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THE PRE-DEVONIAN SEDIMENTS
ON YMERS Ø, SUESS LAND, AND ELLA Ø
(EAST GREENLAND) AND
THEIR TECTONICS

BY

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WITH 19 FIGURES IN THE TEXT
AND 3 PLATES

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**Abbreviations used in the text, the figures,
and the plates.**

B = Barriere fault.
MS (BS) = Multicoloured Series.
EBF = Eleonore Bay Formation.
HV = Main fault.
LDS (KDS) = Limestone-Dolomite Series.
QS = Quartzite Series.
T = Upper Tillites.
t = Lower Tillites.

PREFACE

In the summer of 1947, when the Greenland expeditions under the leadership of Dr. LAUGE KOCH which had been interrupted by the war, were continued again, I had the opportunity as a member of the expedition to carry out geological mapping in East Greenland. I stayed there from August 1947 to September 1948, and in addition for five weeks in July and August 1949.

On such expeditions one depends to a much greater extent on the help and goodwill of others than during field investigations in the inhabited civilised regions. It is therefore only in part a merit of my own that I was able to finish the present work.

I wish to express my cordial thanks to Dr. LAUGE KOCH, owing to whose initiative the many expeditions rich in results to E. Greenland and accordingly also the present work, were rendered possible. He was always ready to offer me all facilities and support imaginable.

The working up of the field observations were undertaken in the Geologisch-palaeontologischen Institut der Universität Basel, under the guidance of Professor Dr. L. VONDERSCHMITT, to whom I am indebted for my education and countless instances of help and consultation.—Professor Dr. M. REICHEL and Dr. E. WENK helped me with various special problems of palæontological and petrographical nature. My most cordial thanks are due to them and to my fellow students.—In the early summer of 1947 Dr. H. BÜTLER, Schaffhausen, introduced me into the problems of my work and gave me much valuable information as to equipment and the work in practice.

I also remember with gratitude my wintering companions, in the first place Dr. W. HUBER, Zürich. On many journeys and excursions I enjoyed his kind companionship. AA. DE LEMOS, the leader of the station Ella Ø, and P. POVELSEN, station assistant, were always ready to offer a helping hand in the preparation of the journeys. Indispensable, notably on the sledge journeys, was the service rendered by the Greenlanders PAVEA and EBA OLSEN, Akunaq, and KRISTIAN KUNAK, Scoresbysund.

My two fellow students from Basel, E. FRÄNKEL and J. HALLER, accompanied me in the summers of 1948 and 1949, respectively, on the

excursions in my working area. Discussions with them as well as their help in the daily work were very valuable to me, and I thank them most cordially. I also remember with gratitude the spirit of helpfulness and comradeship that prevailed during the expedition among the scientists as well as among the ship's company of the "Gustav Holm".

In particular I thank my parents, who at great sacrifice enabled me to fulfill my study, and to whom I dedicate the present work.

INTRODUCTION

The region dealt with in the present paper is situated in the East Greenland fjord district in ca. 73° N. lat., 25° W. long. (see fig. 1). It includes the western part of Gunnar Anderssons Land, the western part of Ymers Ø, the eastern Suess Land, and Ella Ø. To the north the area is bounded by Kejser Franz Josephs Fjord, to the south by Kempes Fjord and Narhvalsund, respectively. Ymers Ø and Suess Land are separated from each other by Antarctica Sund.

The highest elevations in the area described rise to about 2000 m above sea-level. In most cases the land rises steeply from the fjord to 800—1000 m, then more slowly to the point of culmination. Gunnar Anderssons Land and Ymers Ø as well as Suess Land are constantly covered with ice caps, which issue numerous glaciers into the valleys. Many of these glaciers no doubt reached the fjords in the last historic periods. The continued amelioration of the climate in the last decades have, at any rate, given rise to a considerable melting of the ice masses.

The peculiar character of the arctic regions put certain limits to the geological investigations. Apart from the extensive ice caps, sharp-cut valleys or steep rock walls prevent the access to large areas. Tributary rivulets in deep ravines often made whole valleys inaccessible, especially when the investigation had to be made without any companion. In the following text reference will repeatedly be made to this fact, in order that the value of individual observations may be correctly understood.

The earliest geological observations recorded from our area are due to the "2. Deutsche Nordpolarfahrt" under KOLDEWEY in 1870 and NATHORST's expedition about the turn of the century. However, not till 1926 did the systematic investigation of the fjord districts commence. In that year and subsequent years the region was visited by summer expeditions, thus by WORDIE (1930), WHITTARD, PARKINSON (1931), etc. But KOCH, above all, on his sledge journeys in 1926—27, acquired an idea of the numerous problems presented by the fjord area (1929, a, b). The expedition in the summer of 1929 was also led by him, and it was joined by a whole staff of specialists (BACKLUND (1930, 1932), KULLING (1930), POULSEN (1930, 1932, 1937), etc.). In 1931 the

two scientific wintering stations Ella Ø and Eskimonæs were built, which from now on served as bases for extended and intensive investigations. In almost uninterrupted succession, wintering scientists replaced each other at these stations until 1939 (TEICHERT (1933), WEGMANN (1935), BÜTLER (1935 a, b; 1937, 1940 a, b), STAUBER (1939, 1942), VISCHER (1939, 1943), MAYNC (1940, 1942), SCHAUB (1942, 1950), etc., etc.). To us, also, Ella Ø served as wintering station.

Within the scope of a detailed mapping of the pre-Devonian sediments, Dr. L. KOCH and Dr. H. BÜTLER proposed that we were to commence the mapping of Strindbergs Land. However, unfavourable weather, which rendered it impossible for us to circulate, compelled us even at the beginning of the summer of 1947 to an immediate change of program. Dr. KOCH offered us, as a new working field, Suess Land, adjacent to the station, to which, in the summers of 1948 and 1949, Ymers Ø and Gunnar Anderssons Land were added.

As basemaps for our mapping we had the topographic maps of the Geodetic Institute, Copenhagen. The originals of these maps were published on a scale of 1:250,000, but we carried out our mapping on enlarged copies on a scale of 1:100,000. It is no wonder that the map prepared under very difficult conditions at the beginning of the 1930s was encumbered with some deficiencies; if these had any greater influence on the geological mapping, a note is added to the present description.

Of invaluable help to us were the excellent air photographs taken and developed in the summer of 1949 by Mr. MAX BRENNEISEN, M. Eng., and Mr. ERNST HOFER, both of the Eidg. Landestopographie, Bern. To them, also, cordial thanks are due for their always ready fulfilment of the numerous wishes of different kind of the various geologists.

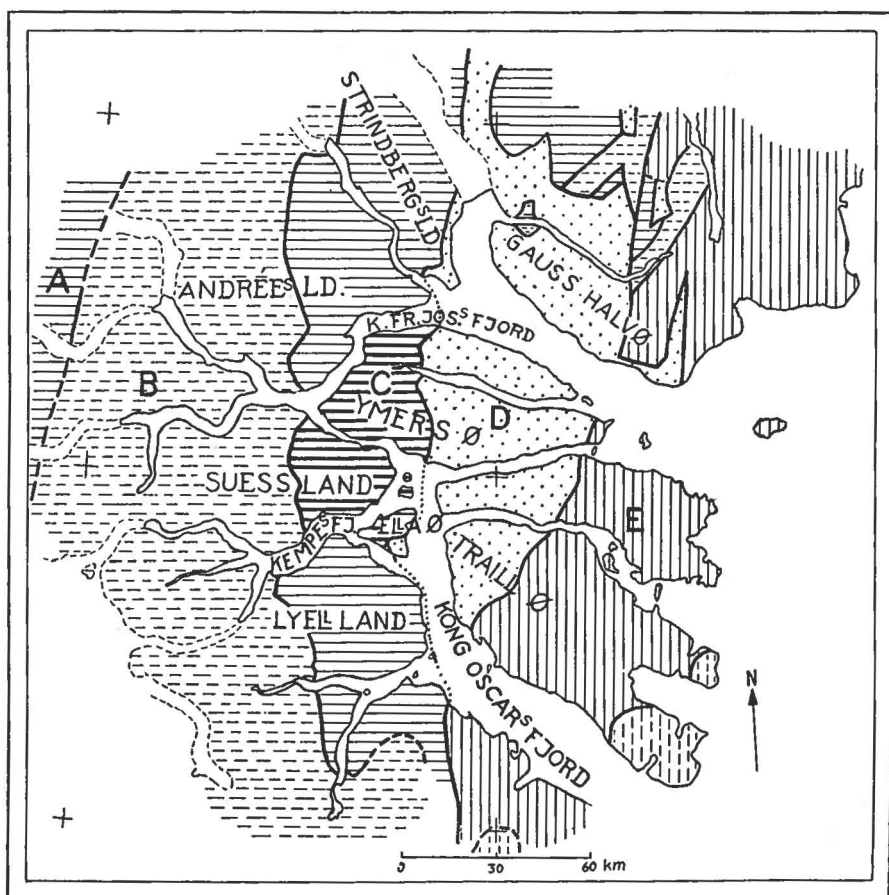


Fig. 1. Geological outline map (after WEGMANN, BÜTLER, STAUBER, etc.). Legend: A = Petermann Series. B = crystalline complex (Vela massif). C = Eleonore Bay Formation and early Palæozoic (heavily ruled: the area described). D = Devonian. E = Carboniferous to Cretaceous. F (the outer part of Traill Ø) = Cretaceous-Tertiary eruptives.

GEOLOGICAL SURVEY

On the basis of the investigations made so far the areas surrounding our working field can be divided as follows (cf. fig. 1):

To the east and west of a NNE—SSW-running zone of exclusively crystalline rocks (B in fig. 1) there occur NNE—SSW-striking stripes of sediments and volcanic rocks (A, C, D, E, and F. in fig. 1).

The individual zones can be characterised as follows:—

Zone B represents KOCH's Vela massif or WORDIE's western crystalline complex. The first detailed investigation of this area was under-

taken by WEGMANN (WEGMANN 1935a); the results were published in a preliminary report. In this summary we therefore confine ourselves to WEGMANN's results. In the present form of this crystalline complex he recognised various zones: a) contact-metamorphic Caledonian-folded sediments, b) slightly migmatised gneisses of sedimentogenous origin (Isfjord zone), and c) a highly migmatised gneiss complex (Sylva Maria and Hagar group). In addition he records, east and west of this crystalline zone, granite which is younger than the Caledonian folded sediments. While WORDIE (1930), PARKINSON (1931), and others regarded the metamorphism as pre-Caledonian, KOCH, BACKLUND, and WEGMANN date the migmatisation as simultaneous with the Caledonian phase; WEGMANN, however, mentions the possibility that the Sylva Maria group may still exhibit traces of an earlier folding. WEGMANN regards the ascertainment that the Greenland Caledonides represent a far-reaching interruption of the ancient Greenland shield as the most important result. This is in opposition to TEICHERT, WORDIE, and others, who were of opinion that these Caledonides should be regarded exclusively as structural elements of a germanotypical mode of folding.

West of the crystalline complex, near the edge of the inland ice, sediments appear again, viz. the Petermann Series (Zone A). Apart from their name and their existence, not very many facts are known. It seems to be an equivalent to Zone C east of the crystalline zone. Whether the supposition of TEICHERT, that it is the marginal facies of the late pre-Cambrian and early Palæozoic trough, is correct, must be decided by further investigations. It is likewise uncertain whether the Petermann Series includes tillites, also.

Zone C (horizontally ruled) is occupied by the late-pre-Cambrian Eleonore Bay Formation (EBF) with the pre-Cambrian tillites and a sedimentary sequence reaching from the Lower Cambrian into the Lower Ordovician. The following publications deal entirely or in part with these rocks: L. KOCH (1929), M. PARKINSON (1929), W. WHITTARD (1929), J. M. WORDIE (1930), H. G. BACKLUND (1930), C. TEICHERT (1933), C. E. WEGMANN (1935 a), H. BÜTLER (1937, 1948), H. P. SCHAUB (1950), and W. HUBER (1950). The Caledonian orogeny is well recognisable in zones A, B, and C.

Our working field is indicated by heavy ruling.

Zone D shows the distribution of the Devonian sandstone and conglomeratic complexes. They have mainly been treated by G. SÄVE-SÖDERBERGH and H. BÜTLER. The beds are mainly of Upper Devonian age, though in recent time Middle Devonian was also found to the south on Gauss Halvø. These beds rest unconformably on zone C, which was subject to Caledonian folding and pre-Devonian erosion. As pointed out by WEGMANN, however, the material removed by erosion is absent.

There is a stratigraphic gap from the Lower Ordovician to some time into the Upper, respectively Middle, Devonian. The Devonian sediments themselves were folded in "late-Caledonian" time. BÜTLER distinguishes between five phases of folding, during which the main mass of Devonian beds were thrown into broad folds. In some places, however, these movements are evidenced by upthrusts and intensive folding. A certain influence of the movements of the underlying pre-Devonian substratum on the folding activity in Devonian times is also ascertainable, especially to the north.

The large post-Devonian main fault¹⁾ of VISCHER (1939, 1943) separates the Devonian area from zone E. This zone contains the sediments from the Carboniferous to and including the Cretaceous period. After the continental deposits of the Devonian and Carboniferous, there commenced, with the marine Permian, a repeated alternation of transgressions and regressions, lasting till Cretaceous time. In addition, as a tectonic characteristic, subdivision into step-faults took place, at times resulting in a westward dip of the individual blocks along the faults. This step-faulting set in after the Devonian time and, after various base-levellings and superpositions, was constantly reactivated (cf. VISCHER, STAUBER, MAYNC).

Three periods of magmatic activity can be ascertained (cf. RITTMANN 1939, 1940):—

- a) Caledonian plutonic activity, and in association with this, the metamorphism of Zone B with its gneisses, various granites, and lamprophyres (see also p. 97).
- b) Palæozoic volcanic and plutonic activity with extrusion and formation of veins in Devonian and Carboniferous times.
- c) Cretaceous-Tertiary volcanism and subvolcanism (SCHAUB 1942). We are here concerned with the gigantic basalt extrusions in the regions near the coast and with young eruptives (these latter indicated as Zone F).

Our own investigations were limited to Zone C, more precisely the area between Kejser Franz Josephs Fjord and Kempes Fjord. In the following chapters the stratigraphic constitution and subsequently the tectonics will be treated.

¹⁾ Not to be confused with BÜTLER's main fault on Gunnar Anderssons Land (cf. pp. 39 and 45).

STRATIGRAPHY

A. General Remarks.

The Caledonian folded, non-metamorphic, pre-Devonian sediment areas on Ymers Ø, Suess Land, and Ella Ø have chiefly been treated, entirely or partially by L. KOCH (1929), KULLING (1929), POULSEN (1932, 1937), TEICHERT (1933), WEGMANN (1935a), and BÜTLER (1937, 1948). They comprise a sequence of strata extending from the late pre-Cambrian till well into the Lower Ordovician (cf. fig. 1 and Pl. 1).

POULSEN subdivided the Palæozoic in most cases on the basis of fossils. As stated by BÜTLER, these horizons can be distinguished lithologically and are accordingly easily distinguishable in the field. The Palæozoic is underlain by conglomerates, which, in agreement with other authors, we will denote tillites (pp. 26—33). They belong to the Eleonore Bay Formation (EBF), named so by L. KOCH (1939a, b), and correlated by him with the Thule Formation of NW. Greenland and referred to the late pre-Cambrian series Greenlandian.

The most detailed stratigraphic records on the Palæozoic were given by POULSEN, those on the tillites by KULLING, and those on the pre-Cambrian by TEICHERT. For some groups, it is true, we might have preferred another subdivision; since, however, the delimitation of the unfossiliferous pre-Cambrian horizons must be purely lithologic, and thus to a certain extent remains arbitrary, we abstained from introducing new names and alterations of the sequence of designations.

Like BACKLUND and WEGMANN, TEICHERT, also, distinguishes within the EBF, from below upwards, a quartzite series (QS), a multi-coloured series (BS), and a limestone-dolomite series (LDS). The recent investigations by FRÄNKEL and KATZ in Andrées Land, respectively Strindbergs Land, have shown that sediments are even present below the QS, they will be called the Eremitdal series. According to information by P. HENDERSON, and as we could ascertain ourselves on a flight over the regions of Alpefjord, even much lower sedimentary elements than the Eremitdal series occur here. A brief stratigraphic description of these Alpefjord sediments by FRÄNKEL (1951) has just

appeared. In addition, a description of Lyells Land and the adjoining areas to the south, which are planned to appear shortly, will provide further data. In the region treated in the present description, however, the lowest horizons are the bed groups of the Lower Quartzite Series.

The total thickness of the late pre-Cambrian sediments is very differently recorded by the various authors: BACKLUND 16,000 m, TEICHERT 3,850—4,050 m, WEGMANN 7,500—8,000 m, possibly 9,000 m. Most striking is the thickness given by BACKLUND, 16,000 m. In this amount, however, BACKLUND not only included the stratigraphic thickness, but also increases of the thickness due to tectonic causes. A close examination of the enormous complexes in Alpefjord (Eremitdal series and deeper sediments), might perhaps show that BACKLUND's figures correspond most closely to the actual conditions.

The total Caledonian sediment sequence, as already mentioned, is overlain by the Upper Devonian—in some regions Middle Devonian is present—transgressing on the eroded post-Caledonian landscape. Thus deposits from the period between the Lower Ordovician and the Middle Devonian are absent, as already pointed out by WEGMANN (1935a). Where the material is to be found, through the removal of which the pre-Devonian surface topography developed, is unknown.

As mentioned above, there is still no certainty as to the lower boundary of the pre-Cambrian sediments. A transgression on older complexes has not hitherto been observed anywhere. Possibly the lower boundary is blotted out everywhere through metamorphism; according to FRÄNKEL, however, the Eremitdal Series of Andrées Land exhibits distinct traces of metamorphism.

As to the Palæozoic our stratigraphic description is based on the subdivision of POULSEN, and as to the EBF on that of TEICHERT. Such boundaries were selected as seemed most suitable for a geological mapping. Consequently some minor alterations in the subdivisions of the two authors had to be made. During the working up of our investigations appeared the paper by H. P. SCHAUB (1950), which, especially for the limestone-dolomite series and the tillites, gives a further subdivision, chiefly on the basis of laboratory work. SCHAUB's investigations will be dealt with in the description of the particular groups of beds¹).

¹) A different subdivision is also used by A. B. CLEAVES and E. F. FOX. Since, however, their division has in the main only reference to the area between Blomsterbugten and the western end of Duséns Fjord, that is to say, it is mainly local, it is not employed by us. Moreover the division according to age erected by them is very problematic. Some fossils figured in their papers are undoubtedly derived from the Cambro-Ordovician of the Barriere region, which they passed on their walk from Kejser Franz Josephs Fjord across the Midnatspas to Blomsterbugten. Reserve is especially to be taken as regards the fossil which is figured in the publication,

It may be expedient to make clear the meaning of some terms. Notably WEGMANN (1935) and BÜTLER (1948) often use the expression "Caledonian sediments" for the pre-Devonian deposits. When we use this term, we use it with the same meaning as these two authors, though actually the Devonian deposits, also, might be regarded as "Caledonian", since according to BÜTLER (1948, p. 115) they were folded in late Caledonian time.

The upper boundary of the Eleonore Bay Formation is placed by us above the tillite series (cf. pl. 1), that is, at the lower boundary of the upper Bastion Formation (upper Lower Cambrian). However, this does not mean to say anything about the age of the tillites. With "pre-Cambrian sediments" we denote the EBF exclusive of the tillites. As "Palæozoic" rocks we denote the beds overlying the lower Bastion Formation (sandstone series). We let it remain an open question whether the tillite series should be referred to the Cambrian or the pre-Cambrian.

We also know that the naming of the formations and beds not by far fulfil the requirements as to a uniform classification of this notion (e. g. the "Eleonore Bay Formation" represents a comprehensive general complex, while the "Ella Ø Formation" covers only a single bed). A new designation of the formations, however, would have rendered a comparison with the existing literature so difficult that it would not be justifiable.

Before commencing the description of the individual beds, a remarkable fact should be pointed out:

Facies differences often occur in the horizontal beds. However, if it is borne in mind that in the area described we are concerned with a complex of strata of a thickness of over 7000 m, the great general constancy of both the thickness and the facies conveys a much stronger impression than the minor differences in facies. Even though, to the north, a bed may be more dolomitic than to the south, or here it may be more bedded, there more massif, it should still be kept in mind that it is a fractional part of a complex which extends over 300 km, and that such minor differences neither indicate radical changes in the deposition nor had any influence on the later tectonic events.

The correct understanding and noting down of all these minor alterations therefore seemed less important to us than the analysis of the tectonic structure. Hence we refrain from recording all the detailed

interpreted as an Arachnide, and named *Sidneyia groenlandica* n. sp. On the basis of this single find the whole sediment series in Noa Dal is dated as not older than the Middle Cambrian. Judging by the figure alone, it seems to us to be more probably damaged ripple marks than the abdominal segments of an Arachnide. Numerous examples of such ripples halfway removed by erosion are found in the QS. CLEAVES and Fox were probably aware of the uncertainty of their determination of this fossil.

sections measured by us, and consider the description of the tillite series and its relations to the substratum to be of greater importance.

B. Description of the Individual Beds.

1. The Eleonore Bay Formation.

TEICHERT (1933), who especially treats the stratigraphy of the Multicoloured Series, introduced a numbering of the beds which we completed upwards and downwards (see pl. 1). The numbered beds are distributed as follows over the series already named by TEICHERT:

Lower Quartzite Series (LQS)	Bed groups	1—3
Upper Quartzite Series (UQS)	—	— 4—6
Lower Multicoloured Series (LMS)	—	— 7—10
Upper Multicoloured Series (UMS)	—	— 11—13
Lower Limestone-Dolomite Series (LLDS)	—	— 14—17
Upper Limestone-Dolomite Series (ULDS)	—	— 18—20

a. Lower Quartzite Series.

This series has its main distribution in Noa Dal north and south of Blomsterbugten. Additional occurrences are found at the southern mouth of Margeries Dal and at the northeastern corner of Suess Land.

The deepest-lying beds of this series are present in the working field in Noa Dal. Here Bed group 1 occurs *in situ*, but its lower boundary is not exposed. The rocks consist in the main of dark rusty-weathering quartzitic shales, alternating with massif white yellow-rusty-weathering quartzite beds. These beds attain thicknesses of 10—40 m, while the shaly complexes are much thicker. The quartzites often exhibit splendid ripple-marks.

Bed group 2. 340 m. The group falls into two parts: Below, well-bedded reddish to violet quartzites, in most cases with distinct cross-bedding and ripple-marks. The upper part consists of more massive white quartzites, which are occasionally interbedded with fine clayey beds and are ca. 100 m thick. They mostly turn yellow or rusty on weathering.

Bed group 3. 500 m. The rocks of Bed 3 are mainly dark olive-green shales, from a distance appearing as a black complex. They are characteristic of the dark rock walls south of Noa Sø, and of a large part of Noa Dal, and consist of quartzitic schists. Interbedded with them are fairly coarse-grained quartzites, which, probably owing to remnants of weathered pyrite, are spotted by brown dots and thus present a

peculiar dotted appearance. Marks of the size of a coin were found in the shales; they consisted of concentric circles and should probably be interpreted as fossil gas outlets.

b. Upper Quartzite Series.

The series occupies large areas of the northern part of Ymers Ø and Gunnar Anderssons Land. It further occurs in vast areas on the north coast of Antarctica Sund. On Suess Land it is restricted to the southeastern part.

Bed group 4. 380 m. It begins at the base with two light-yellow quartzite beds of a combined thickness of 50 m. These quartzites bear a fairly close resemblance to those of the upper Bed group 2. Here they often exhibit beautiful ripple-marks, especially west of Rosinante in Noa Dal, whence they were figured by CLEAVES & FOX (1935). The distance from one wave ridge to another is ca. 50 cm, and the height of the waves is ca. 20 cm.—The basal quartzites are followed upward by black shales alternating with white quartzite beds, so that the quartzite beds decrease in thickness and frequency upwards. In the uppermost 30 m, above uniform black shales, there occur some 1—2 m thick quartzite beds, which are overlain by the deep-red shales of

Bed group 5. This group measures 150 m. To the west it is made up of dark red quartzitic shales, which, however, in the uppermost 50 m exhibit an increasingly green colour. In the landscape the huge red talus slopes are characteristic of Bed group 5, thus near the Rosinante or at the eastern mouth of Carissima Dal (northeastern corner of Suess Land).

Bed group 6. 130 m. At the base of the group occurs a white massive quartzite bed about 30 m thick, which in southern Suess Land and at Kap Alfred on Lyells Land is divided into two parts by ca. 5 m of dark shales. The basal bed is overlain by uniform red, well-bedded quartzites. They turn somewhat more rusty on weathering than the shales of Bed group 5. On Lyells Land the light-red and white portions of this bed group seem to become predominant. In our area, however, the bed group provides an excellent guide horizon, as it stands out by its colour and is very rarely covered by scree.

c. Lower Multicoloured Series.

The best exposures of this series occur on Gunnar Anderssons Land, the southern part of Ymers Ø, at Skildvagten (NE. Suess Land), and in southwestern Suess Land.

Bed group 7. 190 m. The group is characteristic by its yellow, green, and red dolomitic and quartzitic shales. The red colour dominates, and in many places it can be ascertained that, by transformation of Fe^2 into Fe^3 , it has replaced the green colour. Bed group 7 includes no stiff beds; accordingly it is one of those mobile areas which, compared with the relatively stiff Bed 6, often forms intensive disharmonious folds, as at the northern foot of Skildvagten.

Bed group 8. 30 m. The distinction of this bed in the described region was only retained because it was undertaken already by TEICHERT for the northern area (Andrées and Strindbergs Land). While in the northern area it is much more distinctly bedded and hence can be distinguished as a distinct horizon, in our region it might just as well be regarded as the uppermost part of Bed group 7. It consists of yellow dolomite-shales, which, at any rate in most cases, are readily recognisable in the landscape and, as a light-coloured band, accompany the base of Bed 9.

Bed group 9. 150 m. Dark to black bituminous limestones, often rather shaly. Frequently white calcite veins pierce the shales at right angles to the sedimentation plane.

Bed group 10. 150 m. Consisting of rapidly alternating red and yellow dolomitic shales, this group, also, is a striking guide horizon within the Multicoloured Series. In the exposed area it much resembles Bed group 7, though here the colours alternate much more intensively, and green shales are rare. On Ymers Ø and Gunnar Anderssons Land there occurs, in the upper third, a remarkable, almost white band 25—30 m thick, which appears to be more massive than the other rocks and accordingly stands out from the most highly disintegrated portions of Bed group 10.

d. Upper Multicoloured Series.

Characteristic occurrences are found: at Skildvagten, Johnstrups Bjerg (southeastern Suess Land), the northern shore of Kempes Fjord, on Maria Ø, and on Ruths Ø.

Bed group 11. 150 m. Dark, often even black limestones, which are highly bituminous, and emit a strong H_2S smell when struck. Bedded, rather massive portions alternate with thin shales.

Bed group 12. 200 m. This bed is distinguished by its stiffness. It consists of white or light-grey dolomites. While larger tectonic events are manifested in Bed groups 10 and 11 as intensive minor folds, the rocks of Bed group 12 exhibit mostly tranquil forms or react with faulting (Niviarsiat, SW. Blomsterbugten). In the upper part of Bed group 12

we encounter for the first time algal biostromes in a wonderful development (for further description, see below). The uppermost ca. 30 m of the group are intensively yellow-coloured, probably by solutions from the overlying red quartzites of Bed group 13, e. g. algal stocks carrying a brown limonite crust were found, especially in Carissima Dal.

Bed group 13. 300 m. The main mass of the group is made up of very well bedded red and greenish quartzites, red to rusty-brown on weathering. In the uppermost 40 m there occur 2–3 light yellow dolomite bands, each of a thickness of 2–5 m. Owing to these yellow dolomite bands, the group is the most remarkable of the Multicoloured Series. The dolomite bands are further distinguished by abundant algal structures.

When in the general section on the stratigraphy it was mentioned that various differences in facies occurred, the constancies, e. g. precisely in these light-coloured dolomite bands, seemed to us of much greater importance. It is true that these bands are somewhat more numerous, e. g. in the region around Geologfjord, but the fact that the huge rock series of more than 7000 m include 2–5 m which show only a slight change along a distance of 300 km, is no doubt more remarkable than the fact that their thickness and number vary. This must be all the more emphasised since it has reference not only to the single case of Bed group 13, but similar conditions are met with in all the pre-Cambrian and early Palæozoic beds with the exception of the Limestone-Dolomite Series.

The algæ of Bed groups 12 and 13.

The algæ occurring in these beds are readily distinguishable from those found in Bed group 15 and above (cf. p. 22 ff.). Common to both forms, however, is their mode of occurrence: They are typical "Rasen" or, as termed in the English-American literature, "biostromes".

In horizontal section the algæ of Bed groups 12 and 13 appear as concentric shells, and attain a maximum diameter of 15–20 cm. In vertical section only lamellæ, convex upward, are seen, which explains the concentric circles in the horizontal section.

Fig. 2 shows the pattern of this algal structure. It is derived from Ruths Ø, from one of the dolomite bands of the upper Bed group 13. The locality is located about 2 m above the level of the fjord. The structures were therefore beautifully exposed by the surf. The halves of two individuals are visible, and between them the zone where they join. This is, measured on the individual algal columns, rather narrow and probably presented little or no room for the intrusion of sedimentary particles. On the back of the specimen figured it can be seen that the

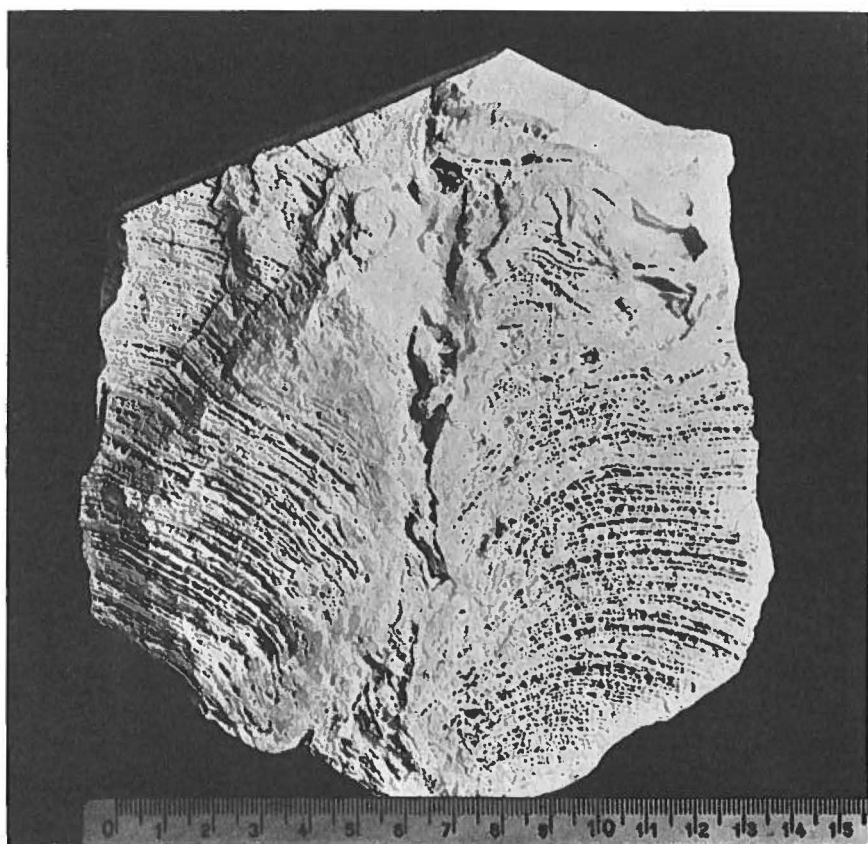


Fig. 2. Dolomitic alga of the upper portion of Bed group 13. Locality: East side of Ruths Ø. Scale in cm.

interspace between the two individuals is only recognisable as a seam 1 mm wide.

In thin section the structures unfortunately show no further details. The dolomitising recrystallisation has blotted out every trace of the delicate organic structures. The macroscopically visible layers and lamellæ appear under the microscope merely as larger and smaller dolomite crystals arranged in bands, but quite probably structural elements of the algæ disappeared during the recrystallisation.—We did not succeed in finding, in the literature, descriptions or figures permitting the identification of this form of algæ with any one previously described.

e. Lower Limestone-Dolomite Series.

Bed group 14. 270 m. Group 13 is overlain by about 25 m of dark grey limestones. Above follow ca. 20 m of bedded, dirty-yellow-weather-

ing grey dolomitic limestones, succeeded by dark-grey massive limestones. Especially at Niviarsiat and Skildvagten some peculiar structures can be observed, which generally follow the bedding planes, but are extremely irregularly delimited. They are light-coloured, nearly purely white, or light reddish portions, which are, so to speak, completely dolomitised. They vary very suddenly in thickness between 5 and 20 m, and end abruptly. At the south coast of Suess Land, also, similar structures are met with in the huge fjord wall between Ørkendal and Lumskebugten. In detail, finely tiger-striped, coarsely crystalline portions, or beautiful dolomite rosettes with red margins occur. In addition to these irregular dolomite structures Bed 14 contains some well delimited, but only 1—2 m thick, dolomite beds.

Bed group 15. 75 m. Combined with Bed group 17, this is characteristic of the Lower Limestone-Dolomite Series in our working field: two white dolomite bands distinguishable from a great distance (cf. Fig. 10), interbedded in the dark, predominantly limy and shaly complexes of Bed groups 14, 16, and 18. The rocks of Bed group 15 consist mainly of a light, nearly white, fine-grained dolomite. They are rich in algal biostromes of the *Collenia* type. The weathered surfaces are very rough, especially where *Collenia* occurs. Easily accessible occurrences are found in Noa Dal, at Kiledal, east of Nanortalikdal, in Antarctica Sund, in Carissima Dal, on Maria Ø, and in the region of Kempes Fjord. The bed group corresponds to the "*Collenia* Member" of H. P. SCHAUB (1950). This author employs a different subdivision of the beds¹).

The algæ of bed group 15.

In these beds algal structures are found which differ distinctly from those described above (figs. 3, 4, and 5). They are not restricted to this group of beds, only, but are also found in some dolomite bands of groups 16 and 18.

The individual specimens have a diameter of 3—6 cm. Between them there are distinct interspaces, which may be up to 3 cm broad. These interspaces are filled with a grey limy matrix and large numbers

¹) In our opinion Kap Oswald (cf. fig. 13) is only little suitable for the measurement of a reliable section. The beds are here, near or even in the core of the antiline, rather much squeezed out. Only upwards from Bed group 18 may the measurements yield reliable values.—SCHAUB's collections and measurements are limited to the surroundings of Kap Oswald on Ella Ø, and for this locality they are extremely valuable. Also the far-reaching conclusions drawn by SCHAUB apply only to Kap Oswald; even on the other side of Kempes Fjord conditions are different. Precisely on account of the far-reaching interpretation, the title of SCHAUB's work, "On the Pre-Cambrian to Cambrian Sedimentation in NE-Greenland" absolutely needs a more precise geographic definition.



Fig. 3. Surface of bed with algal structures, of Bed group 15. Locality: west of the top of Maria Ø. Length of hammer: ca. 50 cm.

of white dolomite fragments (fig. 5), probably algal fragments. The algæ themselves consist of white dolomite, and on a polished surface they exhibit, like the algæ of Bed groups 12 and 13, convex lamellæ in the upper part. Here, also, the dolomitisation has destroyed every delicate structure recognisable in thin sections.

Not rarely is an algal column divided into two individuals. The individual columns may attain heights of 30—40 cm. On the polished surface of the specimen in fig. 4 some light-coloured transverse connections between the algal columns are remarkable. It cannot be clearly ascertained whether they are transverse stripes produced by the algæ

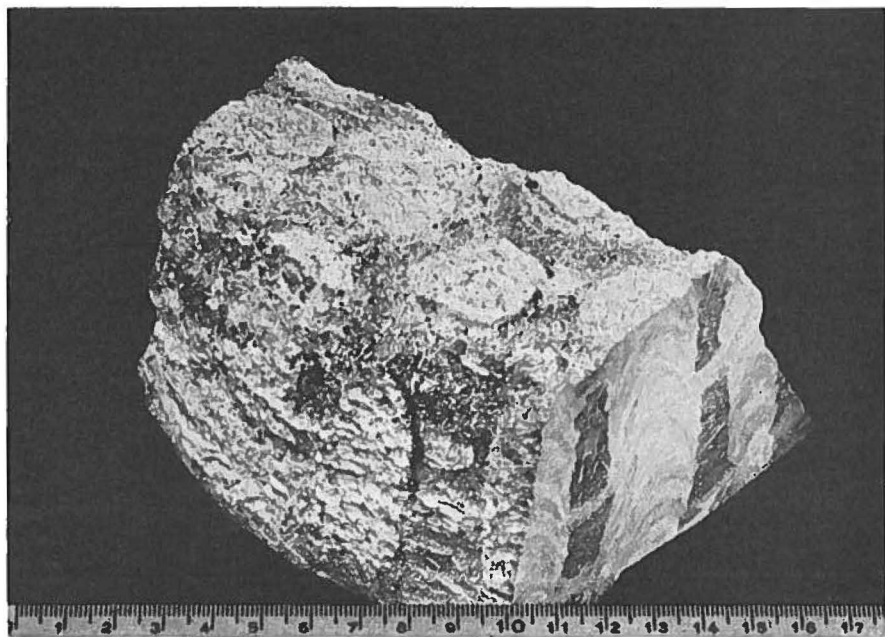


Fig. 4. Dolomitic alga (dolomite) of Bed group 15. The front surface to the right is polished. Scale in cm.

to fortify the whole structure, or they are casual accumulations of the dolomitic clastic material in the interspaces. Fig. 3 shows the surface of a bed of group 15 west of the top of Maria Ø.

The available specimen from group 15 is most closely comparable to *Collenia albertensis* FENTON & FENTON 1935. In the latter, however, the transverse section is oblong, while our algal column is rather circular. Moreover, the narrowing of the column at the base, observed in *Collenia albertensis*, is absent.

Bed group 16. 180 m. Dark grey, often nearly black limestones, highly laminated, especially at the base. They contain organic material, which lends a dark colour to the beds. When the beds, notably the shales, are struck with a hammer, a strong H_2S -smell can be ascertained.

Bed group 17. 75 m. Like group 15, this group consists of light, nearly white, massive dolomites. Unlike the former, however, no algal structures are found here. In the anticlinal crest on Kap Oswald (Ella Ø) Bed group 17 is only rudimentarily present, quite probably owing to strong folding and pressure. Towards the north, however, (Andrées Land) this light-coloured dolomite band seems to disappear entirely, and to be replaced by another equivalent in the Limestone-Dolomite Series.



Fig. 5. Dolomitic alga (dolomite) of Bed group 15. Locality NW-side of Ruths Ø. Scale in cm.

f. Upper Limestone-Dolomite Series.

Bed group 18. 440 m. This group exhibits only few marked and continuous characteristics. It is also well bedded, often laminated, and accordingly forms fewer steep rock walls, but more frequently regions rich in vegetation. Dark grey limestones predominate. At Niviarsiat there occurs, ca. 200 m above the base, a light-coloured dolomite band 15 m thick, and in other places, also, the sequence is interrupted by light-coloured dolomite beds. These light-coloured dolomite beds change their position within the group from locality to locality, and are not suitable as guide horizons. An exception is formed by a yellow-weathering band of light-coloured dolomite, which has a thickness of about 5 m and occurs ca. 100 m below the upper boundary of group 18.

Bed group 19. 140–210 m. This group corresponds to SCHAUB's beds Nos. 18–24. Owing to its multicoloured weathering, the group is valuable as a guide horizon. It consists in the main of a dolomitic shaly complex. The rock-colour is grey to light-grey and red, while the weathering colours are yellow and red, for the upper parts olive-green and purple. The whole group shows intensive colouring and lamination. SCHAUB studied the rocks under the microscope, and ascertained in part

a cyclic sedimentation of the beds, the red laminæ beginning with a sharp contact and consisting of angular quartz grains with abundant pyrite. In the upper beds the quartz grains decrease in size as well as in number, and are increasingly replaced by dolomite in the yellow-weathering beds. The thickness of the red laminæ is 2—5 mm, that of the grey beds 10—15 mm. SCHAUB regards the whole group of beds as pre-glacial varve series, whose approximate age he gives in years (10,000).

For mapping purposes, a reddish-yellow-weathering band, ca. 8 m thick, in this group is very useful.

The thickness of group 19 varies between 140 and 210 m. To the south (Suess Land, etc.) it is thicker, to the north (Kiledal, Noa Dal) thinner. Here in the north, occurs

Bed group 20 (up to 70 m). It is made up of rather massive grey limestones, which turn greyish-brown on weathering. At Kiledal, where it can be observed to be much crumpled in a syncline, its thickness is between 60 and 70 m. This value can neither be verified at Pissevache nor at the north coast of Gunnar Anderssons Land, since in both places the limestones are largely covered by scree. Merely the presence of group 20 can be ascertained here. Thus, the upper beds of group 19 and group 20 are not horizons following upon one another in time, but facies equivalents, for there are nowhere any indications that, to the south, group 19 was followed by an erosion. The introduction of separate designations for the two varieties was, however, justifiable in view of the northern tracts (Andrées Land, etc.), where two facies types occur one above the other.

An occurrence frequently observable in the Limestone-Dolomite Series are the intraformational conglomerates and breccias. These are the outcome of reworking of nearly consolidated rocks, which took place without drainage, and they are not confined to constant horizons.

In the Limestone-Dolomite Series, especially in its upper part, the constancy in the sedimentation from the north to the south is no longer visible. True, there is no actual change in facies; they are still formations of the constantly subsiding geosynclinal trough. However, in this changeable sedimentation the succeeding significant phenomenon heralded its coming: the glaciation.

g. The Tillite Series.

The nature of the conglomerates, which were first described in detail by KULLING (1930), have hitherto been the subject of much discussion (BACKLUND (1930), WEGMANN (1935a), H. P. SCHAUB (1950), HUBER (1950), etc.). They were interpreted as tillites by POULSEN (1930), and

since then this view has had reference to the whole complex of conglomerates, shales, and sandstones occurring between Bed group 20 of the Eleonore Bay Formation and the early Palæozoic. We, also, will retain this view, though some beds might be periglacial or even extraglacial. However, until we have further detailed investigations on this problem at our disposal, we use, with all reserve, the name which at any rate a large part of the complexes is entitled to bear.

The complex of these tillites exhibits within the folded Caledonian sediments very great differences in development and thickness. Thus, in the area of Ella Ø a thickness of 580 m was measured, and in the Duséns Fjord region somewhat more than 400 m (in both areas without the Tillite Canyon Formation). In our area a subdivision into a Lower or Grey and in an Upper or Red Tillite appeared to be expedient. This has reference to the designation of the weathering colour of the matrix of the tillites. The two tillites are separated by a series of sandstones and dark to black shales, which, however, are included in the above-mentioned values of thickness.

The superposition of the tillites.

First, the superposition of the tillites in some localities will be described. On Gunnar Anderssons Land and the northern part of Ymers Ø there is only little opportunity to study this superposition in good exposures. At the north coast of Gunnar Anderssons Land it was only possible to ascertain that the grey limestones of Bed 20 were present in the immediately underlying deposits. Like the greater part of this bed, the actual transition to the tillites is buried below scree and other Quaternary formations.

In the region of Noa Dal the tillites occur in the Arena, which joins the northern bank of Duséns Fjord west of Barrieren. However, here, also, the Lower Tillites are only visible in a few places, and of their substratum only a few square metres can be seen about 500 m west of the lower reaches of Pissevache. Dark-grey to black limestones and shales occur here. The beds underlying these, however, are concealed by scree from the river bank and the overlying beds.

As stated above, the limestones of Bed group 20 are only entirely visible in Kiledal. The boundary towards the tillites is sharp, still it is very doubtful whether intraformational gliding in this extremely sharp synclinal fold did not alter the original conditions.

As a further example of the superposition of the tillites, a place was investigated which was located on the steep eastern side of Nanortalikdal ca. 650 m above sea-level. Here Bed group 20 is evidently replaced by Bed group 19, and the base of the Lower Tillites can accordingly be described: In the mobile shales of Bed group 19, well rounded

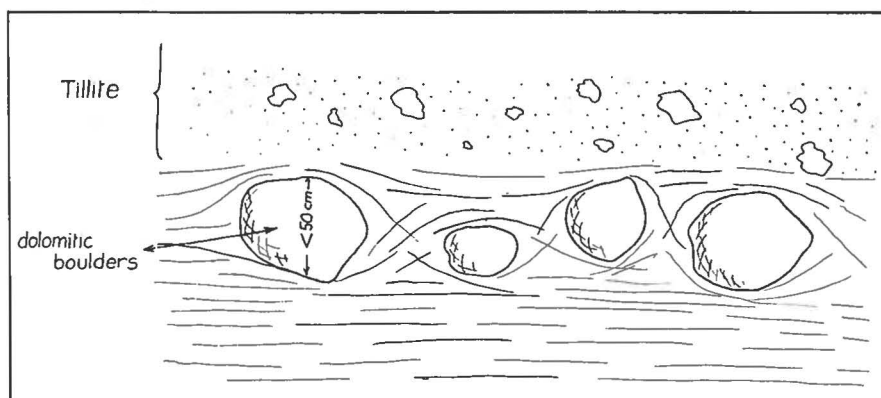


Fig. 6.

yellow dolomite balls of the size of a head are enclosed; the shales pass into the sandy matrix of the tillites, and after about 1–2 m appears the usual picture of the component-bearing tillites, which will be further dealt with below. In the lower 30 m of the tillites there is a certain unmistakable bedding and sorting of the components.

Very instructive is the picture which appears at the slope of Trugbjerg towards the delta of the Langgletscher creek. Here Bed group 20 is again absent. The greyish-green shales are very pure and much compressed, so it is understandable that in many places in the landscape they may give the impression of compact rocks. Enclosed in the shales there occur here, also, yellow, round dolomite balls; but here they attain a diameter of up to 60 cm. The shales “flow” around these balls as if they, on enclosing the balls, had been a tough paste (see fig. 6). Immediately above the balls the increasingly sandy matrix of the Lower Tillites with the pebbly components appears again.

Precisely the same conditions are found near the western mouth of Carissima Dal ca. 100 m above sea-level. But here the dolomite balls grow somewhat smaller again, almost like those in Nanortalikdal.

At the south coast of Suess Land, a little below the top of the mountain with the height figure 1254, the following section through the substratum of the tillites can be observed:

Tillites.

- | | |
|---|---------------|
| Transitional bed with dolomite balls ($\varnothing = 15$ cm) | |
| 30–35 m green clay shales (often black on weathering) | |
| 5 m grey bedded limestone | Bed group 20? |
| 30–40 m green shales, in part | Bed group 19 |
| greenish-brown to purple on weathering | |
| 1 m grey, yellow-weathering dolomite | |

- 5 m yellowish-green-weathering shales
- 1 m limestone bed
- 30 m yellowish-green and purple-weathering shales,
rather massive, often interstratified with silici-
fied beds; interrupted by 3 yellow dolomite
bands, each less than 1 m thick Bed group 19
- 1 m yellow dolomite
- 5 m red } silicified dolomite shales
- 3 m light-grey } ("red-yellow band")
- 50 m grey dolomite shales
- 100 m grey, bedded and shaly limestones

On Ella Ø the substratum of the tillites was investigated in the small creek of Kløftelv, ca. 1 km east of the station. Here, also, Bed group 20 cannot be recognised with certainty. The actual transition from the underlying shales to the sandy tillite matrix is very sudden, and it is not impossible that here, also, near the sharp anticline of Ella Ø, intraformational sliding has taken place. Finely crushed horizons immediately at the transition support such a supposition, so we dare not ascribe any specially typical development to this transition.

The Lower Tillite does not rest on the same substratum everywhere. To the north the substratum is formed by Bed group 20, to the south by Bed group 19 or a facial variation of the two groups. As it was ascertained that Bed groups 20 and 19 are alternating facies types, the difference in the superposition may not be regarded as the effect of erosion prior to the formation of the tillites.

The Tillites.

For a description of the Lower Tillite we may in the main confirm and sum up the detailed observations of KULLING, BACKLUND, SCHAUB, and HUBER: The matrix of the Lower Tillites is a very thick and hard quartz-sandstone with fairly well rounded quartz grains. When fresh, it is dark greyish-green, and greyish-brown on weathering. Most frequently the Lower Tillite is massive, and forms steep rock walls (on either side of Antarctica Sund). However, it may also, especially where the boulders occur in layers, be bedded. In places with increased tectonic influence, a laminated habit can often be ascertained.

The components of the Lower Tillite are, so to speak, derived exclusively from the ametamorphic rocks of the Eleonore Bay Formation. They exhibit very often striation, as is characteristic of glacial deposits. In addition to rounded boulders, angular components occur, also. They vary in size between a few millimetres to 3 m in diameter. Besides the

fairly well sorted portions, such in which big and small boulders are irregularly intermixed are exceedingly frequent.

As suggested above, the two series of tillites are interbedded by a series of sandstones and greyish-black shales. It is easy to trace the development of the sandstones from the Lower Tillite for they simply represent the tillite matrix without boulders. The shales, also, are mostly of a sandy nature.

The Upper Red Tillites exhibit some distinct dissimilarities from the Lower Tillites. Thus we meet here, in addition to the sediments, a multitude of crystalline components. The matrix of the Upper Tillites was investigated by BACKLUND, SCHAUB, and in particular detail by HUBER (1950). In a Fe-containing, predominantly red, but now and then also green cement lie well-rounded, in part even polished quartz grains. In the area around Pissevache the tillite containing crystalline components does not differ from the Lower Tillite as regards the weathering colour. The thickness of the Red Tillites in our area is constantly like that of the Lower Tillite, which at the north coast of Gunnar Anderssons Land, probably also west of Pissevache, measures only ca. 20 m.

The crystalline components of the Upper Tillite are highly diversified. The reader is referred to the detailed petrographic descriptions by BACKLUND and HUBER. Among others, a remarkable red granite is of very frequent occurrence, but so far it has not been found *in situ*. In addition to the crystalline boulders, such of the ametamorphic EBF are also present, all three series are represented. Here, also, the size of the boulders varies greatly, however, a diameter of more than 50 cm is rarely met with in the Upper Tillites.

The habit of the Red Tillite is likewise very massive. To measure its strike will in most cases be problematic. The Upper Tillite, also, often forms relatively marked rock walls, though not to such an extent as the Lower Tillite. On Ella Ø ripple-marks were observed in the Upper Red Tillites, which, however, include sandstone horizons free of components.

The Upper Tillites on Ymers Ø terminate upward by an orange-weathering fresh greyish-green limestone band 3–4 m thick, while on Ella Ø about 10 m green sandstones were ascertained. The limestone band on Ymers Ø is an excellent guide horizon.

Interpretation of the tillites.

In order to investigate the possibility of an interpretation of the tillites, the essential facts will be repeated here:

Lower Tillites: These tillites are known over an area more than 200 km long and 35 km broad. Their substratum is not always the same,

probably owing to facies differences of the latter. The transition itself suggests a superposition on unconsolidated clay, sand- or limestone. The components, derived only from the Eleonore Bay Formation, are very often striated, now rounded, now angular, and of a greatly varying size. In various places there are indications of bedding, and the matrix is often banded. Altogether, however, the conglomerates are massive and hard. The matrix itself is a hard grey sandstone, brown on weathering.

TEICHERT interpreted these tillites as evidence of a glaciation above sea-level, and postulated after the "in relativ grossen Tiefen erfolgten Ablagerung der KDS"¹⁾ a very rapid upheaval with glaciation, after which—in the early Palæozoic—a gradual subsidence may again have taken place. We cannot agree with this view, if only for the reason that it is difficult to imagine such a sudden, almost flash-like, upheaval, and no doubt it would have left considerable facies and tectonic traces. Moreover, during a continued glaciation a subsidence of the area in question might be more likely.

Part of the above-mentioned observations induced HUBER to the following interpretation, and the further facts reported by us seem to support this view: The Lower Tillites are glacial sediments, which, however, were deposited below the level of the sea, either by icebergs or by floating glaciers. HUBER mentions, as a recent example, the conditions at Humboldt Gl. in N. Greenland and at Ross Barrier in the Antarctic. Attention may also be called to the Newfoundland Bank. Independently of HUBER and us, SCHAUB, on the basis of his investigation of the tillites on Ella Ø, arrives in principle at the same interpretation and mentions as an analogue the sediments of the Pleistocene glaciation in the Gulf of Maine.

The components of the Lower Tillites are derived from the sediments of the Eleonore Bay Formation. Accordingly they were somewhere eroded out of those stratigraphic horizons which they now overlie in the tillites. As areas of supply, sediment areas outside our working area may come into question. Their precise location is not yet known. In the first place, the regions of the now metamorphic complex and the Petermann Series come into question. No tillites have been found so far in the latter series, but it should be borne in mind that with the exception of the lowermost beds the material of the Petermann Series is only known from recent glacial boulders. By means of far-reaching sediment-petrographical investigations it might probably be possible to find out the direction of transport of the clastic material.

The shaly complex between the tillites: We agree with SCHAUB in the view that these shales might afford evidence of an interglacial

¹⁾ The occurrence of algæ, however, tells against great depths.

period. The presence of feldspar in the clastic portions of the shales is interpreted by SCHAUB as an indication of the beginning of the denudation of crystalline rocks in the area of supply.

The Upper Tillites: Their distribution corresponds to that of the Lower Tillites. The transition from the immediately underlying sandstones is successive. The components consist of sediments of the Eleonore Bay Formation as well as of crystalline rocks, for the most part of unknown origin. Their diameter rarely exceeds 50 cm; they often exhibit striation. The matrix is a sandstone with well-rounded, in part polished, quartz grains. The quartz grains are enclosed in a Fe-bearing, mostly red cement.

In the interpretation of the Upper Tillites we must be even more reserved than in the case of the Grey Tillites. A far-going similarity in the texture of the two tillites might suggest a glacial contribution to the formation of the Upper Tillites, and the same is true of the striation of crystalline boulders. On the other hand, the polish of the quartz grains indicate aeolic transport. HUBER supposes a desert climate, as it may, indeed, be observed during recent recession of large glaciers in N. Greenland. Definite statements can hardly be made until further areas with red tillite have been investigated, notably to the north.

During a search for the origin of the remarkable red granite boulders in the Upper Tillite, it should be kept in mind that in the neighbourhood of these tillites there may occur rocks which were transformed during the subsequent Caledonian metamorphism. On this assumption the Upper Tillites like a crinicle saved from fire—if the comparison be permitted—would relate to us anecdotes about the original rocks.¹⁾

¹⁾ During our preparation of the present paper, there appeared in the *Zeitschrift der Deutschen Geologischen Gesellschaft* (Bd. 101, 2. Teil 1949, Hannover 1950 pp. 109—120) an article by H. QUIRING, Berlin, on "Ekliptikschiefe, Jahres- und Eiszeiten". According to QUIRING, the moon was forced out of the N. Pacific during the Laurentian revolution at the end of the Archæic. This event might, through diverse reciprocal action, have resulted, finally, in the present inclination of the axis of the earth towards the ecliptic as a prerequisite for the occurrence of the region of the Polar night and the seasons of the year. Computations of the radiation of heat, the radiation of the sun, and the earth's own heat emission in connection with the atmospheric development led QUIRING to the conclusion that a temperature graduation according to latitude set in in early Algonkian time, and that the alternation of the seasons of the year began at the end of the Algonkian.

By means of the distribution of the archæocyathine, for the existence of which a temperature interval of 20—37° is assumed, QUIRING postulates an equatorial annual temperature of 41—42° and a Polar temperature of 16—19° in Cambrian times. An ice age in the early Cambrian would accordingly be entirely excluded. No pre-diluvian ice age has existed at all. "Diesen Tatsachen gegenüber macht es wenig aus wenn von einzelnen Geologen für algonkische und palæozoische Blockströme, Geschiebeschrammen und -schliffe glaziale Entstehung angenommen wird. Solche

Tillite Canyon Formation. 250 m.

With a continuous transition from the Upper Tillites, this formation begins in a sequence of black sandy and clayey shales. The sediment can be characterised as varved. In the upper 70 m of the shales, grey limestone bands yellowish-brown on weathering occur, which gradually become dominant. Owing to the close connection of the Tillite Canyon Formation with the Upper Tillites we referred them to the tillite complex. POULSEN placed them with reservation in the Lower Cambrian. The designation of this bed is somewhat unfortunate, in that a locality was first named after a formation (tillite), and then a new formation was given the name of this locality, although no actual tillites occurred in it.

h. Spiral Creek Formation and the sandstone series.

Spiral Creek Formation. 25 m. The formation consists of sandy shales interbedded with fine clayey beds, and quartzites. Ripple-marks are very numerous, and the same is true, especially in the clayey beds, of beautiful NaCl-pseudomorphoses. Owing to these and to the generally deep-red colour, the Spiral Creek Formation is an excellent guide horizon.

Sandstone series. 50 m. The series is made up of yellow, orange, and violet coarse-grained quartz-sandstones, which exhibit distinct cross-bedding. POULSEN (1930) included this sandstone series in the succeeding Bastion Formation; we prefer, however, to introduce this distinction,

geologischen Erscheinungen können auch durch Trocken- oder Nassgleitungen unter hoher Belastung — pseudoglaziale Blockpackungen des Altkambriums sind über 500 m mächtig — entstehen. Unkritisch hat man Fluss- und Murengeschiebe, ja sogar tektonische Rutschstreifen und Harnische in Verkennung der besonderen Bildungsvorgänge glazial zu erklären versucht. Solange an der algonkischen und palaeozoischen Pflanzen- und Tierwelt keine eiszeitlichen Merkmale erkennbar sind, wird man an gleichzeitigen Vereisungen zu zweifeln oder ihnen nur eine lokale Bedeutung — etwa in damaligen Hochgebirgen — zuzuerkennen haben." (ibid. p. 114).

We think, however, that it is much more uncritical to transfer the ecological conditions of recent reef-building agents into the systematic position of undetermined living beings of the Cambrian time, and on this basis to give almost precisely the climatic conditions prevailing at that time. QUIRING's allusion to the early Cambrian "Blockpackungen" more than 500 m thick might perhaps have reference to the East Greenlandic tracts; a literature reference is unfortunately lacking. At any rate, the term "pseudo glazial", mentioned in this connection, may be an expression originating from QUIRING himself and adapted to his theories. First of all, however, we must reject the attempt, on the basis of theories rich in hypotheses, and without any personal observations in the field, simply to indicate a sediment genetically. The direct investigation in the field is always the primary point and the only decisive factor, and confronted with the "Tatsachen" thus ascertained "macht es wenig aus" if some theories do not agree with them.

for the Bastion Formation is the lowermost bed which can be dated by means of fossils, and this, no doubt, justifies a separation.

We regard the sandstone series as the upper limit of the Eleonore Bay Formation.

The boundary pre-Cambrian/Cambrian.

While POULSEN (1930) with reserve placed this boundary between the tillite series and the Tillite Canyon Formation, TEICHERT (1933) refers the tillites to the Cambrian.

Throughout the whole Caledonian sediment complex, from the deepest beds known so far to way into the Ordovician Narhvalsund Formation there is no sedimentary gap. The lowermost identifiable fossil horizon (Bastion Formation) belongs to the upper Lower-Cambrian. The uppermost portions of the Eleonore Bay Formation ending with the sandstone series must accordingly belong to the Lower Cambrian. We are unable to define the boundary pre-Cambrian/Palæozoic as a distinct horizon. The Eleonore Bay Formation thus contains pre-Cambrian and not accurately datable Lower Cambrian sediments.

2. Lower Palæozoic.

a. Lower Cambrian.

Our dating of the fossiliferous beds of the early Palæozoic is based entirely on the determinations of POULSEN (1930, 1932, 1937). The following two bed-groups are given by POULSEN as Lower Cambrian.

Bastion Formation. 170 m. This formation is a predominantly dark greyish-brown and deep-red complex of shales. The shales contain quartz, mica, and carbonates. The best exposures are found on Ella Ø, especially on the saddle which extends across the island from the NNE to the SSW, as also at the south coast in Narhvalsund.

Ella Ø Formation. 50 m. The formation is built up of brown sandstones interbedded with very fine green clayey beds. Often magnificent structures of subaquatic slidings occur, thus especially at the western end of the northern wall of Bastionen, near the shore of the fjord.

b. Lower Cambrian to Upper Ozarkian.

The two following formations must date back to this period. So far they have yielded no identifiable fossils. The next fossil-bearing horizon (Cass Fjord Formation) belongs to the Upper Ozarkian.

Hyolithus Creek Formation. 280 m. The rocks of this formation are mainly grey, hard, massive, brown limestones penetrated by

many calcite veins and weathering with a pitted surface. Siliceous lenses, also, with a diameter of up to 10 cm are often met with. In the central portions of the formation there appears, in the weathered rock, a distinct bedded structure with beds ca. 5 cm thick.

The name of the Hyolithus Creek Formation (POULSEN 1930) has nothing to do with the fossil content. The type locality, the Hyolithuskløft, is situated in Andrées Land.

Dolomite Point Formation. 300 m. Characteristic of this bed are the well-bedded, yellow-weathering, fresh light-grey dolomites. The dolomite is very compact and often exhibits fine beds of a reddish colour. Intermediate siliceous beds occasionally occur.

c. Upper Ozarkian.

Cass Fjord Formation. 100 m. The formation includes dolomitic marls with flat grey limestone boulders. The limestone boulders are of an irregular shape and size, and are cemented together by a yellow dolomitic matrix. Owing to its marly habit, the formation is very liable to disintegration, with the exception of the more or less vertical walls that occur where the formation is sheltered by the overlying Kap Weber Formation.

The age of the formation was determined by POULSEN on the basis of the fairly frequent fossils occurring in it.

d. Upper Canadian.

Kap Weber Formation. 600 m. This extremely thick complex consists in the main of grey limestones, which turn greyish-brown on weathering. At its base there occur some zones, 5—10 m thick, which resemble the marly Cass Fjord Formation. About 100 m above the base there occur very numerous lenses of chert, which underline the bedding by projecting ledges. In the silicified zones the numerous fossil shells are entirely silicified. The silification is especially beautifully visible on the slope rising from Kap Elisabeth (Ella Ø) towards Fimbulfjeld, but also on the so-called Barriere (Gunnar Anderssons Land) ca. 500 m above sea-level. In the uppermost 100 m of the Kap Weber Formation primary breccias are met with, which consist of dark grey sharp-edged limestone fragments of the size of a nut, cemented together by a yellow dolomitic matrix similar to the cement of the Cass Fjord Formation. A brecciated bed, possibly tectonised, about 5 m thick, was found on the ridge west of Midnatspas, which is surrounded by Devonian sandstones. The intervals between the angular limestone fragments are here occupied by siliceous material.

e. Lower Ordovician.

Narhvalsund Formation.

The formation is made up of limestones or coarse-crystalline dolomite, of a light grey or reddish colour. The dolomitic beds are limited to the lowermost 50—70 m and are overlain by more limy beds. Often the rocks show an intensive red colour, which, however, can in the main be ascribed to the overlying Devonian sediments.

The not very frequent fossils were dated by POULSEN as Lower Ordovician.

The maximum thickness of the formation in our area is about 150 m. The complete profile is exposed at the south coast of Ella Ø in Narhvalsund, where the eastern limb of the boundary anticline dips down below the Devonian conglomerate. Otherwise this youngest portion of the East Greenlandic Caledonian sediments has been removed more or less completely by the pre-Devonian erosion.

C. Conditions of Deposition of the Caledonian Sediment Sequence.

The total picture presented by the immense pre-Devonian sedimentary series was already outlined by WEGMANN (1935a, pp. 16 ff.) In broad features, we agree with WEGMANN's statements.

The main characteristics of the conditions of sedimentation are briefly as follows:

First of all, the sedimentation was entirely continuous and took place in a constantly subsiding basin, which was subject to no great desiccation or remarkable tectonic disturbances. The main part of the sediment material is clastic. Large volumes of chemical sediments only appear with the Limestone-Dolomite series and again in the early Palæozoic. The clastic as well as the precipitated sediments ("algal" fauna) bear witness of a deposition exclusively in shallow water. Nor does the tillite series exhibit actual continental sediments. These glacial sediments were deposited under aquatic, very probably marine, conditions.

Throughout its whole extent, our working area appears as a segment which must have been situated more or less in the centre of the original sedimentation through. With the exception of the Tillite Series, no coarse-clastic sediments occur. Nor were differentiations ascertained anywhere which suggested the proximity of the edge of a basin. As to the direction of supply of the material, our area of investigation gives no suggestion. With all reserve, the following observation must be considered:

In most of the exposures to the west the Lower Tillite shows rather larger components than to the east. Thus, for instance, in the wide surroundings of Nanortalik, extremely large boulders (diameter up to 3 m) were found, while in the region of the Barriere (Gunnar Andersson Land) components with diameters of more than 15 cm were rarely met with. The nature of these marine glacial sediments, however, do not warrant any conclusion on the basis of these facts as to the presence of a clearly westward situated area of supply.

Brecciated horizons are very common in the Limestone-Dolomite Series and in the early Palæozoic sediments. However, they are not constant in the horizontal direction, and rarely attain thicknesses of over 50 cm. Their material consists of components of the overlying and the underlying beds. Accordingly they are not broken-up breccias formed of broken-up beds, but intraformational breccias, which were probably formed by subaquatic gliding of not yet consolidated sediments.

H. P. SCHAUB (1950), also, draws entirely justifiable conclusions from his investigations of the sediments. Since, however, the area at Kap Oswald investigated by him geographically as well as stratigraphically represents only a fragmental area, SCHAUB's upheaval and subsidence phases, in view of the whole enormous sediment volume, become mere episodes.

TECTONICS

As stated in the geological introduction (p. 11 ff.), the Vela massif (zone B in text-fig. 1) forms the "backbone" in the structure of the fjord district. To the west it is flanked by the Petermann Series (zone A), to the east by the Caledonian-folded zone C and the Devonian (zone D).

Zone C (text-fig. 1) was subdivided by WEGMANN (1935a) as follows on the basis of the generally north-south running folds:

- I. Eastern anticline.
- II. Central syncline.
- III. Western anticline.
- IV. Western syncline.

In his paper on the western boundary of the Devonian, BÜTLER (1948), also, used this division. As the western boundary of the Devonian over wide distances runs along the eastern anticline, BÜTLER called the latter the border anticline. In the description of the tectonic structure, we, also, use WEGMANN's classification, following the individual folding zones on their course from the north to the south. (Cf. the geological map Pl. 2 and the structural map fig. 19, p. 89, as also the sections in Pl. 3).

I. The Eastern or the Border Anticline.

Of the various folding systems in our area, this anticline has the largest areal extension. On Gunnar Anderssons Land and in the northern part of Ymers Ø it forms a very broad arch. Towards Antartics Sund it decreases apparently in width, showing subsequently, on Suess Land and Ella Ø, a relatively pointed crest. Further characteristics are the constant dipping of the anticline towards the south, diminished somewhat by various W—E-faults, and the divergence in strike between the two limbs of the anticline; the latter is especially observable on Ymers Ø and Ella Ø. In the following pages the development of the border anticline in the sections produced by the fjords from north to south will be described.

1. Gunnar Anderssons Land and Noa Dal.

In this part of the area only the crest and the eastern slope of the border anticline are found, the western limb is only visible farther south on Ymers Ø. In addition to this eastward-increasing E.slope of the series, the structure of Gunnar Anderssons Land and Noa Dal is characterised by two systems of faults. A large number of E—W-faults intersect the region into antithetically southward-dipping blocks. As a consequence the rise of the anticlinal crest is diminished. The ice cap of Gunnar Anderssons Land and the transgressing Devonian beds east thereof conceal the further course of a large part of the faults. Some of them, however, can be traced as far as Barrieren at the northern shore of Duséns Fjord. The other fault system has almost the direction N—S, and includes the large main fault of BÜTLER (1948).

We will now first describe the E—W-faults and the blocks bounded by them, then the N—S-faults, and in a third section the relation between these two fault systems will be dealt with.

a. The E—W-faults and the corresponding blocks.

The north coast of Gunnar Anderssons Land exhibits the following picture:

From Kap Petersen to about 10 km east of this cape the beds descend slowly towards the east. To the north, bordering on the fault at Kap Petersen—this disturbance will be described below—Bed group 3 forms the coast of the fjord. About 10 km farther eastward the base of Bed group 7 reaches the level of the sea. Thus, there is here a descent of the E-limb of the border anticline of about 6° towards the east (see sections 1 and fig. 19).

In a marked creek ca. 10 km east of Kap Petersen appears the large fault termed the main fault (HV) by BÜTLER (1948). Here the beds of the Upper Multicoloured Series (the base of Bed group 11) are, to the west, in contact with the upper beds of the Cambro-Silurian complex to the east (base of the Kap Weber Formation). Thus the throw of the fault HV here amounts to 3,500 m¹⁾.

In the next 2—3 km of the fjord wall, which joins the main fault to the east, the following picture presents itself:

In the upper part of the fjord wall, especially in the massive limestones of the Kap Weber Formation, a number of minor antithetic faults can be observed, by which the western block was each time

¹⁾ BÜTLER estimates the throw of the fault at 2,500 m. He assumes that the Kap Weber Formation is in contact with the Lower Limestone-Dolomite Series, while we found that it borders on the Middle Multicoloured Series.

lowered somewhat. BÜTLER (1948, p. 142) has already described these faults. Below these beds there appear at sea-level, east of the main fault, after about 500 m of talus slopes, a small flat arch in the rocks of the upper Bastion Formation. At the surface of this formation it has a width of ca. 800 m, and corresponds in its effect to the small faults of the Kap Weber Formation. East of the arch the Cambro-Silurian beds descend slowly towards the east.

The eastern boundary of this Cambro-Silurian complex was likewise described by BÜTLER, it is formed by of a SE fault running obliquely to the coast. Here the eastern side is uplifted in relation to the western side, more precisely, the base of the Kap Weber Formation borders on the upper boundary of the Red Tillite. The vertical displacement accordingly amounts to about 1250 m. East of this SE-running fault the uppermost elements of the Limestone-Dolomite Series occur *in situ* at sea-level, with a general eastward inclination and secondary folding. Thus the tillites and the Cambrian beds gradually reach the shore of the fjord. The limestones of the Kap Weber Formation terminate ca. 7 km east of the main fault at the shore of the fjord. In view of the thickness, however, this observation suggests a fault which likewise, ca. $6\frac{1}{2}$ km east of the main fault, trends southward and shifts the eastern portion to a deeper level.

Thus, in agreement with BÜTLER, we term the Cambro-Silurian complex between the main fault HV and the fault striking SE as a graben (rift valley). In order to be able to answer BÜTLER's question as to the extension and age of this trough, the plateau south of Hammeren and the Midnatspas was investigated. Unfortunately the result did not fulfill the expectations, this region being glaciated and everywhere much covered with Devonian scree. The whole side of the valley east of the ice cap towards Midnatspas consists, so to speak, of grey and reddish Devonian sandstones and their débris. Only a conical hill and a crest rise above the Devonian rocks; both are made up of the Kap Weber Formation, which here, also, dips towards the east. The presence of Devonian sandstones so far to the west is remarkable. It might be explained as a local extension of the Devonian transgression or by a displacement effected by the numerous W—E-faults described below.

Thus, at the north coast of Gunnar Anderssons Land a gradual dip of the beds towards the east can be observed, increasing from the west towards the east. It is interrupted by the large main fault HV, a small arch immediately east thereof, and a NW—SE-striking fault. The latter together with the HV forms the margins of a graben.

A little north of Kap Petersen occurs the northernmost one of the numerous W—E-faults. We call it fault 1. The fault is well visible, especially from the northwest (see sections 21 and 22). The fault plane

dips ca. 75° towards the north. The northern limb occurs about 170 m deeper down than the southern one. In the broad crossing of the Rosinante Pas this fault is distinctly visible in the western as well as in the eastern side of the valley. Here, also, the amount of displacement is the same. Farther eastward the course of the fault is concealed by the ice cap and the Devonian débris of Hammeren

At the east side of the Rosinante Pas, another E—W-fault, fault 1a, only ascertainable along a distance of ca. 1 km, can be observed some distance south of fault 1; it has likewise lowered the northern part in relation to the southern part, at any rate about 100 m. In the western side of the valley it is concealed by scree.

The next W—E-fault, fault 2, runs at the southern foot of Chokoladebjerg. Combined with fault 1 it bounds a block which, seen as a whole, undoubtedly belongs to the eastern slope of the border anticline. However, in the region of Chokoladebjerg the amount of the eastern gradient is less than that of the southern gradient. A measurement on the top of Chokoladebjerg thus gives a ENE-strike.

Fault 2, as already mentioned, runs along the southern foot of Chokoladebjerg. Its topographic intersection is very instructive, for from its course at the slope towards Kejser Franz Josefs Fjord a northward inclination of the fault plane of ca. 60° can, no doubt, be inferred (cf. BOYD 1935 fig. 112). The throw of the fault is fairly considerable, amounting to about 1000 m. Here, also, the character of the fault is the same: the northern side lies deeper. To the east the fault continues below the delta of the drainage stream arising from the glacier that reaches Rosinante Pas, then it ascends the slope, and disappears immediately below the southern edge of the ice cap. Unfortunately this fault, also, is not exposed at Midnatpas on the opposite side of the ice cap.¹⁾

The W—E-faults 3, 4, and 5 exhibit the same mode of movement: everywhere (cf. Pl. 3, profiles 21, 22) the northern block lies at any rate deeper than the southern one. They are concentrated, with fault 2, in Noa Dal, evidently they are altogether responsible for the morphological origin of this valley.

¹⁾ It should be stated here that in the course of our mapping there proved to be some errors in the topographic maps. Where, during the mapping and construction of profiles, they gave rise to discrepancies, we corrected them as well as possible. For this means we had at our disposal a geologist compass and a field glass with per mille graduation. The corrections thus make no demand on accuracy, they are rather hand sketches, and they could not be inserted in the topography of the geologic map on Pl. 2. In the area between faults 1 and 2, Rosinante Pas is therefore indicated ca. $1\frac{1}{2}$ km too far westward. Further, the glacier descending towards Rosinante Pas as also the drainage stream and the lake north of Rosinante are lacking.



Fig. 7 a.

Fig. 7 a, b. Western end of Duséns Fjord, view towards the north (Gunnar Anderssons Land). Legend: IIV = main fault, 2, 2a, 2b, 3, 3a, 4 = E—W-faults. LQS = Lower Multicoloured Series, UQS = Upper Quartzite Series, LMS = Lower Multicoloured Series, UMS = Upper Multicoloured Series, LLD = Lower Limestone-Dolomite Series, ULD = Upper Limestone-Dolomite Series, T = Tillite Series, SD = Spiral Creek to Dolomite Point Formation.—BRENNISEN phot.

Fault 3 extends from Kejser Franz Josephs Fjord to the two small lakes 1 km northwest of the lake Noa Sø, through this, and along the S. foot of Rosinante towards the east. Its further course runs through the small marked valley ca. 3 km north of the head of Duséns Fjord (fig. 7 a, b), whence it meets the main fault in a nearly NE—SE directed ravine. It possibly continues in the talus slopes derived from the overlying beds, north of the narrowing of Duséns Fjord, for which reason we combine it with the fault 500 m north of the waterfall Pissevache. The throw of the fault varies fairly often, according to the relative position of the beds at the N- and the S-side of the fault. To the west the throw of the fault is 700 m, at Rosinante ca. 500 m, north of Lille Noa Sø ca. 1000 m, the same throw is found at the 28° meridian, and finally, north of Point 261 it is about 350 m. The relationship of the E—W-faults to the main fault HV and to the fault at Barrieren will be further dealt with below.

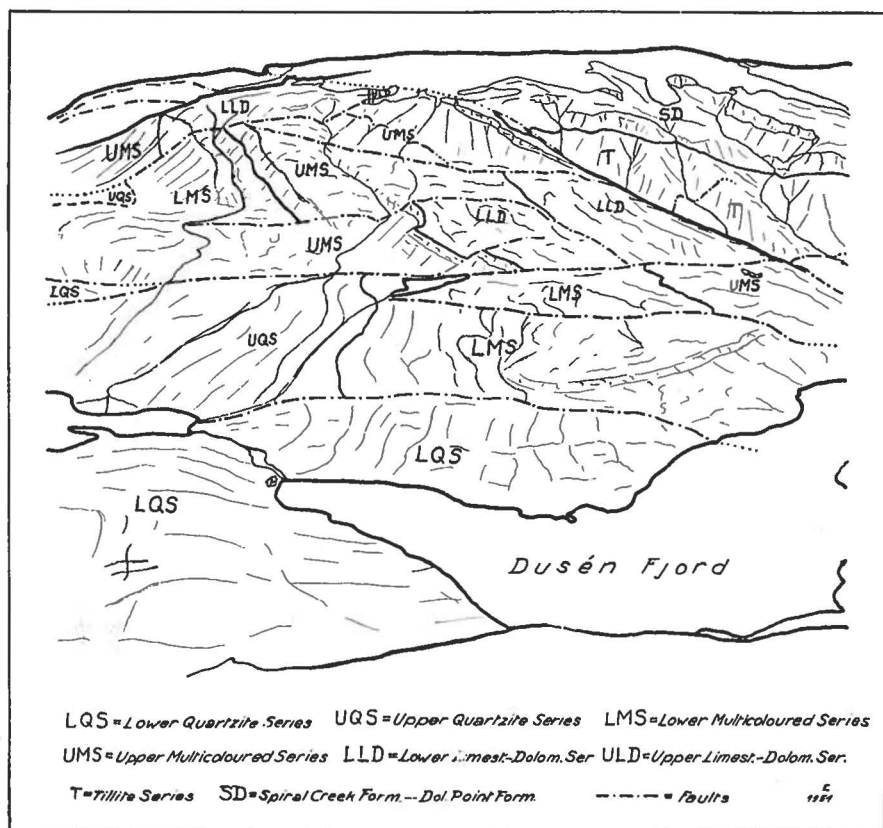


Fig. 7 b.

Fault 4 begins in the west not far from the trapper's hut in Blomsterbugten, extending across Noa Pas to Noa Sø. Its eastern continuation is assumed by us to pass through the river leading towards Lille Noa Sø, after which the fault reaches Duséns Fjord across an evenly sloping ridge. In this latter part it is exceedingly well recognisable in the air photographs (fig. 7 a, b). Beyond the fjord it extends close past Point 261, to disappear entirely in Duséns Fjord. In the area around the 25° meridian its throw is greatest, about 1000 m, but it decreases both towards the east and the west. The intersection of fault 4 with the land area near Blomsterbugten suggests a southward dip of the fault plane of about 70°.

Fault 5 begins almost in the place where the coast of Blomsterbugten bends towards the southwest. Over the whole stretch along the southern bank of Noa Sø, below the delta of the afflux to this lake, as far as the region of Uglesø its path can only be conjectured. Its presence south of the head of Duséns Fjord is evident, though its throw cannot be given owing to the similarity of Bed groups 1 and 3. It is a fact, however, that it dies out about 2 km south of Point 261.

Between these major faults there occur various minor faults, which now run parallel to the major faults, now branch off from the latter. The region on either side of the 25° meridian is especially rich in such minor faults. Here faults 2a and 2b will be briefly described. The former branches off from fault 2, extending eastward to the main fault HV, which it crosses, and is then traceable for a further distance of about 500 m. Fault 2b branches off from fault 3, extending towards the northwest, subsequently bending towards the west. Thus, in a schematic N—S-section of the region around the 25° meridian we get the following picture (fig. 8):

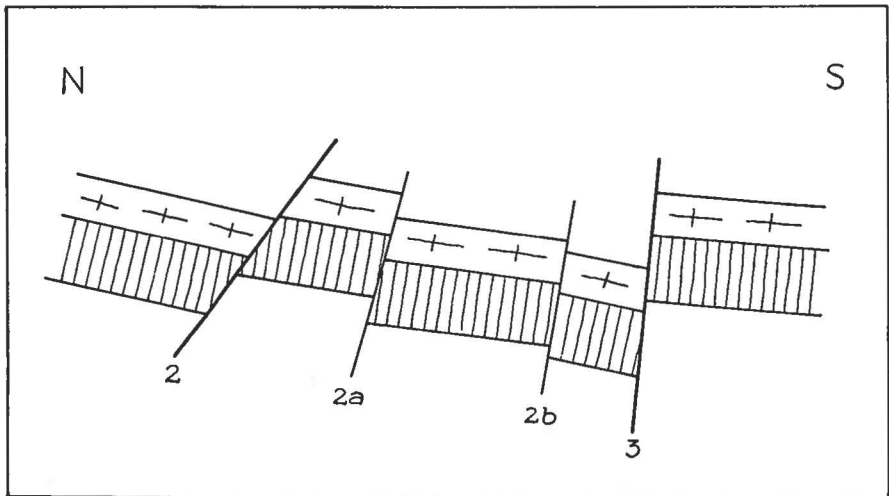


Fig. 8.

How these minor faults behave in the depth, remains an open question. While along the major faults the northern blocks were lowered, along the minor ones the southern blocks were lowered. This effect of the minor faults can be ascertained so to speak wherever such faults occur.

It should further be pointed out that the observer in the region north of fault 3 in the area around the 25° meridian will easily fall a victim to a delusion (cf. fig. 7a, b). Above the remarkable yellow dolomite bands of Bed group 13 there follow to the east, first a dark limestone series, then a light dolomite bed, again a dark limestone, and again a white dolomite. Thus it is largely the same picture as is presented by the normal sequence of the lower Limestone Dolomite Series, Bed groups 14—17. On closer inspection, however, it is readily recognised that we are here confronted with a repetition of Bed groups 14 and 15, precisely owing to the branched-off fault 2b (fig. 7a, b).

That further E—W-faults extend at the foot of Arenaen below the scree, is evidenced by some minor faults in the W. wall of Barrieren.

The direction of movement of the E—W-faults will be discussed in the section after the description of the N—S-faults¹).

b. The N—S-running faults.

As stated on p. 39, the throw of the main fault HV at the north coast of Gunnar Anderssons Land is about 3,500 m. At the fault itself a small slickensided wedge is squeezed in. The fault plane is almost perpendicular. South of the ice cap the HV follows the ravine extending NW—SE, whose stream empties into Duséns Fjord at its narrowest place. The amount of displacement south of the ice cap is about the same, ca. 1,700 m. South of Duséns Fjord the HV is again recognisable at the mountain marked with the height figure 890, Rumpen. Here the throw of the HV, as will be shown below, is again 3,400 m. The apparently highly increased amount of displacement in relation to the northern bank of Duséns Fjord will be further dealt with on p. 47.

Rumpen was already described and figured by BÜTLER (1948, pp. 106, 118, 142, Pl. 8). Here he gives the throw as only a few hundred metres, because he was not quite aware of the detailed stratigraphy of the Caledonian sediments.

Rumpen is chiefly built up of the Caledonian folded sediments of the Palæozoic. At the NW foot of the mountain the Upper Tillites crop out, while the whole series as far as the Kap Weber Formation follows to the top. These rocks dip 40—50° towards the ENE. At the top itself, as already described by BÜTLER, occur the Devonian basal conglomerates, which towards the eastern slope of Rumpen attain a considerable thickness. The thickness of the early Palæozoic sediments decreases somewhat in the proximity of the HV, owing to slickensides. At the HV itself, which extends close along the W foot of Rumpen, conditions are as follows: The beds of the Upper and Lower Multicoloured Series immediately east of Point 1712 show intermittently a steeper eastward dip (cf. profile 4). However, they flatten out again, so on a small spur of Point 1712 towards Rumpen a small erosion remnant still occurs, which consists of the dark limestones of Bed group 11. With the general eastward dip of about 5° the sediments of the Multicoloured Series then reach the main fault HV. Here Bed group 11 to the west borders on the Cass Fjord Formation to the east. Thus the throw of the main fault HV is here 3,400 m. Its further course south of Rumpen corresponds to that of the southernmost

¹) The ENE-direction of the E—W-faults between the HV and Barrieren, as indicated in the map (Pl. 2), does not quite agree with the actual conditions. The erroneous topographical maps which formed the basis of our mapping forced us to give the faults a direction bending somewhat towards the ENE. Actually they run approximately W—E. For it is a fact that the height contours in Arenaen are concave towards the S, not convex, as represented in the original map.



Fig. 9a, b. Duséens Fjord, Barrieren; in the background to the right the Midnatspas, at the left margin of the picture Arenaen is seen. To the right, in the farthest background, Nordfjord and Waltershausen Gl. can be discerned. Legend: ULD = Upper Limestone-Dolomite Series; T = Tillite Series; SD = Spiral Creek Formation; D = Devonian. BRENNISEN, phot.

one of the two small creeks which drain the ice caps on the elevations southeast of Rumpfen. Here the HV separates the beds of the Upper Multicoloured Series and the Lower Limestone-Dolomite Series from the Devonian sandstones.

Another N—S-fault—we call it the “Barriere fault”—extends along the western margin of Barrieren (cf. sections 2, 3, 19). They are only directly visible in the wall west of Pissevache, immediately north of Fault 3 (cf. fig. 9a, b). The Uppermost Dolomite Point Formation, the Cass Fjord Formation, and the lowermost beds of the Kap Weber Formation, all somewhat diminished owing to slickensides, here, owing to an apparently flat fault plane dipping towards the east, rest on the Spiral Creek Formation. The fault line in this rock wall, however, seen from the fjord runs from the left above to the right below. Thus it might be a fault plane which has an almost vertical position and strikes north-westward in the wall. Unfortunately the area between this place and the glacier giving rise to the Pissevache is so highly covered with

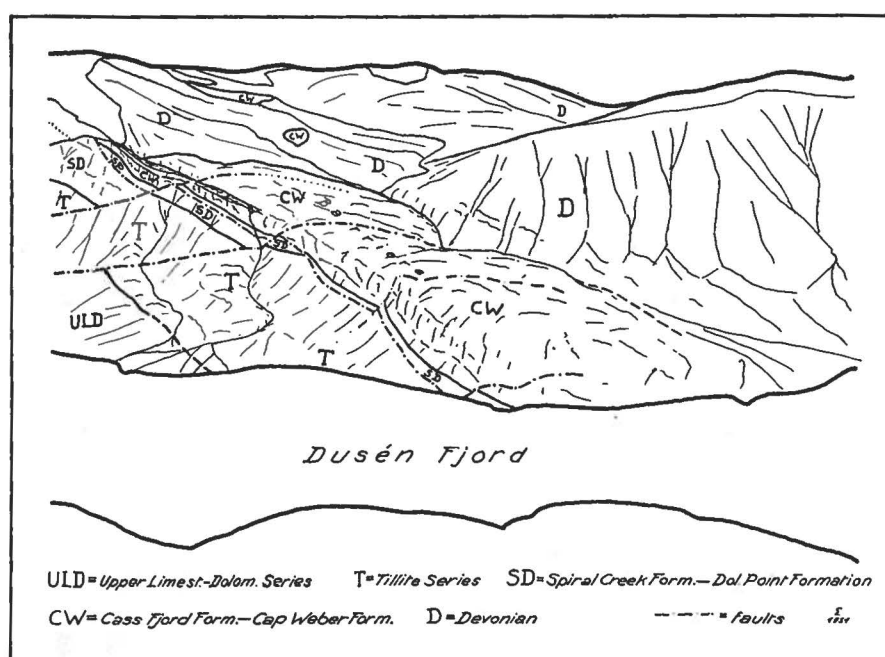


Fig. 9 b.

débris that no clear decision can be made. Along the western margin of Barrieren, whose steep slope depends morphologically on this fault, there are traces of a fault buried below scree from the overlying beds. There is no doubt, however, of its presence: The rock wall of Barrieren is built up of the upper portions of the Dolomite Point Formation, the Cass Fjord Formation, and the lower part of the Kap Weber Formation. On the other hand, the whole river-bed of Pissevache is located in the Upper Tillites, of which some exposures are also found at the foot of the Barriere-wall, near the fjord. The throw of the Barriere-fault must accordingly amount to 900—1000 m.

Hence we regard it as a steep fault, the eastern side of which is lowered about 1000 m. It follows the steep scarp of Barrieren, that is, like the latter, it has been forced more and more northwestward by the various E—W-faults. Possibly it represents a branch of the main fault, for west of the region of Barrieren, this latter has a throw of only about 1700 m, as compared with the 3,500 m at the north coast of Gunnar Anderssons Land and at Rumpfen. The Barriere-fault might thus largely compensate the decreased throw of the main fault HV.

Against BÜTLER's assumption (1948, p. 142), no decrease of the throw of the HV can thus be ascertained along the stretch between the north coast of Gunnar Anderssons Land and the south coast of Dusén's Fjord.

c. Relation between the E—W and the N—S faults.

As mentioned above, faults 2a, 3, and 3a cross the main fault HV without being displaced. At the HV itself, also, which here passes through the ravine described above, there is at first nothing noteworthy, but we will deal with this below. The Barriere-fault is influenced by the above-mentioned three E—W-faults (cf. fig. 9a, b). The west wall of Barrieren, the morphological appearance of which, as demonstrated above, is due to the Barriere fault, obtains its direction, turning successively from NNW to NW, through the various E—W-faults. The latter are accordingly younger than the Barriere-fault. If we consider, now, the course of the HV, we shall find that this, also, in the area of the numerous E—W-faults has been forced towards the northwest. This is true of Duséns Fjord itself and the NW—SE striking ravine. Unfortunately the fault-like displacements at the points of intersection between the main fault HV and the E—W-faults cannot be recognised in the air photographs. Their morphological prominence was evidently blurred by the erosion of ravines. From all this it follows that the Barriere-fault and the HV, which latter (cf. BÜTLER 1948) at any rate was reactivated in post-Devonian time, were displaced by the younger E—W-faults. A further, though modest, contribution to the relative determination of the age of the fault systems in Duséns Fjord is supplied by the lead-bearing quartz veins in the southern wall of Noa Dal, south of the head of the fjord. They will be described in more detail below (p. 49), but we will mention what is of interest here. These quartz veins are older than the E—W-faults. If once it will be possible to combine the quartz veins with a datable magmatic activity, the long interval of time which we must allow for the formation of the fault systems in Noa Dal, might be somewhat shortened. However, it is only possible to give a relative genetic sequence, for the post-Devonian sediments required for an upward limitation of these post-Devonian faults are lacking in this region.

It was described above that the individual E—W-faults sometimes forced the N—S-faults somewhat westward. These small W-components of the E—W-faults, however, were materially superceded by the vertical components, throws up to 1000 m having been ascertained along these faults (p. 43).

d. The south side of Noa Dal.

The whole south side of Noa Dal from Blomsterbugten to the fault HV at the western foot of Nuloq (890) shows a very simple picture: A constant slow dip of the beds (including Bed groups 1—11) towards the east, somewhat accentuated to the east. The mean eastward dip along this stretch is 5°. In our area of observation the core of the border

anticline begins to reflect itself south of Blomsterbugten and fault 5. It is also here that the deepest portions of Bed group 1 are exposed within our area of investigation.

e. Quartz veins.

In the mountain wall south of Uglesøen a few minor N—S-faults occur in the basal bands of Bed group 4, and along these faults the east side is always uplifted. The largest of the faults has a throw of ca. 100 m. These displacements are disharmoniously smoothed out by the underlying quartzitic shales of Bed group 3. In such a place, where the quartz shales even primarily exhibit an intensive folding, a system of quartz veins also occurs. The locality is located ca. 1 km south of the head of Duséns Fjord, immediately at the foot of the steep rock wall. Evidently the quartz veins have still further mobilised the surrounding shales. The ptygmatic figures formed by the quartz veins are also formed by the shales. Associated with these quartz veins is galena, which in places may be up to 20 cm thick. Chalcopyrite is also abundantly present and is often recognisable from a great distance owing to the malachite-green tinge. Such green spots are likewise found at the passage from the trapper's hut in Blomsterbugten to Noa Sø and along the shore of the fjord some 100 m south of the hut. As to the above-mentioned quartz veins, it is of importance that they do not cross the nearby fault 5, and accordingly are older than the E—W-faults.

It may be advisable, here, to throw a glance at the literature on the area of the Noa Dal depression; for the terrain with its rich plant and animal life and easily accessible surface invites investigation.

In his paper of 1939, ODELL uses a brief chapter for the description of this region. He describes some vertical faults as well as some differential slidings within the beds. His investigations, however, seem to be limited to the surroundings of the trapper's hut in Blomsterbugten. ODELL places the main stress on disproving the view held by KOCH, WORDIE, and other geologists, that the main mass of Teufelsschloss (west of Blomsterbugten) is an overthrust rockmass, a disproof which turned out to be right, but this is beyond the scope of our work.

As to the paper by CLEAVES & FOX (1935) the reader is referred to the foot-note in the general part of the stratigraphy (p. 15).

The important and clear descriptions by BÜTLER were constantly taken into consideration in the above description of the details.

2. Ymers Ø.

The development of the border anticline in this part of the region shows a striking characteristic: A divergence of up to 90° in the strike

of the two sides of the anticline (cf. the structural map fig. 19): The west side, somewhat recumbent at Kejser Franz Josephs Fjord and rising steeply from Antarctica Sund, has a direction of 340° . In the gently dipping east side, directions of strike of 60° , often even 90° , can be measured on Munotbjerg. The general strike amounts to about 65° . In Noa Dal the core of the anticline, represented by the Lower Multicoloured Series, at sea-level has a width of ca. 16 km. At the southern mouth of Margeries Dal the corresponding distance amounts to further 5 km. In agreement herewith, no westward or eastward dips can be observed at all along wide distances south of Noa Sø (Section 4). Measurements of the beds here show the amount of the axial dip. Some E—W-faults, among others a very large one, intersect the east side. They differ, however, in character from the E—W-faults in Noa Dal, as will be shown in the individual descriptions.

We subdivide the description into a section on the eastern side with the faults, a second section on the west side, and a third section on the superposed Devonian beds.

a. The east side of the border anticline (Sections 4, 6—10, 22).

aa) Margeries Dal and the region west of it:

The region south of Rumpen as far as the upper one of the two tributary valleys draining into Duséns Fjord, shows no complications. The eastern wall of the mountain with the height figure 1712 in the area with the greatest topographical dip the beds show a somewhat increased eastward inclination of 30° , while at the bottom of the northern part of Margeries Dal and east thereof an inclination of 12° , respectively 10° , was measured.

In the first western tributary valley north of the highest part of the pass of Margeries Dal, a fault occurs in the valley bottom, it crosses Margeries Dal, and is readily visible in the steep eastern side of this valley. Here the base of Bed group 14 to the north meets the base of Bed group 11 to the south, that is to say, the throw is about 600 m. The continuation of this fault to the west especially where the tributary valley bifurcates, could unfortunately not be traced any further. The region south of the fault to about 2 km south of the watershed is entirely dominated by the eastern slope of the border anticline. An occurrence of Devonian conglomerates at the west side of Margeries Dal is surprising; it is located south of the debouchure of the just described tributary valley of Margeries Dal, ca. 2 km north of the highest point of the pass (P. 470). Numerous down-fallen conglomeratic blocks near the main rivulet reveal the presence of this Devonian relict. A remarkable feature about this Devonian relict is its deep position: it occurs

roughly at a 400 m deeper level than the Devonian masses on the plateau east of Margeries Dal.

Along the whole western side of Margeries Dal a light flexure-like accentuation of the east side of the anticline is traceable ca. 1000—1200 m above sea-level (sections 4, 6, 8). This feature evidently causes the existence of the window-like denudation in the first tributary stream to the west south of the watershed. Below the white dolomites of Bed group 12 and the dark bituminous limestones of Bed group 11, the uppermost 10 m of the multicoloured Bed group 10 crop out in the river-bed. Immediately south of this tributary stream the general eastward inclination passes slowly into an ESE to SE-ward inclination. This is also markedly visible in the tectonic map (fig. 19). In the landscape it is indicated by a slow rise of the boundary between Bed groups 10 and 11 along the western slope of Margeries Dal towards the southwest.

After crossing the lowermost western branch valley, this junction of the beds meets a fault, which extends in an ENE-direction across Margeries Dal. This fault is traceable farther westward across the core of the anticline, being especially distinct at the west side of the deep-cut valley which debouches into Antarctica Sund ca. 1 km west of Margeries Dal. Here the base of Bed group 4 to the south meets Bed group 6 north of the fault; thus there is here a difference in level of 530 m. The steeper the position of the beds in the west side of the border anticline, the less is the amount of displacement of this fault, for at the plateau Bed group 6 is in contact with Bed group 9, indicating a dislocation of 400 m. It is a fact, however, that in the valley east of P. 961 this fault is no longer to be seen; hence it must die out somewhere between the two valleys. Observations from the aircraft showed that it dies out in Bed group 12, whose base still shows a very slight displacement, while the upper boundary is undisturbed.—Theoretically the disappearance of an amount of dislocation at a fault may also be explained by the beds having only moved precisely in the direction of dip. However, this possibility must be excluded, as will be shown below.—The further course of the fault towards the east will therefore be described first (Section 22).

At the east side of Margeries Dal the fault has dislocated the base of Bed group 6 to the south so that it is in contact with the lowermost Bed group 14 to the north, that is, to an amount of over 1300 m. This high figure is understandable if it is borne in mind that the eastern limb of the anticline in the area of the east side of Margeries Dal dips eastward to a different degree on the two sides of the fault, the eastward dip being steeper north of than south of the fault. It may be anticipated that this condition changes in the region of Fladedal. Here it is the beds south of the fault that are steepest. A smaller dislocation, amounting

to 800—900 m was ascertained here; it could not be more accurately determined. These regions will be further dealt with below.

With a view to other faults of a similar nature to be described below an interpretation of this fault will be attempted here. The varying amount of displacement, the different dip on the two sides of the fault—a change of the direction of strike north and south of the fault is also readily recognisable in the area of the Munotbjerg plateau (fig. 19) and the sudden dying out to the west, are the most striking characteristics of this fault. In other words, the north and south sides are largely independent of each other. It may therefore be concluded with great certainty that as to its main effects this fault is younger than the folding of the pre-Devonian sediments. In the Devonian complex on the Munotbjerg plateau (P. 1151) no such fault can be observed. The Devonian rests, so to speak, undisturbed over the fault. From the aircraft as well as on air photographs a small graben-like depression can be observed in continuation of the fault. A very inconsiderable post-Devonian reactivation of the fault should not, therefore, be refuted.

Probably it received its main appearance syntectonically with the folding of the pre-Devonian sediments. Its pre-existence is, of course, not excluded, however, the most probable interpretation seems to us to be that the fault was initiated right at the start of the folding, and then subsequently became the boundary between two areas which moved largely independently of each other.

Another fault extends farther northward almost at km 5 in the valley that opens immediately west of Margeries Dal. It was, indeed, only observed in the right side of the valley, as the left side could not be investigated. Here the base of Bed group 11 to the south is in contact with the base of bed group 7 of the north, the throw being accordingly about 520 m.

bb) the area between Margeries Dal and Flade Dal (sections 6, 7, 22).

The observations from this area are incomplete. Visited and thus more closely investigated are:

Margeries Dal from the south as far as the watershed, and from the north to ca. 2 km north of the watershed;

Fladedal from the south to almost km 10 and the corresponding western side of the valley;

Munotbjerg and the plateau, in so far as it was occupied by the Limestone-Dolomite Series. The investigation of the remaining area, viz. the western boundary of the Devonian north and northeast of P. 1151, was planned for the summer of 1949. However, unfavourable ice conditions compelled the expedition to return before the planned date. So we had primarily to map the region north of P. 1151 on the

basis of observations from the air and air photographs. This applies above all to the limitation of the Devonian relicts in this plateau which on the basis of the photographs can only be made with certainty in a few cases. Of the pre-Cambrian rocks in the region west of P. 275 in Flade Dal only the position can be indicated, but not their stratigraphic place within the Limestone-Dolomite Series.

The most striking observation made within the area of the Munotbjerg itself is that here the general strike has the direction SW—NE (cf. the structural map, fig. 19). Thus the limbs of the anticline, which has its core quite near the southern mouth of Margeries Dal, accordingly strike almost at right angles to each other. Along the east side of Margeries Dal this bending is hardly recognisable. However, the higher the beds are located in the stratigraphic sequence, the more distinct is their SW—NE-strike on the Munotbjerg plateau until in Fladedal it practically assumes the direction W—E. The dip, also, increases in this direction, thus e.g. the general dip of Bed 17 northeast of P. 1149 on Munotbjerg was measured to be 40° , while in Margeries Dal the dip of Bed 6 was 15° . At Munotbjerg this southeastward dipping side comprises beds from the Lower Multicoloured Series to the Tillites (sections 7, 22)¹⁾. The tillites occur right south of the Munot stream, and are of a striking red colour. This may probably be ascribed to a former covering by Devonian conglomerates, for it is readily observable that the red-coloured solution has penetrated along cracks and ravines into the tillites.

As suggested above (p. 52), about km 10 in Fladedal this uniform sequence of beds is cut across by the fault which extends westward through Margeries Dal and about 7 km farther towards the WSW. In Fladedal, immediately south of the fault, the bituminous limestones of Bed group 11 strike W—E and dip ca. 30° towards the south. North of the fault the beds of the lower Limestone-Dolomite Series occur *in situ*. Precisely which horizon is represented, it was unfortunately impossible to find out. But the beds are much more horizontal than those south of the fault, they strike approximately SSW—NNE, and dip very slightly towards the ESE. Judging by the aerial photographs, Bed group 17 just appears beneath the ice cap ca. 3 km north of P. 1151; this also seems probable according to observations made at the east side of the northern part of Margeries Dal.

The other fault, which runs about 1.5 km south of P. 1149 of Munotbjerg and parallel with the bank of the fjord, shows the following characteristics (sections 7, 22): stratigraphically its south side has a more elevated position than the north side; the throw is roughly 600 m.

¹⁾ The course of the lowermost western tributary stream to Fladedal, Munot rivulet, is incorrectly indicated on the topographical map. The upper course of the rivulet has the direction W—E, not NW—SE.

The beautiful section exposed along the first 4 km southeast of Margeries Dal along the fjord and extending from Bed group 4 to Bed group 12, is intersected by this fault. After some talus cones follow Bed groups 11, 12, 13, and 14, which here, according to their strike and their bending into the wall of Munotbjerg, show an exaggerated thickness. On top of the plateau, south of the Munot stream, the Lower Tillites are in contact with the dolomites of Bed group 17. The position of the latter is more horizontal than that of the beds north of the fault. Here some additional minor parallel faults are present in Bed group 17.

The interpretation of this fault causes some difficulty, its course being only visible along a comparatively short distance. The side facing the fjord was uplifted about 600 m. The continuation of the fault towards Antarctica Sund might probably suggest a dying out of the fault. To the east every indication of a fault disappears below the Devonian beds. Presumably it is a pre-Devonian fault, at any rate as far as its main effect is concerned. In most cases an interpretation like that given above for the large fault across the Munot plateau and through Margeries Dal would be satisfactory: along this fault the north and the south side were separated during folding and subsequently they moved largely independently of each other.

The stratigraphic position of the bedded reddish quartzites and greenish-grey quartzitic shales at Konglomeratnæs must be left open. They may belong to Bed group 6 or Bed group 5. They strike approximately E—W, and dip 35—40° towards the south. BÜTLER (1948, pp. 103—104) already mentioned this occurrence as well as the overlying Devonian conglomerates. A further exposure of these bedded quartzites is found in a small creek ca. 1.5 km ENE Konglomeratnæs. Here and at the cape the quartzites contain brecciated zones, which we assume to be of tectonic origin. Another small "window" of pre-Cambrian sediments in Devonian sandstones is located a little east of the river in Fladedal about at km 7. It consists of grey, bedded and shaly limestones of the Limestone-Dolomite Series. Some kilometres farther eastward, in the canyon of Jûluts Dal, BÜTLER found an occurrence of pre-Cambrian limestones, which he interpreted as evidence of a slight "Schuppung" of the pre-Cambrian substratum of the Devonian. The exact tectonic and stratigraphical position of all these isolated occurrences is regarded by us as not yet cleared up. They may have been brought into their present-day position by the above-described and some other invisible, W—E-faults.

b. The western limb of the border anticline.

From Blomsterbugten to the place where the shore of the fjord bends from a NE—SW-direction towards the SSW, the beds of the

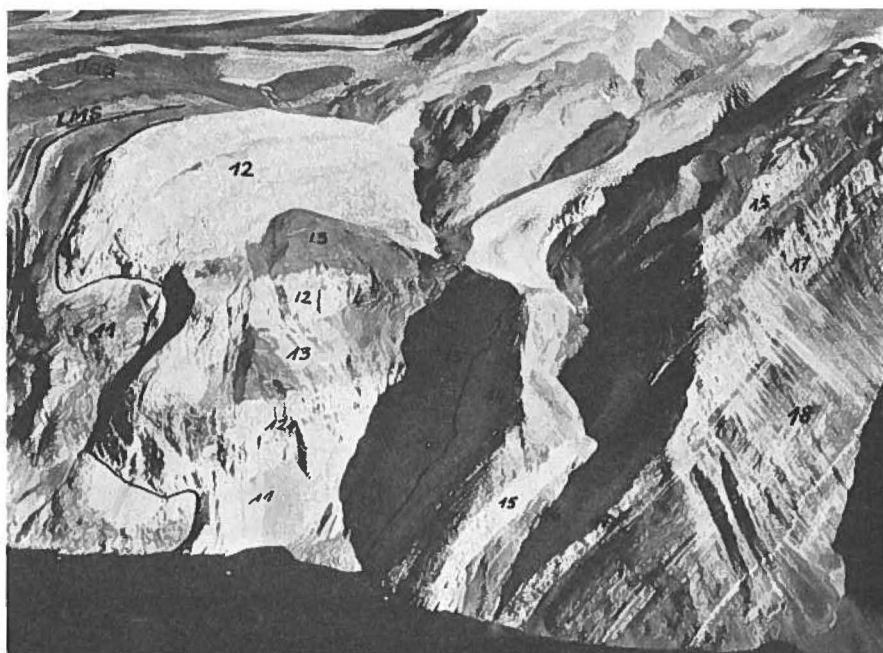


Fig. 10. NW-coast of Ymers Ø; below, right, the mouth of Kiledal, to the left the coast leading to Blomsterbugten. Looking almost eastward.—Legend: UQS = Upper Quartzite Series; LMS = Lower Multicoloured Series; 11—18 = Bed groups 11—18. The boundary of the beds (in heavy line) shows the course of the two recumbent folds (much shortened owing to the oblique view). To the right, successive overturning of the beds; to the left, above, increasingly horizontal position of the beds towards the crest of the border anticline.—BRENNERSEN phot.

Quartzite Series dip increasingly towards the west. From the place where the coast line changes its direction to Kiledal the westward dip increases even more, until the upper Limestone-Dolomite Series at the NE-side of Kiledal grows recumbent and dips ca. 60° towards the ENE (section 5). This increasingly steep position of the limb of the anticline from the slight westward dip to the overtilting is accompanied in the beds of the Multicoloured Series by an impressive zone of disharmonious folding on a large scale (figs. 10 and 11). It begins in Bed group 9 with an inverted fold towards the SSW. Bed group 10 lies in intensive minor folds around this inverted fold, and the same is the case with the bituminous shales of Bed group 11. This upper inverted fold increases its size in Bed group 12, to which comes, below it, a second one, which is indicated by the repeated cropping out of Bed group 10 from the sea. While the stiff Bed groups 12 are folded in gentle waves, those of Bed group 13 show again an increased number of minor folds. However, in the surroundings of Bed group 13 only the lower overturned fold is

recognisable almost 700 m above the level of the fjord, while the upper fold turns away from the fjord already in Bed group 12. In Bed group 14 the intensive minor folds are increasingly smoothing out, and Bed group 15 and higher-lying beds again show gentle folding, beautifully exhibiting the successive overturning of the side of the anticline.

In his account of the coast region, TEICHERT indicates a fault in the place where the overturned fold in Bed group 12 is visible and, bordering on Bed group 13, slightly overturned and crumpled descends into the fjord. This erroneous interpretation is easily understandable, for in this place there is a narrow, steeply rising valley, whose margins, according to the plane of observation, conceal the undisturbed course of the beds.

The slightly overturned western limb of the border anticline is traceable along the whole east side of Kiledal as far as the ice cap. This is indicated by the strike of the beds with gentle crests and hills, which with a strike of about 160° extend towards Antarctica Sund, as could be verified from the aircraft.

The western limb of the anticline is visible again between the southern edge of the ice cap and the northern shore of Antarctica Sund; here it is no longer overturned, but assumes a vertical position, at the most (sections 6, 7, 8). Thus a large part of the west side of Mountain 961 extends approximately along the dip of the beds. This applies in the first place to the Lower Tillites, which form the surface of this side of the mountain. Subsequently, however, the surface cuts into constantly higher-lying beds, the strike of the valley being directed somewhat more westward than that of the beds. In so far as the diagnosis on the basis of the colour of weathering is reliable, the Sandstone series occur *in situ* in the innermost part of the valley, at the southern margin of the ice cap. Here we have only described the upturned and in part even overturned west side of the border anticline; its importance and interpretation will be discussed below in the tectonic summary (p. 90 ff.).

Along the north coast of Antarctica Sund, from mountain 961 to Margeries Dal, a magnificent view of steep walls of the EBF down to Bed group 3 presents itself (sections 8, 9, 10). The more one approaches Margeries Dal, the more an increasingly horizontal position of the beds can be observed. Opposite Niviarsiat a valley rises towards the north, forming the eastern limitation of Mountain 961. At first it runs almost along the base of Bed group 12, then cuts across it, to extend parallel with the strike of the beds at the upper boundary of Bed group 12 until reaching the ice cap. A little above the place where Bed group 12 crosses the small valley, a small pressure fold is observable, which can be seen even more distinctly in Bed group 10 in the fjord wall. This small fold

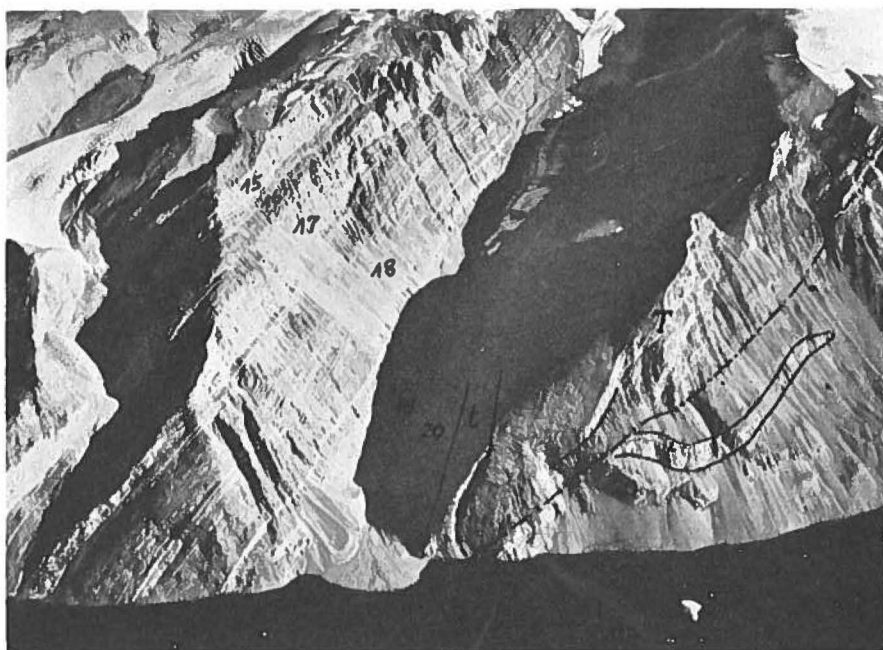


Fig. 11. Northwest coast of Ymers Ø, Kiledal. Continuation of fig. 10. Looking eastward into the overfolded series and the central syncline.—Legend: 15—20 = Bed groups 15—20. t = Lower Tillite; T = Upper Tillite; broken line = faulting in the core of the syncline.

continues as a doubling in Bed group 9; Bed group 8, also, shows some shearing, but this small element seems to have been smoothed out up to this point. From the top of Niviarsiat or of Skildvagten it is easy to observe all these steeply upturned beds striking in the direction 340° in below the ice cap. The small variation in the base of Bed group 12 was already mentioned in the preceding section.

c. The Devonian beds covering the Munotbjerg plateau.

As stated on p. 53, the boundaries of the extensive Devonian relicts in the surroundings of P.1151 can only be indicated. Only in the area south of the large fault cutting across the Munotbjerg plateau were the Devonian conglomerates accessible. KULLING and BÜTLER have already dealt with this region. The boundaries will appear from the map. Through the present investigation it was only ascertained that these Devonian relicts are larger than previously assumed, extending even along the west side of Margeries Dal. It can also be confirmed that the Devonian overlaps a pre-Devonian erosion landscape to the west.

3. Suess Land.

The border anticline here occupies only a small territory, namely a stripe of a maximum width of 5 km in the eastern part of Suess Land. About 2 km south of the north coast the anticline is cut across by an E—W-fault (the Carissima Dal fault). In addition, a number of peculiar N—S-faults occur in the eastern slope of the land.

The description will be given in three sections:

- a. The area north of the fault in Carissima Dal.
- b. The fault in Carissima Dal.
- c. The region south of this fault.

a. The area north of the fault in Carissima Dal (section 12).

As will appear from the description of the other folding zones, a connection free from faults between the tectonic elements on the two sides of Antarctica Sund is quite possible. This view was confirmed during the construction of the tectonic map (fig. 19). KOCH assumed the presence of a longitudinal fault through Antarctica Sund, while TEICHERT postulates a turn of the folding axis from NNW—SSE to NNE—SSW. The tectonic map shows that neither the one nor the other of the supposed complications is necessary for this connection. Only in the region between the southern mouth of Margeries Dal and Skildvagten can a rather slight turn of the folding axis be ascertained.

The two distinctive features of the northeastern corner of Suess Land, Niviarsiat and Skildvagten, belong to the border anticline. The main volume of the former is built up of the Upper and the Lower Limestone-Dolomite Series. Only towards the indentation between Niviarsiat and Skildvagten appear Bed groups 13 and 12.

The depression which extends as a passage south of these two mountains was called Carissima Dal¹⁾. From the northwestern debouchure of the valley into Antarctica Sund to the top of Niviarsiat the uptilting of the beds of the Limestone-Dolomite Series increases steadily. In the western part of Carissima Dal a southwestward dip of about 35° was measured, while at the top of Niviarsiat it is 65°. The highest point of the mountain is made up of Bed group 17.

The deepest-lying rocks, which occur *in situ* at the northern foot of Skildvagten at Antarctica Sund, are the alternating yellow and red dolomitic shales of Bed group 7. They here form the core of the border anticline. Above them follow, with magnificent unconformable folding, the mobile beds to and including Bed group 11. Group 12 forms a

¹⁾ This term covers the pass with the lake including the two valleys leading to the fjord.

broad gentle curve from the indentation between Niviarsiat and Skildvagten to the fjord at the foot of Skildvagten. At the western slope of Skildvagten to the aforementioned indentation, ca. 500 m above sea-level, a small secondary syncline appears at the upper boundary of Bed group 12, which syncline is filled with the lower portions of Bed group 13. In addition, at the northern slope of Skildvagten, on top of the limestones and shales of Bed groups 11 and 10, there lies a considerable block of white algal dolomites, which are a rock-slide derived from Bed group 12.—East of Skildvagten, P.1040, a small valley arises. In the upper part it has almost the character of an inclined plane, which about 900 m above sea-level forms the marked shoulder of Skildvagten. At the middle and lower portions the valley is everywhere much covered with scree. Still, in the perpendicular wall south thereof the yellow dolomite bands of the uppermost Bed group 13 can be recognised, which slope towards the fjord with the same inclination as the valley and are just visible above the talus slopes. The shoulder itself is constituted by these dolomite bands. Over a vast area they here form the surface with their magnificent algal structures.—The bizarre peak of Skildvagten is made up of Bed group 14. In the north wall of the mountain the irregularly dolomitised portions of this bed, as mentioned in the stratigraphic section, are particularly distinct. Further, Bed group 14 here shows not only the eastern limb of the border anticline, but also its axial southward dip, but Bed group 15 also occurs on the south slope of Skildvagten. This latter bed is likewise present in the steep wall rising from the fjord, and is here dislocated by a small N—S-fault (not shown on the map).

b. The fault in Carissima Dal.

At the eastern mouth of Carissima Dal an interesting fault begins. Its course is as follows: Its rise at the fjord is covered by scree and sub-recent accumulated delta relicts, but must be located 500 m north of the mouth of the rivulet flowing through Carissima Dal. Then this fault runs about 200 m above the bottom of the valley, along the southern slope of Skildvagten. Approaching the valley from the east, one encounters, after about 4 km's walk, a barrier formed by the quartzites of Bed group 6. These quartzites extend from the east along the south side of the valley slowly down to the bottom of the valley, where they build up the barrier. Immediately north of the rivulet, which here cuts across the barrier, Bed group 6 is in contact with Bed group 14. In the former, marked slickenside phenomena can be observed, which reveal a westward movement of the southern portion in relation to the north side.—West of this barrier the fault assumes a WNW-direction, to resume later the E—W-direction. The fault then reaches Carissima Gl., which, de-

scending from the south, terminates in two lobes. The further course of the fault of Carissima Dal will be described later under the treatment of the other folding zones (cf. fig. 17). For this part the following features should be pointed out: At the eastern rise of the fault, at the bank of the fjord, the upper Bed group 3 to the south and Bed group 15 to the north are in contact, the throw thus being 2,200 m; At the barrier described above it is Bed group 6 (S) and the upper Bed group 13 (N) that are in contact with each other, that is to say, a throw of 1200 m; at Carissima Gl. Bed group 16 (S) and the Lower Tillite (N) are in contact, the throw being 750 m. As will appear from the map, the fault dies out entirely to the west. In case of a supposed continuation eastward into the fjord, the differences in displacement, at any rate along a certain stretch, might decrease again, for here the north side of the fault dips more gently towards the east than the south side.

c. The region south of the fault of Carissima Dal

(sections 13—16).

The conditions south of the Carissima Dal fault are clear: in the eastern part of Carissima Dal, from about km 2 to km 3, the black shales of Bed group 3 appear as a window beneath the massive quartzite bands of Bed group 4. Bed group 3 here also forms the core of the border anticline. Then, however, about 1 km south of the outfall of the rivulet of Carissima Dal into the fjord, a mass of reddish and white quartzites occur *in situ*, as to the stratigraphical position of which we were long in doubt. However, after a study of the deeper-lying quartz series in Noa Dal and at Kap Alfred, it could be clearly demonstrated that it is the white quartzites of Bed group 2 and the uppermost portions of Bed group 1 that are represented here. These quartzites are separated from the other rocks of Carissima Dal by a marked fault with a throw of about 1200 m and an approximate N—S-direction (section 13). Two smaller parallel faults have lowered the areas near the fjord about 60 m each. Unfortunately the southward continuation of this surprising fault is entirely doubtful. Some faults of this particular direction might occur in the region of the 73th parallel at an altitude of ca. 600 m above sea-level, as well as lamprophyric veins in part associated with them—these veins occur also in association with the other faults of this zone—but large areas of the east side of Suess Land are entirely covered by scree and vegetation. At any rate, neither at the drainage stream from the glacier on Johnstrups Bjerg nor farther south is any trace of a corresponding fault to be found. What can be seen along the bank of the fjord as far as 1 km north of the Eskimo settlement (ca. 73° N. lat., 25° W. long.), are the resistant

quartzite beds of the base of Bed group 4 and Bed group 6, which in their southeastward dip normally fit to Bed group 2.

Southward from Carissima Gl. as far as Johnstrups Bjerg, an undisturbed picture of the west side of the anticline presents itself. The upper edges of the Multicoloured Series and the Limestone-Dolomite Series here form the steep wall extending along the ice cap. The crest and the eastern slope of the anticline are only recognisable in the ravine of Johnstrup Gl. and on the small plateau north thereof. The latter is bounded on the west by the westward dipping beds of the Lower Limestone-Dolomite Series and the Upper Multicoloured Series, on the east by a small ridge of eastward-dipping beds of Bed group 12. In between, Bed group 11 forms the actual surface of the plateau and the crest of the arch.—At the northern ravine wall of the Johnstrups Bjerg glacier the crest of the anticline is likewise clearly visible, and south of the glacier the eastward dip of Bed group 12 is traceable some distance further. In the river-bed of the stream draining the Johnstrups Bjerg glacier the beds—the sequence right down to Bed group 6—are horizontal or dip slightly towards the east.

Along the seven kilometres following this stream towards the southwest the magnificent picture of the wall of Johnstrups Bjerg presents itself. The beds of the Multicoloured Series and the Limestone-Dolomite Series extend down to the fjord, occasionally cut across by small E—W-faults¹⁾. One of these small faults is a fine example showing how the more massive portions of the Lower Limestone-Dolomite Series were faulted, while folding can be seen to have taken place in Bed group 18. The marked peak of Johnstrups Bjerg consists of the uppermost parts of Bed group 18 (cf. WEGMANN 1935a, fig. 3, p. 12).

4. Ruths Ø and Maria Ø.

The beds on Ruths Ø dip 35—40° towards the southeast (section 14). Southeast of the island there is still a remnant of Bed 15, while to the northwest the uppermost Bed group 5 appears above the level of the sea as the deepest member. Small faults intersect the island in a SE—NW-direction, each of them having lowered the southwestern side to some extent. From Ruths Ø a submarine threshold extends to Bed group 6 at the east point of Suess Land.

Maria Ø (fig. 12) is built up of Bed groups 7—16. The strike is almost ENE and the dip averaging 30° towards the SSE. Here, also, some minor faults cut across the series. They have a NNE-direction, and vary in their mode of movement. Characteristic of the landscape is a depression extending along the strike of the beds right across the

¹⁾ These faults are not shown on the map.

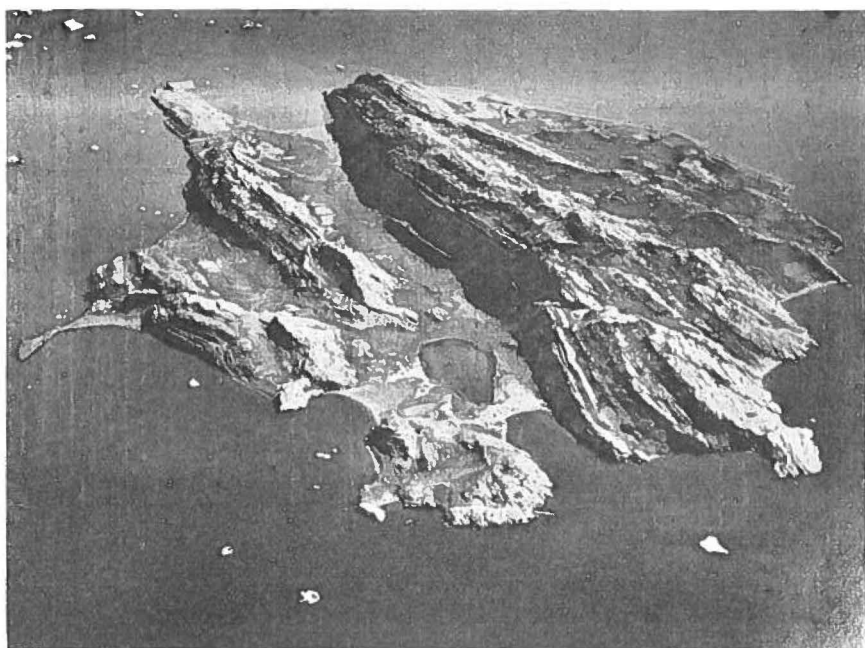


Fig. 12. Maria Ø, seen from the west.—The large depression extending right across the island corresponds to Bed group 13. The smaller one, to the left (north) of it, is formed by Bed group 10. The faults are distinctly visible. A comparison with the map in Pl. 2 shows that with the map at our disposal many interesting details could not be mapped.—BRENNENSEN phot.

island. It owes its existence to the highly bedded quartzites of Bed group 13. The rocks of Maria Ø have their strike and dip in common with those of Fimbulfjeld on Ella Ø. In the constructions of the sections the EBF-rocks of Maria Ø fit in below the Cambro-Silurian beds of Fimbulfjeld. Hence it may be assumed that the beds of Maria Ø and Ella Ø belong to the same tectonic unit. As will be seen from the tectonic map, a connection between Maria Ø and Suess Land is conceivable. This assumption of a connection between Suess Land and Ella Ø is expressed in the structural map (fig. 19), and constructed with a view to the conditions on Ella Ø and Maria Ø.—The tectonic relationships of Maria Ø and Ruth Ø, however, remain mysterious. A large fault must be present between these two islands, which has uplifted Ruths Ø in relation to Maria Ø. This assumption, also, was represented in the structural map.

5. Ella Ø. (Sections 17, 18).

In the course of the years this island has been subjected to various local investigations (BÜTLER, SCHAUB, HUBER, etc.). Owing to the cir-

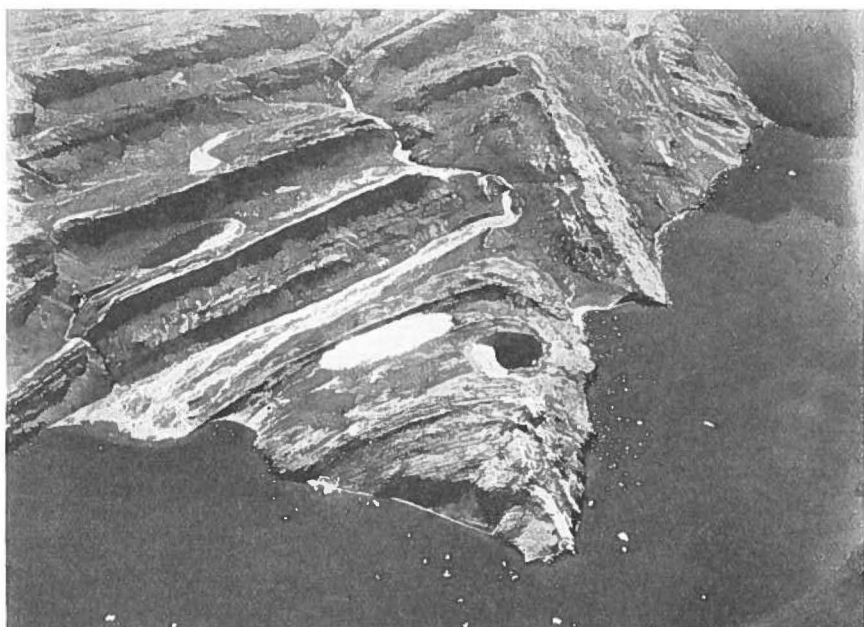


Fig. 13. Ella Ø, Kap Oswald, seen from the north. The anticline and its axial inclination towards the SSW are readily distinguishable. Cf. also fig. 14.—BRENNISEN phot.

cumstance that the expedition station is located on the north side of Ella Ø, the detailed investigations, not only the geological one, were centred here. In order to serve as a kind of excursion guide for the numerous geologists who, no doubt, during the years to come will stay for a longer or shorter time at the station, we treat the region of Ella Ø almost *in extenso*. The tectonical structure in itself, however, justifies a somewhat more detailed description of the island, as it exhibits forms which in regard to impressiveness and easy recognition leave nothing to be desired. Figs. 13, 14, and 15 are distinct illustrations of this fact,

The border anticline here extends from Kap Oswald towards the SSW across the island. In the aerial photograph (fig. 13) the condition are clearly visible. The whole projecting mass of Kap Oswald is built up of the beds of the Limestone-Dolomite Series, that is, of Bed group 14 and overlying strata. They show a beautifully exposed anticline. Rund sø lies in the core of an arch, while Lang sø accurately reflects the strike of the "E"-side of the anticline. For this side has at first a SW—NE-strike, but on approaching Kap Elisabeth turns more and more towards the ENE, to bend, finally, almost towards the E.

Various approximately NNW—SSE running faults intersect this side of the anticline. The westernmost fault begins in the small bay at

the west coast of Kap Oswald. As far as the place where the rivulet debouching at the station turns towards the NE, the course of the fault is concealed below Quaternary rocks and vegetation. In this bend, the rivulet cuts across approximately N—S-striking limestones and limy shales of the uppermost Limestone-Dolomite Series. These limestones form, in the very knee of the rivulet, a hill with the same strike of the beds.—About 100 m east of the limestones, beyond a plain covered with vegetation, the SW—NE-striking dark and multicoloured shales of Bed group 19 occur *in situ*. Above them follow the Lower Tillite, forming the lowermost marked level of the terrain. The SE-dip amounts to 30—35°. The fault accordingly runs below the plain covered with vegetation. Its further course at first coincides with the bed of the rivulet, forming subsequently a readily recognisable creek up the slope towards the SSE. In its lower part the fault has dislocated e. g. the lower portion of the tillites to the west, to meet the upper portion to the east. Farther up the slope, however, the amount of displacement apparently decreases. That it is only an apparent decrease, is due to the intersection with the terrain. The movement mainly followed vertical components. Between the two steps which were formed of the Lower Tillite on either side of the fault, lies a small hill likewise built up of Lower Tillite. We interpret it as a drag-faulted block. It may also be mentioned that the fault about 400 m above sea-level, at the point where the rivulet bends southwest towards Bastionen, issues a small branch towards the western side¹⁾ (to the right in fig. 14).

The next fault to the east is the largest one in Ella Ø. At first it runs through the small ravine eroded out by the stream draining Ulvesø. Here the west side is elevated about 820 m above the east side, the base of the Lower Tillite being in contact with the uppermost Tillite Canyon Formation. In the photograph (fig. 14, left) the further course of this fault is distinctly visible. Just as distinct is the morphological difference between the pre-Devonian sediments and the Devonian conglomerates. At the place where the fault meets the Devonian edge, it can be ascertained that the Devonian, also, has been dislocated, though much less so than the Cambro-Ordovician rocks. The further course of the fault in the Devonian is characterised by a marked slope in the terrain, which extends towards a place some distance north of Kap Harry. From the intersection of the whole course of the fault with the terrain from the station to Kap Harry, an eastward dip of the fault plane of about 70° may be assumed.

Along the following stretch towards the east as far as Kap Elisabeth three more faults, of the same direction and the same character, occur.

¹⁾ Here also, as to the course of the rivulet, the topographical map had to be corrected somewhat.



Fig. 14. Ella Ø. Step-fault landscape of Fimbulfjeld. View towards the SSE. The fault to the right in the photograph is that which begins west of Kap Oswald; the fault to the left is that with the greatest amount of dislocation on Ella Ø. Its continuation in the Devonian is characterised by a marked edge in the terrain. Legend: M = subrecent morainic remnant (cf. p. 100); 18, 19 = Bed groups 18 and 19; t = Lower Tillite; T = Upper Tillite; TC = Tillite Canyon Formation; Sp = Spiral Creek Formation; SS = Sandstone series; B = Bastion Formation; E = Ella Ø Formation; H = Hyolithus Creek Formation; DP = Dolomite Point Formation; CF = Cass Fjord Formation; CW = Kap Weber Formation; D = Devonian.—BRENNEISEN phot.

Their amounts of dislocation, however, are inconsiderable, about 100 m. The strike of the beds curves, as already mentioned, gradually towards the ENE, almost towards the E, the southward dip likewise decreases somewhat in this direction, so the rocks at Kap Elisabeth are only slightly inclined.

The area around Lemmingbugt: This depression evidently owes its existence to a number of trough-faults in almost horse-shoe-shaped arrangement. They are indicated somewhat schematically in the map. It is a remarkable fact, however, that Vega Sund is located in the immediate continuation of the Lemmingbugt depression to the east. It is conceivable, therefore, that the two depressions are genetically associated. However, further support to this assumption can only be supplied by accurate measurements.

The crest of the border anticline extends across a saddle to the steep wall that descends almost perpendicularly to Narhvalsund. WEGMANN (1935a) and BÜTLER (1948) have already described isolated occurrences of Devonian rocks situated west of the crest of the anticline, immediately at the upper edge of the steep wall just mentioned. It remains only to be added that these Devonian conglomerates extend as uninterrupted narrow stripes as far as the northern wall of Bastionen. WEGMANN mentions these Devonian occurrences as an example showing that the inner (upper) beds of a syncline, owing to lack of space at the edges of the syncline, have been thrust across the lower beds (WEGMANN 1935a p. 33, fig. 6 and p. 34). Owing to these processes during the Devonian accentuation of the Bastion syncline the Ordovician limestones may have been thrust across the Devonian relicts. According to our observations, comprising the whole Devonian relict, however, it is just as justifiable to assume that we are here simply concerned with deposition of conglomerates in a ravine eroded in pre-Devonian time, by analogy with observations made by BÜTLER (1948, p. 104) in the E.slope of Munotbjerg.

Fig. 15 shows the anticline as it presents itself in Narhvalsund. The deepest-lying horizons, which here in the core still rise above the level of the sea, are those of the uppermost Tillite Canyon Formation. Thus there is an axial pitch of about 10° for the distance Kap Oswald—Narhvalsund. Owing to the exceedingly acute anticlinal bend, the stiff quartzites of the Sandstone Series have broken in places.

The east side of the border anticline in Narhvalsund grows increasingly horizontal, without complications, to disappear, finally, below the huge Devonian conglomerates. It is in this place, also, that the youngest pre-Devonian rocks so far known in East Greenland occur, viz. the calcareous portions of the Narhvalsund Formation. This fjord wall, also, offers a magnificent view of the acute anticline with its rocks shining in the most intense colours, and also of the very conspicuous unconformity between the pre-Devonian and the Devonian series.

Some geological details on Ella Ø.

In connection with the tectonical description of the greater part of the Ella Ø some further comments will be added. They are not immediately associated with the tectonic description, but in dealing with them, we comply with a request by Dr. L. KOCH to give a more detailed description of the nearest surroundings of the station, where so many geologists have stayed.



Fig. 15. Narhvalsund, south side of Ella Ø.—Border anticline; the dark area in the core of the anticline corresponds to the Tillite Canyon Formation, the next dark soft beds represent the upper part of the Bastion Formation, that is, the lowermost fossiliferous horizon within our area. On top of these beds, the Devonian relict is easily distinguishable on the upper edge (left side of anticline).

In the first place, the tillite exposures and fossil localities will be recorded, and secondly a description of some glacio-geological and morphological observations will be given.

The tillite series on Ella Ø.

As for its stratigraphy and interpretation the reader is referred to p. 26 ff. as also to the papers by KULLING (1930), BACKLUND (1930), HUBER (1950), and SCHAUB (1950). Here only the best accessible exposures will be mentioned. They are located directly southeast of the station, where the lowermost marked rock wall is formed by the Lower Grey Tillite. The small valley in which the Ulvesø is located corresponds to the dark shaly complex occurring between the two tillites. The second rock wall is thus mainly made up of the Upper Red Tillite. A good travelling route for a study of the tillite section is as follows: the station—Kløftelv—Ulvesø, thence along the river Storeelv, which debouches 400 m east of the station, and up the ravine to the second small valley.

Another section is exposed in the western limb of the border anticline. The Lower Tillites here reach the shore in the small bay southwest of Kap Oswald, into which the stream draining Rundsø flows (Tømmerbugt). From this point a good section is likewise accessible along the shore, though it is occasionally covered by soil flow. Walking from the Upper Tillites some distance farther on towards the Bastion wall, one has the opportunity to observe the Spiral Creek Formation. Beautiful ripple-marks are found here in addition to well developed NaCl-pseudomorphoses.

Fossil localities.

We here only give the localities. As to the fossil contents of the Palæozoic beds, the reader is referred to the works by CHR. POULSEN (1930, 1932, 1937) and the collections and investigations which were carried out at the same time as our investigations by a British colleague (COWIE).

Pre-Cambrian algæ.

These structures, described on p. 22, are chiefly found on the Kap Oswald peninsula. The way from the station to the highest point of the peninsula leads across various approximately SW—NE-striking, southeastward dipping upper edges of beds of massive limestones and dolomites. One of these outcropping beds is strewn with silicified algal structures. The beds belong to Bed group 15.

Palæozoic fossils.

The Bastion Formation. The formation is subdivided by POULSEN (1930) into a lower quartzitic and an upper shaly portion. The latter represents the deepest fossiliferous horizon of the Palæozoic in East Greenland. On the basis of his collections, POULSEN referred this upper Bastion Formation to the upper Lower Cambrian. According to verbal information by L. KOCH, about twenty years ago the best exposures in the side of the anticline were met with at the south side of Ella Ø. However, extensive landslides have in the meantime made these places inaccessible over wide stretches, and covered them with scree.—A small exposure is found at the north coast ca. 1 km east of the station in the small valley which meets the largest of the faults about 50 m above sea-level to the east. The easiest way to the place is by the so-called Kløftelv which drains Ulvesø.—Another possibility for collecting fossils in this horizon is offered at Bastionen, more precisely in its northeastern slope about 800 m above sea-level. The dark shaly portions are here very steep, forming the west side of the border anticline. To reach this place, it is easiest to walk along the corresponding

anticlinal limb in the Upper Tillite, climbing as far as the saddle which extends east of the Bastionen across the whole island.

The Ella Ø Formation (Upper Lower Cambrian). In contrast to the upper shaly Bastion Formation, the massive Ella Ø Formation forms a marked horizon in the step-like landscape of the western Fimbulfjeld. It is the third step above the station, that is, the one which follows the plane of Bjørnesø southward. In the down-fallen blocks as well as in the rocks *in situ*, very fossiliferous zones are met with, i. a. such as consist entirely of trilobite shell remains. However, horizons with abundant *Hyolithidae* are also of frequent occurrence. Walking eastward on the plain of the Bjørnesø, one arrives at the largest of the N—S-faults; climbing along this, one soon arrives, at the eastern side of the fault, at the grey, nearly marly beds of the

Cass Fjord Formation (Upper Ozarkian), which is here lowered by the fault about 800 m. Many fossils can be collected in the scree from the overlying beds. Another place in which the formation can be studied, is situated ca. 3 km west of Kap Elisabeth. Here a small valley extends parallel with the bank of the fjord, descending towards Kap Elisabeth, at about 150 m above sea-level. Morphologically it depends on the softer beds of the Cass Fjord Formation. Here, also, there are good opportunities for a study of the formation both *in situ* and in scree. At the northern slope of the small valley there appear in places, in the uppermost bedding planes of the Dolomite Point Formation, concentric structures which are highly silicified. They attain diameters of up to 50 cm, and possibly represent algal structures similar to those described on pp. 20 and 22 from the Eleonore Bay Formation.

The Kap Weber Formation. Upper Canadian (POULSEN 1930). These extremely thick, mostly brown limestone beds, include some very rich fossil horizons. For the most part, the fossils are silicified. Silicification in comparatively thin beds, which often project as ledges from the weathered rock, are of frequent occurrence. The most favourable fossil localities are found at Kap Elisabeth. Climbing westward from this cape towards Fimbulfjeld, we encounter, already after about 500 m, the fossils laid bare by weathering. Many fossils are also to be found on the plateau of Bastionen, the greater part of which is made up of the Kap Weber Formation. But here they are for the most part so highly fused with the rock that no satisfactory and identifiable forms could be found. Good exposures of the Cass Fjord Formation are likewise found at the northern edge of the Bastion plateau.

Narhvalsund Formation (Upper Canadian, POULSEN 1951).—The best exposures of this formation are to be found at the south side of Ella Ø, especially where the eastward dipping side of the anticline disappears below the transgressing Devonian. On the Bastion plateau

the Narhvalsund Formation likewise occupies vast areas, but here it is not always easy to delimit, and moreover the rocks *in situ* are largely concealed by recent formations of frozen soil with block material.

Remnants from the last ice age.

Glacial boulders, e. g. crystalline blocks, are present even on the Bastion plateau. (On Suess Land they were encountered on the plateau of Gravhøjen, more than 1,500 m above sea-level). In the saddle which extends east of Bastionen across the island, almost following the side of the anticline, even morainic ridges from a later stage of the glaciation are preserved.

A little below the Ulvesø plain the river Storeelv, debouching 400 m east of the station, cuts into another sub-recent moraine, parts of which have been washed away. It shows a grey colour of the matrix and numerous boulders of the size of a head. A remarkable feature of the moraine is its absolute agreement in habit with the Lower Tillite occurring *in situ* only about 50 m farther westward. This juxtaposition of two glacial sediments, which are separated by some hundred millions of years, but are otherwise alike, is a rare geological phenomenon.

As remnants of the last ice age may finally be mentioned the roches moutonnée of Kap Oswald and Fimbulfeld.

Morphology.

The whole further surroundings of the station present a beautiful example of the morphological display of all tectonical structures. This is distinctly shown by figs. 13 and 14. The reader is also referred to the excellent description and sections on the geology of Ella Ø published by BÜTLER (1937).—In the proximity of the station, however, there are many interesting morphological details, which will be briefly mentioned below.

Shore-lines.

A general description of the shore-lines is given in the section "Observations on the Quaternary and the morphology". Here only some corresponding phenomena on Ella Ø will be mentioned. In the bay at the station beautiful concentric barrier beaches between the shore of the fjord and the station house are reflected in the shore-gravel. Other indications of an old sea-level are the numerous mussel shells to be met with everywhere on the plain around Langsø and Rundsø. NOE-NYGAARD (1932) devoted a special publication to them. Shore-lines are further found near Kap Elisabeth.

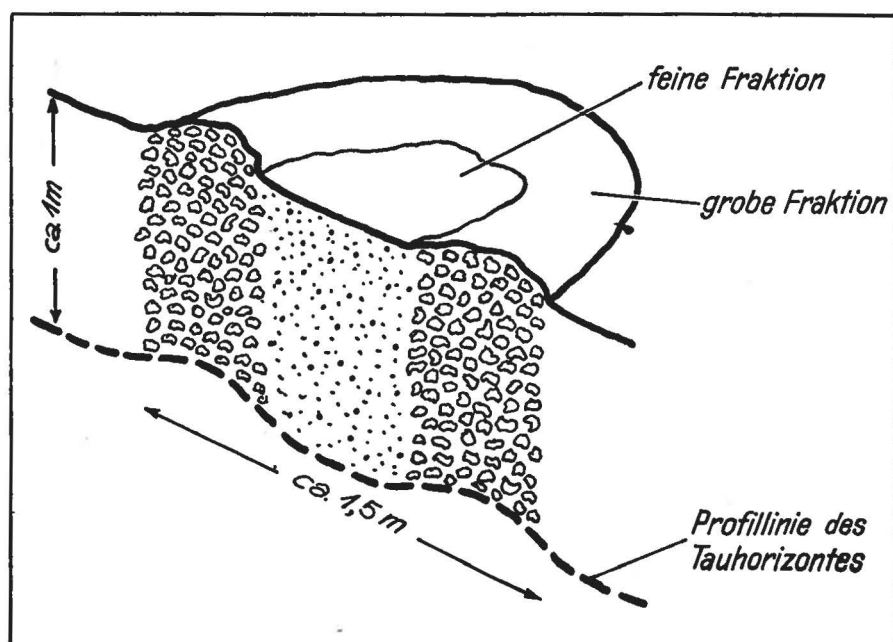


Fig. 16. Schematic sketch of Polygon soil

Polygon soil and frozen soil.

These phenomena, which are widely distributed in East Greenland, can best be studied some hundred metres east of Rundsø. There are here some structures consisting of a circular ridge with pebbles almost the size of a hand, and a centre consisting of much finer-grained gravel, covered with vegetation. The circular ridge has an outer diameter of up to 1.5 m, it may be 10–15 cm high and 40–50 cm wide. In order to examine more closely the differentiation of the grains of the gravel, two similar “frost rings” were examined by digging down to the horizon of thaw. In both cases such conditions were revealed as are represented schematically in fig. 16. The frozen horizon was reached at a depth of 85–115 cm (beginning of July 1948). It is noteworthy that in the area of coarse gravel the frozen horizon was found to mark a distinct threshold. This can probably be ascribed to the increased loss of heat owing to evaporation.

Another striking feature in the region of Langsø and Rundsø is the knolly structure of the ground covered with vegetation. In our opinion, this is likewise due to frost. The areas of fine- and finest-grained gravel were more favourable for the formation of humus than the coarser ones. Thus the vegetation developed here in an intensified degree as compared with the coarse-grained margins.

6. Summary.

On the basis of the tectonic map (fig. 19), the following summary of the observations at the border anticline can be given:

On Gunnar Anderssons Land and in Noa Dal the east side and the crest of the anticline alone attain a width of nearly 20 km. Numerous E—W-faults cut across this area in an antithetic block-faulting, in which the northern block was always lowered and moved slightly westward. The E—W-faults are younger than the general N—S-running main fault, which attains a throw of over 300 m and was reactivated in post-Devonian time. Older than the E—W-faults are likewise some quartz veins at the south slope of Noa Dal.

The west side of the anticline shows, at Kejser Franz Josephs Fjord, a slight overtilting, but in its further course as far as Antarctica Sund resumes a simple vertical position. The east side shows generally a marked strike towards the NNE and NE, most strikingly at Munotbjerg and on Ella Ø. The anticlinal axis is highest on Gunnar Anderssons Land, descending gradually towards Ella Ø. The axial southward dip is reduced successively by the E—W-faults in Noa Dal, by the fault at Munot plateau, and by the fault of Carissima Dal. Especially along the latter two faults this gave rise to a differentiation of the folding intensity of the two sides. From Noa Dal to Ella Ø the inclination of the anticlinal axis is about 6,500 m or somewhat over 8°. The direction of the axis from Noa Dal to the southern mouth of Margeries Dal is almost NNW—SSE. From this place to the fault in Carissima Dal it strikes towards the S. to SSW. Between this fault and Ella Ø the anticlinal axis forms a curve convex towards the east, with a curve radius of approximately 25 km.

II. The Central Syncline.

In contrast to the border anticline, the central syncline north of the fault in Carissima Dal is very narrow, especially in the neighbourhood of the overfolded border anticline. Along the said stretch the synclinal axis rises somewhat towards the south.—South of the fault in Carissima Dal the syncline is much flatter, not rising till about the 73th parallel, showing subsequently an intensive fall of the axis as far as Narhvalsund.

We divide the description according to the three areas as follows:

1. Ymers Ø.
2. Suess Land.
3. Ella Ø.

1. Ymers Ø (Sections 5—9).

As might be expected from the overfolded border anticline in the region of Kiledal, the central syncline here occupies a very narrow space. At the mouth of the valley in Kejser Franz Josephs Fjord the initial synclinal bend (fig. 11) is distinguishable in the Lower Tillites in the southwest side of the valley, immediately above sea-level. A few metres farther southwestward the conspicuous band of in part silicified dolomites of Bed group 19 rises above the level of the fjord. It is intensively folded. Above the following talus slopes it is the limestones of Bed group 20 and the Lower Tillites which, forming two steps, crop out from the scree. They are in contact with northeastward-dipping and tectonically dislocated Upper Tillites. The intermediate fault is well visible. This fault in the syncline is traceable in the wall dipping towards the west, as far as the plateau. There the marked red band of the Spiral Creek Formation belongs to the overfolded west side of the border anticline, separated by the fault from the dark shaly areas overlying the Upper Tillites. In the immediate proximity of the fault, Bed groups 19, 20, the Lower and the Upper Tillites, show a slightly intensified bend owing to drag-faulting.

At the north coast of Antarctica Sund analogous conditions are met with. Thus the rocks on either side of the valley at first extend northward west of Point 961 along a distance of ca. 3 km, then turn northwestward and divide into two branches. The nearly vertical west wall of the valley is made up of more or less horizontal Lower and Upper Tillites. The base of the tillite series crosses the valley at an altitude of about 200 m above sea-level, rising with a gradient of 30—40° in the east side of the valley as the western limb of the border anticline. West of the valley, however, the rock *in situ* is soon covered by an area of débris. The covering is complete, and reaches from the level of the fjord to the SW—NE-directed part of the valley. Still farther westward the upper edge of the fjord wall is built up of Lower Tillites, which here form the limb of the western anticline. Thus, in the zone covered with débris the Lower Tillites must rise from 300 to about 1000 m above sea-level. The topographical picture, however,—especially of the zone of débris—justifies the assumption of a fault which has uplifted the western side. For the sudden interruption of the sharp edges of the territory formed by the tillites on either side of the zone of débris calls for an explanation. Another supposition is that this fault may be connected with that of Kiledal. We regard the fault as an intensification of the west side of the central syncline in the upper Limestone-Dolomite Series and in the Lower Tillites.

2. Suess Land (Sections 11—16).

At the south shore of Antarctica Sund the deepest point of the central syncline is found immediately at the first N—S-running rivulet west of Carissima Dal. From the terminus of Carissima Gl. the Lower Tillites and the immediately overlying beds extend down the valley to ca. 100 m above sea-level. At the eastern margin of the actual valley the multicoloured shales of Bed group 19 accompany the abruptly interrupted tillites throughout their whole thickness. The tillites do not appear again till farther up as a direct continuation of the Upper Tillites, which lie in the just described normal sequence of beds above the Lower Tillite bed descending from Carissima Gl. An analysis of the conditions reveals the following picture: As the normal overlying bed of the Limestone-Dolomite Series at Niviarsiat, the Lower and Upper Tillites form the southwestern side of the border anticline. At the gorge a N—S-fault now elevates the Lower Tillites of the west side to the same level as the Upper Tillites of the east side of the fault (fig. 17). While the Lower Tillites owing to their rigidity are broken rather sharply, the flexible multicoloured shales of Bed group 19 are disharmoniously folded around the "end" of the Lower Tillite formed by the fault. Strong folding is also exhibited by the upper horizons of Bed group 18 farther up the ravine as well as immediately east of the outlet of the gorge into the fjord. Associated with them, the beds of the Limestone-Dolomite Series here rise successively out of the fjord to form, at the next glacier valley, in gentle course the culmination of the western anticline.—Looking at the Tillites descending from Carissima Gl., one should not be deceived by their direction of strike, which is here almost NNE—SSW, and this should be taken into consideration in the analysis of the locality. It will be found that here, also, the central syncline is compressed within a narrow space and, almost at its sole, it was at first not disturbed, analogously to the conditions at the opposite shore of the fjord and at Kiledal.

However, south of the fault in Carissima Dal, conditions are different. This fault, whose course in the border anticline as far as Carissima Gl. was treated above, is also distinctly visible in the central syncline. The mountain between Carissima Dal and the next one to the west is called Mountain 1 (fig. 17). In front of it, towards Antarctica Sund, lies the round platform which is built up of the Upper and Lower Tillites separated by the synclinal fault. Mountain 1, however, shows a fairly broad syncline; its peak is made up of the black shales interbedded between the two tillite series, while the Lower Tillites, rising towards

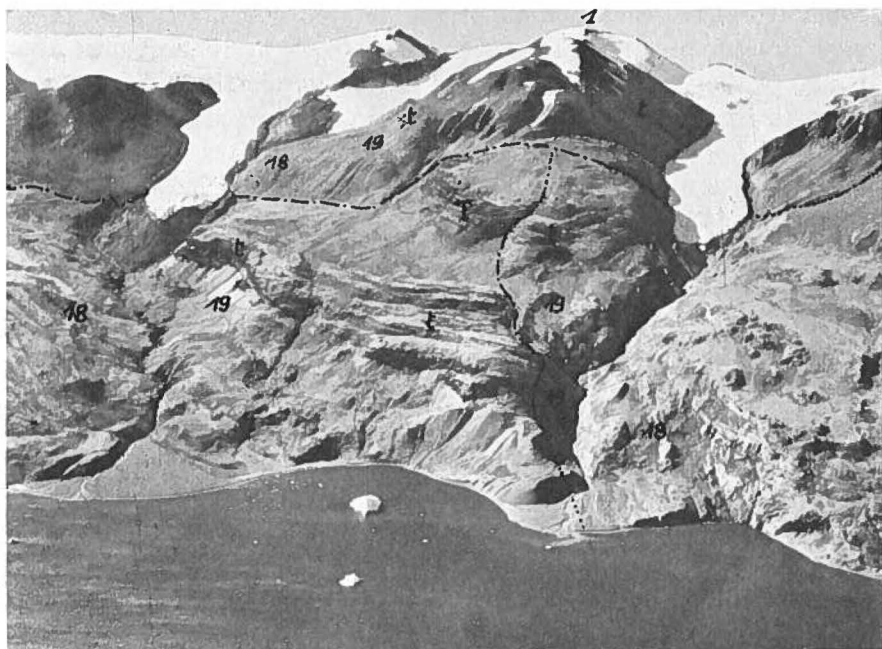


Fig. 17. North side of Suess Land, western mouth of Carissima Dal. Legend: 18, 19 = Bed groups 18 and 19; 1 = Mountain 1; t = Lower Tillite; T = Upper Tillite. Broken lines = faults.—The fault of Carissima Dal extends across the picture from left to right; at right angles to it; the fault cutting through the narrow synclinal core. The general strike of the beds almost follows the diagonal from below, right, to above, left, of the picture.—BRENNERSEN phot.

the south, are visible in both sides of the mountain. Thus it is highly probable that the fault in Carissima Dal is marked by the ledge of the platform of Mountain 1. An escarpment, extending from Carissima Gl. to the platform, confirms this view, and in the slope at the glacier west of Mountain 1 the contact between the Lower Tillites to the north and Bed group 18 to the south is also exposed.

South of the fault in Carissima Dal the central syncline thus assumes a much more gentle shape. At the east side of Mountain 1 the Lower Tillites rise slowly and steadily towards the south. Between Carissima Gl. and the small glacier just west of it they appear again. The narrow ice-free tongue between these two glaciers is occupied by Bed groups 18 and 19. Finally, the Lower Tillites form the steep rock wall of Point 1874 towards the east. At the west side of Mountain 1, also, the Lower Tillites extend, slightly rising, towards the south, 150–200 m above the glacier. The extensive ice-covering of the higher regions renders observation of the structural position of the rocks impossible.

The southern continuation of the central syncline is now to be sought in Ørkendal¹⁾. From the air photographs it is evident that the core of the syncline runs through the mountain ridge that extends northward from P. 1472. At the SW-slope of P. 1472 a minor displacement can be observed, where, within the Upper Tillites, the upper portion has slid down towards the north along a slightly inclined plain.—Moreover the west side of the syncline appears farthest back in the valley, in the area between the two hindmost ramifications of the main rivulet. Here the Lower Tillites, Bed group 19 and the upper part of Bed group 18, occur *in situ*. At the upper end of the western of these two tributary valleys the effects of a fault are evident: east of the valley the just mentioned Bed groups 18 and 19 of the Lower Tillites occur, dipping towards the east, while west of it, in a more or less horizontal position, the Lower Tillites to and including the Upper Tillites are found. However, no indications of such an acute anticline as a fault-free connection of these beds would require, are present. We therefore assume the presence of a N—S-fault with a throw of ca. 300 m. Possibly this fault plays a rôle farther southward, also, namely at the western slope of Ørkendal between the lowermost and the uppermost tributary stream from the west. For here, also, the connection between the horizontal-bedded, slightly southward-dipping Lower Tillites and the opposite eastern side of the valley, where the series constitute the central syncline is not quite obvious²⁾. On following the lower eastern tributary stream upwards from its outfall into the main river, the following section can be observed in the deeply cut wall of the ravine: Uppermost Bed group 18, after ca. 200 m Bed group 19 to a thickness of about 150 m. Above follow the Lower Tillites and the overlying black shales, and finally the lower

¹⁾ With a view to a correct estimation of the reliability of the observations the following remarks may be put forward: The two lowermost tributary streams towards the east and west run in 50 m deep, sharply cut ravines, which extend as far as the glaciers. At the time the area was investigated (first half of June) the streams carried large water masses owing to the melting of the snow. As we were alone, it was impossible for us to push farther into Ørkendal than to the two tributary streams. The geological conditions accordingly had to be analysed solely by means of the telescope and the air photographs. Thus we make all reservations as to erroneous stratigraphical interpretation and areas concealed from view.

²⁾ On the topographical original map there is here an error to be corrected. The lowermost eastern tributary stream debouches well over a kilometer farther down into the main rivulet than the lowermost western tributary. On the map both the streams debouch in the same place. In connection with this, the mutual positions of Points 1254 and 1006 evidently do not agree either. Unfortunately this could not be corrected, as we lacked the necessary observational equipment. The erroneous map had an unfavourable effect on the mapping of the W—E-faults in the lower Ørkendal, since in order to retain the actual direction of these, we had to give them a somewhat incorrect course in the territory.

two-thirds of the Upper Tillites. The whole series dips 15° nearly towards the east. Above follows a tectonically much disturbed zone, which has a vertical position and is about 10 m broad. It indicates a fault, for in connection with it the Lower Tillites form the walls of the gorge of the rivulet, descending with the valley towards the west. The same fault is found higher up in the northwestern slope of P. 1006. Here the deeper portions of the Upper Tillite lie alongside the Lower Bastion Formation. Farther on the fault is visible in the northward indented wall opposite Kap Oswald which extends from P. 1006 to P. 1446. Here it borders on the Spiral Creek Formation to the south, and to the north on the middle portions of the black shales lying between the tillites. Here, about mid-way of the throw, a wedge is squeezed into the fault by dragging. The top of P. 1006 itself is made up of the lower 150 m of the Hyolithus Creek Formation, below which follows the normal sequence as far as the Tillites. To the southwest another small fault (not indicated on the map) occurs in the wall, lowering the south side about 50 m. At the slope of P. 1006 towards Ørkendal an arch is seen in the Ella Ø Formation and the lower portions of the Hyolithus Creek Formation, which, by analogy with that of P. 1472, must be regarded as a bend of the central syncline.—The continuation of the above-mentioned fault towards the west probably coincides with the lower course of the lowermost western tributary streams. For the reasons mentioned in the footnote 2 on p. 76, this is unfortunately not the case on the geological map, as the fault had to be drawn some distance farther northward.

East of the Ørkendal delta, uplifted in post-glacial time, the shore of the fjord is made up of huge talus slopes. Above them appear the eastward rising Upper Tillites. In the immediate proximity of the eastern margin of the delta, however, multicoloured bands of the shaly Bed group 19 may occasionally be recognised, projecting from the scree. They dip towards the east, and form the overlying bed of a small exposure of Lower Tillites near the NE-margin of the delta shingle. Between this Bed 19 and the Upper Tillites there must accordingly be another fault. This is confirmed by the fact that about 400 m north of the aforementioned exposures of visible Lower Tillites the Upper Tillites occur *in situ*, which, undisturbed, are connected with the occurrence in the lowermost eastern tributary ravine. The further westward course of the fault is doubtful, as the whole bottom of the valley is covered with alluvial deposits, morainic débris, and vegetation. A peculiar furrow in the surface of the tillites, which dips towards the east in the slope of P. 1254, may indicate this continuation.

West of the Ørkendal delta the upper portions of Bed group 18 are extremely intensively folded. Farther up the slope the beds show more gentle folding, and, as mentioned above, the Lower Tillites constitute a

large part of the eastern slope of Mountain 1254. At ca. 1000 m above sea-level, however, this side of the syncline is intermittently steeper, and Bed group 19 forms the slope, but subsequently it again adopts a less steep position and forms the mountain ridge.

3. Ella Ø (Sections 17, 18).

The whole gigantic block of Bastionen is dominated by the central anticline. At the rock wall descending towards Kempes Fjord the syncline is readily observable. To the east the Dolomite Point Formation dips more than 60° towards the west, bends sharply into a horizontal position, and rises, after an equally abrupt uptilting, again towards the west. A minor upthrust along a plane dipping about 15° towards the west at the foot of the wall testifies to the space reduction in the core of the syncline.—In Narhvalsund the syncline lacks the flat sole: the east and west sides meet with dips of ca. 30° . Moreover it is here the Cass Fjord Formation which reaches down to the fjord, while on the north side the Dolomite Point Formation forms the lowermost bed above the level of the sea. The syncline accordingly dips towards the S. The lamprophyre veins examined by RITTMANN southeast of the cape of Bastionen will be treated in a later chapter (p. 98).

4. Summary.

A fault accompanies the central syncline from Kiledal to the fault of Carissima Dal. South of this fault the syncline grows much broader. In contrast to that of the border anticline, the axis of the central syncline along this stretch rises towards the SSE, at any rate very slightly. In the area of Antarcics Sund the axis reaches its culmination, subsequently descending somewhat again as far as the fault in Carissima Dal. The latter gives rise to a fault-like elevation of the axis of about 500 m. In the region around the 73th parallel the synclinal axis shows another culmination, south of which it descends steadily towards the south. This descent is further increased by the two W—E-faults in the lower Ørken-dal. The direction of the axis between Kiledal and the CarissimaDal fault is approximately NNW—SSE, from the latter place to Narhval-sund it is N—S.

III. The Western Anticline.

1. Ymers Ø (Sections 5—7).

In the fjord wall southwest of Kiledal the Lower Tillites rise in connection with the central syncline, reaching its culmination in a wide curve ca. 400 m above sea-level. However, this crest is not visible,

extensive talus slopes and talus flows covering the locality. But the tillites are visible on either side of the next creek, where they dip down to the fjord. The Upper Tillites no doubt follow this anticline, even though this is not readily observable owing to the many clefts cutting into the wall. However, if we follow the orange-rose-coloured band which here marks the upper boundary of the Upper Tillites, the arching will be readily recognisable. In the air photographs it can be ascertained that on the plateau, near the steep descent of the fjord wall, there is another small relict of the Upper Bastion Formation. Tectonically it belongs to the western slope of the western anticline, which, as will be shown below, has an axial pitch towards the NNW.

The arching of the western anticline is very beautifully developed at the north coast of Antarctica Sund. The arch is remarkable in the two light-coloured Bed groups 15 and 17, but the Lower Tillites, also, distinctly mark the anticline, at any rate its crest and west side. They build up the uppermost steep rock wall of the fjord at ca. 1000 m above sea-level, subsequently descending gradually towards the level of the fjord, which they reach ca. 8 km southeast of Kap Mohn. During this descent the Lower Tillites show a peculiar folding: only small waves of a length of ca. 600 m and a height of 15 m; these are the maximum amounts; the individual wave lengths naturally vary within certain limits.

2. Suess Land.

The western anticline is likewise well developed at the south side of Antarctica Sund (sections 12, 13). The next mountain to the west of Mountain 1, is marked by the figure 2. On either side of this mountain a glacier descends through deep ravines to the valley. The culmination of the anticline occurs almost in the region of the western of these two glaciers¹). The arch is most strikingly marked by Bed group 17. As the series strike obliquely to the coast of the fjord, the thickness, especially of Bed group 17, appears greater than it is.

Immediately east of the mouth of the gorge, Bed group 15 emerges from the fjord, it likewise reaches its highest point in the gorge, and then begins to descend again. About 1 km west of the gorge, however, the course of the beds is cut across by a N—S-fault. The west side is so much elevated that Bed group 15 lies on a level with Bed group 17. Owing to the fault, the uppermost portion of Bed group 13 with the yellow dolomite bands is just visible at the bank of the fjord. Upward from the fjord the beds normally follow along the north wall of P. 1840.

¹) The direction of the lower part of this glacier ravine is actually approximately SSE—NNW.

At the top the lower two-thirds of the Upper Tillites occur. Descending towards the mouth of Langgletscherdal, the Lower Tillites exhibit the same small waves as in the corresponding section north of the sound.

As stated above, north of Antarcics Sund Bed group 15, and at the south shore Bed group 13, occur *in situ* at fjord-level. Further, the highest beds on the plateau around P.1519 are those of the Bastion Formation, while on Suess Land the Upper Tillites are elevated up to more than 1840 m above sea-level. On this account there is a distinct rise of the axis of the anticline towards the south.

As to the fault in Carissima Dal the following features can be observed in the area of the western anticline: It crosses the glacier between Mountains 1 and 2 at a height of about 650 m above sea-level. Further, it must rise above the talus hills at the eastern slope of Mountain 2, for here, also, there is a similar platform to that at Mountain 1. On the platform the yellowish-orange-weathering dolomite band occurs *in situ*, while immediately at the northern edge of the ice cap, Bed group 17 forms the crest of the anticline, which is somewhat broader here. The amount of dislocation is thus only about 400 m. A creek, which possibly marks the continuation of the fault, runs west of Mountain 2 down to the glacier. At the eastern slope of P. 1840 a minor displacement is visible in the Lower Tillites. We assume that this is the westernmost indication of the fault of Carissima Dal, for neither in Langgletscherdal nor in Nanortalikdal were any traces of it observed.

At Mountain 2 the above-mentioned Bed group 17 forms another distinct arch, though it is much flatter than north of the fault of Carissima Dal. In the region around Points 1875 and 2000 the ice, so to speak, prevents all observation. There is no doubt, however, that the arch of the western anticline is extremely flat here. The fact that the lower boundary of the tillites almost reaches the fjord at the mouth of Langgletscherdal, while in the surroundings of P. 1330 it is elevated about 1200 m above sea-level, distinctly characterises the flatness of the anticline. Thus, in the southwestern part of Suess Land it is difficult to ascertain where the western anticline reaches the coast of Kempes Fjord. In order to give a picture of the structure of this region, the conditions at the coast of the fjord from Ørkendal to Murgangsdalen will first be described (cf. section 16 and Pl. 4).

PARKINSON and WHITTARD (1931) give a brief description and a sketch of this coast of the fjord. TEICHERT was unable to make a close investigation of the area owing to unfavourable weather (1933, p. 95).

The Lower Tillites and their substratum rising at the eastern slope of Mountain 1254 were described above (p. 77). The rise of the beds (Bed groups 6 to and including the Lower Tillites) towards the west is a characteristic feature of the fjord wall.

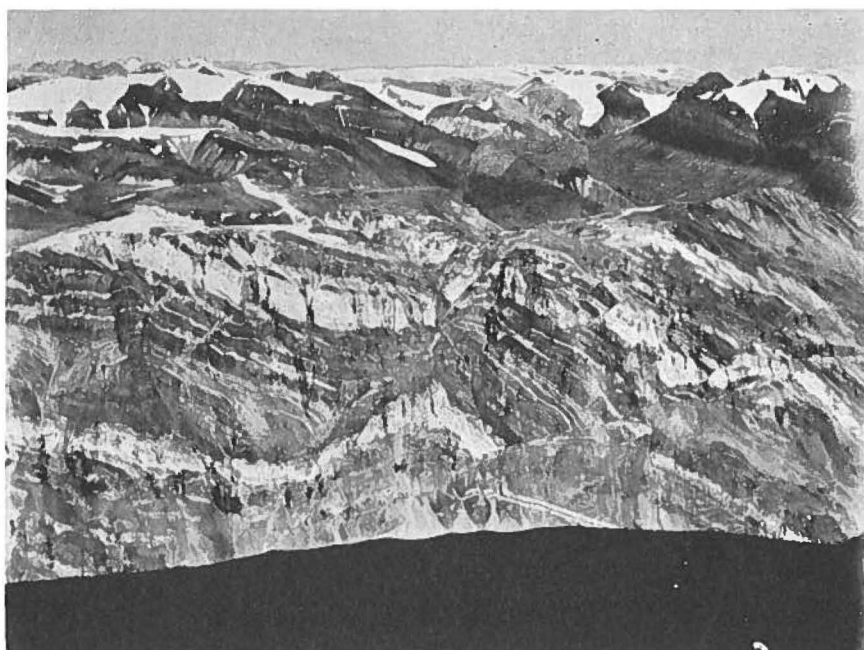


Fig. 18. South side of Suess Land, fjord wall between Ørkendal and Murgangsdalen (cf. Pl. 4); in the background the hindmost part of Ørkendal.—In the left half of the fjord wall there are two distinct light-coloured horizons, the lower of which is Bed group 12, the upper Bed group 15. Not till below the “underthrust” extending obliquely towards the right (p. 81) is Bed group 12 visible again just at the right margin of the picture. The narrow double light-coloured band in the lower middle of the picture is the basal band (here double) of Bed group 6.—BRENNEISEN phot.

About 5 km west of the Ørkendal delta the fjord wall is intersected by a steep and marked creek (fig. 18). Here various faults take their rise.

A fault with a westward dip of 60° runs upward in the creek. It branches out halfway up, in which way the enclosed mass of beds comes to occupy an intermediate position between the two sides meeting at the fault. The eastern of these is elevated about 200 m above the western one. Actually a fault is only present in the upper parts, that is, in the Limestone-Dolomite Series. Below, in the more mobile beds of the upper Multicoloured Series, it vanishes and passes in some measure into a flexure. Thus, in these deeper portions the “flexure” and the lower part of the east side together form a small arch.

The other fault beginning here might be called an “underthrust” towards the east. Its path is distinctly seen in the fjord wall: it emerges from the fjord almost in the core of the aforementioned small arch, rising at first about 20° , subsequently growing flatter again, towards the east. In the south wall of Mountain 1254, at a height of ca. 500 m

above sea-level, it has an approximately horizontal course. The fault shows its greatest amount of displacement in its western part. Here e. g. Bed group 9 rests upon the Lower Limestone-Dolomite Series. In the place where the fault line emerges from the fjord, the basal bands of Bed group 6 forms an arch, which with its crest and west side extends closely below the fault line of these beds.—Farther eastward the displacement apparently decreases, and it may be assumed that the fault dies out below the talus of the southeastern slope of Mountain 1254. Further, it is remarkable that the beds lying below the fault line, especially the Multicoloured Series, dip much more steeply eastward than the Limestone-Dolomite Series separated from them by the fault.

From the two faults together one receives the impression that we are here concerned with a zone of uplift and arching, whose effects are reflected in a nearly vertical fault and in eastward underthrust. It is noteworthy that this deep-seated, fairly considerable, arching loses its effect upwards, owing to the "underthrust". This is an example showing that deeper tectonic elements may be overlain by quite different ones. Moreover this place strikingly illustrates the whole problematic nature of the "folded construction" of the Caledonian sediments.

As this arch is not fully indicated in the contour map (fig. 19), we have completed Section 16 (Pl. 3) downwards as far as necessary to show this feature. It should be pointed out, however, that this was only done in order to illustrate the tectonic conditions, and need not correspond to the actual conditions of the infrastructure.

The continuation of the steeper fault runs, on the peneplaned plateau, in a ravine towards the NNE. Unfortunately its course farther northward could not be followed, owing to the difficulties mentioned above (p. 76); the same applies to the further fate of the underthrust.

About 2 km east of Lumskebugten (in many publications, e. g. in those by TEICHERT and WEGMANN, this bay is called Deceit Bay) another fault is present in the fjord wall. Its plane strikes towards the NNE, and dips 50° , on an average, towards the WNW. The west side lies about 650 m deeper than the east side. In the air photographs the continuation of this fault can be traced as far as the ridge of P. 1690. Probably the throw decreases with increasing height and extension towards the north.

The strike of the series, in so far as it could be inferred from the construction of the sections and the tectonic map (fig. 19), on the plateau of Gravhøjen changes from a NNE to a NNW direction. This direction can also be observed in Murgangsdalen.

The eastward dip is smallest in the region of P. 1068 east of Murgangsdalen. In the very bottom of the valley, however, an eastward dip of 25° can again be measured, which evidently corresponds to the west side of the western syncline. If there is a southward continuation

of the western anticline, the flattening of the beds at P. 1068 might be the last faint indication of it. The assumption seems more correct, however, that this tectonic element dies out in southwestern Suess Land. For it is hardly correct to regard the zone of arching with the "under-thrust" as a southward continuation of the western anticline. Against this tells the extremely wide-spanned arch, distinctly distinguishable in the air photographs, which indicates the Lower Tillites along the whole north slope of the passage which, in part glacier-filled, leads from Ørkendal north of Points 1690 and 1496 into Murgangsdalen.

3. Summary.

The western anticline extends, with a constantly rising axis, as a comparatively broad arch from Kejser Franz Josephs Fjord to Ant-arctics Sund. On Suess Land it is in part somewhat elevated by the dying-out fault of Carissima Dal. Then the anticline grows constantly flatter, dying out in southwestern Suess Land in the west side of the western syncline.—A zone of arching in the south wall towards Kempes Fjord gives rise to a fault dipping about 60° towards the west and an underthrust of upturned Multicoloured Series below less inclined Lime-stone-Dolomite Series towards the east. Near Lumskebugten, along a plane dipping towards the west, the beds are uplifted about 650 m towards the east, respectively lowered towards the west.—The ascent of the tectonic elements towards the south is also unmistakable on Suess Land. Along the whole distance from Kejser Franz Josephs Fjord to Kempes Fjord the axial rise averages $3\frac{1}{2}^\circ$.

IV. The Western Syncline.

The western syncline is not markedly developed in our area of investigation. Only a small fragment of it is present on Ymers Ø. On Suess Land, in the area between Langgletscherdal and Nanortalikdal, the syncline is most distinctly defined, but in Murgangsdalen it merges with the western anticline into one limb rising towards the west.

Our description falls into a section on Ymers Ø and a section on Suess Land. In a third section the area of Kap Mohn and the various N—S-faults are dealt with.

1. Ymers Ø.

The area of Kap Mohn is separated from the remaining part of Ymers Ø by a large fault (sections 5, 6, 7). Both towards the NE (along Kejser Franz Josephs Fjord) and the SE (Antarctics Sund) a com-

paratively broad zone of horizontal beds extends along the fault. In both zones the beds belong to the Tillite Series, but while at Kejser Franz Josephs Fjord the Lower Tillites disappear below sea-level, at Antarctica Sund they are visible in their whole thickness. Thus there is a slight rise of the axis towards the south.

At both coasts of the fjord, in the immediate proximity of the fault just mentioned, the Tillite Series shows a slight upward bend towards the west in relation to the fault (not indicated in sections 6 and 7).

2. Suess Land.

In the mountain marked by the height figure 1044, viz. Trugbjerg, the western anticline is well developed (sections 8 and 9). The top of the mountain lies, as it were, in the core of the anticline, and the beds on either side are bent almost symmetrically. At the bottom of Langgletscherdal runs a faintly developed N—S-fault, which has elevated the west side about 150 m, but seems to die out towards the south. As far as the Nanortalik rivulet the normal beds up to the Lower Limestone-Dolomite Series follow below the tillites of Trugbjerg.

Southward from Trugbjerg the Lower Tillites rise slowly along the east side of Nanortalikdal. This ascent is temporarily interrupted from km 3 to about km 6 in the valley by an area of almost horizontal bedding. At P. 1330 the Lower Tillites disappear, to appear again in the south at the mountain marked by the height figure 1880. South of the glacier descending from P. 1920 to Murgangsdalen the lower boundary of the tillites turns towards the ESE, as a consequence of the continuous rise of the axis. With the exception of a few disintegrated portions, Bed group 17 can be traced from Nanortalik to Gravhøjen. This group, also, marks the rise towards the south. Owing to the diminished eastward dip of the beds in the area of P. 1068, described on p. 82, Bed group 17 here shows an indentation in the territory.

In Murgangsdalen itself the Upper Quartzite series constitutes the lower edge of the eastern wall of the valley, from Lumskebugten to about km 3; then the valley bottom is crossed by some ridges of an approximate direction of NNW—SSE. They are principally made up of the basal quartzites of Bed group 4.

The whole side of the mountain west of the Nanortalik pass and Murgangsdalen was not included in our investigation. To us it was only of importance to ascertain that at the slope that rises west of the highest point of the pass, Bed group 2 could again be identified, dipping eastward approximately with the inclination of the slope. These beds occur without any fault, below Bed group 4 in the second eastern tributary valley to Murgangsdalen.

This is of importance for the fault of Nanortalikdal, which runs almost along the Nanortalik rivulet, and will be described in the next section. Here it will suffice to assume that the portions situated west of the fault, especially at the northern mouth of the valley, have been uplifted to a fairly considerable extent. Immediately west of the Nanortalik delta the white quartzites of the base of Bed group 4 occur *in situ* with an eastward dip of 50° , on an average. The succeeding beds to the west, of the Lower Quartzite Series, show distinct indications of contact metamorphism. In the NE-wall of Scheeles Bjerg, near the mylonite zone bordering the actual crystalline rocks, the partially reddish quartzites of Bed group 2 can be identified with some certainty.

As to the contact phenomena found west of the depression Nanortalik-Lumskebugten, the reader is referred to the investigations carried out by BACKLUND (1932), WEGMANN (1935a), and HUBER (1950).

3. Kap Mohn and the N—S-faults.

As stated on p. 83, a large NNE—SSW-fault (sections 5, 6, 7, 8) runs east of the Kap Mohn mountain, P. 828. To the south its course is concealed by a N—S-extending ravine, it crosses the saddle west of P. 828, and reaches Kejser Franz Josefs Fjord a short distance beyond the first small stream draining the glacier northeast of Kap Mohn. The fault plane is almost vertical. At Antartics Sund the throw amounts to ca. 1500 m, the Lower Tillites to the east lying at the level of Bed group 12 to the west. This throw decreases rapidly towards Kejser Franz Josefs Fjord. In the northern part of the depression west of P. 828 the Lower Tillites are still distinguishable. Along a small creek which marks the course of the fault northward as far as the small glacier stream, the multicoloured shales of Bed group 19 show repeated dragging. North-east of the fault, as stated on p. 83, the lower part of the dark shales occurring between the tillite horizons, is found at sea-level. Thus the throw of the fault is here only about 400 m.—A small, almost due N—S-striking fault runs at the western slope of the Kap Mohn mountain. It dips ca. 60° towards the east, and elevates the west side nearly 100 m higher.

The rocks building up Kap Mohn—they reach from the Lowermost Multicoloured Series to the Lower Tillites—have to a great extent been transformed by contact metamorphism. Especially at the spur which projects into the sound 2 km southeast of the actual cape, magnificent marbled coarsely crystalline dolomites and limestones occur. TEICHERT (1933, p. 91) erroneously regards these rocks as quartzites. Which beds are actually the deepest-lying beds at Kap Mohn, is not quite clear, but probably they belong to Bed group 9 or 10. Unfortunately we were unable to visit the tillite locality just described west of the large fault.

It is quite probable that this is likewise slightly metamorphosed.—A large number of lamprophyre veins (cf. RITTMANN 1940) penetrate the rocks at Kap Mohn, now running parallel with, now unconformably with the bedding.

It was shown above (p. 85) that the ravine of the Nanortalik rivulet harbours a large fault. At the western margin of the delta the base of Bed group 4 occurs *in situ*, at the eastern margin probably the lower part of Bed group 16. The throw thus amounts to nearly 2300 m. Almost at km 3 a tributary stream debouches, coming from the west (the second one reckoned from Nanortalik). At its junction with the Nanortalik rivulet, yellow dolomites with algæ and reddish quartzites, highly tectonically disturbed, are found in the river bed; no doubt they are those of the uppermost part of Bed group 13. However, some distance farther up the tributary stream the eastward dipping quartzites of Bed group 4 are again met with. Thus the fault has left the bed of the Nanortalik rivulet, and here attains an amount of dislocation of 1700 m. As stated above, no indication of a continuation of this fault could be ascertained in Murgangsdalen at the place where it trends towards the west. It seems to have died out here.

The direction and mode of moving of this Nanortalikdal fault allow us to combine it with the larger ones of the Kap Mohn faults. Thus this fault has its greatest throw at Nanortalik and dies out both towards the north and the south.

4. Summary.

On Ymers Ø only the bottom of the western syncline is visible. On Suess Land, east of Nanortalik, however, the syncline is distinctly developed, but then apparently vanishes towards the south during the general western rise of the series at Lumskebugten.—A marked fault separates the Kap Mohn mountain from the remaining part of Ymers Ø; the contact-metamorphic rocks of Kap Mohn have been uplifted to a 1300 m high level. The fault continues in Nanortalikdal, reaching its greatest throw, 2300 m, at the debouchure of the valley into Antarctica Sund. The fault seems to die out both in Kejser Franz Josephs Fjord and towards Murgangsdalen.

V. General Remarks on the Tectonics.

1. Summary of the tectonic observations.

We have now traced the course of the individual folding zones distinguished by WEGMANN through our working field and examined their form. The general idea thus obtained is represented on the structural map (fig. 19).

If we look at this map, one feature will be remarkable: the character of the folding zones in their vertical projection offers not the picture of mountains formed by folding in the usual sense. The border anticline shows as it were, throughout its whole area a sharp divergence in the strike of its two sides. The isohypsometric lines of this anticline spread out in an almost fan-shaped arrangement, often forming an angle of over 90° . To the north (Noa Dal), the axis of the arch lies, about 6,500 m higher than on Ella Ø.

The central syncline is very narrow in its northern section, corresponding almost to a furrow between two uplifts. It is not till south of the fault of Carissima Dal that it increases in width. It culminates in Antarctica Sund, subsiding both northward and southward.

The western anticline likewise flattens towards the south and is here united with the western syncline into a common westward rising isoclinal limb. Both elements rise considerably and increasingly towards the south.

Thus, there is one single line along which the structures over a wide distance have a common direction. This line extends from Kiledal right across Ymers Ø to Niviarsiat, and thence in a slightly eastward convex curve to Kap Oswald and Narhvalsund.

The faults are very significant in regard to their effects. The faulted blocks forming northward descending steps along numerous E—W-faults are characteristic of Noa Dal. This lowering by steps, reaching from the south side of Noa Dal to the coast of the fjord north of Chokoladebjerg, amounts to nearly 3,500 m. The northernmost level of the base of Bed group 18 would accordingly lie at 6,500 m (cf. Structural map fig. 19).

On p. 48 it was demonstrated that the horizontal component of the E—W-faults at Barrieren north of Duséns Fjord and at the course of the main fault can be ascertained. This horizontal component is inconsiderable, attaining an order of magnitude of possibly 100—200 m. The chief movement of the E—W-faults in Noa Dal was no doubt vertical. The amount is often altered according to the dip of the blocks in contact with each other, still all the E—W-faults show a tendency to die out towards the east. Their reactivation is younger than the main fault and the Barriere fault.

The main fault (HV) has almost a NNW—SSE-direction, and at the north coast of Gunnar Anderssons Land and at Rumpen lowers the eastern part about 3,500 m. Between these two points its throw in places amounts to only ca. 1,700 m. Parallel to the main fault runs, to the east, the Barriere fault with a corresponding throw of almost 1,000 m. It seems to be connected with the main fault, and to complement this in the area of the diminished throw.

BÜTLER already demonstrated that the main fault cuts across the Devonian, also. On his geological map (KOCH 1950, pl. 7) the Devonian basal conglomerate is indicated at P. 1723 on either side of the main fault. This shows a minor amount of displacement of the main fault in the Devonian series. The main fault is accordingly to be regarded as initiated in pre-Devonian time and reactivated in post-Devonian time. This reactivation may have taken place in one of BÜTLER's early Devonian phases. The small E—W-thrust along the E—W-faults in Noa Dal may probably be referred to BÜTLER's last phase, the Ymers Ø phase. However, these E—W-faults also attained their main dislocation in pre-Devonian time.

Another kind of E—W-faults occur repeatedly south of Noa Dal (Margeries Dal—Fladedal, south of Munotbjerg, Carissima Dal, south part of Ørkendal). It is, in regard to the actual folding (cf. p. 52), syn- or early tectonic faults along which the folding zones change their directions of strike, heights, and widths. Their presumed age is determined on the basis of these characters. However, for these, also, or for part of them, a post-Devonian reactivation is not excluded.

Connected with the westward bend of the western syncline is a fault which extends from the region of Kap Mohn southward into Nanortalikdal. It attains its largest throw at Nanortalik itself, and seems to die out both towards the north and the south.

In the detailed description, the part played by the various faults at the north coast of Kempes Fjord, between Lumskebugten and Ørkendal, was also treated. They testify to an expansion and a secondary arching of the whole eastward dipping series. In view of their position in the total picture of the folding, we regard these faults, also, to be of pre-Devonian age.

The crowd of N—S-faults in the east side of the border anticline on Ella Ø was likewise more closely dealt with and figured in the detailed description (pp. 63—66). Here, also, there is a difference in their effect

Fig. 19. Structural Contour Map. The structures indicate the upper boundary of the lower limestone-dolomite series. They were constructed on the basis of the sections in Pl. 3, but additional sections, not represented in Pl. 3, were necessary. The object of the structural map is first of all to show the style of the folds, accordingly it does not give the actual course of the horizons selected. Thus, all the folds and faults were projected upwards, respectively downwards,—often several thousand metres. These structures might actually, owing to disharmonies of the interposed series, show a very different picture.—In order to provide a better general picture, the synclinal fault extending SSE-ward from Kiledal (see Pl. 2) was likewise omitted. The faults on Ella Ø, which dip slightly towards the east, were, for the same reason, drawn as vertical fault planes. Legend: B = Barriere fault; HV = main fault; 1, 2, 3, 4, 5 = E—W-faults in Noa Dal; I, II, III, IV = border anticline, central syncline, western anticline, western syncline. Heights in hundreds of metres.

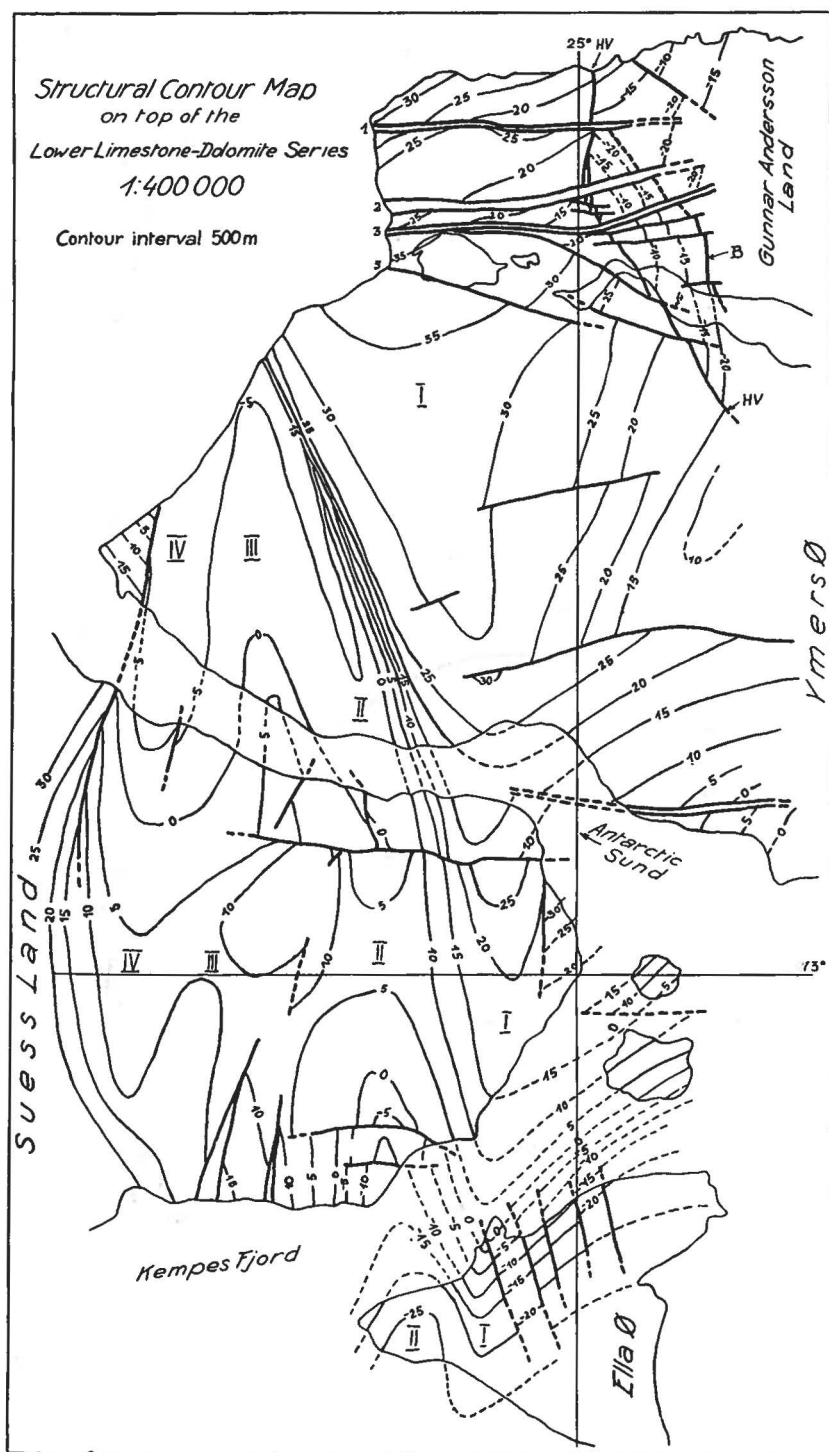


Fig. 19.

in the pre-Devonian sediments and in the Devonian conglomerates. Thus, for them, also, a pre-Devonian initiation and a post-Devonian reactivation may be assumed.

The lamprophyric veins (p. 97) were determined by RITTMANN as associated with the Caledonian metamorphism. Such veins were found which are associated with faults. Thus, for instance, at the N—S-fault south of the eastern mouth of Carissima Dal. This fault may accordingly be regarded as Caledonian.

It is obvious that most of these age discussions as regards the faults lack the only conclusive criterium: the covering by younger sediments. Where this is the case (Devonian conglomerates and sandstones), one is still confronted with great difficulties; for the Devonian sediments often form extensive talus slopes. Owing to the lack of continuous and clearly definable horizons no measurable throws were met with.

2. Analysis of the tectonic events.

BÜTLER (1948, p. 91) compares the tectonics of the Caledonian sediments with the thrust folds of the Jura Mountain type. We do not think, however, that BÜTLER would employ this comparison with all its consequences, but that he merely compared the outer features of the folds with certain areas of folded beds like those of the Jura Mountains (in contrast to e. g. overthrust folds on a large scale).

To gain an idea of the shortening of the sediment complexes solely due to folding, we measured some sections. Sections extending, as far as possible, through all folding zones were selected. The following sections were measured:

Section No.	Length of section l	Length of section f ¹⁾	$\Delta = f - l$
	in km	in km	in km
7.....	33.5	36	2.5
8.....	31	33	2.0
12.....	22.4	23.9	1.5
13.....	23.3	24.5	1.2
14.....	31.9	32.8	0.9
15.....	22.4	23.1	0.7
16.....	23.1	23.5	0.4

A striking feature of this table is that the shortening decreases from north to south. In order, however, to obtain shortening values, which are possibly independent of the length of the sections, we calculated the quotients of the shortening Δ and the section length l. This gave for the individual sections:

¹⁾ Measured at the base of Bed group 18.

Section No.	$\frac{\Delta}{l}$
7	0.0746
8	0.0645
12	0.06695
13	0.0515
14	0.0282
15	0.03125
16	0.0173

From this table the diminished southward shortening of the series owing to folding is evident for Ymers Ø and Suess Land. The noteworthy difference between the two sections 12 and 13, only 5 km distant from one another, can be ascribed to the effects of the fault of Carissima Dal.—From a mean length of the sections of 27.2 km we get a mean shortening quotient of 0.05262 or an average shortening of 1.43 km. Thus, for a section of a length of 25 km the shortening is, on an average 1.316 km ($\approx$ 5.3 per cent.)¹⁾.

The decreasing southward shortening on Ymers Ø and Suess Land, in connection with the folding structure, however, in our opinion is an indication of the influence of post-Caledonian movements on the folding of the beds within our area, as will be shown below.

BÜTLER has treated the connection between the Devonian E—W-thrusting and the deformation of the border anticline (1948, p. 134). According to his observations, the thrusting was slighter north of Duséns Fjord than south of this fjord. It also affected the pre-Devonian substratum. The result of this in the area of Noa Dal is the slighter westward dislocations along the E—W-faults. South of Noa Dal the border anticline was tilted towards the west (Kiledal). With the axial descent towards the south the deformation of the border anticline was likewise successively diminished. In the area east of Ella Ø, the deepest axial depression, the Devonian thrusting first of all gave rise to an overthrust

¹⁾ Quite quantitatively there is here a distinct difference from the folding structure of the Jura Mountains: Buxtorf ("Prognosen und Befunde beim Hauensteinbasis- und Grenchenbergtunnel und die Bedeutung der letztern für die Geologie des Jura gebirges". Verhandl. Naturf. Ges. Basel, Bd. XXVII, Basel 1916) estimated the shortening of the folded Jura Mountains in the area of Grenchenberg at 22.6 per cent. in a section 36 km long.—CADISCH ("Die Entstehung der Alpen im Lichte der neuen Forschung". Verhandl. Naturf. Ges. Basel, Bd. LIV, Basel 1943) gives the following figures: Section Blochmont-Grenchen: shortening 25 per cent.; Section Rheinfelden-Aarburg (extensive overthrusts): shortening 38 per cent.—In other respects, however, the Jura differs so greatly from the East Greenland Caledonian (thickness of the folded series, tectonics of the infrastructure) that these two mountain chains cannot be regarded as comparable phenomena.

in the Devonian series itself without strongly influencing the border anticline. This much according to BÜTLER's statements.

It is no doubt correct that the greater part of the shortening of the folding, initiated already in pre-Devonian time, was due to the Devonian thrusting. This late and post-Devonian shortening cannot be estimated in figures. In our opinion, however, much more than half the amount of the total shortening is due to it. We base this assumption on the differences in shortening in the sections given above.

Unfortunately the measurements of the shortening in the E—W-sections through Ella Ø give no reliable values which can be compared with the above measurements. For in sections 17 and 18 only two of the four tectonic elements (border anticline and central syncline) are included. Moreover section 17 runs east of Point 402, largely following the strike of the beds. It is a fact, however, that the folds of the pre-Devonian sediments on Ella Ø were reactivated by late- to post-Devonian thrusting. We call attention here to the basic vein near Bastionen in Narhvalsund, which after the intrusion was dislocated by thrust-folding at the bedding planes. Moreover BÜTLER (1937) states: "In den unteren Teilen der Devorkonglomerate findet man am Narhval Sund Schieferlagen, die wie die Konglomerate bis zu 40° gegen E geneigt sind. Glätten wir die Inselantiklinale um diesen Betrag aus, so bleibt ein breites, ziemlich flaches Gewölbe übrig: die ältere, kaledonische Antiklinale. Wir müssen annehmen, dass die starke Aufrichtung der Falte erst während der Ablagerung der unteren Devonschichten erfolgte. Der grobe Schutt, der in den Konglomeratmassen an der Basis der Devonablagerungen angehäuft wurde, lässt ebenfalls vermuten, dass zu dieser Zeit, wahrscheinlich anfangs Oberdevon, ein grosser Teil der ostgrönländischen kaledonischen Falten eine Verjüngung erfuhr. Die oberen Lagen der Konglomerate dagegen liegen ziemlich flach und überdecken bereits wieder den abgetragenen Gewölbescheitel".

We agree with these statements. It appears from them that the pre-Devonian folds on Ella Ø were more influenced by the Devonian movements than those on Suess Land. The aforementioned (p.91) steady decrease of the shortening of the folds from the north (about the centre of Ymers Ø) towards the south (southern Suess Land) is thus interrupted in the region of Kempes Fjord.

On the basis of these considerations we arrived, in contrast to BÜTLER, at a somewhat modified conception of the reciprocal action of the Devonian E—W-thrusting on the western Devonian border areas and the adjoining Caledonian folding structures. As shown above, BÜTLER mentions the N—S-descending axis of the border anticline as the cause of the following facts: Overtilting of the border anticline owing to the transfer of part of the Devonian thrusting to this anticline; overthrusting

within the Devonian east of the deepest axial depression of the border anticline with only slight deformation of the anticline itself.

Furthermore, BÜTLER writes (1948, p. 134): "Der Widerstand, den das kaledonische Vorland dem spätdevonischen Zusammenschub bot, beeinflusste weitgehend die Entwicklung der Strukturen am W-Rand der devonischen Senke". This is undoubtedly true. In our opinion, however, the height of the axis alone was not responsible for this development. It may be difficult to explain why just east of the deepest axial depression, that is, where the slightest resistance is offered, the overthrust in the Devonian attains its greatest width. We are much more inclined to think that the divergence of the Caledonian initial folding had a decisive influence on the neighbouring Devonian areas and on the varying shortening of the folds in the pre-Devonian rocks.

For if we consider the structural map (fig. 19), it will be ascertained that wherever this divergence of the border anticlinal limbs assume extreme forms, some reaction has set in.

Gunnar Anderssons Land: West of the main fault the beds for the most part strike ENE. Thus here, also, the east side of the border anticline presented obstacles to the Devonian thrusting from the east. At any rate, this was not particularly intensive here (BÜTLER 1948, p. 134), still it was able to activate minor westward movements along the E—W-faults (cf. p. 48ff.). (The SE—NW-directions of strike between the main fault and the Barriere fault resulted from the inclination of this "Scholle" (block)).

Ymers Ø and Suess Land: A marked divergence occurs in the area of Margeries Dal and Munotbjerg. The Devonian thrusting activity resulted in the overtilting of the border anticline in Kiledal, and had one further consequence. The fault extending right across Munotbjerg and through Margeries Dal, acts as a relaxation for the beds which south of it strike almost asymptotically towards the east. Of the same character are the faults at the southern edge of Munotbjerg and that of Carissima Dal: the northern, diverging, side is constantly forced westward. Still, it is here rather the faults which reacted to the Devonian thrusting; the border anticline has only suffered little deformation.

Ella Ø: Here, again, the limbs of the border anticline move markedly away from each other. The Devonian thrusting towards the west here had somewhat different consequences from those farther northward. Part of it transferred the huge mass of Devonian conglomerates in the southeastern area of the island to the border anticline, thus causing a number of phenomena as described by BÜTLER (1937): the thrusting had a considerable SE—NW-component. This caused a depression and accentuation of the arch. Further, the pre-Devonian N—S-faults were the subject of a certain reactivation in regard to a slighter transverse

movement. This is the influence of the Devonian thrusting activity on the pre-Devonian area. In the Devonian sediments themselves, however, appear the up- and overthrusts of Angelins Bjerg, Rødebjerg, Svedenborgs Bjerg and Kongeborgen. Thus, this upthrust does not seem to be a consequence of a slighter resistance of the border anticline, but a very marked counter-action of the ENE-striking beds of the anticlinal limb to the Devonian thrusting.

Quite mechanically considered, it seems likewise logical that the divergent folding structures are most intensively effected by a uniform thrust-front in places where they are first (in regard to time) reached by this thrusting, that is, in the places of the greatest divergence. In addition these eastward striking beds stiffen the whole structure against the W—E-thrusting and thus offer an increased resistance to it.

3. Discussion of the age of the tectonic events.

As repeatedly pointed out, it is not possible to give a clear, absolute dating of the tectonic events within our area of investigation. A covering by younger sediments is lacking. We can only make inferences as to the probable sequence of the movements from the manifestations of their influence on each other.

The following deformations belong to the proofs of earlier movements:

1. The main fault on Gunnar Anderssons Land and Ymers Ø.
2. The step-fault in Noa Dal.
3. The faults at Kap Mohn and Nanortalik.
4. Dislocations in Fimbulfjeld.
5. N—S-faults at the eastern mouth of Carissima Dal.
6. Formation of an approximately N—S-running zone of arching (north-western part of Ymers Ø, east side of Suess Land), which led to the formation of the later border anticline. This arching attained its maximum in the north. On the east side of this zone of arching the beds strike almost W—E. On the west side, however, a pronounced line developed, running from Kiledal SSE-ward to Niviarsiat. This line—the later central syncline—is actually only an immense flexure extending from the intensive high area of the arch to the depth of the region southwest of Kiledal—the later western anticline (cf. WEGMANN 1935a, p. 29). With the subsidence of the eastern arch towards the south and the simultaneous slight rise of the western deep area this flexure decreased in "throw".
7. In the western part of Suess Land some accessory folds and tumours are met with, which for instance resulted in the expansion phenomena between Ørkendal and Murgangsdalen.

8. KOCH gave to the whole area of depression between Liverpool Land, Gauss Halvø, and the Vela massif the name of "Royal Bay". The most strongly developed depression of the western part of this immense syncline is found in the region of Ella Ø. This fact is very distinctly illustrated by BÜTLER in a N—S-section (1948, p. 85, fig. 1, I). Although a detailed stratigraphy of all the pre-Devonian sediments south of Ella Ø is still lacking, it can be definitely demonstrated that the Royal Bay depression did not yet exist at the time of the sedimentation of the EBF. The corresponding increases in thickness and transgression phenomena are lacking in these sediments. Still, the westward advance of the exceedingly thick Devonian conglomerates on Ella Ø seems to us to indicate that the outline of the Royal Bay was suggested here even before the deposition of these conglomerates.

The phenomena mentioned here under points 1—8 are not enumerated according to age. It is quite possible, and even probable, that these early tectonic elements are not of the same age. However, we have found no satisfactory reasons for a further distinction according to age within this group of early movements. It can be stated very definitely, however, that the early movements were to a very great extent vertical movements.

In addition to these early movements we think it possible to distinguish a group of younger tectonic events within our area. To this group we refer the following elements:

1. Slight westward thrusting along the E—W-faults in Noa Dal.
2. Formation of the faults in Margeries Dal and the Munotbjerg plateau, the southern part of Munotbjerg, the fault in Carissima Dal, and the faults at the mouth of Ørkendal.
3. Reactivation of the step-faults in Fimbulfjeld.
4. Considerable accentuation of the folding zones, especially of the border anticline (e. g. overtilting in Kiledal and accentuation on Ella Ø).

Now, are there any facts on the basis of which these two age groups can be limited as to time?

As to the first, the older group there are only very few and, above all, no distinct indications. RITTMANN (1940) considered the basic veins repeatedly occurring in the pre-Devonian sediments to be associated with the Caledonian metamorphism of the Vela massif. Such basic veins were found by us at the N—S-faults at the eastern mouth of Carissima Dal. The dyke rocks are here entirely fresh and undisturbed tectonically. —The faults at Kap Mohn and at Nanortalik might be associated with the elevation of the Vela massif or the subsidence of the sediment zone

(cf. WEGMANN 1935a, p. 35).—To us it seems possible that the older, especially vertical movements may likewise be traced back to the effects of the Caledonian metamorphism in the floor of our sediment area.

As to the second, younger, group of tectonic elements, it can be stated. The N—S-directed step-faults in Fimbulfjeld also intersect the Devonian conglomerate, though with a less amount of dislocation than in the pre-Devonian sediments. The same applies to the fault running right across the Munotbjerg plateau and through Margeries Dal. Now, the Devonian conglomerates here to the west, as also pointed out by BÜTLER (1948), are the youngest of their kind. They are the marginal conglomerates, which in the inner Devonian basin correspond to thick sandstone series. BÜTLER's investigations established the existence of quite a series of Devonian phases of movement. Now, it is quite possible that the most intensive results of the younger group of movements in the pre-Devonian sediments (accentuation and shortening of the folding zones, E—W-faults in the Munotbjerg plateau, the southern part of Munotbjerg, and Carissima Dal) were caused by earlier Devonian phases. The westward advancing and spreading Devonian sedimentation of conglomerates on the one hand, and the simultaneously decreasing Devonian movements on the other hand thus also explain the concrete case—of the fault of Munotbjerg plateau—Margeries Dal, where there is a great difference between the throws in the pre-Devonian and the Devonian sediments.

As to the second, younger, group of tectonic events it may be supposed that the events were successive and not spontaneous and strictly limited as to time. The ascertainment of the sequence of their individual elements is therefore much more difficult.

To sum up, there are thus two large categories of movements:

1. The older, chiefly vertically directed and at the same time in places expanding phase. It may be more or less closely associated with the Caledonian metamorphism of the Vela massif.

2. The younger phase, which chiefly manifested itself by a thrusting movement from the east towards the west. It gave rise to a strong accentuation of the folding zones and various faults reacting to these westward thrusts. It represents the influence of the various late-Caledonian movements on the pre-Devonian sediments, movements which document themselves in the Middle and Upper Devonian Molasse sediments (BÜTLER 1935a, 1940a, 1940b, 1948). As to how far the Hercynian and later movements made themselves noteworthy within our area, every indication is lacking. According to the above statements, their influence would be limited to regional, very wide-spread uplifts and subsidences.

THE VEINS IN THE AMETAMORPHIC ROCKS

These veins, which are of fairly common occurrence in the whole pre-Devonian sediments area, were subjected to a detailed petrographic description by RITTMANN (1940, pp. 146—151). RITTMANN calls these veins laggards of the same plutonic activity as gave rise to the metamorphism of the Caledonian sediments.

Our investigation of the veins was mainly limited to a mapping of their occurrence and an analysis of their relation to the tectonics. However, the literature on magmatic carbonaceous rocks that has appeared since RITTMANN's investigations (VON ECKERMANN 1948) rendered it justifiable for us to compare our veins with these descriptions. We are greatly indebted to E. WENK, Ph. D., Basel, for his guidance in this respect.

BRÖGGER (1921) called the medium- to coarse-grained biotite-carbonatites by the name of Sövite. VON ECKERMANN (ibid.) defined this term more closely, designating the corresponding calcite-rich dyke rocks as alvikite and the dolomite-rich types as beforsite. If the content of carbonaceous matter is below 50 per cent., the rock is called alvikitic (beforsitic).

Our dyke rocks can be subdivided into three groups; the large majority of the veins found belong to the first group.

1. Alvikitic, biotite-bearing lamprophyres, often somewhat tectonically affected.

This kind of dyke rocks were found in the following localities from north to south:

a) As an erratic at the outlet of the northernmost western tributary valley to Margeries Dal. A noteworthy feature of this block are some round inclusions, up to the size of a fist, of coarse-grained red granite. The mineral content of the dyke rocks is as follows: quartz, as a fine-grained mass and as larger crystals with undulatory extinction; biotite: brownish or, if weathered, greenish, in most cases accompanied by ore-grains; chlorite; calcite; apatite. This mineral content is the same in all dyke rocks of the alvikitic group.

b) At the north coast of Antarctica Sund a lamprophyre vein penetrates the Lower Tillites. The place is located ca. 8 km southeast of Kap Mohn, at the point where the lower boundary of the Lower Tillites emerges above the level of the fjord and rises eastward in the west side of the western anticline.

c) In the region of Nanortalik some of these veins occur *in situ*. About 1 km southeast of the trappers' hut three small occurrences, only 10–20 cm thick, are found as conformable sills in Bed group 16. Following the coast of the fjord southeastward from the mouth of the Nanortalik rivulet, we reach, after ca. 2 km, another vein, which, likewise following the strike of the sediments, makes its way into the shaly limestones of Bed group 18. It is almost 1 m thick. The sediments were somewhat transformed at the contact edges during the intrusion, and a white band, 1–2 cm broad, of comparatively coarsely crystalline calcite can be observed.

d) At the east coast of Suess Land, south of the eastern mouth of Carissima Dal, three N–S-faults occur. Associated with two of these, lamprophyres of the above-described type were found. They attain a thickness of 40–50 cm, and are not noticeably tectonically disturbed. Farther south, almost in the area around the 73th parallel, a lamprophyre vein occurs, likewise at a fault. It is only ca. 20 cm thick. Owing to covering by scree, neither the fault nor the vein could be traced towards the north; we assume, however, that it may be connected with the fault south of the mouth of Carissima Dal just described.—In a horizontal direction these veins are distant 25 km from the nearest crystalline beds (Nanortalikdal). Thus they are not entirely associated with the surroundings of the metamorphic EBF (RITTMANN, *ibid.* p. 141), at any rate not in a horizontal direction.

e) A very interesting occurrence of lamprophyres is found southeast of Bastionen, at the northern shore of Narhvalsund. Enclosed in the dyke rock there occur round components, up to the size of a nut, of carbonaceous rocks and sedimentary quartzite. Evidently they have been forced up from the depth. In places the rock thus shows almost a conglomeratic habit. Further, in the higher parts of the fjord wall the vein is post-intrusively displaced at various bedding planes (cf. p.92). The place is located in the west side of the Bastion syncline. Peculiarly enough, however, at this dislocated vein the nearest portion above has been shifted towards the south, that is, in the direction towards the core of the syncline. This is so in spite of WEGMANN's sketch (1935, p. 33, fig. 3), where the nearest bed above in a syncline should have slid outwards and in this way also have enclosed the Devonian remnant at the south slope of Bastionen. Thus, WEGMANN's generalisation, other-

wise entirely logical from a mechanical point of view, cannot be inferred from the special case of the Bastion syncline (cf. also H. CLOOS, "Gang und Gehwerk einer Falte", Abb. 1; Z. Deutsch. Geol. Ges., Bd. 100, Hannover 1950, p. 291).

2. Calcite-bearing pyroxene-biotite-diorite-porphyrite.

This dyke crops out at sea-level in the quartzites and quartz-shales of Bed group 1 in the Blomsterbugt, immediately below the trappers' hut. It strikes N—S, it is ca. 40 cm thick, and subsequently branches out. Mineral content: plagioclase, hedenbergite-like augite, biotite, calcite, and a little quartz.

3. Pyrite-bearing veins with quartz, chalcedony, muscovite.

These veins pierce Kap Mohn in a fairly large number. Now they cut the beds unconformably, now they appear as conformable sills. A similar vein was observed at the lower end of the Nanortalik creek, ca. 800 m south of the delta. It is here associated with a small local fault. According to verbal information from Dr. WENK, these veins should be interpreted as hydrothermal formations.

OBSERVATIONS ON THE QUATERNARY DEPOSITS AND THE MORPHOLOGY

Just as to the geologist, some objects of quite text-book-like distinctness present themselves to the morphologist within our working area. A careful investigation of the Quaternary sediments and the morphology was not included in the task entrusted to us. Still some particularly remarkable phenomena will be mentioned. As to Ella Ø, some observations with relevance to this chapter were reported on p. 66 ff.

1. Latest glaciation.

Evidence of the youngest glacial period are to be found everywhere within the area investigated by us. Crystalline erratics may be met with even on the highest ice-free plateaus (ca. 1,500 m above sea-level). In the depression east of Bastionen on Ella Ø (ca. 800 m above sea-level) some morainic walls are still preserved. Moraines of later glacial stages, in addition to those on Ella Ø mentioned on p. 70, are to be found especially in Ørkendal. Flanking the gorge of the main rivulet, they here rise to a height of about 250 m above sea-level.—Roches moutonnées can be observed everywhere, naturally especially in the deeper regions, where the glacial striæ, also, mostly remain preserved.

2. Elevation of land.

In post-glacial time, that is, constantly following in time the retreat of the ice, an elevation of the mainland took place in East Greenland. This fact is chiefly manifested by two phenomena: old shore-lines and elevated deltas.

Old shore lines occur at various levels. In our area the highest well recognisable shore-lines occur at the eastern slope of Munotbjerg at a height of ca. 700 m above sea-level. Ancient shore-lines are likewise distinctly visible at the west side of Nanortalikdal. One of them is readily traceable from Scheeles Bjerg towards the highest part of the pass of Nanortalikdal. Owing to the great abundance of ancient shore-lines it

would take us too far to mention them all with indication of altitude. Nor is it possible to follow the same level over wide distances, since now one, now another level is most conspicuous, or disappears entirely. Thus it cannot be ascertained, either, whether during the elevation of the land regionally unequal amounts of elevation occurred.

One ancient sea level has, at any rate, left particularly distinct traces. Naturally, it is one of the last, i. e. deepest ones. Thus the smooth rock walls of the coast in the northern part of Kempes Fjord west of Lumskebugten show, ca. 50 m above sea-level, a distinct change in the reflection of the light. This is especially striking, if one looks westward from the mouth of Ørkendal about 4 o'clock in the afternoon. The 50 m high seam appears in the reflecting light somewhat darker than the high rock walls. We ascribe this phenomenon to the difference in the intensity of weathering. To the same level (ca. 50 m) belong some ancient wave-cut notches. They are not readily recognisable in summer, but in winter, with the selective snow-covering, they are fairly distinct. One of these old wave-cut notches extends over a great distance around Kap Alfred (Lyells Land), another was observed on either side of the southern mouth of Margeries Dal.

To the same level (50 m above sea-level) belong, with certain variations, the numerous raised deltas. Only the most beautifully preserved of these will be mentioned: northern mouth of Margeries Dal, Fladedal, Nanortalik, Ørkendal, etc. In all these cases we are not concerned with one single level; above, as well as below, there occur now better, now more poorly preserved delta levels. At the outfall of the lowermost western tributary rivulet into the major rivulet in Ørkendal we observed a still well preserved delta surface at an altitude of 330 m above sea-level.

Especially in bays with gravelly or coarsely gravelly bottoms the very youngest fjord levels are still marked in the form of concentric wave-like or step-like rising structures, thus e. g. on Ella Ø in the bay at the station at Kap Elisabeth, and on Maria Ø.

3. Frozen soil and soil-flow.

These two phenomena are typical of the arctic regions. Frozen soil occurs in our area as polygon soil in a diversity of forms. We refer the reader to the detailed description of an occurrence on Ella Ø on p. 71. Noticeable is also a kind of polygen soil occurring on the high-peneplaned plateaus (e. g. Gravhøjen, Bastionen). The polygonal structures here attain a diameter of up to 2 m. The sides of the polygons are often formed of components of the size of a head, in the interior angular fractions a grain-size of 2—10 cm are often cemented out. Walking

across this polygon soil, one will notice, especially during the melting of the snow, that even extremely coarse polygons with components of over 10 cm in diameter appear to be very elastic and soft.

Soil-flow appears wherever finer morainic material, in addition to the previously coarse material of earlier coast-lines, is mixed with the scree of the mountain slopes.

4. Formation of fjords.

Opinions differ as to the genesis of the fjords. BACKLUND, KULLING, WORDIE, TEICHERT, and WEGMANN have already discussed these questions. We do not venture, considering the limited space of our working area, to take a stand as to any hypothesis in this respect, and we therefore confine ourselves, as a contribution to the further discussion, to report an observation.

Noa Dal has developed along the numerous E—W-faults. Here, also, the formation of the fjord, respectively the valley, has followed tectonic lines. The morphological continuation of Noa Dal towards the west is found in the depression south of Teufelsschloss (Andrées Land).

While in this case, also, the formation of the fjord is evidently associated with a fault scarp, the courses of other fjords seem to be little affected by tectonic lines. WEGMANN (1935 a, p. 44) rightly mentions Narhvalsund, which in part cuts across a horst. Thus, in the question as to the origin of the fjords we will have to look out carefully for all factors of the individual cases. Generalising theories only attain a small degree of probability.

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PLATES

Abbreviations on Pls. 2 and 3.

HV = "Hauptverwerfung" (main fault).

B = Barriere fault.

1, 2, 2a, 2b, 3, 3a, 3b, 4, 4a, 5 = Numbering of the E—W-faults in Noa Dal and in Gunnar Anderssons Land.

Indication of place names on pls. 2 and 3.

- 1 = Midnatspas.
- 2 = Barrieren.
- 3 = Arenaen with Pissevache.
- 4 = Rosinante.
- 5 = Rosinante Pas.
- 6 = Rumpen.
- 7 = Chokoladebjerg (ODELL 1939).
- 8 = Kiledal.
- 9 = Fladedal.
- 10 = Trugbjerg.
- 11 = Langgletscher Dal.
- 12 = Carissima Dal.
- 13 = Nanortalik pass.
- 14 = "Mountain 1".
- 15 = "Mountain 2".

Plate 1.

Stratigraphic column 1:15,000.

We have employed the symbols of LE ROY, "Subsurface Geologic Methods", 2nd edition, Golden, Colorado 1950, p. 859. Different symbols were combined and adapted according to their meaning.—With a view to the condensed representation, the symbols within the individual bed groups are often somewhat schematic. A lithological detail-horizon does not, accordingly, give the exact thickness.—The figures indicating thicknesses give the mean value for the whole area investigated.

Thickness m	Subdivision	Age	Thickness m	Subdivision	Age
100 70 140/210	Lower Tillite			Devonian	Upper Devonian
	20/19			Narhval Sund Form.	
440	18	Upper LDS	600	Kap Weber Form	
75	17				
180	16	Lower LDS	100	Cass Fjord Form.	
75	15		300	Dolomite Point Form.?	
270	14				
			280	Hyalithus Creek Form.?	
300	13	Upper MS	500	Ella Ø Form.	
200	12		170	Bastion Form.	
150	11		50	Sandstone Series	
150	10		25	Spiral Creek Form.	
150	9		250	Tillite Canyon Form.	
30	8				
190	7		140	Upper Tillite	
130	6		340		
150	5		100	Lower Tillite	
380	4				
500	3				
310	2				
	1				

Plate 2.

The pre-Devonian sediments on Ymers Ø, Suess Land, and Ella Ø. 1:250,000.

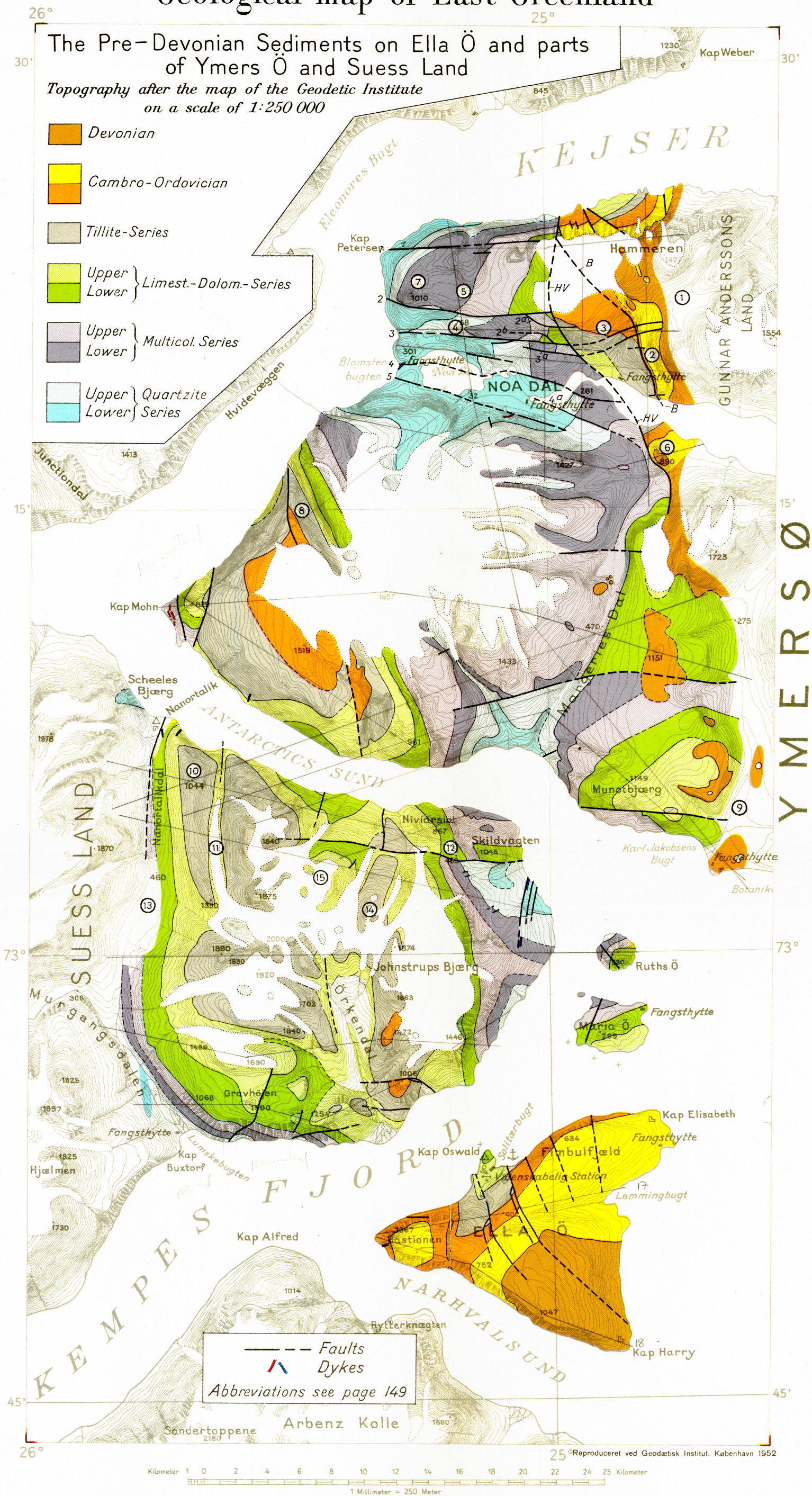
Sections of maps 2 Ø 72 and 2 Ø 73 of the Geodetic Institute, Copenhagen, served as a topographic basis. We express our cordial thanks to this institute for kind help. —Errors in the maps, if they meant obstacles to the indication of the geological feature, were pointed out in the text.—The geological map is “abgedeckt” (solid), that is, the Quaternary deposits and the vegetation are not indicated with the exception of the ice-covering.—The directions of strike, if not appearing from the map, may best be obtained from the Structural Contour Map (fig. 19, p. 89).

Geological map of East Greenland

The Pre-Devonian Sediments on Ella Ö and parts of Ymers Ö and Suess Land

Topography after the map of the Geodetic Institute on a scale of 1:250 000

- Devonian
- Cambro-Ordovician
- Tillite-Series
- Upper } Limest.-Dolom.-Series
- Lower }
- Upper } Multicol. Series
- Lower }
- Upper } Quartzite Series
- Lower }



--- Faults
 / \ Dykes
 Abbreviations see page 149

Plate 3.

Geological Sections 1:150,000.

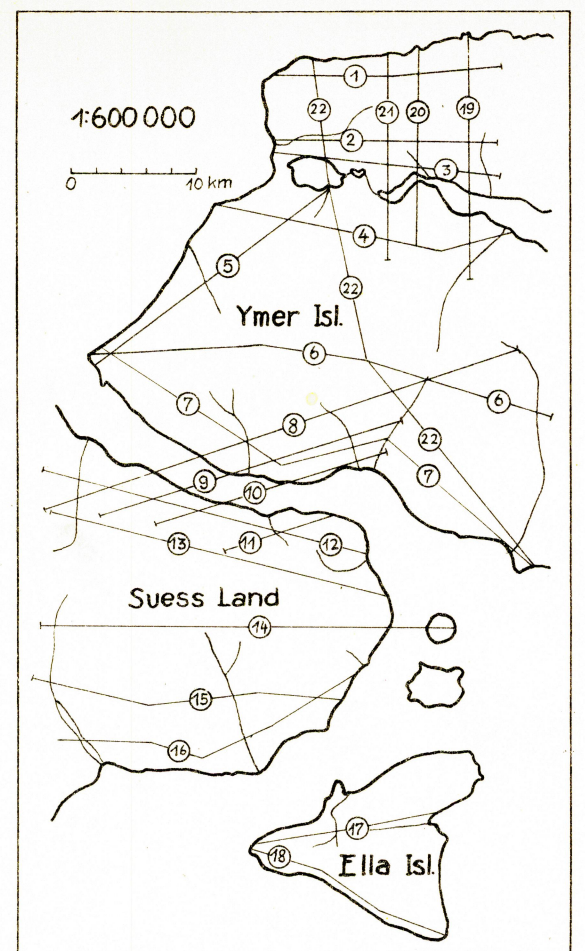
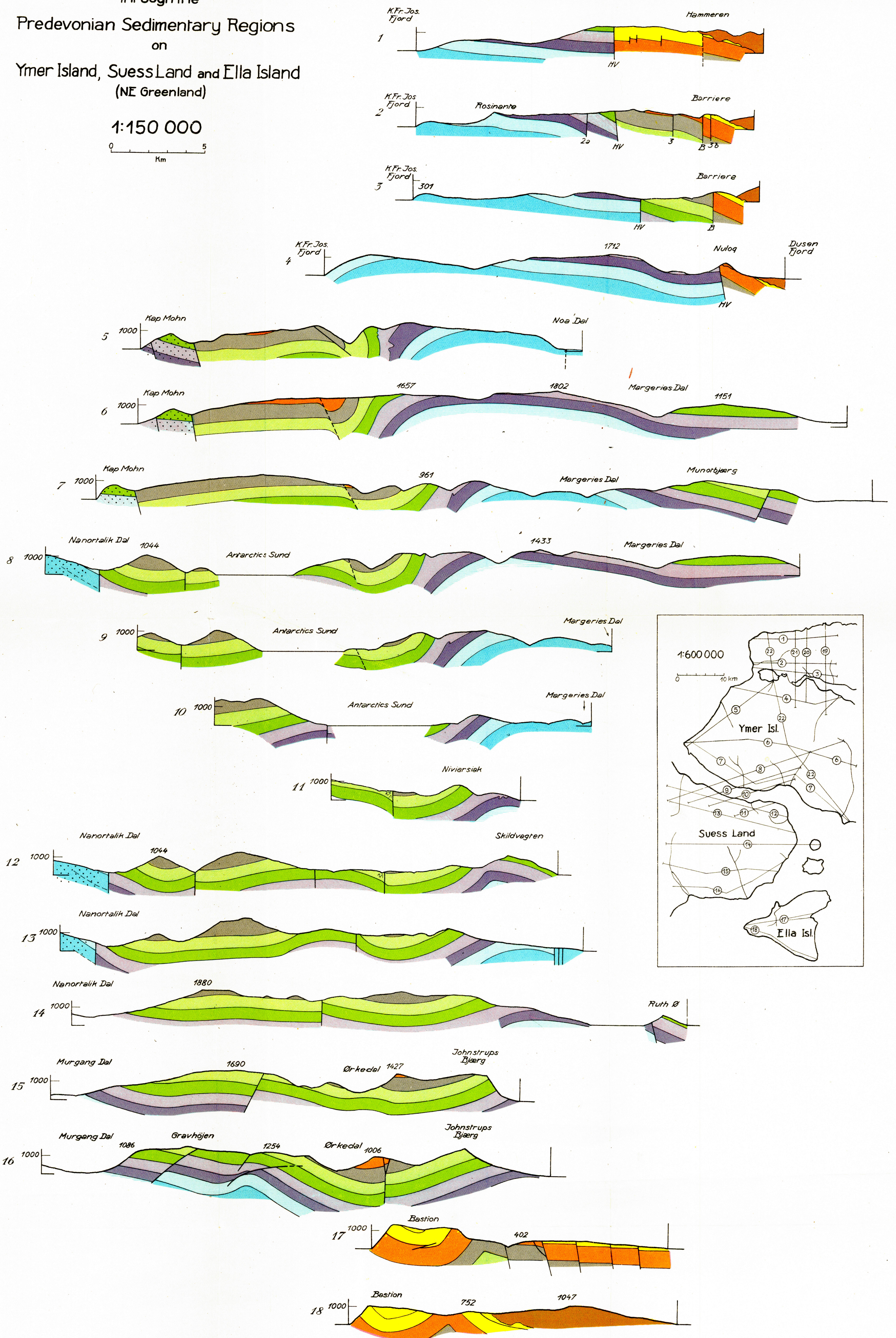
Owing to some errors in the topographical maps, a somewhat deviating thickness of some series had to be given in various sections (e.g. Nos. 2, 6, 7, 13).—Section 16 is too wide towards the depth. This was done in order to give a better representation of the nature of the zone of arching at P. 1254. Complications which might perhaps arise at these depths (metamorphism) were not taken into consideration.—The abbreviations and figures in sections nos. 1, 2, 3, 4, 19, 20, 21, and 22 are the same as in Pl. 2.—The Spiral Creek Formation and the sandstone series are indicated by the same colour as the Lower Palæozoic (orange).

Geological Sections through the Predevonian Sedimentary Regions on Ymer Island, Suess Land and Ella Island (NE Greenland)

1:150 000

0 5
Km

W-E Sections



N-S Sections

