to 50 m over the last 5000 years.

Thus, the present boundaries of the Devonian Graben have been determined by vertical displacements along rejuvenated faults both of Devonian origin and of post-Devonian age. The most important were the submeridional faults (graben-bordering faults) and the Pretender fault zone with a NW strike.

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Stratigraphic subdivision of the Devonian deposits of Spitsbergen

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Abstract

Devonian deposits of the north-western part of the island of Spitsbergen are dealt with in the paper. A new stratigraphic scheme is proposed, based on a formational division of earlier recognized successions and formations. A brief lithological and palaeontological characterization of the formations is presented, and their interaction and area distribution is given.

Introduction

Devonian grey-red continental and partly marine deposits totalling 8000 m occur within an extensive graben in northern Spitsbergen. Local outcrops of Devonian rocks are also known in the Hornsund area. The Devonian of northern Spitsbergen is subdivided into six groups: Siktetjellet (GEE and MOODY-STUART 1966), Red Bay, Wood Bay, Grey Hoek, Wijde Bay, and Mimerdalen (HOLTEDAHL 1914; VOGT 1938). Four of these groups were subdivided into formations.

At present, due to studies carried out by geologists from different countries, the Devonian section in Spitsbergen is rather well known.

Soviet geologists JU. P. BUROV, L. G. MURAŠOV, JU. I. MOKIN, A. I. PANOV) began to study the Devonian deposits of Spitsbergen during 1964. The studies and surveys carried out provided rich material allowing refinement of the existing stratigraphic section, and the recognition of a number of new formations. Fossils collected also enable more accurate dating of stratigraphic units.

On the basis of the material collected and the data available, a standard stratigraphic section of the Devonian deposits for northern Spitsbergen was compiled and is presented below.

Lower Devonian

GEDINNIAN

The Gedinnian continental, red-grey deposits are subdivided into two groups — Siktefjellet and Red Bay — occurring only in north-western Spitsbergen and in a narrow horst parallel to the main basin, adding complexity to the Devonian graben in the west.

Siktefjellet Group

The Siktefjellet Group was recognized in 1966 by GEE and MOODY-STUART in the Raudfjorden and Liefdefjorden areas and subdivided into the Lilljeborgfjellet and Siktefjellet Formations.

Lilljeborgfjellet Formation. These deposits extend in a narrow band from Rabotdalen in Raudfjorden to Siktefjellet in Liefdefjorden. They rest with a sharp angular unconformity on metamorphosed Upper Proterozoic rocks. This contact is evident on Siktefjellet, Högeloftet, Fränkelryggen, Lilljeborgfjellet, and in Rabotdalen. The conglomerates are overlain by the Siktefjellet sandstone.

The Formation consists of grey fine to medium pebble conglomerates with well rounded pebbles of schist, granites, gneiss, migmatite, quartzite and less common metamorphosed limestones. The matrix is grey coarse-grained polymict sandstone and gritstone containing calcareous material.

The greatest thickness (400 m) was measured on Lilljeborgfjellet. Northward and southward the thickness decreases to 100 m and less.

Siktefjellet Formation. The deposits on Siktefjellet in northern Liefdefjorden also extend in a narrow band southward to Bockfjorden. They rest without evident angular unconformity on conglomerates of the underlying formation and are overlain erosively by the Red Bay conglomerates.

The formation is represented mainly by fine- and coarse-grained polymict sandstones with gritstone lenses, and thin bands and lenses of siltstone and mudstone. The grain size decreases up the section.

Plant remains such as *Taeniocrada decheniana* (GOEPP) KR. et WEYL, *Prototaxites psigmophilloides* KR. et WEYL, *P.sp.* and *Hostimella* sp. were collected from three localities on Siktefjellet (north-eastern, north-western, and southern slopes). They indicate, in N. M. PETROSJAN's opinion, a Lower Devonian age.

Table 1

Sys-tem	Se-ries	STAGE	Group	After FRIEND 1961 and GEE & MOODY-STUART 1966		After L.G. MURASOV 1975	
				FORMATION	FORMATION MEMBER		
D E V O N I A N	UPPER	Famennian	MIMERDALEN			Plantekløfta 100	
						Planteryggen 180-400	
	Frasnian	Plantekløfta 100					
		Planteryggen 400					
	Fiskekløfta 400						
	Esteriahaugen 90-130						
	MIDDLE	Givetian	WIJDE BAY	5 0 0		Tage Nilsson 600	
				Forkdalen 630			
		Eifelian	GREY HOEK	1 0 0 0		Tavlefjellet 300	
				Gjelsvikfjellet 250		Skamdalen 150	
	LOWER	Emsian	WOOD BAY	Stjørdalen 500		Stjørdalen 400	
				Keltiefjellet 600-900		Keltiefjellet 600-900	
		Siegenian	Kapp Kjeldsen 1000 - 1500		Kapp Kjeldsen 1500		
			RED BAY	Ben Nevis 900		Ben Nevis 900	
		Fränkelryggen 600		Fränkelryggen 600-750			
		Andréebreen 100-200		Andréebreen 200			
	Red Bay conglomerate 600			Princesse Alice 300			
	L	Gedinnian	SIKTEFJELLET	Siktefjellet 1400		Siktefjellet 350	
				Lilljeborgfjellet 100-400		Lilljeborgfjellet up to 400	

All these remains were reported in 1942 by HØEG from the Fränkelryggen Formation.

According to GEE and MOODY-STUART (1966), the thickness of the Formation is 1400 m, but the present authors consider it to be 350 m. This large disagreement is accounted for by different views on the geology of Siktefjellet.

Red Bay Group

The Red Bay Group was recognized by HOLTEDAHL (1914) and subdivided into four formations: Red Bay Conglomerate, Andréebreen, Fränkelryggen, and Ben Nevis. In 1966 two lithologically different rock-units — Rabotdalen Sandstone and Princesse Alice Conglomerate — were reported from the Red Bay Conglomerates by GEE and MOODY-STUART. The present authors propose a tripartite subdivision for the Red Bay Conglomerate into the Wulfberget, Rabotdalen and Princesse Alice Formations because they differ in composition, have substantial thicknesses, distinct contacts, and a wide distribution.

Wulfberget Formation. These deposits are most widespread on Wulfberget as well as on Siktefjellet, Högeloftet, and Fränkelryggen. They rest unconformably on the Upper Proterozoic Signehamna and Generalfjellet formations and grey rocks of the Siktefjellet Formation. The upper boundary is drawn at the base of the overlying sandstones.

The Wulfberget Formation consists of red, large-pebble conglomerates with pebbles of metamorphosed limestone, quartzite, and metamorphosed microquartzite. Large blocks of angular, metamorphosed limestone occur in places at the base. The matrix is coarse-grained sandstone containing calcareous material. The thickness of the Formation is 200 m.

Rabotdalen Formation. These deposits occur mainly in the Raudfjorden area. They rest without an evident break on the underlying conglomerates and consist of coarse-grained polymict sandstones with lenses and bands of mudstone, siltstone, and silty calcareous rocks weathering (bright yellow). There are gritstone lenses in the sandstones.

In the Raudfjorden area plant remains such as *Taenioocrada? spitsbergensis* HØEG, characteristic of the Lower Devonian, were found for the first time, as were the ostracod *Clavofabelina? sp.*, A. ABUSČEK considers these also to be of Lower Devonian age. The thickness of the formation is 200 m.

Princesse Alice Formation. These deposits also occur mainly in the Raudfjorden area where they rest on the erosion surface of the underlying Rabotdalen Sandstones. The formation consists of red fine pebble conglomerates composed of quartz and quartzite. The conglomerates contain gritstones of the same composition and bands of coarse-grained sandstone. The thickness is 300 m.

Andréebreen Formation. This formation was named in 1961 by FRIEND (FRIEND 1961), but it was known as a rock unit earlier. It outcrops mainly in the Raudfjorden and Liefdefjorden areas and in particular on Wulfberget and Pteraspistoppen. The Andréebreen Formation rests without evident unconformity on the erosion surface of the Princesse Alice Formation. The upper boundary is drawn at the base of the Fränkelryggen redstones.

The Andréebreen Formation consists of coarse parallel bedded and large-

scale cross-bedded grey and greenish grey polymict sandstones with numerous siltstone and mudstone bands and lenses (up to 2.5 m thick). Mudstones contain carbonized detritus. The thickness is 200 m.

Frænkelryggen Formation was recognized by KLÆR in 1916 (KLÆR 1918). Until 1974 it was considered as the lowermost formation yielding fossils and plant remains. A narrow band of outcrops can be traced along the eastern coast of Raudfjorden up to Liefdefjorden, and may also be present on the southern coast of Liefdefjorden. The formation rests conformably on the Andréebreen sandstones and is overlain by micaceous grey sandstones of the Ben Nevis Formation.

The Frænkelryggen Formation consists mainly of red sandstones and mudstones with greenish grey sandstone and mudstone bands.

Abundant remains of fish, pelecypods, arthropods, and plants were collected from these deposits. Osteostraci (*Cephalaspis*), Heterostraci (*Cyathaspis*), and *Acanthodii* were identified. The thickness is 600 to 750 m.

Ben Nevis Formation. This Formation was recognized for the first time by KLÆR in 1916. The type section occurs on Ben Nevis, and outcrops are also present near Pteraspistoppen (northern Liefdefjorden) and on the northern side of Vonbreen. The lower boundary is drawn at the base of the first thick bed of grey-green micaceous sandstones. The upper boundary is drawn at the base of red beds of the Kapp Kjeldsen Formation.

The formation consists of greyish green, cross-bedded, fine- and coarse-grained polymict micaceous sandstones. A thick unit of violet-red sandstones and mudstones is present in the middle of the section. Abundant fossils of fish, ostracods and pelecypods occur throughout the sequence. Also the arthropod *Mesostomata* is found.

Fish include Osteostraci (*Cephalaspis*), Heterostraci (*Poraspis* and *Traquairaspis*), and *Acanthodii*. The thickness is 900 m.

SIEGENIAN

Wood Bay Group

The Wood Bay Group was recognized by HOLTEDAHL in 1914 and subdivided by FRIEND and HEINTZ in 1943 into three formations: Kapp Kjeldsen, Lykta (renamed Keltiefjellet in 1961 by FRIEND), and Stjørdalen. In 1966, FRIEND et al. proposed not to subdivide the group into formations because of the lack of good marker horizons and the uniform lithology. Instead of formations they proposed independent separate sequences differing in rock composition such as the Austfjorden Sandstone. However, geological works and surveys carried out by Soviet geologists showed that these deposits, despite some difficulties, may be subdivided into the earlier recognized formations which are even mappable.

Kapp Kjeldsen Formation. This formation is very widespread, extending from Liefdefjorden to Dicksonfjorden in the west and from Kartdalen (Andrée Land) to Nathorstdalen (Dickson Land) in the east. The beds of the Kapp Kjeldsen Formation rest conformably on the grey-coloured rocks of the Ben

Nevis Formation. The contact was described by FØYN and HEINTZ (1943) and by BUROV and MURAŠOV (1967) in the Woodfjorden area on Kronprins-høgda and Sigurdfjellet. The top of the so-called “pale beds” was regarded as the upper boundary of the formation.

In the western areas of the Devonian graben the Kapp Kjeldsen Formation consists largely of a complex alternation of red siltstones, mudstones and fine-grained sandstones with rare thin layers of grey-green siltstones and sandstones yielding plant remains. The section is capped by the “pale beds” which consist of interbedded green, greenish-yellow, crimson and bright brown sandstones, siltstones, mudstones, and silty limestones. A gradual increase in sand content takes place south-eastward along the Bockfjorden–Austfjorden line, and in the Austfjorden area the Formation consists mainly of coarse-grained green, greenish-grey, and greenish-yellow cross-bedded and massive micaceous sandstones and sandy siltstones, with lenses and bands of gritstone and fine-pebble conglomerate beds along the boundaries. The red beds, very characteristic of western areas, occur only in the upper part of the formation to the east. The “pale beds” in this area consist of coarse-grained cross-bedded and massive green and yellowish-green sandstones with rare thin bands of crimson sandstone, siltstones and mudstones. A notable feature of the Kapp Kjeldsen Sandstones in the Austfjorden area is the presence of rounded clasts of quartz and orthoclase and numerous inclusions of black, very dense, coaly, sandy mudstones and siltstones.

The contrast in lithology and mineral composition of the rocks in the west and those in the east, implies different conditions of sedimentation. It is evident that the Austfjorden area was much closer to the source area, and the presence of quartz and orthoclase pebbles in the sandstones suggests that the deposition of the Kapp Kjeldsen Formation took place during the erosion of older crystalline rocks.

Osteostracii, Heterostraci, Arthrodira, Crossopterygii and Charophyta are the guide fossils. Pteraspids such as *Gigantaspis*, and Arthodires such as *Arctaspis* indicate a Siegenian age. Abundant ostracods and plant remains suggest a wider age range of Lower-Middle Devonian.

The fossil content of the formation varies between the western and the eastern areas. The western sections contain numerous bands with remains of fish and ostracods, while plant remains are much less common. However, the eastern sections contain a high percentage of bands with plant remains; fish and ostracods are less common, implying contrasting environments of sedimentation. The thickness of the formation is 1500 m.

Keltiefjellet Formation. This was first described by FØYN and HEINTZ (1943) as the Lykta Formation, later renamed by FRIEND in 1961. Deposits are widespread in Andrée Land and Dickson Land where they rest on the “pale beds” of the Kapp Kjeldsen Formation. The first thick (25–30 m) bed of green sandstone above the “pale beds” is considered as the lower boundary. This sandstone bed is ubiquitous and forms a good marker horizon. The upper boundary is marked by the change from the grey-green and brown beds of the Keltiefjellet Formation to the crimson-red siltstones of the Stjørdalen

Formation. The last thick (up to 15 m) bed of green cross-bedded sandstone of uniform occurrence is regarded as the upper boundary. Both contacts are readily traced in Andrée Land.

Sections of the formation in north-western Andrée Land and in Dickson Land are different. They are characterized by the south-eastward increase of sand grain size. In Andrée Land the formation consists generally of brick-red siltstones and fine-grained sandstones alternating with rare bands of coarser brown-green and grey-green sandstones, sandy siltstones, and less common calcareous gritstones. In Dickson Land coarse-grained sandstones and gritstones considerably increase in abundance.

The most common fish are *Doryaspis* and *Arctolepis*. Less common are *Homostius*, *Actinolepis*, *Porolepis* and *Arctolepis* suggesting a Lower Devonian age (Siegenian-Emsian). The flora is represented by *Hostimella*, *Psilophyton*, and *Aphylopteris*. The thickness of the formation is 600 to 900 m.

Stjørdalen Formation. This formation was first documented by FØYN and HEINTZ (1943). The deposits occur both in Andrée Land and in northern Dickson Land. The lower boundary is marked by the change from the Keltiefjellet Formation rocks to the more argillaceous cherry-red and crimson-red deposits of the Stjørdalen Formation. The contact is conformable. The upper boundary is drawn at the base of the calcareous rocks known as the Verdalen Member.

The Stjørdalen Formation consists mainly of mudstones and cherry-red and calcareous siltstones with pelletoid structure. Thin (0.5 to 2.5 m) red-yellowish-grey, micaceous, flaggy, fine-grained sandstones occur at intervals over the entire section. Individual bands (up to 0.5 m) of violet-brown, fine-grained sandstone and light grey, quartzitic sandstone occur in the lower part of the section. The middle part is characterized by a slight increase in number of bands of greenish-grey, fine-grained sandstone, and the appearance of calcareous gritstones, greenish-grey in colour. In the upper part the number of greenish-grey bands abruptly decreases. Cherry-red pelletoid siltstones and mudstones predominate.

The most common forms of fossil fish remains include Nectaspids, Monaspids, Osteostraci, Arthorodira and Crossopterygii. In V. N. TALIMAA's opinion (personal communication) the most probable age of these deposits is Emsian. The thickness of the Formation in northern Andrée Land is 400 m, gradually decreasing southward; in northern Dickson Land (Lancasterryggen) the thickness does not exceed 200 m, on Bulmanfjellet it is 70 to 100 m, on Watsonfjellet 50 m, and farther south the Stjørdalen Formation dies out entirely.

Middle Devonian

EIFELIAN

Grey and partly Eifelian deposits occurring largely in Andrée Land and in places in Dickson Land and the Hornsund areas, are subdivided by the present authors into three formations: Gjelsvikfjellet, Tavlefjellet, and Forkdalen.

The *Gjelsvikfjellet Formation* is subdivided into two members: Verdalen and Skamdalen.

The Verdalen Member has been recognized from the Upper part of the Stjørdalen Formation. It occurs in central Andrée Land resting conformably on the Stjørdalen redstones. The lower boundary is drawn at the base of the first bed of violet-grey, yellow-weathering, silty limestone. The upper boundary is characterized by an abrupt transition from red to grey rocks.

This member is made up largely of violet-grey and grey silty limestones, alternating with violet-red calcareous siltstones, with bands and lenses of fine-grained violet-grey and greenish-grey polymict sandstones in the upper part.

Fish remains include *Homostius arcticus*, *Herasmius granulatus*, *Heimenia ensis*, *Amaltheolepis winsnesi*, characteristic in ØRVIG's opinion of the Lower Eifelian (1969).

The thickness of this member at the type locality (Woodfjorden) is about 100 m, in the Vestfjorden area up to 60 m, and on Dickson Land it is absent.

The Skamdalen Member has been recognized from the lower Grey Hoek Group. The type section lies on the left of Skamdalen in Andrée Land. The member is widespread in Andrée Land (south of Jakobsenbukta latitude) and can be traced southward at least to Nathorstdalen (Dickson Land). It rests unconformably on the Verdalen deposits in Andrée Land and on the Stjørdalen red beds in northern Dickson Land. The lower boundary is marked by an abrupt change from red to grey rocks.

The upper boundary in Andrée Land is drawn at the change from calcareous rocks to almost black mudstones. It is impossible to trace it in Dickson Land.

The deposits of the Skamdalen Member are represented by dark grey and grey micaceous calcareous siltstones with bands and lenses of dense parallel-bedded siltstones. In the lower part there is a 27 m thick unit of almost black highly calcareous mudstone. South of Skamdalen the grain size of the sandy material increases and thin sandstone bands appear. In the Nathorstdalen area the deposits generally consist of grey-green siltstones and quartz sandstones with very scarce thin bands of grey arenaceous limestone.

Fish remains include *Heimenia ensis*, *Homostius arcticus*, *Amaltheolepis winsnesi* and *Herasmius granulatus*. In V. N. TALIMAA's opinion (personal communication) this assemblage is nearly identical to the fish fauna described by ØRVIG (1969) from the Verdalen Member and is Lower Eifelian in age. The Verdalen and Skamdalen Members may therefore be considered to be approximately coeval and similar in lithology. This, and their recognition over long distances enables them to be united into the Gjelsvikfjellet Formation totalling 250 m thick.

Tavlefjellet Formation. This formation was first recognized from the lower part of the Grey Hoek. The type section lies on the southern slope of Tavlefjellet west of Wijdefjorden.

This Formation may be traced in central Andrée Land from Mushamna in the north to Kartdalen in the south. A thick sequence of mudstones cropping out on the northern side of Hornsund probably belongs to the same formation.

The boundary with the underlying deposits is conformable and is marked by the change from the Skamdalen calcareous siltstones into mudstones. The upper boundary is drawn at the top of the mudstones which are conformably overlain by massive siltstones. Both contacts can be traced in several places on the southern coast of Andrée Land from Forkdalen to Kartdalen.

Within the type section these deposits are represented by two sequences. The lower sequence is composed of dark grey to black fragmentary mudstones, with bands of lighter calcareous siltstones forming scarps up to 2–3 m distinctly reflected in the topography. Individual siltstone beds are highly calcareous and sometimes are replaced by silty limestones. Carbonate nodules up to 10 m in diameter with fragments of fish and pelecypods occur throughout. The thickness of the sequence is 170 m.

The upper sequence is composed mainly of dark-grey to black calcareous mudstones with bands of mats and loaf-like silty carbonaceous nodules (up to 0.5 m in diameter) with fissures filled with brown calcite. The thickness of the sequence is 130 m.

Fish fossils include *Wijdeaspis arctica* HEINTZ, *Heimenia ensis*, *Porolepididae* gen. idn., *Crossopterygii* fam. et gen., *Arthrodira* fam. et gen. The following pelecypods were found: *Chenodonta* ex gr. *maureri*; BEUSH, *Prosocoelus* (?) sp., and *Nucula* sp.

V. N. TALIMAA and O. V. LOBANOVA concluded that the fossils found suggest a Middle Devonian (Eifelian) age. The thickness is 300 m.

Forkdalen Formation. This Formation is separated by the authors from the upper Grey Hoek Group. The type section lies on Tavleljellet and a complete section was found on the northern flank of Forkdalen.

The Forkdalen Formation occurs mainly in northern Andrée Land where it composes the eastern and part of the western flank of the Forkdalen syncline and its periclinal zone.

The lower boundary is drawn at the first siltstone band conformably overlying the Tavleljellet mudstones.

The upper boundary is drawn at the base of the first bed of light grey quartzitic sandstone of the Wijde Bay Group.

The Formation consists mainly of interbedded grey and dark-grey siltstones, black fragmentary mudstones and polymict sandstones. An increase in the number and thickness of sandstone units as well as in sand grain size is observed up section. "Loaf" carbonate nodules 1.0×0.4 m occur throughout the section. In the upper part large pelecypods of genus *Myalina* were found.

Within a fault zone in north-western Andrée Land the Forkdalen Formation is strongly folded and represented mostly by light-grey and grey quartzites and quartzite-like sandstones.

Fish fossils from the Forkdalen Formation include *Arthrodira*, *Heimenia*, *Arctonemia*, *Homostius*, *Antiarchi* and *Brachythoraci*. Pelecypods are represented by *Carditomantea* ex gr. *spinata* QUESNT., *Prosochama*, *Prosocoelus*, *Myalina*, *Myaphoria*, and *Montonaria*. The flora contains *Hostimella*, *Protocephalopteris*, *Psilophyton*, *Pseudouralia*, *Enigmophyton*, *Arctophyton* and *Taeniochrada*.

These fossils suggest a Middle Devonian (Eifelian) age for the Forkdalen Formation. The thickness of the Formation is 630 m.

GIVETIAN

Both the Wijde Bay Group, which the authors propose to consider as the Tage Nilsson Formation, and the Esteriahaugen Formation in the Billefjorden area, belong to the Givetian.

Tage Nilsson Formation. This formation occurs in the Tage Nilssonfjellet area in north-eastern Andrée Land, at the locality of the type section. The lower boundary is drawn at the base of the first quartzitic sandstone. The upper contact is not exposed.

The Tage Nilsson Formation consists of closely intercalated quartzitic sandstones, massive siltstones, and mudstones. A notable feature is the intense jointing of the quartzitic sandstones usually accompanied by iron mineralization of magnetite-hematite type. Lenses of gritstone with abundant fish remains were observed at the base of sandstone bands. The siltstones yielded abundant fossils of strongly deformed pelecypods and floral remains.

Fish include *Arthrodira*, *Holomena*, *Homostius*, *Herasmusius*, *Antiarchi* and *Asterolepis*. The following pelecypods were found: *Myalina*, *Avicula*, *Puella*, *Concoardium*, *Pterinea*, *Solenomorpha*, and *Laiopectinella*.

The flora is represented by *Protocephalopteris*, *Enigmophyton*, *Hostimella*, *Psilophyton*, *Taeniocrada*, and *Barrandeinopsis* allowing these deposits to be assigned in N. M. PETROSIAN's opinion, to the Late Eifelian-Givetian. The observed thickness is 600 m.

Esteriahaugen Formation. This unit is not very widespread. Outcrops extend in a narrow band in the interfluvium of Munindalen and Mimerdalen. The lower contact is tectonic. This formation is faulted against variegated Lower Devonian Reuterskiöldfjellet sandstones. However, VOGT (1938) has observed the Esteriahaugen deposits resting on the red-coloured Lower Devonian deposits with an unconformity representing a gap.

The upper boundary is drawn at the base of the 1.7 m thick unit of black fragmentary mudstones.

In the lower part of the formation fragmentary mudstones dominate, with less common bands and lenses of light-grey polymict quartzitic sandstones. A large number of rounded argillaceous nodules and thin bands of coals and coaly rocks occur in the mudstones. The upper part is composed mainly of quartz sandstone containing numerous rounded pyrite nodules which when oxidized show rusty patches.

The following fossils were collected from this Formation: Plants: *Platiphyllum*, *Protocephalopteris*, *Pseudoporochnus* and *Svalbardia* (determined by N. M. PETROSIAN); ostracods: *Hogmochilina* (identified by ABUSHEK); pelecypods: *Myalina* and *Pteria*.

Identified spores include *Camarozonotriletes*, *Archaeotriletes*, *Stenoronotriletes*, *Archaeoperisaccus*, and *Archaeozono* (determined by G. K. VAITEKUNENE).

Thus, these deposits may tentatively be assigned to the upper Givetian, but they may be Frasnian. The observed thickness is 100 m.

Upper Devonian

FRASNIAN

Frasnian rocks are represented by a single formation and were observed only in the Mimerdalen area (Billefjorden).

Fiskekløfta Formation. This formation occurs mainly in the upper and middle reaches of Mimerdalen. The type section lies in the Fiskekløfta gorge. The lower contact is conformable and is drawn at the top of a plant-bearing, grey-green sandstone.

Generally the deposits consist of interbedded black fragmentary mudstones, grey-green, fine-grained quartz sandstones with carbonized plant detritus, and dark-grey siltstones. Flattened carbonate-iron nodules (up to 5 cm in diameter) occur throughout the section.

Fish and plant remains as well as spores were collected from the deposits.

Fish: *Asterolepis scabra* (WOOD), *A.sp.ind.*, characteristic in V. N. TALIMAA's opinion of the lower Upper Devonian (Frasnian).

Flora: *Aulacopteris vulgaris* GRAND EURU., *A. vulgaris* HØEG, *Anarthrocanna gopperti* NATH., *Heteraugium* sp., *Leptofloeum rhombicum* DAMS. (*Bergenia spitsbergensis* sp. nov., and *Rhizomopteris nordenskiöldi* NATH.). N. M. PETROSIAN considers these to be indicative of an Upper Devonian age.

Spores: *Archaeozonotriletes* cf. *notatus* var. *asper* TSCHIBR., *A.sp.*, *Acinosporites* sp., *Densosporites lyssi* var. *spinatus* TAUG-LANTZ, *Calamospora* cf. *microrugosa* (IBR.) BALME, *Hystricosporites porcatus* (WINSON) ALLEN, *Geminospora* sp., *Lophotriletes ungnatus* NAUM., *Punctatisporites* sp., and *Retusotriletes greggsii* MCGREGOR. suggest an Upper Devonian (Frasnian) age. The thickness of the formation is 145 m.

FAMENNIAN

Planteryggen Formation. The formation is assigned by the present authors to the Famennian stage.

The Planteryggen Formation was first named by FRIEND in 1961. Deposits occur in the upper Munindalen and Mimerdalen. The lower boundary is drawn at the base of a bed of sugary sandstones containing large fragments of tree trunks.

The lower part of the formation consists of grey-coloured sandstones with bands of mudstones and siltstone. The upper part comprises sandstones, siltstones and mudstones. The section is capped by a unit (40 m thick) of red conglomerate with pebbles of quartzite, microquartzite, sandstone and siltstone.

Floral impressions include *Bothrodendron* sp., *Leptophleum rhombicum* DAMS., *Cyclostigma Kilterkense* (HAUGHTON) NATH., *Knorria* sp., *Lepiodendron spitsbergensis* NATH. and *Lepiodendrops* sp. In N. M. PETROSIAN's opinion this assemblage suggests an Upper Devonian age.

In the upper reaches of Odindalen a large fragment of a vertebra was found. E. VOROBYEVA believes it to belong to the crossopterygians, however it may also have belonged to the oldest stegocephalians. The thickness of the formation is 180 to 400 m.

CARBONIFEROUS

These deposits are discussed because they have previously been considered to be Upper Devonian. These rocks belong to the Mimerdalen Group and are represented by the Plantekløfta Formation developed in the area of Plantekløfta creek and along the west side of Munindalen. These deposits rest with an angular unconformity on different horizons of the Planteryggen Formation. The upper contact is not exposed. In Munindalen the upper beds of the formation are faulted against the Lower Devonian Reuterskiöldfjellet sandstone (Wood Bay Group).

The Formation consists of interbedded fine- to medium pebble conglomerates, dark-green sandstones and siltstones. The pebbles consist of 95% violet and greenish-greyish sandstones and 5% grey siltstones. Gritstones cementing the pebbles contain 60% quartzite and 40% sandstone and siltstone.

PETROSIAN determined the flora to be *Cyclostigma kilterkense* (HAUGHTON) NATH., *Bothrodendron* sp., *Leptophlocum rhombicum* DAWs., *Bergenia mimerensis* HØEG and *Lepidodendropsis theodoty* (ZAL.) JOGUM.

In addition remains of lycopods were found with fine cushions similar to lycopods of the lower Carboniferous. However, poor preservation does not allow accurate identification. PETROSIAN claims that they may be younger than Upper Devonian, namely lower Lower Carboniferous. The thickness of the formation is 100 m.

Conclusions

The study of the Devonian deposits of Spitsbergen by Soviet geologists resulted in the further subdivision and dating of the Devonian section:

1. Three formations, i.e. Wulfberget, Rabotdalen, and Prinsesse Alice, earlier known as lithological horizons, were recognized within the Red Bay Group.
2. The Grey Hoek Group was subdivided for the first time into three formations: Gjelsvikfjellet, Tavlefjellet, and Forkdalen.
3. On the basis of plant remains the ages of the formations of the Mimerdalen Group were refined.
4. New paleontological data became available which enabled more accurate determination of the age of the Devonian deposits in Spitsbergen.

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The Franklinian Geosyncline in the Canadian Arctic and its relationship to Svalbard

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Abstract

Development of the Franklinian Geosyncline began, perhaps earlier, but certainly by late Proterozoic time, with the deposition of clastic and carbonate rocks in the region of northeastern Ellesmere Island. Sedimentary units thicken northward, away from the exposed Aphebian

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crystalline basement. Certain metamorphic rocks of the north coast of Ellesmere Island may have been part of this early geosynclinal sequence; both volcanic and sedimentary origins are inferred for granitoid gneisses from which late Proterozoic isotopic ages have been obtained. Cambrian to Late Devonian clastic and carbonate sedimentary rocks subsequently were deposited in the geosyncline. Carbonates and some evaporites and clastic sediments dominated in the southeast, and immature clastic sediments with volcanic rocks, carbonates, and chert in the northwest. The sediments in the northwest evidently were derived from a geanticlinal welt, Pearya, which lay in the present offshore region. A distinctive basin of flysch deposition in the axial region, the Hazen Trough, received sediments from early Middle Ordovician to early Devonian time. From sole markings it is clear that sediment-bearing currents entered the trough from the northwest and were deflected to the southwest, along the trough.

Widespread deposition of clastic sediments in Middle and Late Devonian times was heralded in the Early Devonian by the development of three northerly trending, positive structural belts, of which the Boothia Uplift is the most prominent, with adjacent troughs and basins. Intrusions of ultramafic to granitic and syenitic composition possibly were emplaced in the northernmost region at this (Early) time, judging by K-Ar age determinations.

Tectonic activity in the northern geanticline in Middle and later Devonian time is inferred to have advanced southward to terminate the normal geosynclinal sequence and provide a southward-directed flood of clastic sediments. This, the Ellesmerian orogeny, involved regional folding, metamorphism, local quartz monzonite and quartz diorite intrusion, and widespread uplift. The synkinematic intrusions and metamorphism were restricted to the northernmost region, whereas folding to the south affected both earlier and later, synorogenic clastic beds. Ellesmerian structures conform to the present shape of the craton: a markedly sigmoidal pattern in the Canadian Arctic Islands lies between the related orogenic belts of northern Greenland and north shore, arctic Alaska.

Certain tectonic elements of the orogenic belt, such as the successor, Sverdrup Basin, are younger than Franklinian or Ellesmerian features but are geographically closely coincident with them. The younger elements thus appear to owe their origin to reactivation of tectonic processes that gave rise to the older features.

A comparison of the sedimentary and tectonic features of the Innuitian region and Svalbard shows that the tectonic histories of the two regions were mainly unlike before Devonian time but distinctly similar during certain periods since then. Taking into account ocean spreading, it appears probable that Svalbard and the Franklinian Geosyncline were once adjacent. A model is proposed for the earliest (Precambrian-late Paleozoic) times, in which rudely matched sedimentary basins were separated by the geanticlinal ridge, Pearya. A linear zone of younger basins formed in Carboniferous and Permian time, and by mid-Tertiary time the region was ragmented by the opening of the Atlantic and "neo-Arctic" oceans.

Introduction

The Innuitian Orogen is a continent-bordering tectonic region only somewhat less grand in scale than the better-known Cordilleran and Appalachian regions (see Figs. 1, 3). The Franklinian Geosyncline, a major component of the Innuitian Province, was a long-lived sedimentary-tectonic feature that provided a framework for subsequent events.

The purpose of this paper is to explore the history and disposition of the Franklinian Geosyncline¹ and to consider its possible relationship with the geosynclinal succession of Svalbard. The relationships between younger components of Svalbard and the Canadian Arctic orogen are also of prime interest and, although these basins will not be described here in detail they will be included in a suggested model.

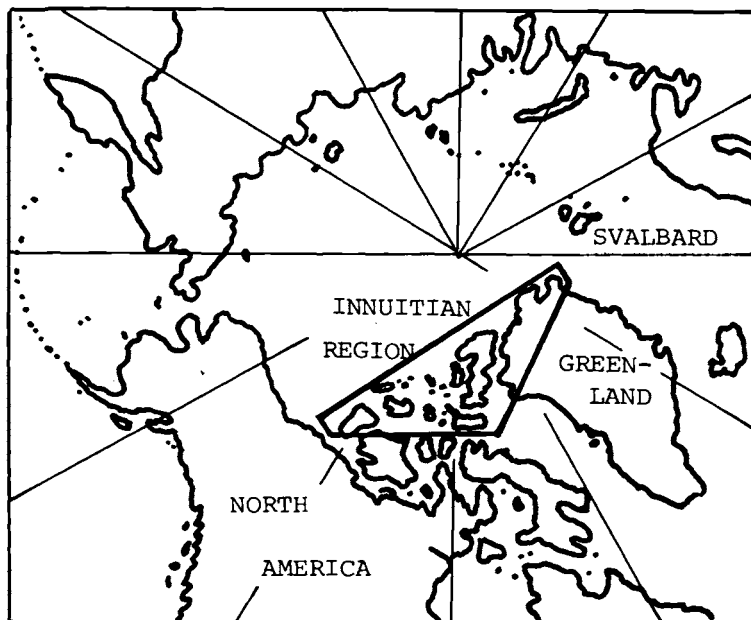


Fig. 1. Index map of north polar regions.

The manuscript for this paper was critically read by H. R. Balkwill and U. Mayr of the Geological Survey of Canada; discussions with them have contributed greatly to the paper presented here.

The Franklinian Geosyncline

The Franklinian Geosyncline is well exposed in northernmost Ellesmere Island and Greenland, but nearly disappears to the southwest, hidden by younger sedimentary basins, by coastal plain sediments, and by marine water. From exposures available, the geosyncline appears to encompass a nearly complete suite of characteristics now considered to be typical of mountain belts that border large cratonic masses. The important aspects of the Franklinian Geosyncline have been described recently by THORSTEINSSON (1974, p. 6-10; and in THORSTEINSSON and TOZER 1970), TRETTIN (1973, 1972, and other papers), and by KERR (1967b, 1968, 1976).

GENERAL TECTONIC PATTERN

The most complete cross-section of the Franklinian Geosyncline is exposed between central Ellesmere and the north coast of the island. Trends in this region are fairly consistently northeast, and the structures pass without major apparent dislocation into northern Greenland. To the southwest, the Franklinian

¹ The Canadian part of the geosyncline is considered here. An account of the Franklinian Geosyncline in northern Greenland was given, at the Svalbard symposium in Oslo, by P. R. DAWES, who has recently compiled a full review of the geology of that region (DAWES 1976).

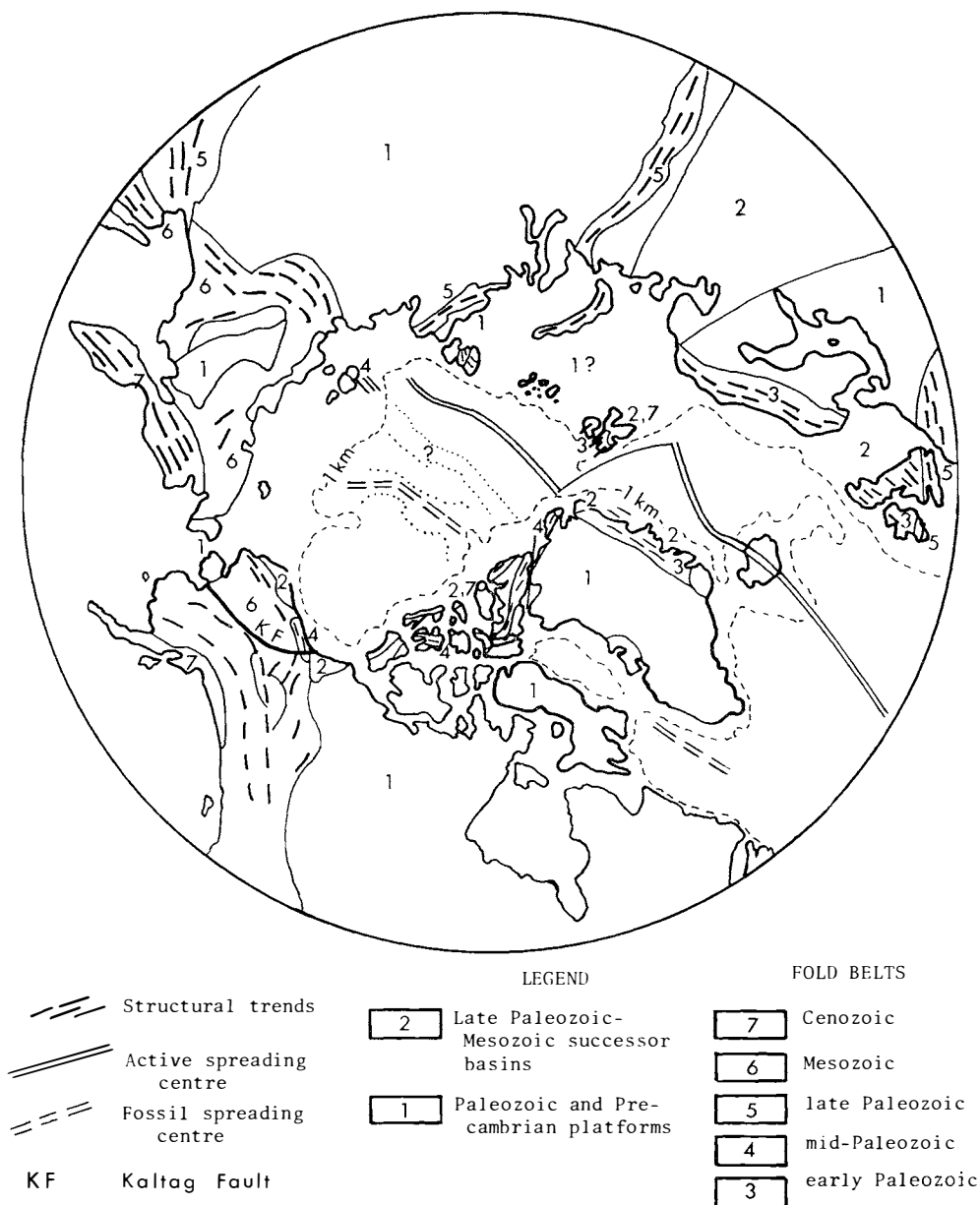


Fig. 2a. Major tectonic features of the Arctic (modified from CHURKIN 1972; OSTENSO 1974).

ian rocks are covered in the axial region by a large successor basin, the Sverdrup Basin. Miogeosynclinal parts of the Franklinian basin are exposed south and east of the Sverdrup Basin, and Franklinian sedimentary and structural trends there trace a marked sigmoidal or double bend, swinging south, then west. Structural trends at the westernmost exposures, where they disappear beneath the Arctic Coastal Plain, are heading approximately toward the northern tip of Alaska, across the Beaufort Sea.



Fig. 2b. Tectonic and geographic features of the Arctic. For legend, see Figure 2a.

STRATIGRAPHY

Geosynclinal development began, perhaps earlier, but certainly by late Proterozoic time, with the deposition of more than 1700 m of clastic and carbonate rocks now exposed on eastern Ellesmere Island (Kennedy Channel and Ella Bay Formations).¹ These rocks are miogeosynclinal in character: fine-grained, dark-coloured clastic rocks with some limestone and dolomite, passing toward the craton into cleaner, fine-grained sandstone. Upper units are sugary, slightly shaly dolomite.

The geosynclinal sequence is broken by an unconformity, in the miogeosynclinal sections studied, and formations containing Cambrian fossils and organic-appearing markings (Ellesmere Group, Scoresby Bay and Parrish Glacier Formations) overlie the Proterozoic beds. The Cambrian units, up to

¹ Certain formation names are noted in this discussion to aid those who may read other accounts of the same sequences.

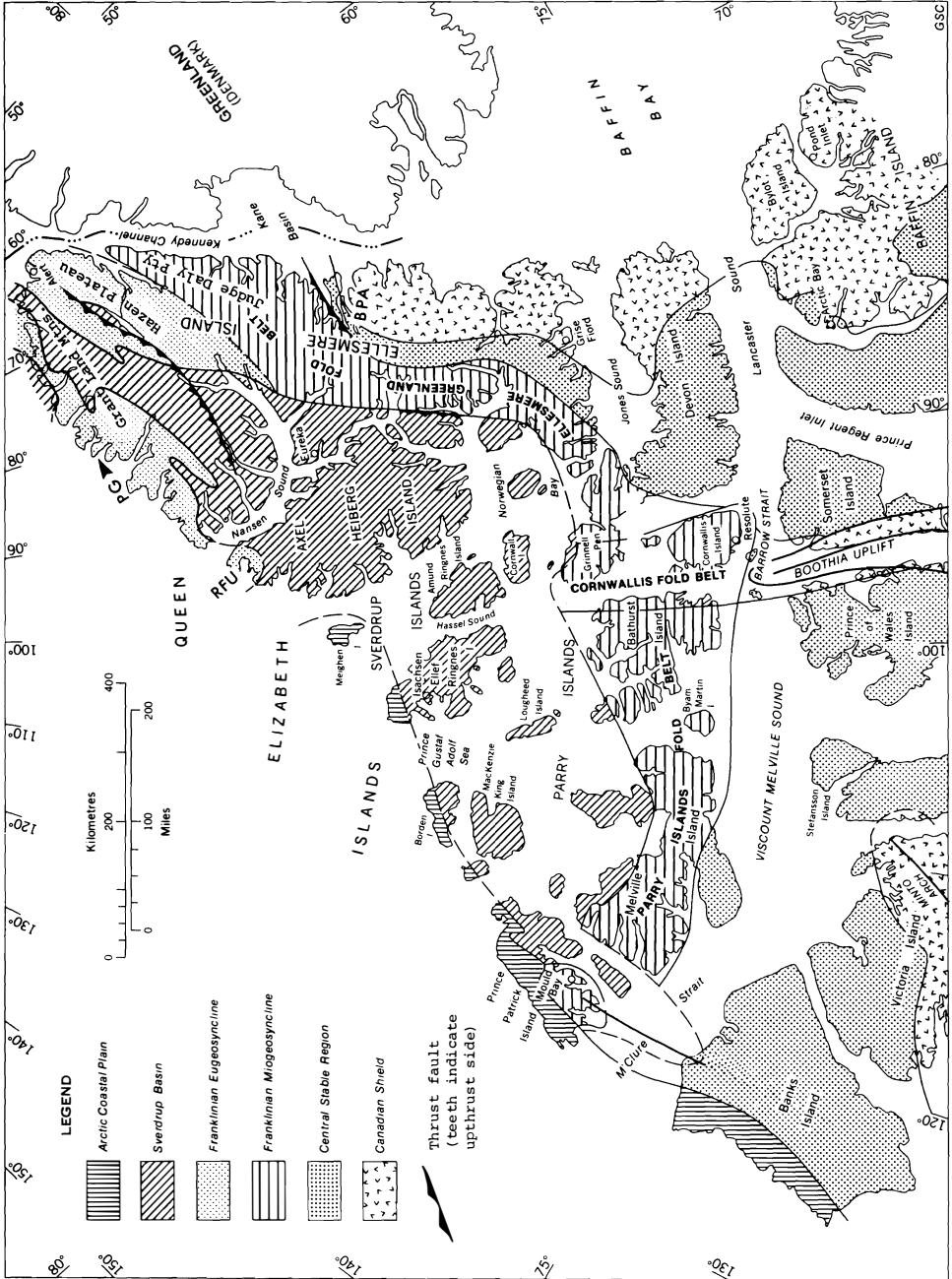


Fig. 3a. Structural Provinces, Canadian Arctic Islands. BPA: Bache Peninsula Arch. RFU: Rems Fiord Uplift. PG: Peary Geanticline.

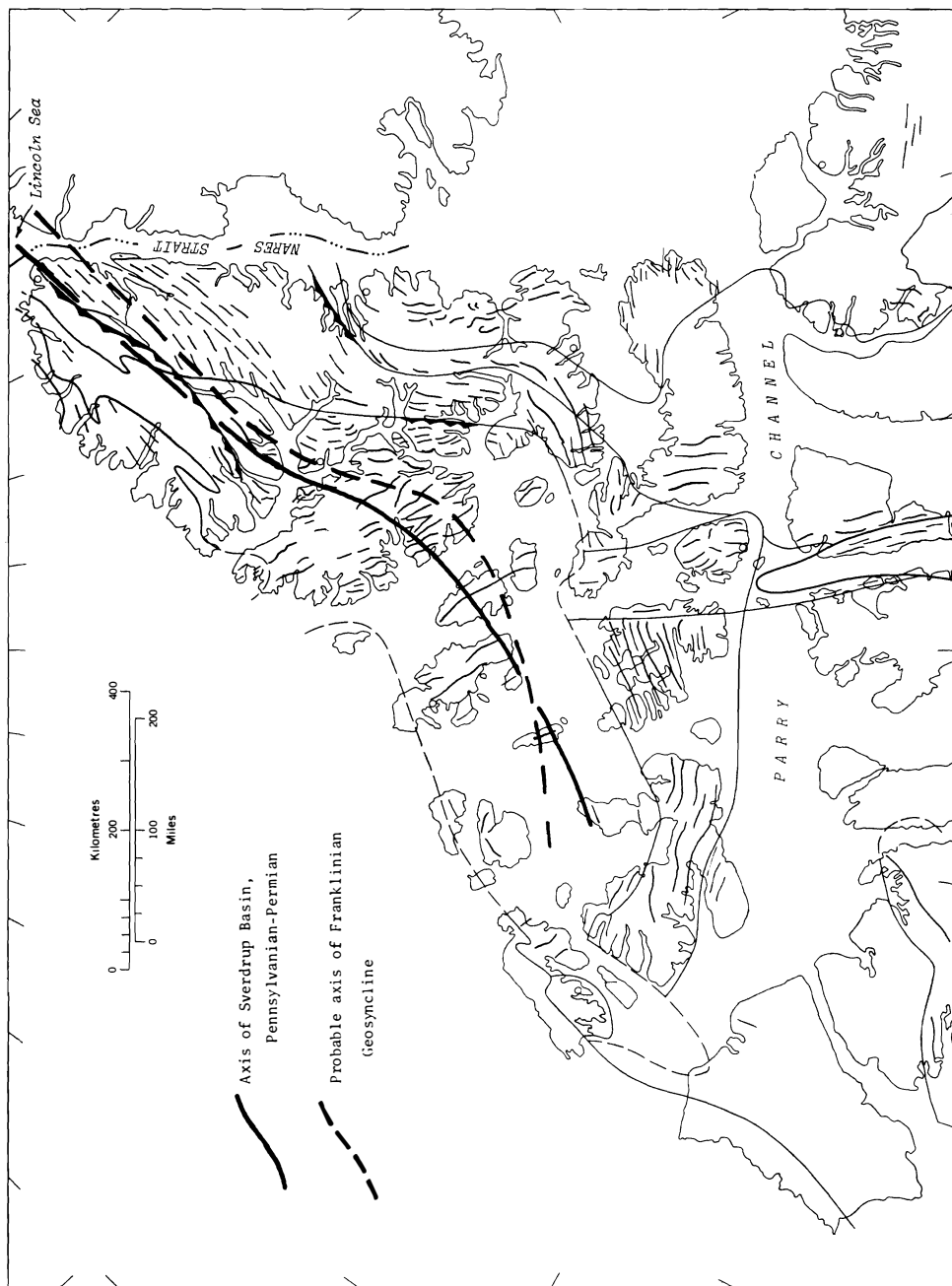


Fig. 3b. Structural trends, Precambrian to Tertiary, Canadian Arctic Islands. Structural provinces of Figure 3a are outlined. Axis of Sverdrup Basin after THORSTENSSON and TOZER 1970, p. 570 (facies boundary); probable axis of Franklilian Geosyncline conforms to exposed sedimentary and structural trends.

2400 m thick, are almost entirely land-derived clastic rocks that were deposited in a marine environment, and rather pure sandstone units of the platform grade northwestward into more shaly sands, then to black phyllite of the geosyncline (KERR 1967b, p. 19). There appears to have been onlap south-eastward, with Proterozoic units perhaps thin or even absent toward the craton.

A Cambrian age is proposed for a clastic unit (Grant Land Formation) more than 1100 m thick in the northern coastal region of Ellesmere Island (TRETTIN 1971). This unit consists of quartzose and feldspathic sandstones, red, green, and grey-weathering siltstones and shales, and some pebble conglomerate and carbonate rocks. The detrital mineralogy and some paleocurrent determinations suggest a northern, gneissic provenance. The presence of this widespread clastic unit and the evident northern bordering lands (Pearya) complete the picture of a geosyncline bordered by cratonic platform on one side and by tectonic lands on the opposite side (see Fig. 4B).

Deep subsidence took place by early Middle Ordovician time to form a tectonic trough, the Hazen Trough, in the axial region of the geosyncline (TRETTIN 1971, 1972, 1973). Deposited in this elongate trough were distinctive, deep-water sequences comprising a lower, “starved basin” graptolitic shale facies and an upper, clastic and calcareous flysch facies. The lower unit, about 500 m thick, consists of graptolitic shales, chert, carbonate rock, and minor amounts of breccia (Hazen Formation). These beds conformably overlie the older clastic rocks. Carbonate rocks of this unit are thin-bedded to laminated, are partly graded, and contain redeposited carbonate material, probably derived from adjacent shelves and carried into deep water of the Hazen Trough by turbid flows. Conformably overlying the graptolitic rocks is a uniform succession, locally more than 2 700 m thick, of alternating calcareous greywacke, calcareous siltstone, and calcareous shale with minor amounts of conglomerate and breccia (mainly Imina Formation). Many bottom features, such as ripple marks, flute casts, longitudinal furrows and ridges, and groove marks are present, characteristic of sediments deposited in deep water under turbid conditions. From more than 2200 determinations of directional structures, TRETTIN (1971) has shown that the turbid flows entered the trough from the northwest and were deflected along the trough to the southwest.

Meanwhile, clastic deposition was taking place in the present north coastal region, which was then a coastal plain and shelf environment. Clastic sedimentation was interrupted by brief intervals during which carbonate accumulation prevailed (see Fig. 4C). Repeated floods of clastic material (in which the proportion of debris from metamorphic rocks increases upward in the stratigraphic succession), and the presence of siliceous to intermediate volcanic rocks and sedimentary suites derived from volcanic rocks, suggest tectonic activity in an anticlinal welt. This tectonic zone has been named the Pearya Geanticline (see TRETTIN 1969b, 1971, 1972).

Unfossiliferous, metamorphosed rocks are overlain unconformably by beds containing late Middle Ordovician (Wildernessian) fossils in northern Ellesmere Island and, in northern Axel Heiberg Island, Middle Silurian rocks are overlain with angular unconformity by Lower Devonian beds. These features

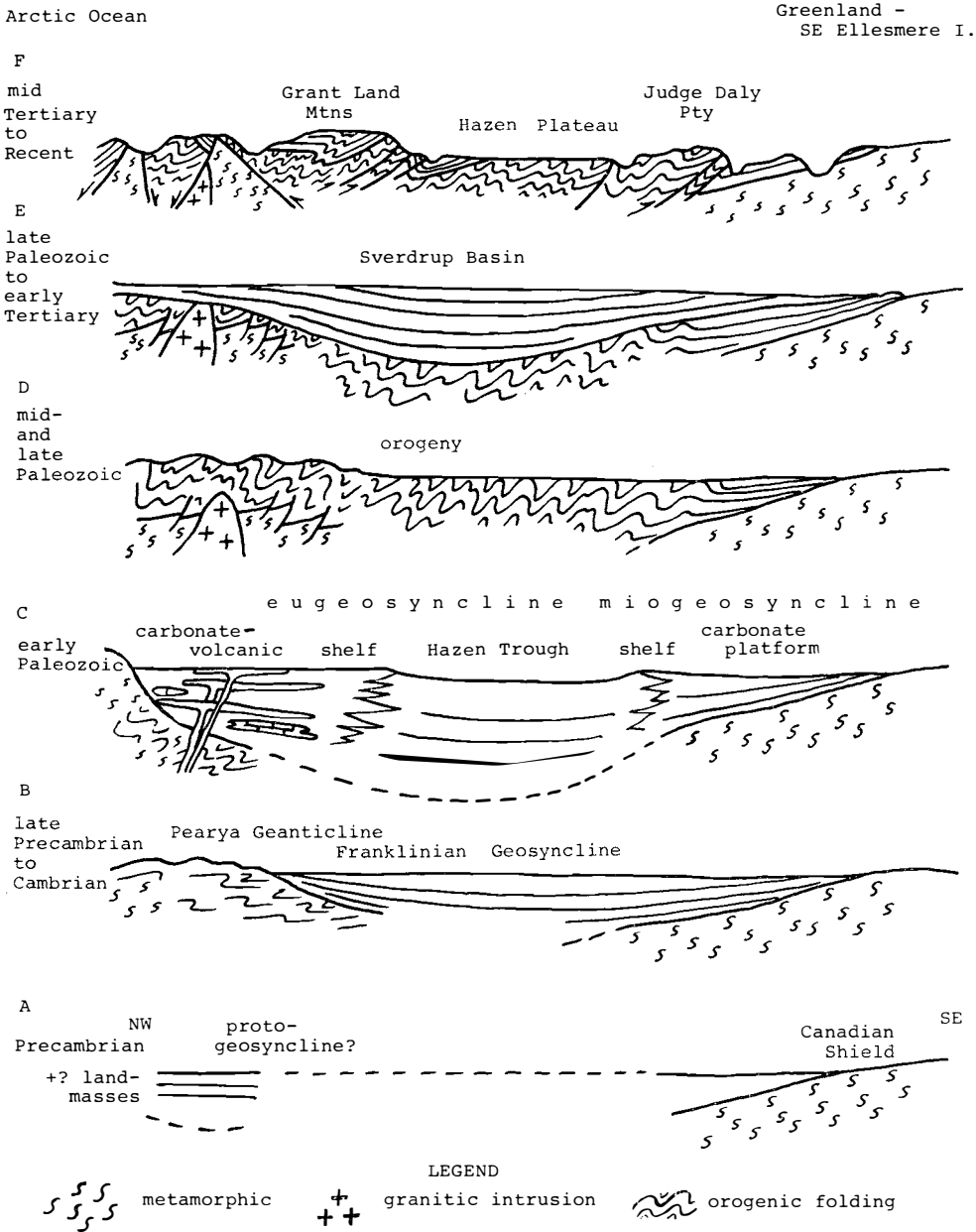


Fig. 4. Innuitian sedimentary and tectonic episodes, northern Ellesmere Island.

show that uplift in northern regions at times advanced southward so that areas of northern Ellesmere and Axel Heiberg Island were exposed to erosion.

The Hazen Trough expanded to both the north and south until about mid-Silurian time, then evidently shifted to the southeast. Silurian reefs in the southeast (e.g. in Greenland, see NORFORD 1972) appear to have been inundated by shaly, flysch facies sediments. The sedimentary pattern of the trough in later

Ordovician and Silurian time appears to have been: graptolitic shales, siltstones, and limestones on the flanks, and the thick, flysch-like suites of calcareous rocks (Imina Formation) in the axial parts. Great thicknesses of flysch deposits may be present, but efforts to measure these are frustrated by complex folding. A thickness of 4500 m is estimated at a locality not in the axial region.

Deep-water (siliceous shale, argillaceous chert) sedimentary rocks of middle Early Devonian age have been found in the subsurface on Banks Island, and this area is considered by MIALL (1976, p. 57, Fig. 11) to be part of the Hazen Trough. Sedimentary trends (but not later, structural trends) of the Franklinian Geosyncline thus may swing southward from Melville Island, at least for this late stage of the geosyncline's history.

In Early Devonian time, coarse flysch deposits (Imina Formation, upper part) accumulated in the northeastern part of the Hazen Trough. Similar sediments are absent, apparently, in the southwestern regions, though they may be hidden by younger cover. About 3000 m of Lower Devonian clastic red beds (Stallworthy Formation) on northern Axel Heiberg Island appear to represent a delta complex that prograded into the Hazen Trough (TRETTON 1969a).

The account, to this point, emphasizes the historical geology of the section of the Franklinian Geosyncline exposed on northern Ellesmere Island. Ordovician and Silurian sedimentation in the miogeosynclinal part of the basin, widely exposed from central Ellesmere Island through Grinnell Peninsula and Bathurst Island to Melville Island, is represented mainly by carbonate and evaporite facies (Copes Bay, Baumann Fiord, and Eleanor River Formations, Cornwallis Group, Allen Bay and Read Bay Formations). These facies interfinger northward with graptolitic shale and siltstone units (Cape Phillips, Ibbett Bay Formations). Both the carbonate-evaporite and the shale belts are overlain by clastic rocks (described below) representing changing tectonic conditions in Early Devonian and later times.

The main orogeny closing the history of the Franklinian Geosyncline began after rocks as young as Early Devonian age were deposited (in northern Axel Heiberg Island), but other disturbances took place at earlier times, as determined by the presence of unconformities and fossiliferous clastic formations. Certain of the earlier uplifts, which took place in Late Silurian to Early Devonian time, have resulted in anomalous or interfering patterns. Other structural trends may have been produced, of course, but now are masked by younger structures. Three such northerly-trending structural-stratigraphic regions have been identified: i) Boothia uplift, ii) Rens Fiord Uplift; iii) central eastern Ellesmere Island, including the Bache Peninsula Arch (Fig. 3a, b).

The Boothia Uplift and Cornwallis Fold Belt (Fig. 3a) form an elongate, horst-like structural high that extends northward some 1000 km from well within the Arctic Platform region in the south to cross the Franklinian miogeosyncline. The northern continuation of the structural high is hidden by the younger, Sverdrup Basin. Late Silurian to Early Devonian movements of the uplift in the platform region are recorded in syntectonic red sandstone and conglomerate beds of the Peel Sound Formation, a unit assigned a Pridolian to

Gedinnian age (THORSTEINSSON pers. com.). Northward, in the Franklinian Geosyncline, Early Devonian formations (Bathurst Island, Stuart Bay Formations) contain clastic tongues derived from the uplift zone, and other formations of the same age comprise clastic and carbonate conglomerate rocks that lie with local unconformity or disconformity on older beds, the unconformities disappearing away from the uplift (KERR and CHRISTIE 1965).

The Rens Fiord Uplift is a north-trending structural high of early Paleozoic rocks on northern Axel Heiberg Island. Steeply dipping faults separate a structurally complex core of the uplift from flanking rocks, which include beds as young as Early Devonian. An angular unconformity separates upper Middle Silurian strata from Early Devonian or older red beds. Thus, stratigraphic evidence suggests that the Rens Fiord Uplift became a positive feature during Late Silurian to Early Devonian time (TRETTIN 1969a, 1972).

The uplift region of central eastern Ellesmere Island is marked by a narrow, south-trending belt of probably early Middle Devonian red beds, (Vendom Fiord Formation), including conglomerate, of both syntectonic and post-tectonic origin. The red beds grade basinward into siltstone and shale, and indicate a source to the east, where now lies Precambrian crystalline terrain. The Bache Peninsula Arch at the north end of this region forms a northwest-trending projection into the Franklinian miogeosyncline; a local angular unconformity at the base of the Vendom Fiord Formation indicates contemporary tectonism in that area (KERR 1967c, 1967d). The early Devonian uplift may have trended parallel to the local (and regional) Franklinian depositional trends; if so, then unlike the Boothia and Rens Fiord uplifts, the eastern Ellesmere uplift lay parallel to the regional Franklinian trends. As with the Boothia Uplift, however, the younger structures of central eastern Ellesmere Island apparently conform to a Precambrian structural “grain”.

The youngest phase of sedimentation of the Franklinian Geosyncline is that represented by sequences, up to 2900 m thick, of clastic rocks of late Middle Devonian to Late Devonian ages. These rocks are nonmarine to shallow marine, commonly deltaic sandstone, siltstone, and shale with coal; the widespread units are named the Okse Bay Formation and the Melville Island Group. Sedimentological evidence indicates northern and northwestern sources for the clastic deposits,¹ which are thus apparently a “clastic wedge” derived from northern tectonic lands (TOZER and THORSTEINSSON 1964, p. 206). (These positive, tectonic lands that presumably lay along the northern edge of the continent constitute “Pearya” of SCHUCHERT 1923, or the Pearya Geanticline of recent literature – see TRETTIN 1971. The geanticlinal belt will be discussed later in this paper.) The thick clastic sections pass southward and southeastward into thinner carbonate and clastic sequences of the platform; a syn-form belt of foreland basins (see KING 1959, p. 28) thus lies along the southeastern boundary of the folded part of the Franklinian Geosyncline. These basins have been referred to as the Melville and the Devon Basins, respectively, west and east of the Boothia Uplift (see FORTIER, McNAIR, and THORSTEINSSON 1954; CHRISTIE 1972).

¹ But see also EMBRY 1976, and a later footnote in this paper.

A volcanic and clastic sequence (Svartevaeg Formation) some 3200 m thick overlies the lower Devonian red beds of northern Axel Heiberg Island. The overlying unit includes volcanic arenite, tuff, siltstone, conglomerate, volcanic breccias, and keratophytic and spilitic flows. Turbidites and submarine slides appear to be represented; the region evidently lay near the margin of the Pearya tectonic borderland (TRETTIN 1969a).

TECTONIC EVENTS

The Ellesmerian orogeny in Middle Devonian to Early Mississippian time produced the most widespread and intensive deformation in the Franklinian region (Fig. 4D), but other periods of deformation both preceded and succeeded it, and the resulting structural picture is complex. Earlier tectonic events took place in: a) late Precambrian time (north coast of Ellesmere Island); b) Middle Ordovician or earlier time (northern Ellesmere Island); c) Late Silurian to Early Devonian time (Boothia and Rens Fiord uplifts, and central Ellesmere Island); and d) Middle Devonian time (northern, eugeosynclinal terrain). The principal subsequent tectonic period is that of the latest Cretaceous to middle Tertiary, Eureka Orogeny, the main event of which took place in mid-Eocene time (see THORSTEINSSON 1974, p. 6-9).

a) *Late Precambrian orogeny*

Gneisses and schists of the north coast of Ellesmere Island have resulted from regional metamorphism to greenschist and amphibolite facies. From field structural and stratigraphic studies it is clear that the metamorphic rocks are older than Middle Ordovician beds. Several K/Ar isotopic age determinations that have been obtained are too young, apparently having been reset by later events, but whole-rock Rb/Sr age determinations recently carried out indicate a late Precambrian (minimum 742 ± 12 m.y.) age of regional metamorphism. This metamorphic terrain probably was the source of detritus that makes up the clastic formations in the lower part of the Franklinian geosynclinal sequence (SINHA and FRISCH 1975; FRISCH 1974).

b) *Middle Ordovician or earlier orogeny*

An orogenic episode of this age in northern Ellesmere Island is suspected from the presence, as noted earlier, of fossiliferous upper Middle Ordovician beds unconformably overlying weakly metamorphosed beds from which no fossils have yet been obtained. The underlying rocks, quartz-muscovite-chlorite schists, are part of a complexly deformed area of metamorphic rocks of presumed sedimentary and volcanic origin. The precise age, extent, and importance of this orogenic event are unknown (TRETTIN 1969b; 1972).

c) *Late Silurian to Early Devonian orogeny*

Caledonian orogenic activity in the Canadian Arctic Islands is represented by the stratigraphic records of the Boothia and the Rens Fiord uplifts and the synorogenic red beds of central Ellesmere Island. Additional evidence is the presence of volcanic rocks of Late Silurian age in northern Ellesmere Island.

Supporting evidence for the Caledonian age of these events has been provided by isotopic age determinations that cluster about 390 m.y. for several plutons on north coastal Ellesmere Island.

Uplift occurred repeatedly throughout the long history of the Boothia Uplift. The structural relief on this cross-trending zone of uplifts and basins increases from about 500 m in the platform region to 5300 m in the geosyncline (KERR and CHRISTIE 1965; KERR 1977). (The relief is due primarily to high-angle faulting in the platform region; to the north, in the miogeosyncline, KERR (in press) suggests a vertical complex of faults and folds). The Boothia structural zone appears to have acted as a “buttress” and to have resisted younger (Ellesmerian) deformation, structural trends of which lie nearly at right angles to the older structures.

The Rens Fiord Uplift, as noted earlier, has a northerly trending, structurally complex core. Existing structural trends are dominantly those of the later, Ellesmerian orogeny, but complex folds peculiar to the Silurian and older rocks are attributable to the Late Silurian – Early Devonian (Caledonian) deformation (TRETTIN 1969a; 1972, p. 129).

Both the Boothia Uplift and the Rens Fiord Uplift are overlain by a thick successor basin, the Sverdrup Basin, structural trends in the central part of which are aligned with structures in the older uplifts. Thus, the latest Cretaceous and Tertiary structures in parts of the Sverdrup Basin appear to represent a reactivation of earlier structural trends, and perhaps of earlier tectonic processes. KERR 1977 describes an arch and an anticlinal structure lying north of the Cornwallis Fold Belt (Fig. 3b) as the youngest, or highest structural expressions of the Boothia Uplift. The presence of Tertiary strata in small grabens within the boundaries of the Boothia Uplift provides some record of movement continuing into later times.

d) *Middle Devonian?, Acadian orogeny*

The precise age or ages of the climactic orogeny or orogenies that affected the Franklinian Geosyncline are not known directly from biostratigraphic data, particularly in the eugeosynclinal part of the sedimentary basin. In northern Axel Heiberg Island, for example, the basal beds of the Sverdrup Basin (Early Carboniferous, or Viséan) unconformably overlie rocks as young as Early Devonian. An Acadian (that is, pre-Ellesmerian) age of deformation in this region is possible, and is suspected because of the widespread presence of a clastic wedge of late Middle Devonian to Late Devonian age to the southeast, in the miogeosynclinal region. Supporting evidence for an early orogeny in the northern regions is provided by isotopic age determinations of about 360 m.y. for plutonic rocks, and the apparent absence of sediments of Middle Devonian age throughout the northern, eugeosynclinal part of the Franklinian basin. Further evidence, recently discovered in northern Ellesmere Island, is biostratigraphic: palynomorphs recovered from relatively undisturbed clastic bed-overlying the folded geosynclinal sequence have been assigned a Frasnian (early Late Devonian) age (U. MAYR pers. com.). This assignment supports the supposition that the cycle of orogeny was completed earlier in the northern

parts of the Franklinian region. (However, some caution might be used here until corroborating evidence shows that the palynomorphs were contemporary and not recycled).

A southeastward expansion of the Pearya Geanticline in Late Silurian time has been inferred by TRETIN (in press) from changes of detrital mineralogy upward in stratigraphic sections. Such a shift of exposed tectonic lands would have coincided with the southeastward shift, noted earlier, of the Hazen Trough after mid-Silurian time.

e) *Latest Devonian to Early Mississippian, Ellesmerian orogeny*

The principal orogeny that deformed the Franklinian Geosyncline, and particularly the miogeosynclinal part, is the Ellesmerian orogeny (THORSTEINSSON 1970). The elastic wedge deriving from the earlier, Acadian orogeny in the northern, eugeosynclinal regions was folded during the Ellesmerian event and later was overlain with angular unconformity by the basal beds of the Sverdrup Basin. The youngest beds of the essentially concordant geosynclinal sequence are Faménian in age, and the oldest beds of the successor basin, Viséan.¹ That Ellesmerian activity occurred also in the northern regions is suggested by isotopic age determinations of 335 ± 25 m.y. from plutonic rocks there.

The sedimentary history of the Franklinian Geosyncline thus appears to have been closed by tectonic episodes that extended over a rather wide interval of time: as early as early Middle Devonian to as late as Early Mississippian. Tectonic activity probably commenced in the north and advanced southward so that the detritus from early phases was incorporated in younger fold structures. Precise limits of areas affected by the older and younger climactic orogenies are not known; certain areas may have been affected by both.

Structural features of the Innuitian orogen

The Innuitian orogen² comprises the Franklinian Geosyncline, Pearya Geanticline, and Sverdrup Basin (see FORTIER, McNAIR, and THORSTEINSSON, 1954; TRETIN 1972). The Franklinian Geosyncline, already described, and the Pearya Geanticline form older elements of the orogen, and the superimposed Sverdrup Basin, a younger part. The older elements appear to range in age between late Precambrian and Devonian, and the younger, Mississippian to Tertiary. Orogenies affected the orogen, as noted, in late Precambrian time, in mid-Paleozoic time (Ellesmerian), and in late Cretaceous to early Tertiary time (Eurekan). The existing orogenic belt thus incorporates structural features of many ages. Younger structural trends conform to older in most parts of the orogen, and it is often unclear, locally, either how many events have taken place or to which event should be assigned the principal deformational features.

¹ The Frasnian beds noted in the preceding section appear to underlie the Sverdrup Basin with slight angular unconformity.

² The Innuitian Province includes northern Greenland; as noted in an earlier footnote, however, this paper is concerned mainly with the Canadian portion of the orogen.

The structural pattern of the Franklinian Geosyncline appears to be that of an open sigmoid; the axial line of the superimposed Sverdrup Basin is also sigmoidal, if perhaps slightly less pronounced (Fig. 3b). The Innuitian Orogen thus appears dominated by Ellesmerian (in the broad sense, including earlier, Acadian) northeastern trends. Certain structural zones, however, such as the older Boothia and Rens Fiord uplifts and those of the central part of the Sverdrup Basin are marked by trends that lie across the regional (northeast) pattern.

The principal or dominating structural features are described first in the following paragraphs in order to establish clearly the regional structural framework. Older and younger features are discussed later, rather than in chronological order (see TRETTIN 1972, 1973; THORSTEINSSON 1974). As with the stratigraphic account, the following description of structural features will emphasize relationships evident between central and northern Ellesmere Island, where the most complete cross-section of the Franklinian basin is exposed.

THE PRINCIPAL STRUCTURES: ACADIAN-ELLESMERIAN

The dominating Ellesmerian structures, with their sharply sigmoidal pattern, conform closely to the sedimentary depositional trends of the Franklinian Geosyncline. Thus, the structures trend southwest in the north, then swing southward, then westward. The structural style of features of Ellesmerian age varies from northwest to southeast, across the orogen as exposed on northern Ellesmere Island. The change in structural style may be due in part to differing position within the Ellesmerian orogen but undoubtedly also relates to differences in competence among sedimentary facies (see TRETTIN 1972, p. 129). Axial planes of folds in the axial region are generally steep, dipping northwest, although some dip towards the craton (CHRISTIE 1964; TRETTIN 1971). Low angle thrust faults are not a conspicuous feature; where these are present, they appear mainly to be of Tertiary (Eurekan) age, although some may be re-activated older faults. Some thrusts in northern Ellesmere Island are directed to the northwest. (The major, northwest-dipping Lake Hazen Fault Zone only is shown in Fig. 3b). Concentric, faulted folds prevail in the northwestern, volcanic eugeosynclinal region, where massive carbonate and volcanic units are present. However, the structural pattern in this region is complicated by metamorphic complexes and anomalous structural trends of uncertain origin. Isoclinal folds – that is, complex, tight folds with near-parallel limbs – prevail in the axial, flysch belt. Both normal and reverse faults are present in the axial region (TRETTIN 1971, 1972).

Folds are systematically spaced and concentric in the foreland fold belt to the southeast, with intensity of folding decreasing toward the craton. From the evident concentric limbs and flexural slip it appears that the folding extends to limited depths and must be of a décollement type, presumably having slipped on one or more of the thick and widespread lower Paleozoic evaporite deposits of the southwestern region, and on deeper, unknown surfaces in the northeastern region (TRETTIN 1972).

Regional metamorphism to low grades is characteristic of the northern regions, and the degree of alteration decreases southeastward from the north coast of Ellesmere Island. Greenschist and amphibolite facies are developed extensively in the metamorphic belt; metamorphism can be described as Barrovian in type, and there is evidence of widespread cataclasis and retrograde metamorphism. Metamorphism on a regional scale has occurred at different times and different places, however, so that the various metamorphic terrains do not form a single entity (FRISCH 1974).

OLDER STRUCTURAL ZONES

A zone of older folded and metamorphosed rocks of the north-coastal region of Ellesmere Island is broadly concordant with Ellesmerian structures. The Boothia and Rens Fiord Uplifts, on the other hand, are transected by the younger trends.

The northernmost, coastal exposures of the metamorphic terrain are amphibolite-facies granitoid gneisses and amphibolite, and extensive areas of greenschist facies rocks lie inland (see FRISCH 1974). Metamorphic grade thus appears to increase northward. However, cataclasis and retrograde metamorphism are widespread, and certain gneisses are overlain with sharp contact, in domal structures, by altered sedimentary rocks of distinctly lower metamorphic grade. The metamorphic rocks thus apparently form a complex. The southern limit of the main Precambrian orogeny is as yet uncertain because of difficulty distinguishing older from younger structures.

Foliation and banded structures, and the boundaries of metamorphic facies, trend mainly west and southwest, thus conforming approximately to the regional pattern.

The gneisses and schists, from the presence of repetitive layering and from their chemistry, are probably derived from volcanic and/or sedimentary rocks in part. From the ratio of strontium isotopes, a crustal rather than a mantle origin is suggested by SINHA and FRISCH (1975) for gneisses that may have been intrusive, granitic rocks. It is now clear from isotopic age determinations (minimum 742 ± 12 m.y., noted earlier) that these rocks, and an early orogeny that raised their metamorphic grade to the amphibolite facies, are older than the lower Paleozoic sediments of the Franklinian geosynclinal sequence. Steep attitudes of the gneissic structures presumably are due to the early orogeny, structural features of which are otherwise difficult to distinguish from those of later periods of deformation.

The gneisses, schists, and phyllites thus form the infrastructure of the Pearya Geanticline (see TRETTIN 1972, p. 125): although older than the Paleozoic part of the geosynclinal sequence, as just noted, they still may be equivalent to the oldest known Franklinian strata, or to conformably or disconformably underlying, non-metamorphosed strata not yet recognized.

The north-trending Boothia-Cornwallis structural belt lies nearly at right angles to the later, transecting Acadian-Ellesmerian structures. The Cornwallis Fold Belt is interpreted as a broad anticlinorium that overlies the Boothia

Uplift; the fold belt flanks the exposed core of the uplift in the Central Stable Region and continues northward beyond it, into the geosynclinal region. The north-trending structures in the geosyncline thus probably formed in response to vertical movements in the basement. The trends of high-angle reverse faults, probable horst-and-graben structures and related folds, mainly of Late Silurian–Early Devonian age, are about parallel to the gneissic structures of the Precambrian core over much of the length of the uplift (KERR 1974, 1977).

Northwesterly-trending structures are present in the Ordovician or older core of the Rens Fiord Uplift. These trends differ from the predominantly southwestward trends of northwestern Ellesmere Island. Dips are steeper and structure more complex in the core of the uplift, but the structures are difficult to distinguish in detail from later, Ellesmerian structures. From stratigraphic evidence (noted earlier), the complex structures in the older rocks are attributed to Late Silurian–Early Devonian deformation (TRETTIN 1969a, 1972).

YOUNGER STRUCTURES

A few folds and faults of Early Permian age have been identified at scattered localities from Ellesmere Island to Yukon; these structures represent the Melvillian Disturbance (see TRETTIN 1972, p. 132). Such structures in the Arctic Islands are superimposed on the earlier, Ellesmerian trends, so their true extent and origin are uncertain.

The large, successor sedimentary basin, the Sverdrup Basin,¹ with its partly conforming structural pattern, is an intriguing feature of the Innuitian Orogen. The overall conformity of structural trends of different ages naturally leads to speculation that major structures in the Sverdrup Basin reflect underlying features, in spite of the considerable, blanketing thickness of rock in the younger basin.

The Eureka Orogeny (see THORSTEINSSON and TOZER 1970, p. 585) ended sedimentation in the Sverdrup Basin and imprinted a widely spread pattern of folds and faults on the Innuitian Orogen in latest Cretaceous to mid-Tertiary time. Eureka structures include a wide variety of folds, thrust faults, and normal faults. Tectonism was most intense on Ellesmere and Axel Heiberg Islands. In the western islands, in contrast, folds are of low amplitude. Eureka and the older, Ellesmerian structures are concordant; however, Eureka folds and thrust faults extend into miogeosynclinal parts of the Franklinian Geosyncline that had not been affected by the Ellesmerian Orogeny.

Thrust faults of probable Tertiary age trend mainly northeast, with motion directed southeastward; some of these faults occur on the southeast sides of tectonically high, mountainous areas and apparently bound major uplifts. Northwesterly-directed thrust faults, normal faults, and elongate grabens also are present. Some of the faults probably represent rejuvenation of motion along Ellesmerian structural trends.

¹ General accounts of the Sverdrup Basin have been written by THORSTEINSSON and TOZER (1970), PLAUCHUT (1971, 1973), and NASSICHUK (1972).

A REVIEW OF SOME TECTONIC FEATURES OF THE INNUITIAN REGION

Certain tectonic features and events of the Canadian portion of the Innuitian region, although of uncertain origin, appear potentially significant in achieving an understanding of the tectonics of the region. Among these are: coincidence of basinal axes of different periods; parallelism of structural trends of different ages; the sharply sigmoidal form of the fold region; and transection of the orogen by a fracture zone of a wide range of ages.

The Franklinian Geosyncline now is structurally complex and deeply eroded. In contrast, the successor, Sverdrup Basin is relatively simple. But the two basins have some features in common: near coincidence of depositional axes, and comparable depth of sedimentary fill (see Fig. 3b; Table 1).

The Sverdrup Basin, although evidently at no time larger than the Franklinian Geosyncline, may have been nearly co-extensive and contained an almost equal thickness of sediments. Estimates of maximum thickness for the Sverdrup Basin range from about 10 000 m to 14 000 m; for the Franklinian Geosyncline, 9000 m to 15 000 m. Such repetition of episodes of downsincking and accumulation of major volumes of sediment in one linear belt may have been due to repeated operation of a single mechanism. A thinned crust beneath the axis of the Sverdrup Basin is inferred from gravity data (SANDER and OVERTON 1965), and speculation based on the magnetic pattern of the Sverdrup Basin questions whether a Precambrian basement is present beneath the basin, or whether the basement is of a different character (RIDDIHOUGH et al. 1973). A geotectonic model that allowed repeated thinning of the crust would be favoured for the Innuitian region.

Table 1.

Some recent estimates of total thickness and volume of sediments in Sverdrup and Franklin basins.

	Sverdrup Basin	Franklinian Geosyncline
THORSTEINSSON 1974	12,000 m (40,000 ft)	15,000 m (50,000 ft) Late Precambrian to Late Devonian
DRUMMOND 1973, p. 450	12,000–14,000 m (40–45,000 ft)	14,000 m (45,000 ft)
	1,500,000 km ³ (350,000 mi ³)	1,100,000 km ³ (273,000 mi ³)* * Cambrian to Devonian of the “Franklinian Fold Belt”; vol- ume underlying Sverdrup Basin and northern Ellesmere I. eugeosynclinal belt not included.
THORSTEINSSON and TOZER 1970	10,000 m (35,000 ft) 600,000 km ³ (150,000 mi ³)	12,000 m (40,000 ft) Cambrian to Devonian

Thinning of the crust and subsidence of the axial region of a basin are not provided for specifically in the recently developed theory of plate tectonics although subsidence of a *continental shelf* does figure in the formation of a “flysch-molasse basin accumulating continent-ward directed sediments” (see DEWEY and BIRD 1970a; 1970b, p. 2639, Fig. 10C). However, this does not appear to be a good description of the Sverdrup Basin.

The Sverdrup Basin may be an unusually well-developed, fault-bounded trough, perhaps owing its size and depth to massive reduction of the root zone of the Franklinian Orogen by convection in the mantle (see LAMBERT 1973, Fig. 5). McCROSSAN and PORTER (1973, p. 618, 668–673) have classified the Sverdrup Basin as a “rift basin” from its trough-like form, thick sedimentary fill, and extremely abrupt thickening from the edges to the basin centre. Alternatively, the Sverdrup Basin might be related to a “new” continental edge, a result of rifting that preceded it, and its coincidence with the Franklinian Geosyncline may be more or less fortuitous. This possibility is discussed in a later section of this paper.

Structural trends of different ages are markedly parallel in many parts of the Innuitian region. For example, folds in the eastern margin of the Ellesmerian fold belt parallel gneissic trends in the adjacent Canadian Shield. Similarly, Eureka trends in the eastern part of the Sverdrup Basin lie parallel with those in the adjacent Ellesmerian Orogen, and approximately conform to the trend of otherwise transecting structural zones such as the Boothia Uplift. This uplift, in turn, follows trends in its gneissic core, at least in its northern and greater part.

It is possible that the prominent Boothia structure, which is clearly an expression of a basement feature, extends beneath the Sverdrup Basin, and one might suppose (as KERR 1977 does – noted earlier) that basement structures have influenced or controlled structural trends in the successively younger basins. TRETIN (1972), p. 165) describes as a “transverse ridge” four or more en echelon uplifts, including the Boothia Uplift, of various ages that cut across the depositional axes of the Franklinian and Sverdrup sedimentary basins. These are taken (op. cit., p. 127) to be expressions, possibly, of a region of basement-controlled, Siluro-Devonian block faulting.

Can older structural trends reappear through a blanket of sediment at least 10 km thick? The northern part of the Canadian Shield is characteristically heterogenous and variable in structural grain, and the large, linear sedimentary depressions (both Franklinian and Sverdrup) apparently cross the basement structures, little affected by them. The parallelism may be more apparent than real: trends perhaps conform only at basin-edges, where the younger cover is thin, or a degree of convergence may result from, say, a sharp flexing of the regional structure.

A notable feature of the sharply flexed, southeastern boundary of the Franklinian sigmoidal form is the absence of late Precambrian sediments along the margin of the Franklinian sedimentary basin. Some 3000 m of late Precambrian sedimentary rocks are exposed in northern east-coastal Ellesmere Island, and late Precambrian rocks also underlie early Paleozoic beds on

northern Victoria Island, in the platform region. (The western extension of the pre-Paleozoic Franklinian basin edge is hidden beneath a Paleozoic cover). Between Bache Peninsula and the Boothia region, however, the Precambrian gneisses of the Canadian Shield are overlain directly by Cambrian beds. This absence of late Precambrian beds could be due to non-deposition – that is, simple onlap of the basal Paleozoic beds onto the crystalline shield; or, it could result from uplift and erosion of older beds in latest Precambrian – earliest Paleozoic time. In either case, the absence of the late Precambrian beds may signify an early expression of tectonic activity possibly relating to the “sigmoid”. It can be observed that the flexure is directed like a broad arrow down Baffin Bay. Or, put another way: the coincident, southeast-trending embayments of the Franklinian and Sverdrup basins are aligned with the very large scale Baffin Bay-Labrador Sea embayment. The lower Paleozoic rocks of the Innuitian “embayment” region are of platform- and limited-circulation basinal type (carbonate and evaporite). Deeper basinal facies of the Sverdrup Basin, however, are known to extend far to the southeast, into the “flexure” of the sigmoid (W. W. NASSICHUK pers. comm.). From the alignment of the Innuitian Paleozoic and Mesozoic embayments with Baffin Bay it seems possible that the latter feature may have been tectonically low at those times and that Paleozoic and Mesozoic sedimentary rocks may be found there. This possibility was suggested earlier by FAHRIG, IRVING, and JACKSON (1971, 1973), who reasoned that a northwest-trending, late Precambrian dike swarm of Baffin Island is evidence of crustal tension of that age.

It appears, indeed, from geophysical data and dredging samples recently obtained, that remnants of a sedimentary basin, or a series of basins, of Mesozoic and possibly Paleozoic age may occur along the continental shelves of Baffin Bay, Davis Strait, and Labrador Sea (JOHNSON et al. 1975; GRANT 1975; and see Fig. 2).

BEH (1975) has proposed a sedimentary trough in Baffin Bay and Davis Strait that received sediments as early as Jurassic time, the age based on the probable time of rift-opening of the seaway. VOGT and AVERY (1974) have speculated on a possible connection between the Alpha Ridge of the Arctic Ocean and the Labrador Sea basin, the ridge being perhaps a “fossil” mid-oceanic ridge. Such a connection would of course pass through the sigmoidal embayment; no field evidence is known, however, to indicate a spreading ridge at this locality, and this line of speculation apparently is unsupported.

The lack of evidence just noted could, however, be ascribed to *incipient* development of a spreading centre or ridge in late Proterozoic time; thus, a domed or uplifted zone might explain the absence, either through non-deposition or erosion, of late Precambrian beds in the “flexure” region of the Innuitian sigmoid. Such an early tendency to ocean spreading in Baffin Bay, if it occurred, presumably was transformed to early transcurrent faults along Nares or Parry Channels.

The youngest major structural features expressing orogenic processes in the archipelago are the rifts, evidently of late Mesozoic or Tertiary age, that now are followed by large seaways and channels. The two largest rifts, Parry Chan-

nel and Nares Strait (Fig. 3b), cut obliquely across the overall northeasterly Innuitian trend.

Parry Channel, comprising Lancaster Sound, Barrow Strait, Viscount Melville Sound, and M'Clure Strait, nearly follows the 74th parallel of latitude. From geomorphological and geophysical data it appears that the feature is graben (M'Clure Strait) and graben or half-graben (Lancaster Sound) in form (GREGORY, BOWER, and MORLEY 1961, p. 24; BARRETT 1966; DAAE and RUTGERS 1975; M. J. KEEN et al. 1972). A northwestward extension of this rift feature appears to transect Ellesmerian trends, and certainly transects – at right angles – the southwestward extension of the Hazen Trough suggested by MIALI (1976, p. 51, Fig. 11). The relationships here, however, are obscured by young deposits of the Arctic Coastal Plain and by widespread inundation by sea water. Few geophysical or subsurface data are yet available; so that much remains speculative.

Nares Strait, comprising Kane Basin, Kennedy Channel, and connecting straits to north and south, trends north-northeastward from the head of Baffin Bay, and on a recent bathymetric chart (HEEZEN and THARP 1975) the lineament can be seen to extend northeastward along the eastern margin of Lincoln Sea. The Nares Strait lineament will cross the termination of the Lomonosov Ridge if that ridge terminates against Greenland as shown on the recent chart.¹ Northeastward, north of Peary Land, the lineament is represented by the western edge of the Morris Jesup Plateau. A distinct magnetic “low” follows the western side of Nares Strait but is broken by a “high” near latitude 79°30' N. The “high” crosses the lineament, seemingly unaffected by it (RIDDHOUGH et al. 1973).

The Nares Strait lineament is relatively straight and narrow; this is the Wegener Fault of WILSON (1963). The lineament probably does mark a zone of transcurrent movement, but the net amount of translation that has occurred is unknown. Major stratigraphic features such as the margins of the Franklinian Geosyncline and the Thule Basin appear very little, if at all displaced (KERR 1967a; CHRISTIE and DAWES in prep.), but uncertainty arises due to the acute angle (about 25°) at which the lineament crosses stratigraphic and structural trends: possible offset trends are accommodated easily in mapping. Both horizontal and vertical components of displacement on coastal faults along the northern part of the lineament were recognized in the field (CHRISTIE 1964, 1974), but no evidence was found suggesting horizontal displacement of more than a few kilometres. The continuity across Nares Strait of the magnetic “high” mentioned above also argues against significant horizontal displacement, as RIDDHOUGH et al. (1973) have noted.

The Nares Strait lineament, a younger, major structure of the Innuitian Orogenic system, is of interest in the present context if, as is the case for other structural features discussed previously, it was a locus of repeated tectonic

¹ The new chart, unlike earlier ones, shows the Lomonosov Ridge approaching western Peary Land, Greenland, as a straight and prominent, rising topographic ridge. The bathymetry of this region, however, is based on sparse data and may not be reliable (E. F. Roos pers. com.).

activity. The Nares Strait lineament must, in fact, be an older and more complex structure than first appearances might suggest; it forms, at least, a dividing line between contrasting structural regions that record differing patterns of mid-Paleozoic tectonic events: tectonic transport in folds and faults is dominantly southward west of the strait, and northward to the east (see KERR 1967a; DAWES 1973). This dividing line is inferred to have existed as some linear structural feature, probably a compressional fault zone with incipient strike-slip movement, during an earlier, presumably mainly Ellesmerian period of compression. The history of Nares Strait thus appears to be a long and complicated one. (For a review of the difficulties in comparing the geology on either side of the strait and of the restraints on overall movement see DAWES 1973, p. 938–939.) A line of fracture cutting diagonally across a major orogenic zone and expressed at more than one time suggests a deep-seated feature.

Little agreement has been reached on the positions and history of Greenland in the spreading opening of the North Atlantic Ocean, and some proposed positions cannot be reconciled with landward geology. Possibly, movement in both directions has taken place along the Nares lineament; that is, the lineament may be a zone of both sinistral and dextral transcurrent “adjustment” between Greenland and North America, and the net movement may thus be incidental rather than a measure of the importance of the zone.

Tectonic development of Svalbard and the Innuitian region

Both Svalbard and the Innuitian region lie at the edges of large continental masses: Svalbard at the northwestern “corner” of the Barents Shelf, and the Innuitian region along the northern edge of the North American craton. Modern theories of ocean-floor spreading invite hypothetical reassembly of earlier or supposed “supercontinents”; this exercise requires a review of the essential geological features of two regions that may have been juxtaposed.

Some possible tectonic settings of Svalbard and the Innuitian region will first be examined. Following this, the geology of the two regions will be viewed with the aim of determining connections that may have existed in the past.

TECTONIC MODELS TO ACCOUNT FOR THE ARCTIC OCEAN BASINS AND THE GEOLOGY OF SVALBARD

The Arctic Ocean Basin is small among ocean basins, and surrounded by continents. Various physical features are present, some of which clearly indicate rifting and sea-floor spreading whereas others are of less certain origin. Contrasting hypotheses have been formulated to account for the basin, and these have been described and appraised by CHURKIN (1973) and by CLARKE (1975), among others (see VOGT and AVERY 1974, p. 84).

The Arctic Ocean Basin is divided into two major regions separated by the Lomonosov Ridge: the Eurasian Basin off the Barents and Kara Shelves, and the Amerasian Basin off Canada and Siberia (see Fig. 2). The Eurasian Basin is elongate and marked by a median ridge, the Nansen Cordillera (or Gakkel Ridge); the Lomonosov Ridge and the opposing continental edge appear to

form a good geometric fit. The Amerasian Basin is roughly triangular, with a straight margin along the Canadian Arctic Islands and an irregular one on the Siberian side. The irregular Alpha Ridge (comprising the Mendeleev Ridge and the Alpha Cordillera) divides the Amerasian Basin asymmetrically with the large and deep Canada Basin lying north of Alaska.

An early concept of subsidence of a continental crust (that of SHATSKY and others, 1935 and later; see CHURKIN 1973) to explain the Arctic Ocean basins has given way to hypotheses applying one concept or another of rifting of continental masses or of geosynclinal development and continent-margin tectonics around a very ancient or “proto-” Arctic Ocean basin.

An orocline-and-rift theory proposed by CAREY (1955) interpreted the oroclinal bend of southern Alaska as complementary to a wedge-shaped opening of the Arctic basin. In a sea-floor spreading theory (VOGT and OSTENSO 1970), the Alpha Ridge is interpreted as a “fossil” spreading centre and the Nansen Cordillera as an active extension of the mid-Atlantic Ridge. Different geometric schemes have been incorporated with the seafloor-spreading concept to provide variations on the theory; e.g. early spreading at right angles to the present basin margins (TAILLEUR 1973) to account for the deep part of the Canada Basin north of Alaska; and Paleozoic collision of a “Kolymski Block” (of eastern Siberia) with North America and subsequent backward drift in Mesozoic time to open the Amerasian Basin (HERRON et al. 1974).

The existence of a Proto-Amerasian Basin in early Paleozoic or even Precambrian time was postulated by CHURKIN (1973), who showed that a series of continent-margin geosynclines may have existed in early Paleozoic time around the circumference of a Proto-Canada Basin. OSTENSO and WOLD (1973) also proposed a primeval sea, the “Hyperborean Sea”, to account for the Amerasian Basin, in this scheme an ocean area remaining throughout the collision of three continental plates: the North American, Russian¹, and Siberian platforms. In all variations of the sea-floor spreading theory, the Eurasian Basin is accounted for by the splitting away of a continental sliver (the Lomonosov Ridge) during Cenozoic time, so that the eastern Arctic Ocean (as viewed from Canada) is, in origin, a northward extension of the North Atlantic Ocean and Svalbard has been removed from a position north of Ellesmere Island or Greenland.

Svalbard, in a pre-drift position north of Greenland, lies close to the junction of East Greenland (Caledonian) and Innuitian (Ellesmerian) structural trends. In its geology, too, Svalbard may reflect the tectonic history of both structural regions (see Fig. 5). The age of the main orogeny, Caledonian, suggests a close relationship with the folded East Greenland geosyncline, whereas the structural trends (in the reconstructed position) conform most closely to Ellesmerian trends. Svalbard evidently lies near a major junction of continental masses, and near the centre of a possible former “supercontinent”; the relationships, therefore, between structures of Svalbard and those of formerly surrounding regions may be critical to an understanding of the tectonic history of the polar regions.

A model proposed by HARLAND (1965, 1966, 1975b) begins with an essentially Caledonian Svalbard (that is, relating to the East Greenland geosyncline)

¹ Eurasian platform in the present discussion (see Fig. 2b).

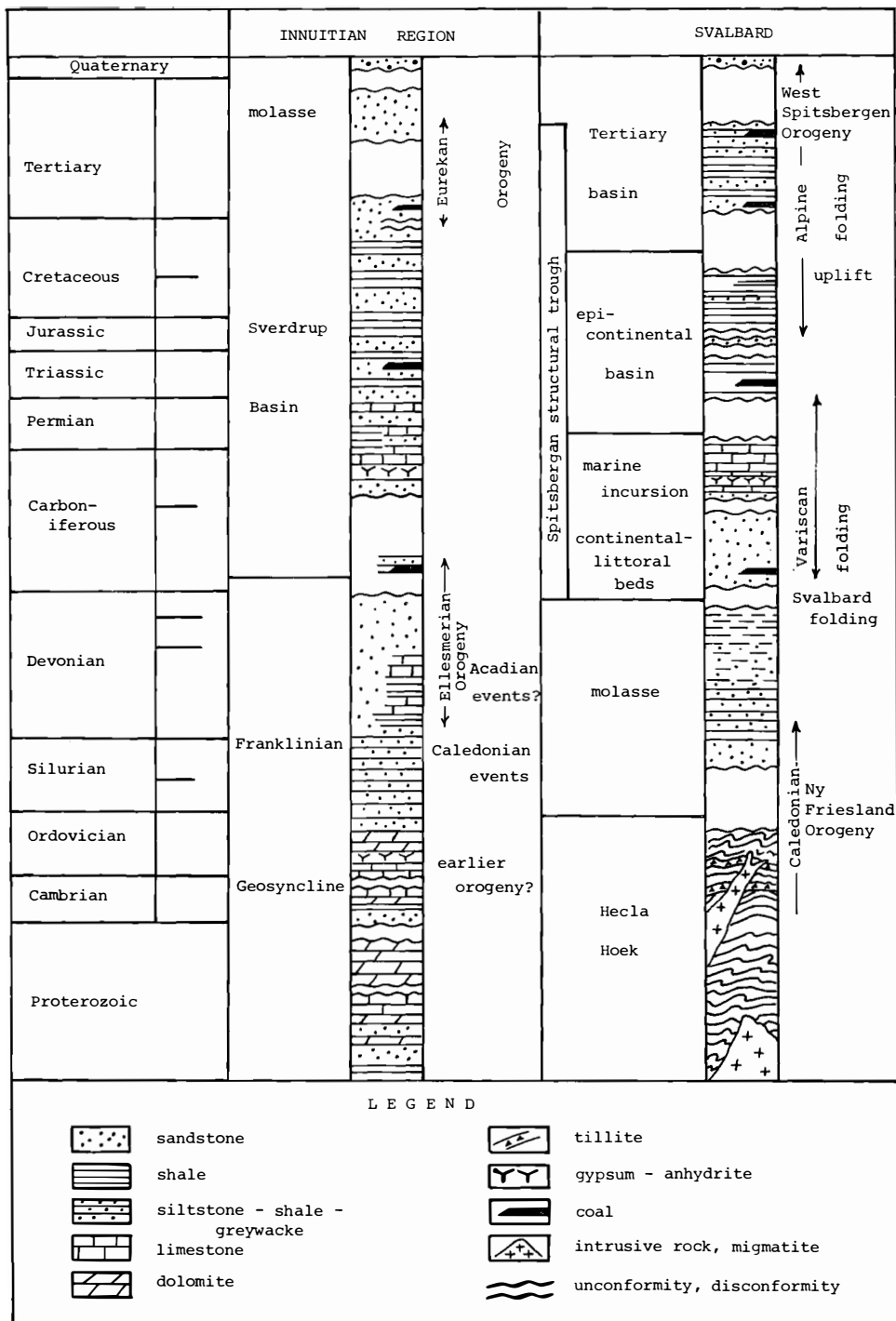


Fig. 5. Comparative columnar sections, Innuitian region and Svalbard (Svalbard from ORVIN 1940, Pl. IV).

then, through Late Devonian sinistral plate motions, juxtaposes Svalbard and northeast Greenland. The Billefjorden Fault Zone, a major fault zone separating eastern and western Svalbard, is proposed to mark a line of substantial transcurrent displacement such that western Svalbard derived from a position adjacent to Peary Land while eastern Svalbard was related to central east Greenland. Svalbard and Peary Land then separated in late Phanerozoic time by dextral transcurrent movement along the de Geer line.

This scheme, however, requires substantial changes in directions and styles of motion (and tectonic processes?) along rather closely related lines of shear (e.g. HARLAND 1966, Fig. 4, 6).

The model proposed in this paper, though in some ways no simpler, also may accommodate the data (see Fig. 6): in this case, the Svalbard and Franklinian geosynclines are presumed to have been a “pair”, divided during certain periods by an intermittently positive tectonic welt, the Pearya geanticline of TRETTIN (1972, 1973). The paired geosynclines were thus separate but related. A geological history for the combined region will be proposed after a review of certain geological features.

The geology of Svalbard and the Innuitan region

An overview of the geology of the Innuitan region and Svalbard discloses both marked similarities and substantial differences. Some sedimentary and tectonic features are compared in Table 2. It seems clear (Table 2; Fig. 5) that the tectonic histories of the two regions can be divided into two main periods: an earlier, pre-Late Devonian period during which the histories differ, and a later, post-Devonian period during which the sedimentary and tectonic styles become distinctly similar. From late Precambrian time, when the known sedimentary record begins, to early Devonian time, the tectonic histories of the two regions appear to have little in common. A convergence in styles began during the Devonian; then, in early Tertiary time, certain differences reappear. These features have been recognized, and certain tectonic reconstructions are based on them (see HARLAND 1965, 1969a, 1969b, 1975b; HARLAND and GAYER 1972).

PRE-CARBONIFEROUS TIME

Some sedimentary-tectonic features of the two regions for this period appear, at first glance, to be rather similar: thicknesses of 15 and 19.8 km, respectively, of geosynclinal sediments pre-dating the main closing orogenies, or 15 and 28.8 km, respectively, of geosynclinal and molassic sediments before Carboniferous time (see Table 2a; Fig. 5). A major break in sedimentation took place in both regions by early Carboniferous time. However, although the complex geology of Svalbard is not yet explored thoroughly, it is evident that differences of a major order exist; the two geosynclines, while possibly connected by seaways, were nevertheless of different sedimentary-tectonic regions. Dissimilarities in sedimentary style appear when the sedimentary columns of the two late Precambrian–early Paleozoic geosynclines are compared: only

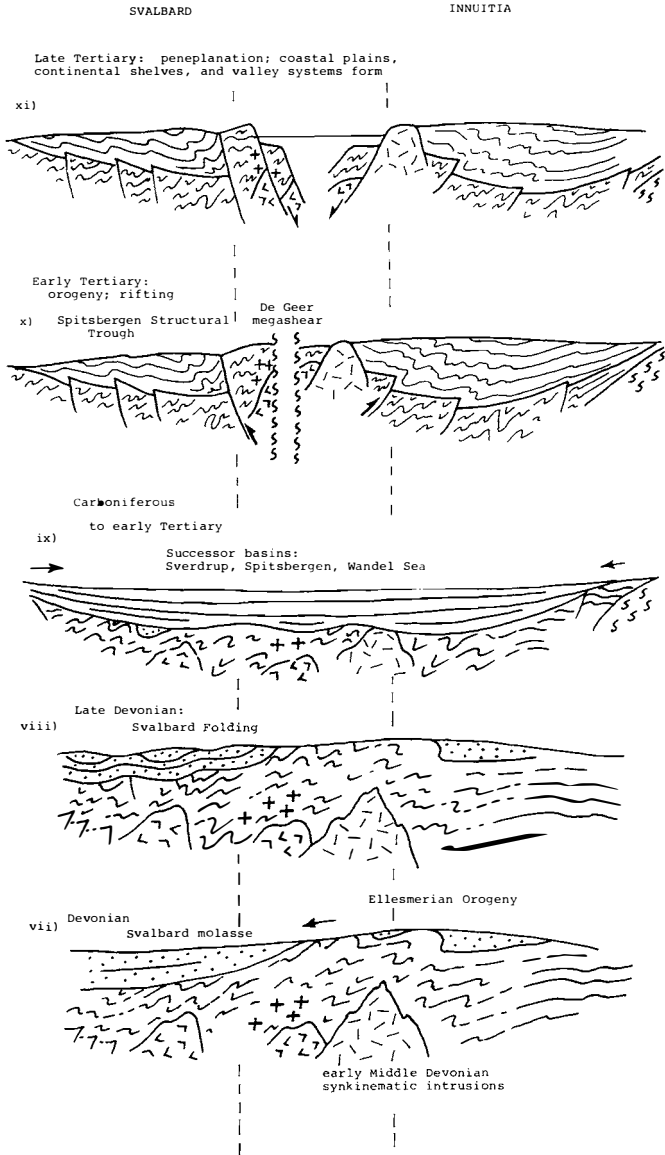


Fig. 6. Diagrammatic representation of possible stages in the development of Svalbard and Innuitian geosynclines. Arrows indicate directions of sedimentary transport.

1.7 km of late Precambrian beds are known in the Franklinian basin, whereas 18.6 km have been measured in Svalbard (see Table 2b; Lower and Middle Hecla Hoek). The proportions are reversed in early Paleozoic time: 8.3 km for Cambrian to Devonian time in Canada vs 1.2 km for Cambrian through Ordovician time (Upper Hecla Hoek) in Svalbard, with most of Silurian time not represented in Svalbard due to elevation of the region during the Caledonian orogeny.

A detailed comparison of sedimentary characteristics (Table 2b) might be interesting, but probably is not justified until more certain or complete

Fig. 6 cont.

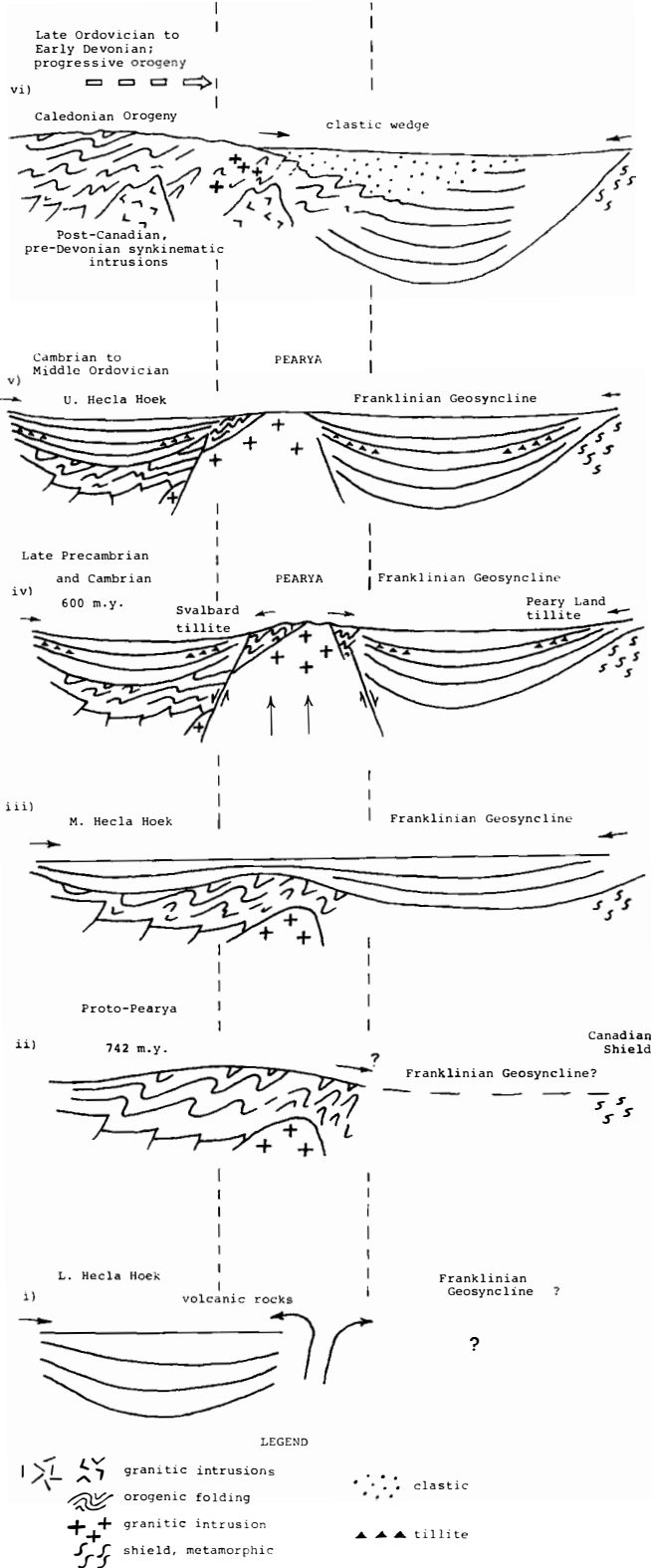


Table 2.

Sedimentary and tectonic features of the Inuitian region and of Svalbard compared. Superscript numbers refer to authors listed in table below. K-Ar, Rb-Sr: isotopic age determinations in millions of years (m.y.). “>” signifies “greater than”.

	Inuitian Region		Svalbard	
a) Thickness and ages of sediments	Arctic Coastal Plain and late Cenozoic basins	2.5 km?	Paleocene-Eocene	1.5 km
			(up to 3.5 km) ²⁵	
	Sverdrup Basin (L. Carb.-Eocene)	12 km	L. Carb.-L. Cret.	4 km
			Old Red ss (L. & M. Dev.)	9 km
Franklinian: (late Prot.-Dev.) refs: 1,3	15 km	Hecla Hoek (late Precambrian-M. Ord.)	20 km	
		refs: 2,9		
b) General lithology Thickesses are maxima; many units range from zero thickness	Arctic Coastal Plain and late Cenozoic basins: sand, gravel, silt, wood, peat refs: 3,6	0.1-3 km?	Continental basin Paleocene-Eocene: sandstone, shale, coal	2 km
			(3 km ²¹)	
	Sverdrup Basin Early Cenozoic conglomerate, sandstone, shale, coal	3 km	Epicontinental basin L. Trias.-L. Cret. shale, sandstone, limy sandstone, limestone, coal, conglomerate	2.8 km
			refs: 8,9	(2.4 km ²⁴)
	Triassic -Cret. sandstone, siltstone, shale, basalt	7.6 km	Continental beds and marine incursion	
			M. Carb.-L. Perm. limestone, sandstone, conglomerate	0.7 km
	U. Perm. sandstone, chert, limestone	0.8 km	refs 8,9	
	U. Carb.-L. Perm. conglomerate, sandstone, shale, limestone, evaporites	2 km	L. Carb. sandstone	1.2 km
L. Carbonif. siltstone, sandstone, coal refs: 1,3,6,7	0.3 km	Molasse (Old Red) Middle Dev. shale	2 km	
		Late L. Dev. sandstone, shale	5 km	

Table 2 cont.

	Innuitian Region	Svalbard
	Franklinian M. & Late Dev. sandstone, siltstone, limestone, shale	Downton-Ditton sandstone, conglomerate, siltstone
	5 km	refs: 8,9 2 km
	Camb.-Early Dev. limestone, dolomite, shale, siltstone, sand- stone, greywacke, chert, conglomerate	Hecla Hoek U, carbonates Camb.-Ord.
	8.3 km	1.2 km
	Late Prot. sandstone, dolomite, shale, limestone	M, tillites, carbonates, greywacke, quartzite
	1.7 km	7.1 km
	refs: 5,6,7	L, meta-volc., acidic & basic; quartzite, marble, meta-tilloid ref: 4 11.5 km

c) Provenance of
sediments

late Cenozoic	(Beaufort Fmn) ¹⁴ SE, E	
Oligocene- Miocene?		
Eocene		W (S) ²¹
early Cenozoic	“local” ¹⁴	E (N) ²¹
Paleocene	(L. Cret. Isachsen Fmn: S ²³)	E (N), NE (NW) ^{21,27}
Mesozoic	E, S (except certain Lower Triassic to Jurassic rocks: N, NW ¹⁴) Triass-Jur.: S, E ¹⁶	Jur.-Cret.: S?, SE? (E?) ⁸ N (W) ¹⁷ FREBOLD in 8 Mes.: W (S) Mes.-Tert. W (S) ^{19,8}
Permian, Carboniferous	in NW: NW ^{1,3} in SE: SE (“local”) ¹ “no source to N” ¹⁶	
Devonian	W, NW ⁶ E, NE ²²	W (S) ¹⁷ , E? (N?) ¹⁷ S (E) ²
Late Silurian Early Devonian		W (S) ⁸

Table 2 cont.

	Innuitian Region	Svalbard
Ordovician– Silurian	flysch: NW ¹⁵	
Cambrian late Precambrian	NW ¹⁵ SE margin: SE ⁵	Hecla Hoek: “possibly a flanking . . . volcanic ridge”: W (S) ref: 17
() – direction adjusted 90° to conform with meridian of northern Ellesmere Island – Peary Land.		
d) Tillite and “tilloid”	absent or not recognized	(c) In Polarisbreen Formation, ca. 800 m below L. Cambrian fossil horizon; late Pre- cambrian (Vendian). (b) = Varangian tillite (15 km separate tillites) (a) meta-tilloid (tillite?) in Rittervatnet Fmn.; earlier late Precambrian. ref: 8,9
e) Main orogenies and peak metamorphism	Eurekan (mid-Eocene to Oligocene) ^{14,28} Ellesmerian (post-Famennian, pre- Viséan) ⁶ K-Ar determinations inconsistent ³	Alpine (“West Spitsbergen”) ²⁰ Orogeny) E-W (N-S) compression; overthrusting, folding to E, NE (N, NW) ⁸ Miocene? post-Paleocene or -Eocene ⁹ Subhercynian or ⁸ Laramian? uplift; Late Cretaceous Svalbardian Folding (Late Devonian) ⁸ Caledonian ⁸ (Ny Friesland Orogeny) ⁹ (Silurian ⁸ ; post- Canadian, pre-Downtonian ⁹) K-Ar } 430–420 my ¹² Rb-Sr } 390–380 my ? Pre-Caledonian event: ? early Cambrian or late Precambrian ^{9,12} K-Ar 600 my ¹²

Table 2 cont.

	Innuitian Region	Svalbard
	Late Precambrian	
	Rb-Sr 742 my ¹¹	
f) Volcanic rocks	Basalt flows:	“Tuff-conglomerate”:
	Upper Cretaceous ⁶ ;	Paleocene-Eocene ²¹
	mid-Lower Cretaceous ⁶ ;	mid-Lower Cretaceous ⁸
	lower Lower Cretaceous ⁶	Barremian (?) volcanism ²⁶ (mid-L. Cret.)
and/or	Bentonitic and tuffaceous layers: U. Cretaceous ²⁹	Diabase dikes:
	Diabase dikes: Mid-Cretaceous:	Berriasian (?) sills ²⁶ (early L. Cret.)
		Early to Late Cretaceous? ⁸
	K-Ar 102-110 m.y. ³	mid-Cretaceous (100 m.y.) ²⁵
late, basic intrusions		Mid-Mesozoic: K-Ar ca 150 m.y. ¹²
(diabase, dolerite, gabbro)		(Probably late Jurassic or early Cretaceous, from agreement between radio-metric and structural-stratigraphic data) ¹²
		Distributed along S, SE (E, NE) trending fracture lines? ⁸
	Post-Ellesmerian (early to mid-Mississippian?; trends NE, E) ³	
		Caledonian (Silurian) ⁸
		2 generations of metamorphosed dolerites in Hecla Hoek succession ¹⁰
g) Lamprophyre dykes		K-Ar 309 m.y. ¹²
h) Post-tectonic intrusions (granite)	Late Middle to Late Devonian? K-Ar 360 ± 25 m.y. ³	Silurian-Devonian boundary, Lower Devonian
		K-Ar } 400 ± 20 m.y. or ? Rb-Sr } slightly younger ¹²
i) Synkinematic intrusions	Early to Middle Devonian ^{3,13}	

Table 2 cont.

	Innuitian Region	Svalbard
		Post-Canadian, pre-Devonian
(granite, migmatite)	K-Ar 376 ± 16 to 390 ± 18 m.y. ¹³	K-Ar Rb-Sr } ca 450–400 m.y. ¹²
j) Basement	Northern boundary: gneiss $> 742 \pm 12$ m.y. predates overlying meta- sediments and may be basement ¹¹	Existance of exposed basement doubted, or contentious ¹⁰ ; 'Archean' not known.
	Southern boundary: Aphebian, Canadian Shield (K-Ar ca 1735 m.y.) ¹⁸	

Source references for comparative table

- | | |
|---|---|
| 1. THORSTEINSSON 1974. | 15. TRETIN 1971. |
| 2. HARLAND 1969a. | 16. MENELEY et al. 1975. |
| 3. TETTIN 1972. | 17. HARLAND 1959. |
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| 8. ORVIN 1940. | 22. EMBRY 1976. |
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| 11. SINHA and FRISCH 1975. | 25. LIVŠIĆ 1974. |
| 12. GAYER et al. 1966. | 26. HARLAND 1973. |
| 13. FRISCH 1974. | 27. BIRKENMAJER 1972. |
| 14. TOZER In : THORSTEINSSON and TOZER 1970. | 28. BALKWILL, BUSTIN, HOPKINS Jr. 1975. |
| | 29. MIALI 1975. |

knowledge of the facies distribution in the two geosynclines is available. On the other hand, certain distinctive rocks such as tillite (see Table 2d) or gypsum-anhydrite would be most useful if they could be identified in the two areas. Tillite has not been recognized in the Canadian islands, however (although it occurs in Peary Land, northern Greenland – see DAWES and SOPER 1973, p. 121), nor evaporites in early Svalbard rocks.

Major orogenic events in adjacent regions should be reflected in the stratigraphic record of each region. On this basis, it appears that Svalbard and the Innuitian Region were not contiguous (Fig. 5; Table 2e): except for the Svalbard episode of folding, orogenies in the two regions have been consistently out of phase since late Precambrian time. The Svalbard folding does coincide with part of the Ellesmerian Orogeny and these tectonic events in each area ushered in similar periods of mixed marine and continental basinal deposition (see Fig. 5).

The emplacement of intrusions (diabase, gabbro, granite, etc.) similarly might coincide in time in related regions, although it can be argued that because intrusions are known to vary in time of emplacement they are not a reliable criterion. In any event, the apparent ages of syntectonic and post-tectonic intrusions (Table 2h, i) suggest lack of contemporaneity, although limits of analytical error for certain ages allow slight overlaps. Late basic intrusions (diabase, gabbro), however, appear to have been contemporaneous (Table 2f), although the dates of the intrusive events in Svalbard are uncertain.

Basement rocks are recognized or suspected in both flanks of the Franklinian Geosyncline, but the existence of an exposed basement in Svalbard is questioned (Table 2j) so that comparison of such ancient terrains is not yet possible.

CARBONIFEROUS AND LATER TIME

Uplift and erosion took place in latest Devonian time both in the Innuitian region and in Svalbard. In both regions, this was followed by accumulation of continental-littoral sediments and by a marine incursion in Carboniferous time. Basal clastic rocks in each basin are succeeded by mixed sandstone, limestone, and evaporite beds, then by thick shale-sandstone sequences with some coal. Thus, the sedimentary columns are broadly similar although the upper parts differ in details of continuity and in thickness (see Table 2a, b). Particularly notable are similarities in the Lower and Upper Carboniferous successions; indeed, certain units in the two areas have been described as almost identical on a bed-by-bed basis in terms of lithology, age, and thickness (NASSICHUK 1975, p. 4). HARLAND (1965, 1969a, 1969b) and others (WILSON 1963) have proposed, on the basis of the evident similarities, juxtaposition of Svalbard and Ellesmere Island – northern Greenland in late Paleozoic and Mesozoic time.

Basalt flows, which can be contemporaries of basic intrusions, are present in the sedimentary column of the Sverdrup Basin, where early Early, mid-Early, and early Late Cretaceous times are represented. In Svalbard, “tuff-conglomerate” is mentioned by ORVIN (1940) in beds of about mid-Early Cretaceous age.

Perhaps especially significant is a similarity in Mesozoic tectonic history in Svalbard and the Sverdrup Basin: in both areas, a Cretaceous period of basic intrusion and (?) volcanic effusion was associated with a widespread marine inundation, in effect a dramatic widening (spreading of the shorelines due to depression of the land) of the area of the Mesozoic basins (see Fig. 5, Table 2b, 2f). The crustal tension and regional downsinking represented by the rocks noted presumably have a common origin: the beginning of rifting that eventually separated the basinal regions.

TECTONIC CONNECTIONS BETWEEN SVALBARD AND INNUITIA

The remarkably close correspondence of Carboniferous and younger units in the two regions serves to emphasize the differences between older successions. The similarities of the stratigraphic sections of one geological period suggest a close linkage of the basins at that time; by the same token, the contrasting thicknesses and lithologies of the older periods suggest that the earlier geo-

synclines of the Innuitian and Svalbard regions occupied distinctly different sedimentary-tectonic sites.

The suggestion of close connection in later times but distinct separation in earlier times appears enigmatic. The enigma perhaps may be resolved, however, by supposing that the two geosynclines were a rudely matched “pair” of sedimentary basins separated by a geanticlinal ridge. Such a positive anticlinal belt may have been the “borderland” required by several authors for paleogeographical and sedimentological reasons; a northwestern, rising orogenic welt in the Innuitian region was named the “Pearya geanticline” by TRETIN (1971) after SCHUCHERT’s earlier (1923), hypothetical borderland, “Pearya” (see Fig. 4B, 6 iv, v).

A relatively positive orogenic welt lying between the Hecla Hoek and the Franklinian basins would have been a source of clastic sediments for both basins during certain times. Deduced directions of sources of sediments, as reported in the literature (see Table 2c) and noted below, tend to support such a supposition.

The comparison of source directions in the two regions is complicated by three main factors: a) the change in nominal azimuth direction when Svalbard is moved to the meridian position of northern Ellesmere Island (for example); b) uncertainty or disagreement among various authors in the interpretation of the direction of the paleo-sources; and c) uncertainty as to disposition of the two basins before drift and whether adjacent parts of the basins are being compared. The first factor is easily adjusted, while the second and third must await refinement or agreement among researchers.

The third factor is a critical one: Svalbard probably was adjacent to Peary Land and not to Ellesmere Island. This is apparent from the geometry of the landmasses (see Fig. 7) and from evidence of sedimentary provenance noted by KELLOGG (1975) (see Table 2c). The question then arises: how consistent is the geology along the axial direction of the Franklinian trough (that is, from northern Greenland to Canada)? In answer: there are structural differences, already noted; but the sedimentary trends do continue from Canada through Peary Land (from personal observation; and see KERR 1967a; DAWES 1973, 1976; DAWES and SOPER 1973). From the consistency and continuity of both sedimentary and structural trends in the northeastern part of the Franklinian–Ellesmerian region it appears that the hazards of speculation here are at worst no more severe than those typically weathered in such exercises.

The apparent relationships between basins of deposition and source areas of sediments are reviewed in the following paragraphs (see Table 2c). Note that directions for Svalbard are adjusted counterclockwise 90° to conform to the longitude of northern Ellesmere Island–Peary Land region; thus, a “westerly source” reported in the literature is stated here as *southerly*.

i) In late Precambrian time, a source of sediment lay to the south (for marginal deposits) in the Innuitian region, and to the south (for volcanic and clastic geosynclinal deposits) for Svalbard. These directions apply to different facies of the basins and fit the model proposed.

ii) In early Paleozoic (pre-Carboniferous) time, directions of transport

evidently were opposed: that is, from the north or northwest for the Franklinian Geosyncline and from the south or east for the Hecla Hoek succession.¹

iii) Sedimentation in Permian and Carboniferous time began in both regions in scattered, small basins. Source areas in the Innuitian region were to the northwest in the northwestern part (although this is uncertain) and to the southeast or “local” in the southeastern region. No reported directions have been found by the writer for Svalbard.

iv) Sediments were derived mainly from the south in Mesozoic time in Svalbard, with some from the west, and possibly from the east. In the Innuitian region, source directions to the east and south are reported, as are sources to the north and northwest for certain Lower Triassic and Jurassic rocks. H. R. BALKWILL (pers. comm.), however, considers that there were contributions of sediment from the northeast and west, and possibly from the northwest, throughout Mesozoic time in the Sverdrup Basin.

v) Sedimentary sources were clearly in the north or northwest in Svalbard in early Tertiary time. In the Innuitian region, source directions are often reported as “local” or “unknown”. Recent studies by BALKWILL (pers. comm.), however, suggest that early Tertiary beds on western Ellesmere and eastern Axel Heiberg Islands had sources to the north and west.

If one assumes that the two regions were juxtaposed, a picture emerges of a linear Tertiary basin with sources in the north or northwest (including the Barents Shelf).

A new sedimentary cycle in Svalbard, however, beginning in late Eocene time, approximately, gives evidence of a southerly source, opposite to that of the underlying beds. This event can be taken to represent the main tectonic phase of Tertiary time for this region (see KELLOGG 1975).

Can these various directions of source or direction of transport of sediments be accommodated in a single tectonic model?

A TECTONIC MODEL FOR SVALBARD AND INNUITIA

The model proposed here, as already noted, assumes that the late Precambrian and early Paleozoic geosynclines of Svalbard and Innuitia were adjacent, or “paired”, separated by an intermittently positive tectonic welt. A geological history based on such a supposition might be somewhat as outlined below (see Fig. 6). The relationships between the Svalbardian–Franklinian geosynclines and others, such as the nearby Caledonian geosyncline, will be taken into account in subsequent paragraphs.

i) Lower Hecla Hoek time: beds of Lower Hecla Hoek deposited, with the region of the present Barents Shelf providing most of the sediment; volcanic vents along the future Pearya orogenic welt may signify an island arc and are the earliest recorded indication of a major linear structural weakness; state of Franklinian Geosyncline uncertain (Fig. 6i).

¹ Sources to the east and northeast are proposed by EMBRY (1976) for Devonian rocks of the Franklinian Geosyncline, based on paleocurrent indicators and on the distribution of sedimentary facies. It is here considered that these paleocurrent directions must represent basin-longitudinal transport.

Was the Barents Shelf a shield area? Geophysical data show the crustal structure to be continental, with a heterogeneous basement of folded structures of various ages (OSTENSO 1974, p. 759). ТКАЧЕНКО et al. (1973, p. 342) describe the “Barents stable region” as outcropping in Nordaustlandet (northeastern Svalbard), where a granite-gneiss complex is considered pre-Riphean (pre-latest Precambrian). Uncertainties exist, but the entire “Barents plate” is assumed, as a working hypothesis, to be pre-Baykalian (older than the latest Precambrian orogeny). For the purposes of this paper, the Barents plate is assumed to be an isolated Precambrian platform, or microcontinent.

ii) Proto-Pearya: rising isotherms (=orogeny) result in intrusion, deformation(?), and metamorphism of Lower Hecla Hoek succession; uplift to provide early source of detritus to nascent Franklinian Geosyncline. Radio isotope age, ca. 742 m.y. (Fig. 6ii).

iii) Middle Hecla Hoek time: middle part of the Hecla Hoek and lower part of the Franklinian sequence deposited; proto-Pearya perhaps subsides, and the shield areas of the North American and Barents plates provide detritus. (Fig. 6iii).

iv) Late Precambrian to Cambrian time: Pearya rises as an orogenic welt shedding debris to north and south, respectively to the Svalbardian and Franklinian basins. Continental glaciation in the (mountain?) regions of northern Greenland, Svalbard, and? Pearya results in deposition of tillite (now preserved in Svalbard and in southern Peary Land; see Table 2d; Fig. 6iv).

v) Cambrian to Middle Ordovician time: continuing subsidence of the Svalbardian and Franklinian geosynclines and accumulation of clastic and carbonate sediments and evaporitic deposits (Fig. 6v).

vi) Late Ordovician to Early Devonian time: orogeny begins, and advances from north to south; Caledonian orogeny begins in Svalbard, during which syntectonic intrusions (post-Canadian, pre-Devonian) are emplaced; uplift results in deposition of a clastic wedge southward into the Franklinian Geosyncline (Fig. 6vi).

vii) Devonian: continued orogenic activity; Ellesmerian orogeny results in deposition of molasse (Old Red) deposits in Svalbard from debris shed northward from a rising “Franklinia”; early-middle Devonian synkinematic intrusions emplaced in Franklinia. The recently identified Frasnian clastic beds of northern Ellesmere Island probably were continuous with the molassic beds of Svalbard (Fig. 5, 6vii).

viii) Late Devonian: Ellesmerian orogeny widens to fold Svalbard molasse slightly (=Svalbard Folding) (Fig. 6viii).

ix) Carboniferous to early Tertiary: regional downsinking along the axes of earlier geosynclines; eventual general down-sinking resulting in widespread onlap of Cretaceous sediments. *Either* several separate basins exist, somewhat as beads on a string, *or* a linear depression is formed to receive sediments of this period (see Fig. 6ix; Sverdrup-Wandel Sea-Spitsbergen basins). If the several Carboniferous and younger successor basins were connected (as seems likely), they nevertheless probably were divided by “sills” or “highs” such as that determined for the northwestern margin of the Sverdrup Basin (see MENELEY

et al. 1975). The “highs” are represented schematically by basinal highs in Fig. 6ix).

The Carboniferous–Permian linear depression, it is suggested, crossed the Barents Shelf to the Urals–Moscow Basin–Donitz Basin of the U.S.S.R. (see DUTRO and SALDUKAS 1973, NALIVKIN 1973, p. 192). Former connection between these polar basins and the Atlantic (East Greenland–North Sea) linear Carboniferous to Cretaceous troughs is uncertain; perhaps they were not joined until a younger stage.

It is possible that the Canada Basin opened during an early (latest Paleozoic or early Mesozoic?) part of this stage (discussed in a later section). According to MENELEY et al. (1975), the northwestern margin of the Sverdrup Basin, earlier a platform region, in Triassic time was overwhelmed by clastic debris from the south. The termination of the “carbonate barrier” or platform stage of the northwestern margin may signify the birth of the Arctic Ocean in that region.

x) Early Tertiary: orogeny and rifting along the de Geer line. En echelon folds and strike-slip faults with compression to produce overthrusts; a “strike-slip orogenic belt” (as termed by LOWELL 1972) results (Fig. 6x). The resistant Pearya Complex is thrown up as horsts and in a later stage, when Greenland and Svalbard are separated, grabens form. The opening of the North Atlantic began in late Paleocene time (63 m.y. ago, according to PITMAN and TALWANI 1972); this marks the beginning of motion between the Eurasian and the Greenland–North American plates, and is about the time of the early phases of the Eureka Orogeny in northern Ellesmere Island (see CHRISTIE 1976). The main phase of the Eureka Orogeny took place somewhat later – usually described as mid-Cenozoic time (see THORSTEINSSON and TOZER 1970, p. 585); the model of PITMAN and TALWANI (1972, p. 638) suggests a “collision” between Svalbard and northern Greenland at around 47 m.y., or mid-Eocene time. Both pre-orogenic (as young as Paleocene) and syn-orogenic (late Eocene or Oligocene, or even Miocene?; see LIVŠIĆ 1974; HARLAND 1975a) clastic beds were deposited in Svalbard. Eocene beds in northern Ellesmere Island have been affected by thrust faults, and molassic beds at the top of the Tertiary section may be equivalents of the syn- and post-orogenic clastic beds of Svalbard (see CHRISTIE 1976).

The late Precambrian to mid-Paleozoic sequence of events listed above is proposed to provide a plausible or possible model for the Franklinian and Svalbard geosynclinal rocks. Subsequent events, such as the deposition of strikingly similar strata and fauna in the Carboniferous–Permian basins of the Arctic, are better documented. DUTRO and SALDUKAS (1973), for example, show by paleogeographic maps (for Early and Late Permian) that a linear sea joined the Sverdrup Basin and the “Perm Basin” (Urals–Moscow Basin–Donitz Basin) of western U.S.S.R. Similarly, VANN (1974, Fig. 2), in demonstrating a sophisticated “fit” of Greenland and western Europe, has shown that a linear Mesozoic basin may have extended northwest from the North Sea; such a basin would almost certainly be continuous with the Wandel Sea Basin (DAWES and SOPER 1973), probably as early as Carboniferous time.

LIVŠIĆ (1974), however, concluded from a “block-faulting” model and sedimentological evidence that Svalbard was an archipelago in Paleogene time and that the Polar Tertiary sedimentary province comprised relatively small, separate, or “segmental” basins.

THE DE GEER LINE AND OTHER LINEAMENTS

The supposed dextral translatory movement of some 750 km along the major shear zone and transform fault, the de Geer Line (see WEGMANN 1948, p. 21) is an important element in the proposed model for the Franklinian and Svalbard geosynclines. Is there direct evidence for such a major shear zone in Tertiary or earlier time? Such evidence might be geomorphological, geophysical (the line is submerged in ocean water), or sedimentological. If the shear zone extends onto continental areas, then major displacement of landward structures may be identifiable.

The chief geomorphological evidence for transcurrent movement is the large-scale lineament itself: the alignment of continental edges along a small circle of the globe (see Fig. 2). This feature was noticed by de Geer and named after him by WEGMANN (1948).

The de Geer Line has long been considered a possible path of continental displacement, and Wegmann, in reconstructing pre-drift land masses, suggested a northern Greenland–northern Ellesmere source for the Tertiary sediments of Svalbard. De Geer’s Line, on a globe or a polar projection, indeed conforms convincingly to the movements proposed. The line, at least in the Greenland Sea, is a dextral transform fault (WILSON 1965) rather than a simple transcurrent displacement, in terms of contemporary tectonic theory.

Geophysical data confirm the supposed transform fault: earthquakes have been reported along the fault but not along the physiographic extensions, the now quiescent traces of early breaks between continental masses (see WILSON, 1965, p. 344).

The sedimentological evidence for former juxtaposition of Svalbard and Peary–Grant Land has been reviewed in preceding paragraphs: a westerly (southerly, at the longitude of Peary and Grant Lands) source of sediments in Svalbard in late Eocene time is required from field evidence.

The conspicuous linear feature, Nares Strait, which cuts across the Innuitian region and is an important structural element, was described earlier. Displacement along this lineament of up to 150 km since early Tertiary time is required in certain reconstructions based on geophysical data: for example, that devised to accommodate an oceanic crust in northern Baffin Bay (C. E. KEEN et al. 1972). From other geophysical evidence, however, one is cautioned against the assumption that oceanic crust floors certain parts of Baffin Bay; Davis Strait, an area particularly critical to “pre-drift” reconstructions, may be underlain to a considerable extent by continental crust (GRANT 1975). As suggested earlier, the net movement along the Nares lineament may be small, and it is presumed, in the context of the model proposed in this paper, that the lineament is unimportant.

Two submarine plateaus, the Yermak, off Svalbard, and the Voring, off

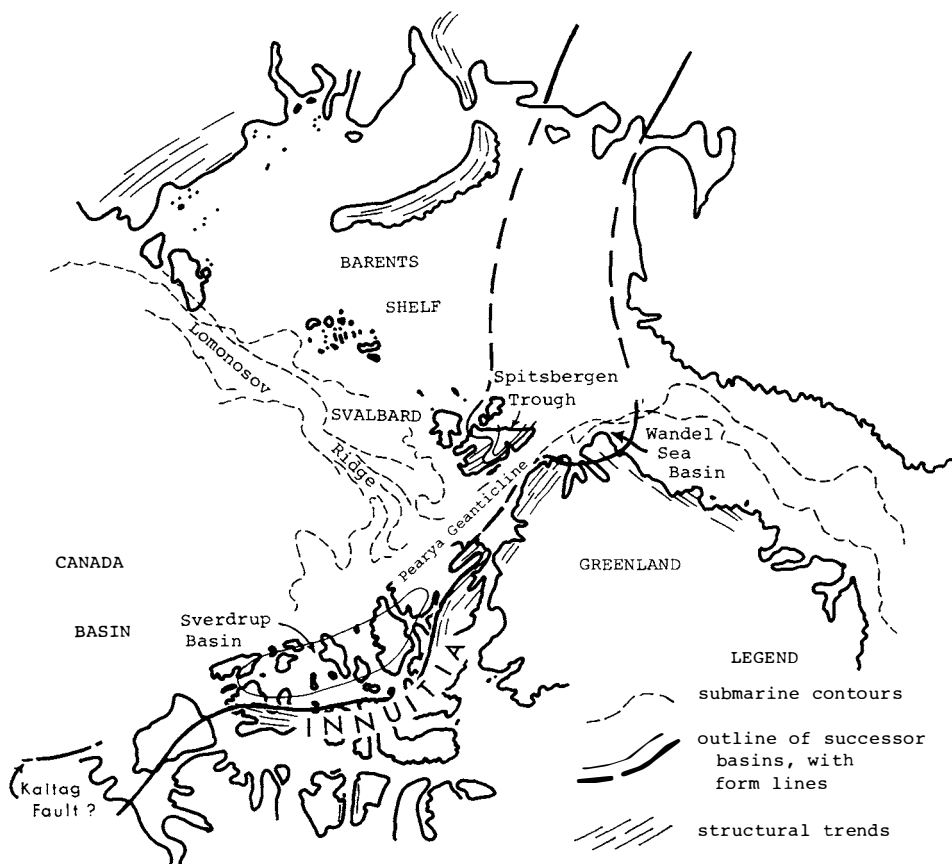


Fig. 7. Possible pre-drift configuration, Inuitia and Svalbard.

central Norway, were noted by BIRKENMAJER (1972, p. 214) as “not fitting the jigsaw puzzle” of continental drift and therefore as being, probably, of late Tertiary age. The close spatial relationship of both plateaus to fracture zones (transform faults) was taken to suggest a probable presence of volcanic rocks.

The Yermak Plateau appears to have a mirror partner on the other side of the Nansen Ridge (Fig. 2): the Morris Jesup Plateau off the north coast of Greenland (see HEEZEN and THARP, 1975). The composition and structure of these plateaus are critical in an interpretation of the history of this part of the Eurasian Basin. For example, the presence of continental crust beneath the plateaus would be evidence that Greenland had, after all, moved northeastward along a Nares Strait transcurrent fault. More probable is the possibility that a superficial “pile” of material was deposited in Tertiary time, during or after partial opening of the Eurasian Basin; continued spreading of the sea floor could separate such a deposit into two “plateaus”. The volcanic rocks of Kap Washington (northern Greenland) are probably Tertiary in age (from isotopic age determinations – approximately 35 m.y. indicated – and structural relationships) and certain geophysical data may indicate an offshore extent (DAWES and SOPER 1971; DAWES 1973); such a suite of rocks may represent a

young, oceanic volcanic pile. The site of the eruption may be related to the former junction of Nares Strait, the Arctic Mid-oceanic Ridge, and the Nansen Fracture Zone (of the de Geer Line). This supposed former junction would have been a “triple junction” of the FFR (Fault-Fault-Ridge) type (McKENZIE and MORGAN 1969). If such a junction existed, it evidently became unstable and movement ceased on the Nares transcurrent fault at an early stage. Both the petrographic character and the tectonic siting of the volcanic rocks of Kap Washington are similar to those of Iceland: basalt, andesite, rhyolite, and breccias at the junction of oceanic ridges or lineaments (see DAWES and SOPER 1970; EINARSSON 1973).

A possible westward extension of the de Geer lineament figures in tectonic models that provide for an “opening” of the Arctic Ocean basin, as does the relationship of the fold belts of northern Alaska and Yukon to the Cordilleran and Franklinian Geosynclines. Several hypotheses have been erected to explain the geometry of these regions; examples are: subsidence of a boreal landmass into the Arctic Ocean (Eardley); and rifting and rotation of Alaska away from the Canadian islands (Carey). CHURKIN (1969, 1972, 1975) showed that the Canada Basin may be an ancient ocean, rimmed from Greenland to eastern Siberia by an early Paleozoic geosynclinal belt. The Alpha–Mendeleev Ridge as an inactive mid-ocean ridge was discussed by HALL (1970); three successive periods of arctic ocean-floor spreading, none older than Jurassic, were suggested by TAILLEUR (1973); and folding of the Franklinian Geosyncline in mid-Paleozoic time by collision of an Asian, “Kolymski plate” with the North American plate was proposed by HERRON, DEWEY, and PITMAN (1974).

A possible westward extension of the de Geer lineament is the Kaltag Fault of northern Yukon and Alaska (NORRIS 1974). This major fault zone has been traced northwestward from Bering Sea to the Canadian border, where continuations of the fault, curving northward, have been mapped. A possible extension of the Kaltag Fault across the Beaufort Shelf to the southern edge of the Canada Basin (see Fig. 7) was suggested by NORRIS, who noted a marked kink in the Bouger anomaly field of the shelf. NORRIS suggested the Kaltag Fault may be a junction between continental plates: it cuts through the Cordilleran orogenic system from Bering to Beaufort Seas, it has a displacement measurable in several tens of kilometres, and its sense of movement is right-lateral and thus compatible with the supposed differential motion of the plates in Tertiary time. Large-scale motion on the fault, however, is supposed, from geometric data, to have ceased by the end of the Cretaceous; subsequent westward drift of the by-then fused North American plate was presumably accomplished through right-lateral movement along the de Geer fault. Northeastward extension of the Kaltag structure has been suggested on the basis of a belt of elliptical-shaped, strong, positive Bouger and free air anomalies that lies along the continental margin northwest of the Canadian islands. TRETIN (1972, p. 151) noted that the belt continues, apparently uninterrupted, past the Nares Strait lineament. In a recent analysis, however, SOB CZAK (1975a; 1975b) suggested that most of the “elliptically-shaped, positive, free-air gravity anomalies” can be explained by assuming the presence of prograded wedges of

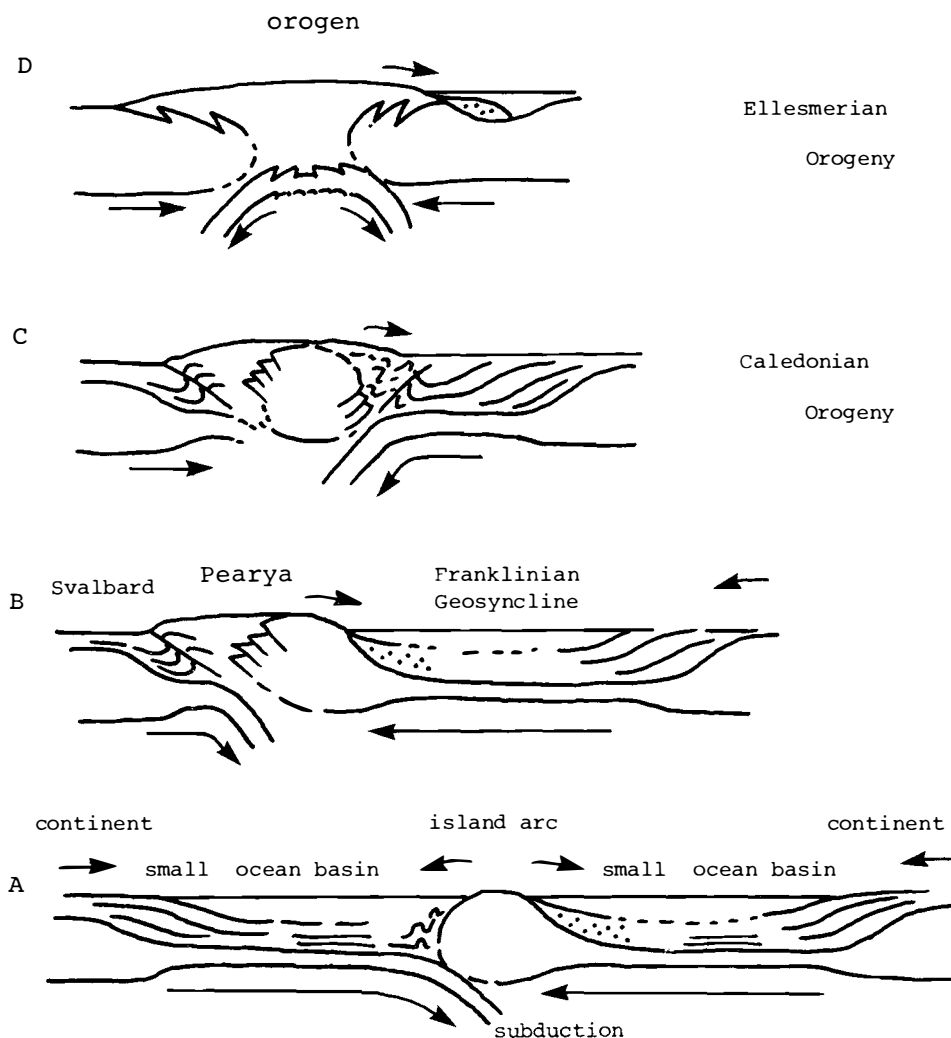


Fig. 8. Diagrammatic representation of two small ocean basins flanking an island arc (Diagrammatic style after DEWEY and BIRD 1970a, 1970b). Arrows indicate directions of sedimentary and tectonic transport.

Tertiary and younger sediments acting as uncompensated loads on the crust at the continental margin. The edge of the continental crust becomes distinctly more abrupt in the new model, which is thus the more compatible with a faulting or rifting origin for the edge.

The Ellesmerian folds of the western Queen Elizabeth Islands are mainly buried near the continental margin, but from their limited western exposures (on Melville Island) they are west-trending and appear to be truncated by the continent-margin. Were northern Alaskan structures perhaps once continuous with those of the Canadian Islands? The early Paleozoic rocks of northern Alaska have long been compared with those of the Franklinian Geosyncline (MARTIN 1959; CHURKIN 1969, 1975) and interpreted as part of a circum-Canada Basin geosynclinal belt. An approximation of the Paleozoic position of

northern Alaska might be obtained by northeastward translation and rotation of a “Bering Sea Plate” along some zone such as the Kaltag–de Geer lineament¹ to reverse sinistral movement that may have taken place in post-Ellesmerian (late Paleozoic or Mesozoic?) time (Fig. 9D, E, F). Such shearing would be an early stage of opening of the Arctic Ocean, with a probable spreading centre along the Alpha Ridge (see CHURKIN 1972; GALLAGHER 1973). Northwestward thickening of Carboniferous carbonate sediments over the northwestern, marginal rim of the Sverdrup Basin (MENELEY et al. 1975) may be evidence that a Canada Basin was present at that time.

THE GEOSYNCLINAL CONCEPT AND THE MODEL FOR SVALBARD–INNUITIA

The model here proposed for Svalbard and the Innuitian region is one in which paired geosynclinal belts are separated by a geanticlinal belt. At least substantial parts of the combined belts lay either within a continent (now broken up) or between two continental masses.

How well does the proposed model for Svalbard–Innuitia conform to the recently elaborated theory of plate tectonics (see DEWEY and BIRD 1970a; 1970b)? TRETTIN (1972, p. 165) has remarked on the general plausibility of the plate tectonic explanation for the Innuitian region, and VOGT and AVERY (1974, p. 109–112) have reviewed the possibilities and probabilities of various forms of plate-tectonic origin, in the light of available geophysical data, for the Canada Basin and the continental margins of the Arctic. TRETTIN noted that the early history of the mobile belt may be an example of plate tectonics in a “small ocean basin” (a term applied by DEWEY and BIRD to small oceanic areas lying between island arcs, between arcs and continents, or between two continents). Among the models figured by DEWEY and BIRD (1970a, p. 633), the “small ocean basin” lying between two continents appeals most to the present author for the Svalbard–Innuitia regions; the “paratectonic orogen” arising from this arrangement of tectonic features (op. cit., Figs. 4E, G) should have an inherent symmetry that conforms well to the model proposed in this paper. An improvement on this model might be that of an island arc lying between two continents. In this case (Fig. 8A, this paper), two small ocean basins flank the island arc. Orogenic deformation of one or the pair of geosynclines would occur when one of the continental masses collides with the arc (Fig. 8B); later, collision of the second continental mass with the older orogen would result in a second, or progressive orogeny (Fig. 8C). This model provides: a) an early tectonic welt; b) progressive orogeny; and c) symmetry in the resulting final, paratectonic orogen.

A yet more complex variation of small ocean basins might be suggested: that of two island arcs, eventually caught between approaching continents. Collision of the island arcs could result in the formation of an early orogenic welt (a proto-Pearya Geanticline) while the paired geosynclinal basins still existed,

¹ The identified movement on the Kaltag structure is right-lateral, as noted, opposite to that required by the proposed reconstruction of Figure 9. Left-lateral movement may have taken place at an earlier time, or along some similar but unrecognized zone.

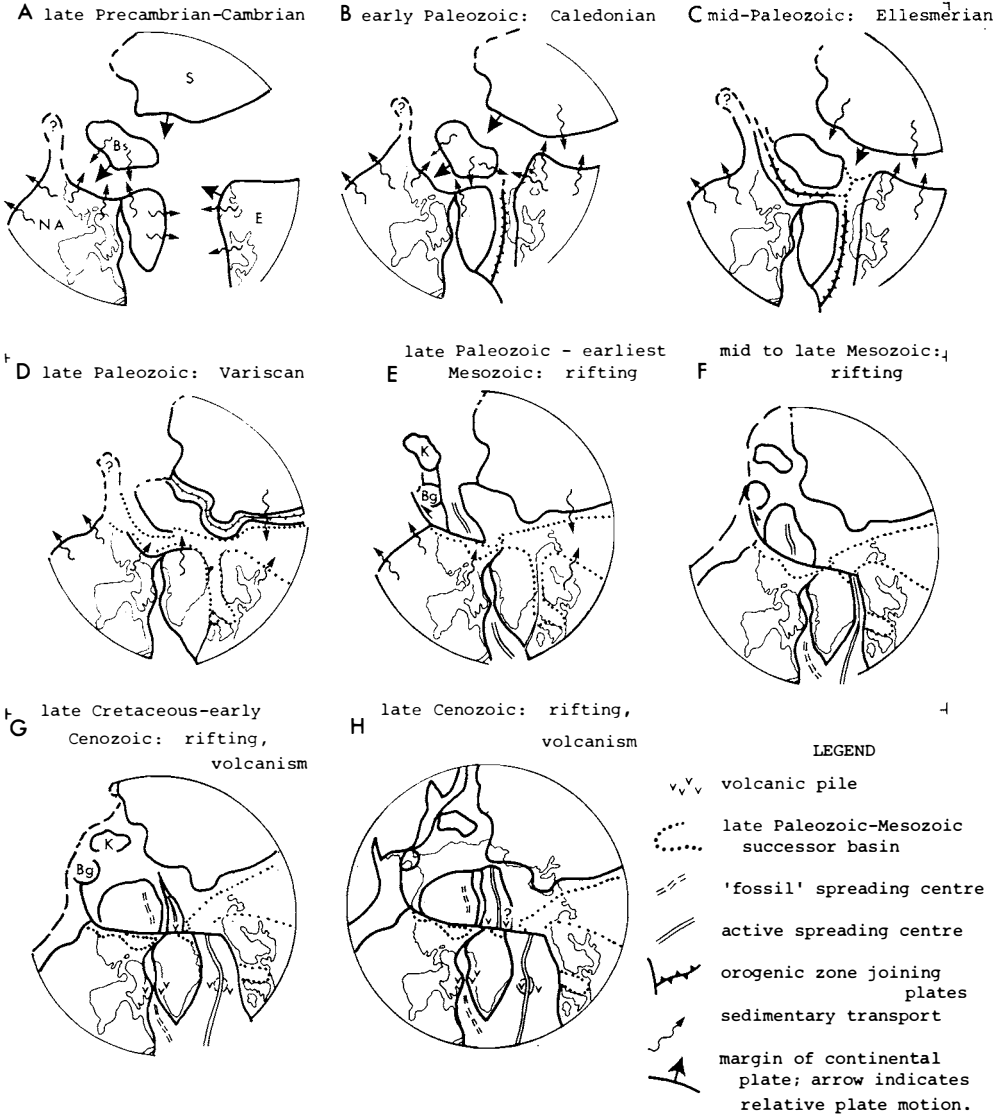


Fig. 9. Stages in a hypothetical reconstruction of continents of the boreal region. For geographical nomenclature see Figure 2b. NA - North American platform; S - Siberian platform; E - Eurasian platform; Bs - Barents splinter; Bg - Bering splinter; K - Kolymski splinter.

relatively undeformed, to receive sediments and volcanic rocks. Subsequent collapse of first one geosyncline, then the second of the pair, would result in the final, extensive, symmetric paratectonic orogen.

A review of the history of thought on geosynclines shows that both the concept of linear, geosynclinal belts separated by intermittently positive tectonic welts, and the concept of mobile belts between continental masses have recurred since about the turn of the century (see GLAESSNER and TEICHERT 1947). HAUG (in 1900), influenced by the tectonics of the Mediterranean area, defined geosynclines as mobile belts between rigid continental masses. HAUG

evoked a “median geanticline” as a “tectonique embryonnaire”, or earliest indication of crumpling of the geosyncline. Geanticlinal ridges (lying between “furrows” within the geosyncline) formed a prominent part of AUBOUIN’s (1965) careful review of the concept of geosynclines. He also noted and figured some tectonic characteristics of the “complex geosynclines” that may develop when two continents with marginal geosynclines are sufficiently close: a “median zone” of eugeosynclinal ridges may form. This zone would contain the central massifs, or “*zwischengebirge*” earlier described by L. KOBER. The median zone of the geosynclinal complex, it was supposed, would be affected early by orogenesis, the orogenic activity “diverging” in both directions toward the miogeosynclinal and foreland regions. It was not considered that geanticlinal ridges are necessarily continuous in the median zone: where the ridges die out, then a median, eugeosynclinal furrow may remain. Such geosynclinal complexes, AUBOUIN stated, give rise to the “classic (mountain) chains of bilateral symmetry”.

The model here proposed for the combined Franklinian Geosyncline–Pearya Geanticline–Hecla Hoek geosyncline contains features, including bilateral symmetry and a median geanticline (see Fig. 6v, vi, vii), that recall some of the ideas alluded to above. The Franklinian–Svalbard region was, according to this model, one of early continental collisions and of “Mediterranean” orogenic type.

The preferred models, described above, require small ocean basins between continents and rifting of the subsequent orogen. This scheme contrasts with the often-suggested model in which the Arctic Ocean (an ancient ocean) was rimmed by several subduction zones (e.g. HALL 1970, and see VOGT and AVERY 1974, p. 112). Rather, as with the present margins of Greenland, Norway, and the Barents Shelf, the existing borders of the Canada Basin are taken (after VOGT and AVERY, p. 112, and GALLAGHER 1973) to be rifted margins formed in the initial stages of sea-floor spreading at ancient times. The continent-margin from about northern Ellesmere Island east is matched by the western edge of the Barents Shelf (including Svalbard) (see Fig. 2), while that to the southwest is presumed to be represented by northern Alaska and the “Bering plate” (Bg in Fig. 9F).

The question remains of the relationship of the successor, Sverdrup Basin, with its large size and substantial depth apparently along the axis of the Franklinian Geosyncline. This basin and its extension or contemporary basins appear to conform to KAY’s (1951) *epieugeosyncline* (“above” a eugeosyncline: elongate, relatively non-volcanic; sediments derived from the older, relatively immobile orthogeosynclinal belt). However, some aspects of the geometry and of the provenance of sediments of the Sverdrup Basin make it a partial misfit in this category: a large part of the sediments may have been derived from the craton, and the depth of fill is of an order equal to that of the earlier geosyncline. A *taphrogeosyncline* or rift-basin assignment was suggested by McCROSSAN and PORTER (1973) on the basis of the deep fill and the very rapid basinward thickening at the margins. Although there are no identifiable major, basin-margin faults (that is, that could account for the basin’s depth in a few simple

steps), there are two lines of evidence that tend to confirm that faulting has played an important part in the development of the Sverdrup Basin: i) the northwestern “rim” of the Sverdrup Basin appears as a faulted or horst-form structural “high” in seismic sections of that part of the basin margin (see MENELEY et al. 1975); ii) early deposits of the Sverdrup Basin include substantial amounts of evaporite and marginal units of red-weathering, coarse, terrigenous clastic rocks; such a suite of rocks is generally taken to be characteristic of fault-bounded basins with restricted circulation. (Such a fault-origin for the entire series of successor basins, noted below, is not implied at this time, and no faulting is shown in Figure 6ix).

The Sverdrup Basin is a substantial and long-lived sedimentary basin that subsequently was more or less deformed by an important (the Eureka) orogeny. In spite of the distinctive characteristics of great size and longevity, however, the Sverdrup Basin can be taken as a “successor basin” in the Innuitian orogenic system, and can be grouped with similar, if smaller, successor basins between Svalbard and Alaska: e.g. the Carboniferous to Tertiary basin of Svalbard (KELLOGG 1975), the Wandel Sea Basin of northern Greenland (DAWES 1973, p. 131), certain successor basins with “molassoid” successions in the Mackenzie region (YORATH and NORRIS 1975), and the Colville Geosyncline of northern Alaska (GRANTZ et al. 1975) (see Fig. 8D). These basins are perhaps best described merely as “post-orogenic basins” in the spirit of P. B. KING’s simple terminology (KING 1959).

Certain distinctive tectonic characteristics of the basins here described as successor basins of the Innuitian region may be noted (see Fig. 3a, 3b): i) marked depocentres lie inside the continental margin; ii) a prominent basement high or arch lies near or along the continental margin (the arch, however, is covered by at least the younger beds of the basinal fill); iii) the detritus was mostly transported from the craton, evidently filling the basin and prograding across the continent-margin “high”. Thus, the successor basins of the Innuitian region are tectonic depressions within the continent, and there is an apparent positive tendency along the continent-margins separating the successor basins from the adjacent ocean. The formation of the cratonic basins may be due to crustal distension. In tectonic position and relationships the successor basins are thus similar to the grabens, figured by DEWEY and BIRD (1970a, Fig. 1B, C), that lie within the newly-formed edge of a continent being separated by an expanding ocean. Faulting, however, apparently played a part only in the early stages in the case of the Sverdrup Basin, the trough perhaps having formed through faulting at depth, with “drape” or warping of beds taking place in the upper levels. Yet unexplained is the scale of the downsinking in the Sverdrup Basin – locally matching that of the earlier geosyncline (see Table 1). One can suppose that massive crustal thinning took place, although the mechanism for this (mantle convection?) is unknown.

Essentially this model for basins of “flyschoid and molassoid” sediments along the northern continental rim was proposed, and tested against geophysical results, by YORATH and NORRIS (1975, Fig. 13). In their model, however, another series of (smaller) linear basins lies to seaward (but still landward of a

continent-edge uplift) of the rim of the Sverdrup Basin. The northern rim of the Sverdrup Basin is considered by these authors to be a positive tectonic feature related to rifting along the northwestern edge of the Queen Elizabeth Islands. The rifting began perhaps as early as Triassic time. The foundering of the Sverdrup Basin is supposed to have followed a period of relatively slow subsidence in the Carboniferous, during which time subcrustal mantle transfer may have been initiated.

A tectonic history of the boreal continental margins as conceived in the preceding sections will be summarized below. Tectonic reconstructions to account for a history of northernmost North America and the Svalbard regions, however, inevitably necessitate suppositions for adjacent regions of the North Atlantic and Arctic oceans. These suppositions are stated below in order to complete a picture but, for economy of words, justification or support for all of them will not be attempted here.

i) A setting for late Precambrian time may have been as follows (see Fig. 9A): three large platform areas and several smaller continental “splinters” are separated by oceanic areas (the splinters perhaps derived from a preceding continental breakup); the fragments of continental material include the North American, Siberian, and Eurasian platforms and the Barents, Bering, and Kolymski splinters, or smaller plates. The proto-oceans between the fragments are presumed to have been shrinking, and the three platforms approaching one another. (The question of the time of separation of the subcontinent, Greenland, from the North American platform is not considered here and the Labrador Sea–Baffin Bay “opening” is left, unclosed, in the diagrams.)

ii) The Caledonian and the Franklinian geosynclines accumulate sediments from the continents and, as the continental platforms approach one another (Fig. 9B), sediments and volcanic material accumulate as proposed in the scheme of plate tectonics of DEWEY and BIRD 1970b). The Caledonian Orogeny represents the collision of the North American and Eurasian plates in early Paleozoic time. The fit of these plates follows that of VANN (1974), which is a modification of the earlier proposals of BULLARD et al. (1965). Sedimentation continues in the Franklinian Geosyncline; however, earlier collisions of island arcs and continents had resulted in the rise of a proto-Pearya (see the tectonic model for Svalbard and Innuitia, described earlier). Caledonian structures evidently end at about the latitude of northern Norway and do not cross the Barents Shelf (TKACHENKO 1973, p. 342), a contention supported by geophysical data (VOGT and OSTENSO 1973). The Caledonian Geosyncline thus may terminate northward as a relatively neutral area in which sediments were deposited but little deformed by the continental collisions (Fig. 9B).

iii) Collision of the Barents plate and the North American platform (Fig. 9C) results in the Ellesmerian Orogeny and the close of sedimentation in the Franklinian Geosyncline. Svalbard now lies adjacent to northern Greenland.

iv) Collision of the Siberian platform with the agglomerated North American–Barents–Eurasian continental masses forms the Uralides (Fig. 9D) in late Paleozoic (Variscan) time (HAMILTON 1970). The boreal successor basins (earlier described) form a linear belt.

v) Rifting in the Arctic Ocean region begins in latest Paleozoic or earliest Mesozoic time (Fig. 9E). Note that the Bering and Kolymski plates are assumed to have been part of the North American platform; their westward movement resulted in formation of a proto-Arctic Ocean (the Canada Basin).

vi) The North Atlantic Ocean opens in mid- to late Mesozoic time; the proto-Arctic Ocean is an enclosed sea, with the Kolymski plate colliding with the Siberian platform (Fig. 9F).

vii) Ocean-floor spreading continues into Cenozoic time with the splitting off of a continental sliver, the Lomonosov Ridge, and the opening of the neo-Arctic Ocean (Fig. 9G). Shearing, but with a compressive component, occurs along the de Geer Line transform fault. The compression is represented by the Eureka and West Spitsbergen orogenies. Molassic sediments accumulate at the top of the succession in the Sverdrup Basin and the Spitsbergen Trough.

viii) The platforms assume about their present disposition by late Cenozoic time (Fig. 9H); the Franklinian–Hecla Hoek basins are now torn asunder and represented by rocks in Svalbard, the Queen Elizabeth Islands, and northern Alaska. The belt of successor basins similarly is dismembered and represented by scattered remnants between Alaska and western U.S.S.R.

Conclusions

The sedimentological and stratigraphic data of the Innuitian and Svalbard regions can be reconciled by a tectonic model in which paired, late Precambrian to early Paleozoic geosynclines are separated by an intermittent tectonic welt, the Pearya geanticline. Subsequent downsinking of the entire orogen may then have resulted in deposition of late Paleozoic to early Tertiary beds in a polar linear basin extending from Alaska to the Moscow Basin region of the U.S.S.R. Rifting and ocean-floor spreading in the Arctic Ocean (in late Paleozoic or early Mesozoic time) may have resulted in separation of the western extension of the Franklinian Geosyncline along the Kaltag Fault. Tertiary rifting and ocean-floor spreading in the North Atlantic and neo-Arctic Oceans appear to have separated the northern basins – now exposed on Svalbard – from those of northern Canada and Greenland. The de Geer Line then must be a major shear or transform fault.

Close similarity of the post-Carboniferous successions of the Innuitian and Svalbard regions, in this model, is explained by their being connected or continuous, while dissimilarity of stratigraphy but conforming structural trends of the pre-Carboniferous rocks are presumed to be due to separation of geosynclinal basins by an orogenic welt and by progressive orogeny, from north to south, in the combined sedimentary-tectonic region.

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The Upper Proterozoic of Timan and the Kanin Peninsula

By A. A. KRASIL'SČIKOV and V. N. SOKOLOV

Introduction

The Precambrian of Timan and the Kanin, Rybačij, and Varanger peninsulas forms a continuous fold belt across the northern Russian platform (Timanides, Baikalides, Hyperborean zone, "Timan corridor"). Despite some similarities in lithology there is no commonly accepted stratigraphic standard for subdivision and correlation. This is because of the relatively monotonous lithologies (mainly terrigenous sequences), the absence of reliable marker beds, the extremely poor fossil content, and a complex tectonic setting.

On the basis of geologic-geophysical data, ŽURAVLEV and GAFAROV (1959), proposed a tectonic zonation of the Timan-Kanin region parallel to the margin of the Pre-Baikalian Russian Platform. Recent geological studies have proved that the western (outer) and eastern (inner) zones of the folded basement differ both in rock type and in grade of metamorphism and dislocation.

The most complete section through both zones was recorded on the Cetlaskij Kamen', Čil'menskij Kamen', and Vymskaja Range in Central Timan. The stratigraphy is unclear because the only fossils occur in a carbonate sequence (the Bystrinskaja Formation) within tectonic blocks along the boundary separating the two zones. There are two interpretations of this section: Some geologists suggest contemporaneous deposition of the different facies within each zone (KAL'BERG 1948; RAZNICYN 1962, 1972; GECEN 1970) while others claim an older ("Pre-Bystrinian") age for the lower metamorphosed terrigenous formations of the western zone, and a "Post-Bystrinian" age for the shale complex of the eastern zone (ŽURAVLEV and OSADČUK 1962; ČERNYJ 1965) stratigraphy.

Three formations have been recognized within the western zone: the Svetlinskaja, Četlaskaja, and Džežimskaja Formations (ŽURAVLEV and OSADČUK 1960, 1963; GECEN 1970). They are separated by unconformities, which are, however, unlikely to be of regional extent.

The Svetlinskaja Formation (over 1850 m thick) forms the observed base of the Precambrian section of Timan and is composed of light, fine- to medium-grained quartzites with thin bands of quartz-sericitic shales.

The *Četlasskaja Formation* (2500 to 2700 m thick) has continuous horizon of quartz conglomerates at the base overlain by intercalated dark quartzitic sandstones and mica-quartz and quartz-chlorite slates. The upper part of the formation is characterized by rhythmic alternations of quartz-sericite, quartz-sericite-chlorite, and coaly-argillaceous shales with feldspathic quartzite, quartzitic sandstone, and siltstone.

The *Džežimskaja Formation* (up to 850 m thick) rests unconformably on the *Četlasskaja Formation*, and has a gritstone unit with conglomerate lenses at the base (GECEN 1970). It is divided into two members. The lower member (up to 350 m) is made up of intercalated grey quartzites, siltstones, and shales similar to those of the *Četlasskaja Formation*. The upper member (500 m thick) consists largely of light, feldspathic quartzites and arkosic quartzite-sandstones.

The upper *Četlasskaja Formation* (up to 700 m thick) and the entire *Džežimskaja Formation* (up to 900 m thick) are exposed in southern Timan (ŽURAVLEV and OSADČUK 1963). Above this an observed thickness of 350 m, of grey or red dolomitic limestones and algal dolomites was measured. These were assigned to the *Bystrinskaja Formation*.

Algal dolomites form several horizons from which RAABEN (ŽURAVLEV et al. 1966) has recognized Upper Riphean stromatolites: *Gymnosolen giganteus* RAABEN, *G. ramsayi* STEINM., *Inseria dzejimi* RAABEN, *Conophyton miloradoviči* RAABEN, *Tunquissia* sp. SEMIKH., *Kassiella enigmatica* RAABEN, *Parmites concrescens* RAABEN.

The *Bystrinskaja Formation* marks the boundary between the western and eastern zones and is usually faulted against the terrigenous sequences. According to GECEN (1970) in southern Central Timan, the terrigenous sequence overlying the carbonates of the *Bystrinskaja Formation* is equivalent to the *Svetlinskaja*, *Četlasskaja*, and *Džežimskaja Formations* of the type section. However, ŽURAVLEV and OSADČUK (1960, 1973) think that the *Bystrinskaja Formation* lies stratigraphically above the *Džežimskaja Formation* and they subdivide it into two members. The Lower carbonate member (up to 2000 m thick) contains the same assemblage of Upper Riphean stromatolites as in Southern Timan. The Upper member (2000 to 2200 m thick) consists mainly of shales with thin dolomite bands. In northern Central Timan (Cil'menskij Kamen') the stromatolite *Poludia polymorpha* RAABEN was found in the dolomites, suggesting an upper Upper Riphean age. However, farther south shales assigned to the Upper member are intruded by diabases dated at 1200–1220 m.y. (MAL'KOV 1969). This supports GECEN's (1970) proposal of different ages for geographically scattered rock-units assigned to the Upper member of the *Bystrinskaja Formation*.

The *Kisloručejskaja Formation*, described and named by ŽURAVLEV and OSADČUK (1962) outcrops in the eastern zone of Central Timan and consists largely of strongly folded shales metamorphosed to the biotite-chlorite subfacies of the greenschist facies.

The lower part of the formation (500 to 600 m thick) is composed of variegated quartz-micaceous and sericite-chlorite-quartz slates with subordinate

bands of mica quartzites. Above is a sequence (up to 1500 m thick) of phyllitic quartz-biotite shales, the upper part containing units of magnetite-bearing calcareous shales. The section is capped by a sequence of thin-bedded quartz-sericite and chlorite shales often with graphite and bands of feldspathic quartzites.

The outcrops by the well 200 km east of Central Timan exposures and Northern Timan and Kanin Kamen' Ridge are also thought to belong to the Kisloručeskaja Formation (ČERNYJ 1965; ŽURAVLEV and OSADČUK 1962).

The Barminskaja Group (up to 3700 m thick) includes the shales of Northern Timan and is subdivided into three formations: the Cape Rumjaničnaja, Maločernoreckaja River, and Jambozerskaja Formations (ČERNYJ 1965; GECEN 1968). A timegap and angular unconformity are indicated between the Cape Rumjaničnaja and Maločernoreckaja River Formations. The Cape Rumjaničnaja Formation (500 to 700 m thick) is made up of garnet-biotite and sericite-biotite shales. The Maločernoreckaja River Formation (2000 to 2500 m thick), consists of alternating dark garnet-quartz-bimicaceous slates, sometimes calcareous, and graphite-bearing green albite-quartz-chlorite shales (meta-volcanics?), and feldspathic quartzites. The Jambozerskaja Formation (1000 m thick) is composed mainly of dark-grey quartz-sericite shales and light-grey quartzite-sandstones with flysch-like rhythms.

Within the strongly folded Precambrian complex of the Kanin Kamen' Ridge three groups are recognized: the Mikulkinskaja, Tarachanovskaja, and Tabuevskaja Groups (GECEN 1971).

The Mikulkinskaja Group consists of amphibolite facies rocks outcropping in two areas on the Kanin Peninsula, the South-East, and along the North-West coast. In the South-East the Cape Mikulkinskaja Group consists of rhythmically alternating dark-grey porphyroblastic garnet two-mica plagiogneisses and lighter equigranular garnet two-mica quartzite shales. Conformable and non-conformable pegmatite veins are common. The boundary with the overlying series is tentatively drawn at the base of a marker bed of garnet-bimicaceous shales with inclusions close in composition to greisenized granites (metaconglomerates) (?). The thickness of the gneiss-shale sequence is more than 500 m.

In the north-west of the Kanin peninsula a sequence (700 to 800 m thick) of alternating micaceous quartzites, garnet-staurolite two-mica slates, and quartz orthoamphibolites was assigned to the Cape Mikulkinskaja Group. Lenses of quartz conglomerate-breccias were found in quartzites of lower (?) beds. The quartzite-amphibolite sequence is intruded by small bodies of granite-syenites, lamprophyre dykes, and pegmatite veins. IVENSEN (1964) and ČERNYJ (1965) thought that the presence of pegmatite veins substantiated the theory that contact-metamorphism was responsible for the high metamorphic grade of the country rocks. However, the presence of staurolite in crystalline schists opposes this.

The Tarachanovskaja Group (4300 to 5300 m thick) is developed along the south-western slope of the Kanin Kamen' Ridge and is represented mainly by quartz-micaceous slates with units of quartzites in the lower and upper

parts of the section. These rocks are metamorphosed to epidote-almandite and biotite-chlorite subfacies of the greenschist facies. A bed of peculiar thin-banded forsterite-diopside-dolomite and diopside-calcite “skarnoids” as well as lenses of poikiloblastic garnet-amphibole hornfels, resulting presumably from selective metamorphism of marls and sand-carbonate nodules, is present in the lower part of the series. Gabbro-diabase bodies altered into amphibolites, intruded prior to folding, are associated with rocks of the Tarachanovskaja Group.

The Tabuevskaja Group (up to 5000 m thick) begins with a flysch-like shale sequence (up to 1500 m thick) with units of feldspar-quartz sandstones at the base. The middle part (1500 to 2000 m thick) is characterized by a rhythmic alternation of metamorphosed carbonate siltstone and mudstone and includes a 100 m thick marker bed of metamorphosed calcareous magnetite-bearing siltstones. Argillaceous silty shales with numerous units of quartzite-sandstones and coaly shales dominate in the upper part (up to 1700 m thick).

The *Ludovatye Mysy Formation* is a 700 m thick dolomite sequence exposed in an isolated narrow horst. It yields the Upper Riphean stromatolites: *Gymnosolen giganteus* RAABEN, *G. ramsayi* STEINM., *Insertia dzejimi* RAABEN, *Parmites con-crescens* RAABEN (ŽURAVLEV et al. 1966). The bore-hole data (GECEN and NAUMOV 1973) suggest that the Ludovatye Mysy dolomites are overlain by a variegated shale sequence (up to 1700 m thick) tentatively assigned to the Vendian. Similar sequences assigned by different authors to the Vendian-Lower Cambrian (or even to the Ordovician) are known from bore-holes of other areas of the Timan–Pechora region.

Conclusions

In summary, the Precambrian of the Timan–Pechora region has the following general features. It consists of lithologically uniform metasediments of mainly terrigenous origin, showing facies variation and rhythmicity, with an almost complete absence of coarse-clastic deposits and visible evidence of extensive unconformities. Carbonate sequences containing Upper Riphean stromatolites (the Bystrinskaja and Ludovatye Mysy Formations), and a sequence with magnetite-bearing varieties of shale assigned by different authors to different parts of the section, may be used as regional markers.

K-Ar ages for shales and postfolding intrusions reported from different areas of the Timan and Kanin Peninsula (IVENSEN 1964) do not aid the subdivision of the section. The oldest ages (790–687 m.y.) were yielded by shales of the (?) Kisloručejkskaja Formation in southern Central Timan; farther north phyllite-like shales yield 660–640 m.y. (whole rock) and 560–483 m.y. in the Kanin Peninsula; similar values, 600–550 m.y., were determined for biotite from crystalline shales of the Mikulkinskaja Group. K-Ar ages of intrusive rocks (on mica from pegmatite bodies inclusive) show an irregular scatter over the range of 665–475 m.y. All these values imply only Pre-Vendian (prior to 680 m.y.) age for the shale complex deposition.

The authors believe that the shale complex of the eastern zone of Timan lies at the base of the Precambrian section and is Pre-Upper Riphean in age. The oldest are rocks of the Mikulkinskaja Group (Kanin Peninsula) metamorphosed under amphibolite facies; they may belong to some other structural complex regenerated by the Baikalian folding. The area where the “Kislyj Ručej” shale complex is developed, includes the Kisloručejskaja Formation of Central Timan, Barminskaja Group of Northern Timan, as well as the Tarachanovskaja and Tabuevskaja Groups of the Kanin Peninsula, can be in general outlined on typical magnetic anomalies confined to units of magnetite-bearing shales. On the basis of stratigraphic position, the Kislyj Ručej complex may be correlated with the Middle Riphean Jurmatinskaja Group of the type section on the western slope of the Urals and with thick clastic sequences of Rybačij (the Rybačij Group) and Varanger (The Barents Sea and the Raggo Groups) peninsulas. These terrigenous complexes are very thick and intensely folded. They formed within the inner zone of the late Precambrian (probably mainly Middle Riphean) trough and are separated from the outer zone by a series of steep thrusts en echelon.

The outer zone includes terrigenous formations of Southern and Central Timan as well as carbonate formations (the Bystrinskaja and Ludovatye Mysy) dated, on the basis of the stromatolite assemblages, as upper Riphean (Min’jarskaja Formation of the Southern Urals). The Kil’dinskaja Group of the Kola Peninsula and the Tanafjord Group of North Norway also belong to the outer zone. This trough may have begun to form after the Upper Riphean and orogenic formations of Baikalides (the Volokovaya Group of the Srednij Peninsula, Vestertana Group of Finnmark) deposited in inherited or superimposed depressions in Vendian time.

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Precambrian and Palaeozoic development of northern Greenland

By PETER R. DAWES

Abstract

Regional Proterozoic and Palaeozoic structural and stratigraphic trends in northern Greenland form a subrectilinear pattern flanking the crystalline Greenland shield (Fig. 1). The shield forms a core that is bordered on three sides by sedimentary basins; the intra-cratonic Thule basin of Proterozoic rocks on the west, the North Greenland geosyncline (Proterozoic to Devonian) on the north and the East Greenland geosyncline (Proterozoic to Silurian) on the east. The regional trends in each basin conform to the general configuration of the subcontinent.

The Thule basin and the North Greenland geosyncline have westerly extensions in adjacent Ellesmere Island and this indicates the close juxtaposition of Greenland and Canada north of 76° N in late Precambrian and Palaeozoic time. Recent geological and geophysical data from the Nares Strait region places restraints on overall transcurrent movement between Greenland and Ellesmere Island. Pre-drift fits that show Ellesmere Island and Greenland widely displaced are not in agreement with the regional geology.

Post-Archean earth movements of three main ages affected northern Greenland: late Proterozoic, Palaeozoic, and late Phanerozoic (Fig. 2). The Proterozoic movements produced graben and block fault tectonics in the Thule basin and folding and metamorphism in East Greenland. The Palaeozoic movements deformed the North and East Greenland geosynclines into orogenic belts of contrasting styles defined by a dominant northerly sense of tectonic transport away from the platform in the northern belt and a westerly directed thrusting towards the platform in the eastern belt (Table 1). The orogenic movements were initiated at various times: middle to late Ordovician in central East Greenland, Silurian and ?younger in northern East Greenland, and Devonian and ?younger in the North Greenland geosyncline.

The North Greenland fold belt is regarded as a marginal compressional structure of a Palaeo-Arctic ocean involving interaction of oceanic and continental crust. The East Greenland fold belt, as part of the circum-Atlantic Caledonian orogenic system, is generally thought to be the result of contraction of a Palaeo-Atlantic ocean, involving collision of opposing continental margins with their accreted geosynclinal tracts. The E-W compression causing intense folding and thrusting in East Greenland spanned some 60 m.y. from late Ordovician to latest Silurian and Devonian. This diastrophism transgressed northwards as indicated by the contrasted stratigraphical record between the segments of the fold belt north and south of latitude 76° N. The stratigraphical record and timing of orogenesis in the northern segment of the East Greenland fold belt and the North Greenland fold belt show certain similarities and it is suggested that both orogenic belts (which both show certain geological similarities to Svalbard) may be part of a single orogenic system linked with the Lomonosov Ridge.

The junction of the two orogenic belts in the north-east corner of Greenland became the site of a regional depression, the Wandel Sea basin, in late Palaeozoic time. This basin contains a late Palaeozoic, Mesozoic and Tertiary sedimentary sequence and is now a NW-trending tectonic zone.

Table 1

Summary diagram illustrating the nature and timing of the mid-Palaeozoic orogeny in the North and East Greenland fold belts.

		North Greenland fold belt	East Greenland fold belt	
			76° N to 82° N	70° N to 76° N
Sedimentary regime	Age of youngest known geosynclinal strata	Lower Devonian (Gedinnian)	Upper Silurian (Ludlovian)	Middle or ?Upper Ordovician
	Age of oldest known post-orogenic strata	Upper Carboniferous (late Pennsylvanian)	Lower Carboniferous (Mississippian)	Middle Devonian (Givetian)
Tectonic regime	Direction of main tectonic transport	North towards the Arctic Ocean	West away from the Atlantic Ocean	
	Character of fold belt margin	Autochthonous – gradual incoming of structural elements	Allochthonous – thrust and nappe front	
	Tectonic status of platform	Hinterland	Foreland	

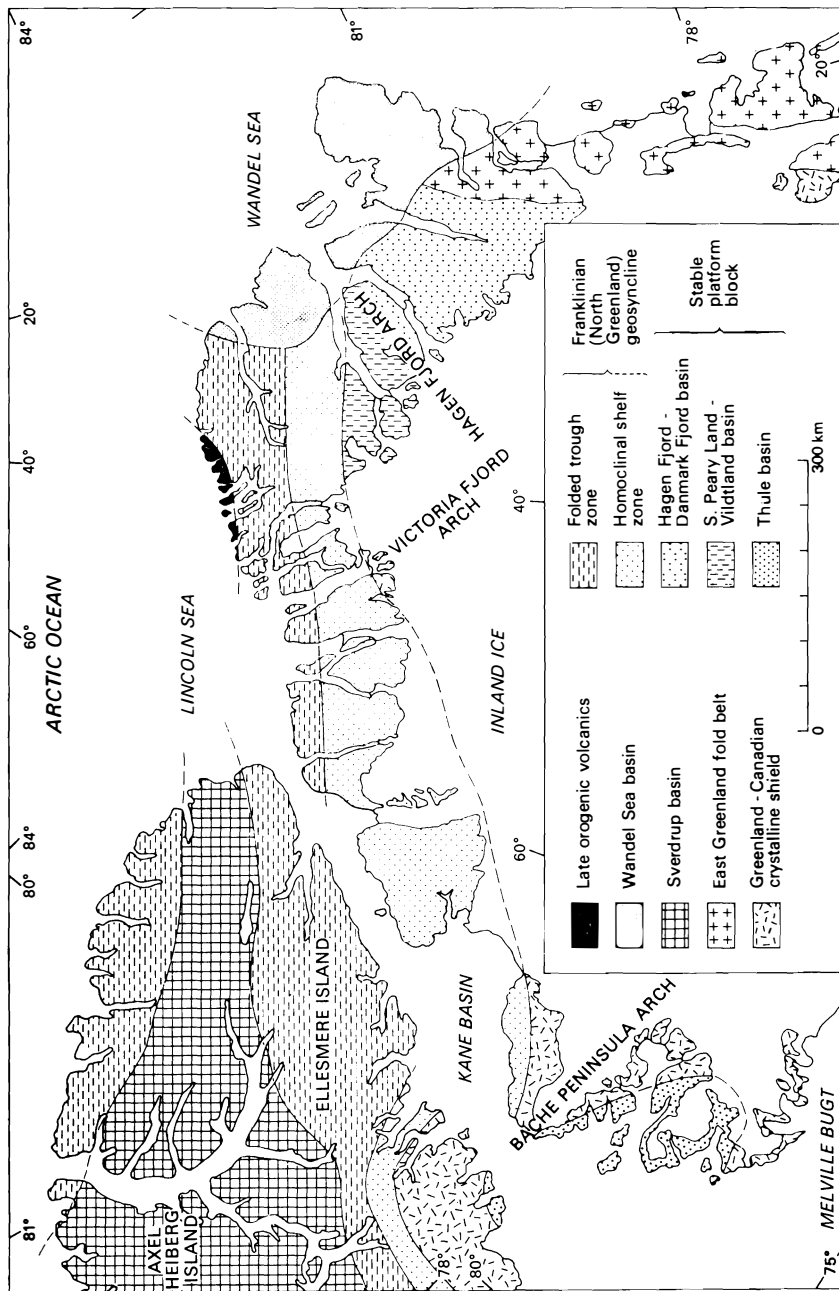


Fig. 1. Main stratigraphical-structural units of northern Greenland and adjacent Canada.

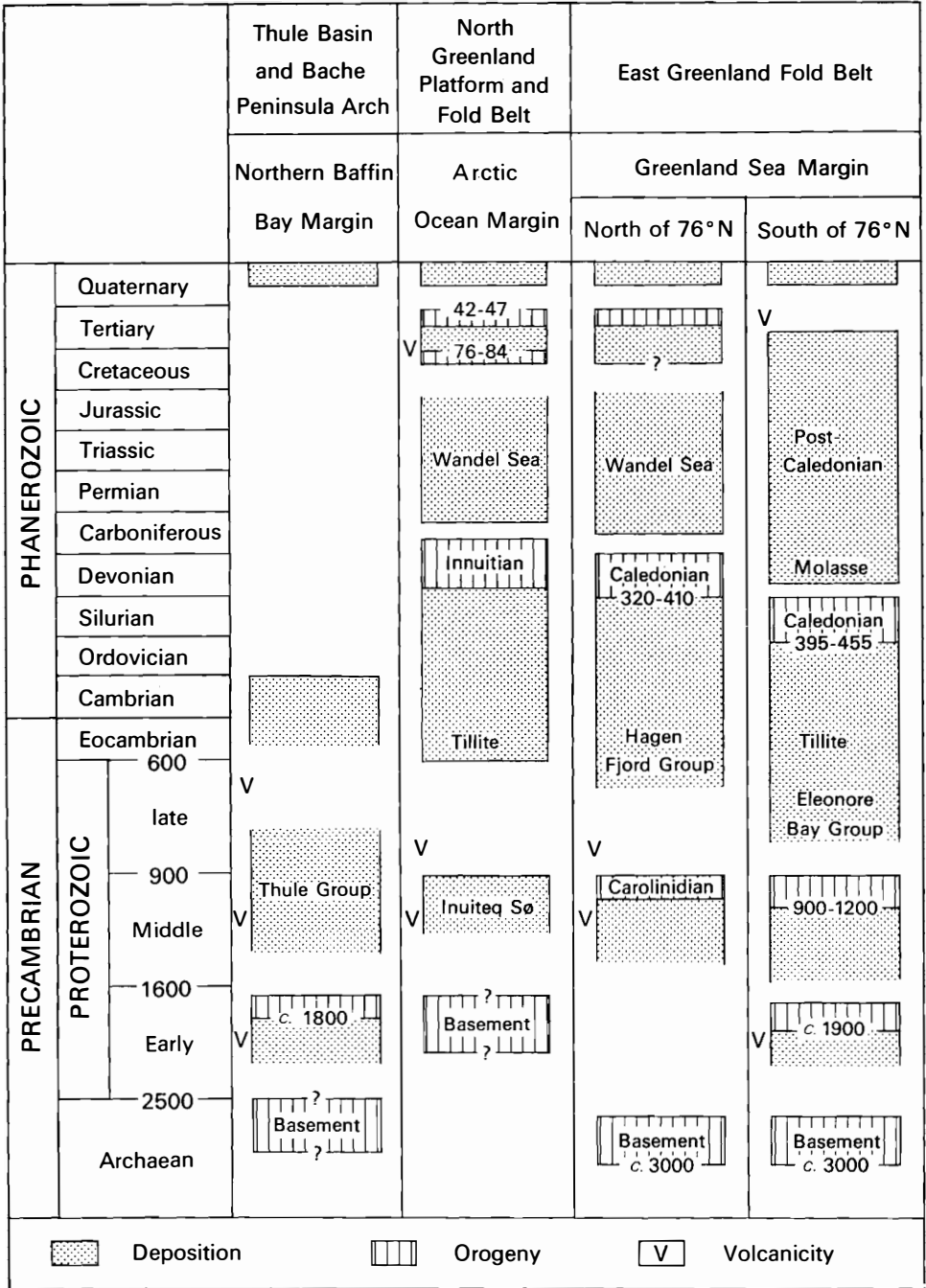


Fig. 2. Chronological chart illustrating the main depositional, orogenic and volcanic events of northern Greenland.

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