THE GEOLOGY OF THE DALSKOG DALS-ROSTOCK REGION
DALSLAND, SWEDEN

BY

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PREFACE

A cordial and generous invitation on behalf of "SVENSKA INSTITUTET FÖR KULTURELLT UTBYTE MED UTLANDET" brought me for the first time to Dalsland, together with 14 fellow students in geology of Leiden University, in the autumn of 1945. During the field work carried out there under the supervision of the late Prof. Dr T. KROKSTRÖM and Prof. Dr L. U. DE SITTER, I started to survey the region north of Billingsfors. I returned to Dalsland in the summer of 1946. After having consulted Prof. Dr P. GEYER, director and Dr W. LARSSON, geologist of "Sveriges Geologiska Undersöknung", the latter thought it better to abandon the area I had started to work on in 1945 and proposed the Dalskog Dals-Rostock region as being the most interesting part to investigate. I finished the survey of that area in the summer of 1948.

For the financial aid granted to me for the field work by the "N.V. BATAAFSCHE PETROLEUM MAATSCHAPPIJ" and by the "VOLLENHOVEN FONDS" I wish to express my sincere gratitude.

To Prof. Dr L. U. DE SITTER and Prof. Dr E. NIGGLI I am profoundly indebted for the interest they took in the progress of my work and their countless and highly valued advice.

In 1948, for one month and a half, I obtained in the field the assistance of the following students: J. H. ALLAART, H. A. GROEN, T. VAN DER HAMMEN and C. W. SPIELE, in mapping the region north of the line Södra Halängen—Dalskog—Årbol. Miss COR ROEST has drawn the maps and the sections with the greatest care, as well as the figures in the text. The photo-technical knowledge of Mr W. F. TEGELAAR has been of great benefit to me. The numerous thin sections required, have been prepared with much skill by Mr M. DEYN. The rock analyses have been made most accurately by Miss B. HAGEMAN. The "N.V. BATAAFSCHE PETROLEUM MAATSCHAPPIJ" obligingly provided the enlargement of the topographic map. With my colleague Dr A. J. A. VAN OVEREEM I had stimulating discussions about the problems of the geology of Dalsland. The translation of the manuscript into English is due to Mr A. J. DE WITTE, whose knowledge of English geological nomenclature has been of great value to me. Messrs. J. SMITH, J. CARR, C. HICKSON, P. LEICESTER, and Miss E. BRANDON have been so kind to search the English manuscript for linguistic errors. My wife, Mrs. C. W. HEYBROEK—SPIELE and my mother, Mrs. M. HEYBROEK—MEETER, have spent much care and time in preparing the manuscript for print.

Finally, the population in the area of my field work made my stay in Sweden very pleasant by their readiness to help and their hospitality. In this respect I wish to remember particularly: Kapten K. BEURLING and the LINDEBERG family at Dalskog, Mrss. JOHANNSON and LINDBERG at Dals-Rostock and the LINTSKOG family at Molkom.

Without the kind help of all these persons and institutions I could not have achieved this work and to all of them I wish to express here my profound gratitude.

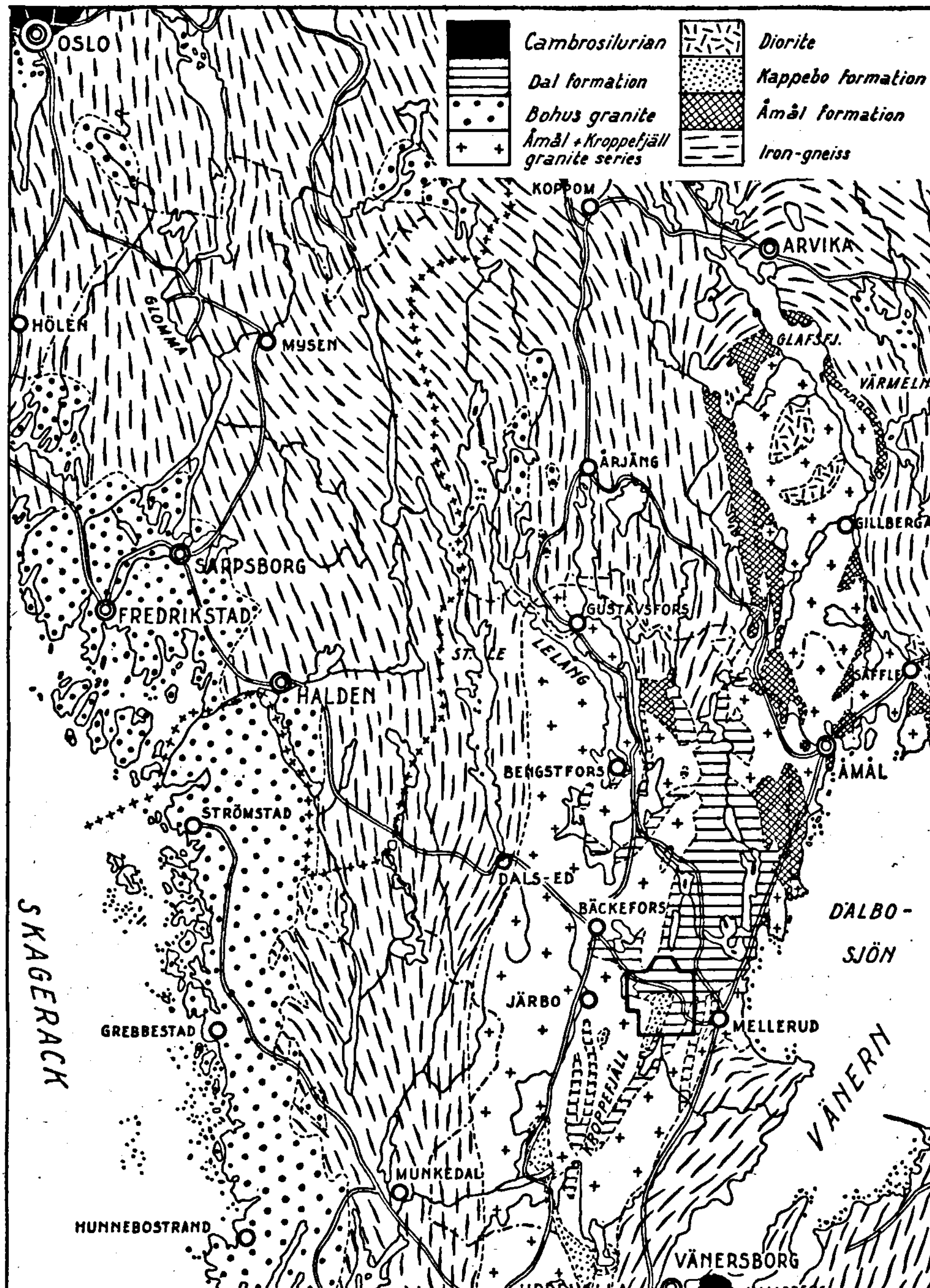

Fig. 1.

Geological map of the region between Vänern and Skagerack after Gavelin and Magnusson (1935), slightly modified. Scale 1: 1.000.000. The Dals-Rostock region is indicated.

INTRODUCTION

Situation and geological setting of the region

The Dalskog Dals-Rostock area lies in the Swedish province of Dalsland, to the west of lake Vänern. It lies entirely within the Upperud sheet of both the topographic (1926) and geological map (1870) and comprises parts of the parishes Gunnarsnäs, Dalskog and Ör.

As shown by the outline map (fig. 1), the investigated region is situated in an area of gneiss-granites and supracrustal formations, which lies to the west of lake Vänern as an island in the great, highly metamorphic complex of gneisses of southwestern Sweden. In the adjoining table the geological events wich left their marks in the rocks of the Dalskog Dals-Rostock area are listed in chronological order. For the sake of clearness the table has been completed with data known from the adjoining regions, but these are placed in parentheses.

The gneiss complex of southwestern Sweden, also called järngneiss (= iron gneiss) complex, is likely to have been formed originally from rocks of different genesis (1) [1]) (MAGNUSSON 1946, p. 191 f. and LARSSON 1947, p. 331—333). As a result of a very strong gneissification (2), in which recrystallization played a prominent part, they were transformed into veined, banded and other gneisses. These rocks will not be further discussed here, as the järngneisses are not exposed in the investigated area. It only may be noted that at a distance of 7 km more to the east, on the shore of Vänern, due north of Mellerud, a banded gneiss was found which is similar to the banded gneiss described by MAGNUSSON (1929b, p. 20) from the gneiss complex of the Säffle region.

In northern Dalsland and southwestern Värmland it may be concluded that, following the gneissification (2) which was probably an effect of mountain building, the gneiss complex of southwestern Sweden was subjected to a long period of denudation (3). After that came the deposition on the gneiss of the volcanic and sedimentary Åmål formation (4). If one feels doubtful about the validity of the age determination of this formation, as I do (cf. p. 136), a study of the literature fails to dispel these doubts. I still think it possible, therefore, that the Åmål formation will prove to be younger than the Åmål-Kroppefjäll granite series.

Provided that the position of the Åmål formation is as indicated on the chronological table, this formation and the underlying gneiss complex were intruded by the Åmål-Kroppefjäll granite series (5), in large areas of Dalsland and adjoining provinces. According to LARSSON (1947, p. 325—326) the series is composed of differentiation products of one magma, viz.

[1]) The numbers in brackets refer to the corresponding numbers of the table.

CHRONOLOGICAL TABLE OF THE GEOLOGICAL EVENTS IN THE DALSKOG DALS-ROSTOCK REGION

23.	Slight erosion; formation of peat	
22.	Regression of the Yoldia sea; strong erosion of loose sediments; formation of peat	
21.	Melting of the ice; the ice recedes before the Yoldia sea; sedimentation of glacial deposits	
20.	Pleistocene ice cap	
19.	Erosion from 17 to 21	
18.	Renewed movements along the big faults in the Vänern area between 17 and 20, Caledonian?	(Kållandsö line) (Dalbobergen fault) Kroppefjäll fault
17.	Sedimentation of the Cambro-Silurian	
16.	Denudation: forming of the sub-Cambrian peneplain	
15.	(Transcurrent faults in northern Dalsland); (Billingsfors dolerite dyke)	
14.	Dislocations along major north-south trending faults in the Vänern area between 13b and 16	(fault separating eastern and western Sweden) (mylonite zones of central Värmland and the Kållandsö line) Kroppefjäll fault (Svinesund-Kosterfjord overthrust?)
13.	Dal orogeny	c. Bäckedal and Lysesjö overthrust b. Extra compression in southern Dalsland a. Simple folding
12.	Sedimentation and volcanism of the Dal formation	f. Liane layers e. Quartzitic layers d. Upper slate layers c. Spilite-bearing layers b. Lower slate layers a. Basal layers
11.	Denudation	
10.	Tectonic movements	
9.	Faulting; sedimentation and volcanism of the Kappebo formation	slate, conglomerate, graywacke, tuff and agglomerate, rhyolite, greenstone, albite-rhyolite
8.	Denudation	
7.	Intrusion of dykes and stocks of diorite	
6.	Gneissification; mountain building	
5.	Intrusion of the Åmål-Kroppefjäll granite series	e. Teåker aplite-granite d. Kroppefjäll granite c. (intermediate granite) b. (Åmål granite) a. (diorite)
4.	(Sedimentation and volcanism of the Åmål formation?)	
3.	(Denudation)	
2.	(Gneissification; mountain building)	
1.	(Formation of the rocks of the järngneiss complex of southwestern Sweden	

from old to young: diorite, Åmål granite, intermediate granite and Kroppefjäll granite. Of all these, only the Kroppefjäll granite is exposed in the area under discussion and further the Teåker aplite-granite which must be considered as the fifth and youngest member of the Åmål-Kroppefjäll granite series.

After the intrusion and the solidification of the granites orogenic forces must have operated, which changed the granites into gneiss-granites (6). This gneissification is distinguished by extreme cataclastic effects in the rock, whereas recrystallization is only subordinate. During this process the Kroppefjäll granite has been transformed into an "augen-gneiss" and, in places, even to a granite-schist. The age of the gneissification of the granites follows from the fact, that the younger rocks of the Dalskog Dals-Rostock area are less metamorphosed and also that pebbles of gneiss-granite in the conglomerate of the Kappebo formation have a random orientation with respect to their schistosity (cf. p. 114). The gneiss-granites are discussed in chapter I.

In and near the investigated area occur dykes and stocks of diorite (7) in the gneiss-granite which, judged from their degree of metamorphism, must be younger than the gneissification of the granite, but older than the Kappebo formation which overlies the diorite (p. 82).

After a long period of denudation (8), which brought the gneiss-granite of the Åmål-Kroppefjäll series over large areas to the surface, another period of volcanism and sedimentation sets in, in which the Kappebo formation was built up. The sedimentation and the volcanism have probably been initiated and accompanied by fault movements. The volcanites of this formation consist of albite-rhyolite, rhyolite and greenstone, with their tuffs and agglomerates. Above all, their chemical composition indicates that they constitute a spilitic series. The sediments consist of poorly sorted graywackes, conglomerates, breccias, siltstones and a few slate lenses. They are essentially composed of detrital fragments of the gneiss-granite. A small amount of material derived from the volcanites is there too and fragments of earlier sediments of the formation itself being also found. A more detailed description of the rocks of the Kappebo formation is given in chapter II.

There do not occur definite layers of a regional extent in the Kappebo formation. All sorts of sediments were deposited right from the beginning and are, in an irregular way, mixed up with the volcanites, as the volcanism and sedimentation started simultaneously. This irregular way of deposition, as well as the poor assortment of grain size and composition of the sediment and the angular form of the grains and fragments are indicative of great topographic contrasts of altitude. These conditions must have arisen from young tectonic movements at the time of deposition. The sedimentation was probably continental. Along and near the few faults, which could be dated as prior to the deposition of the Dal formation, occur dykes of albite-rhyolite and greenstone, which suggests that the tectonic movements, postulated above, were fault movements (9). The Kappebo formation as a whole is treated in chapter III.

Unconformably overlying the gneiss-granite and the Kappebo formation is the Dal formation. This indicates that after the deposition of the Kappebo formation tectonic movements (10), followed by denudation (11), have taken place. Not much can be said about the extent of both events. The Dal formation (12) is constituted by a series, of 2000 m thickness, of mainly sedimentary rocks, in which are intercalated some spilitic effusives. As more to the north in Dalsland, it consists of the following strata:

Liane layers	> 520 m
Quartzite layers	280 m
Upper slate layers	200 m
Spilite-bearing layers	200—300 m
Lower slate layers	350 m
Basal layers	370 m
	> 1920—2020 m

This development shows that the Dalskog Dals-Rostock region is situated in the western part of the sedimentation basin of the Dal formation, where the layers are much thicker than in the eastern part (VAN OVEREEM 1948, p. 105). Compared with the section of the Dal formation between Lilla and Stora Ärven, 12 km more to the north, which has also been deposited in the western sedimentation basin, the lower layers are thicker and the upper ones thinner. The sediments of the Dal formation are fairly well sorted as to content and grain size, which indicates a transport over a long distance. The regularly observed ripple marks, intraformational breccias, and "mud cracks" indicate that a great part of the sediments were deposited in shallow water. The rocks and the sedimentation of the Dal formation are discussed in chapter IV.

The latest orogeny which has affected the region, the Dal orogeny (13), pushed the layers of the Dal formation out of their horizontal position. Three stages which are consecutive, but by no means sharply divided in their occurrence, may be distinguished in this orogeny. After a simple folding (13a) which produced, especially in middle Dalsland, symmetrical anticlines and synclines, southern Dalsland suffered an extra compression (13b). This brought about a number of new anticlines and, more to the south, overturned the layers over large distances. This is probably due to a westward movement of an eastern block (see p. 161). In a subsequent stage the big recumbent Bäcke valley and Lysesjö overthrusts came into existence (13c). They have their roots to the west of the discussed region and tend from there in an eastward direction over a large part of middle Dalsland. From the data now at hand it may be concluded that the length of the front is 28 km and the amount of thrust 14 km.

The curved and approximately north-south trending Kroppefjäll fault, with a throw of 2000 m, cuts through the structures produced by the simple folding (13a) and the extra compression (13b) in southern Dalsland. Its relation to the overthrusts is not known, because it is not in contact with them. Therefore, it cannot be said whether it originated in a late stage of the Dal orogeny or afterwards. The Kroppefjäll fault belongs to the system of faults of the Vänern region (14), of which the Kållandsö line with its prolongation in the mylonite zones of central Värmland and the fault between the geologically different territories of eastern and western Sweden, are the most important. Possibly the Svinesund-Kosterfjord overthrust on the shore of Skagerack belongs also to the same system (SUNDIUS 1944).

In northern Dalsland transcurrent faults (15) are known, striking northwest-southeast, which are decidedly younger than the Dal orogeny, as is the case with the dyke of dolerite near Billingsfors. All the differences of elevation in the topography caused by the Dal folding and later fault movements, must have been levelled by a long period of denudation (16), as a result of which vast areas were degraded to a peneplain. On the peneplain thus formed, the Cambro-Silurian sediments were deposited. For this reason it is known as the sub-Cambrian peneplain.

The Cambro-Silurian sediments (17) have probably been deposited all over Dalsland and the adjoining regions. They are still preserved in the Oslo graben and in the Halleberg and Hunneberg near Vänersborg, whereas at the shore of Vänern near Vingershamn, 20 km northeast of our area, dykes of fossiliferous sandstone of lower-Cambrian age have been found (GAVELIN 1908a). Except in the localities just mentioned, all the Cambro-Silurian has been eroded (19), but the sub-Cambrian peneplain is still plainly recognizable as a "Gipfelfluhr" or in rocky plateaus. However, it has been broken up by renewed movements along the fault system in the Vänern region (18) and as a consequence shows a number of steps. Thus, the ancient peneplain slopes gently upward from the shore of Skagerack in an eastern direction to the Kroppefjäll fault, where it attains a height of 230 m. To the east of this fault the peneplain is cast down 150 m. Similar steps are found on the southwestern shore of Vänern (Dalbobergen fault) and along the Kållandsö line (S. DE GEER 1910). All we can say about the age of the renewed movements along the Vänern fault system is that they must have occurred after the Cambro-Silurian and before the Pleistocene. LJUNGNER (1927—'30 and 1937) supposes that these movements occurred in connection with the Caledonian folding, but other ideas have also been put forward on this subject. The tectonics of the Dal orogeny and of the later movements are dealt with in chapter V.

The erosion (19), which started after the Cambro-Silurian, has probably been active, though in general with moderate intensity, until the retreat of the ice sheet which, in the Pleistocene, covered the land (20). That the ice cap had an important share in the erosion can be seen in the typical, glacial morphology of the bedrock. In chapter VI the relation between the geology and the morphology of the bedrock is briefly discussed.

The retreating land ice made way for the Yoldia sea (21) which, with a level 150 m higher than at present, reached up to the glacier and covered large parts of the land. Only high rock masses and plateaus stood out above the water-level (cf. fig. 28, p. 186). Large quantities of material derived from the glacier have been deposited during the retreat of the ice in the lower parts of the country. Subglacial in origin are probably the coarse deposits of boulders and pebbles alongside north-south trending steep cliffs. Possibly these are praecrags, but they might as well be considered as eskers. Marginal formations, such as terminal moraines and marginal deltas are beautifully developed in the Dalskog Dals-Rostock area. The ridges of terminal moraines owe their origin to a stationary stage of the retreating ice, which marks the limit between the Gotiglacial and the Finiglacial phases. The Yoldia sea soon withdrew again from the area (22) and from then onwards brooks with a capacity many times as great as the present ones, must have had a strong erosive action upon the loose sediments. Its traces are still visible at every turn. In the shallow excavations of the bedrock and in places on the quaternary deposits, where the drainage is poor, a formation of peat (23) started after the regression of the sea, which continues yet today. The present erosion is very weak (23). Chapter VII deals with the quaternary deposits and their morphology.

It will now be briefly indicated how the geological events, outlined above, can be fitted into a more general chronological scheme according to the opinions of the Swedish geologists. For this purpose two different subdivisions of the pre-Cambrian are available, viz. that of Högbom (1910, p. 2) which has been generally adopted for many years, and that of von Eckermann (1936 p. 155) who was led to a new classification of the pre-Cambrian era by his studies of the Loos-Hamra region.

Högbom subdivides the pre-Cambrian as follows:

Palaeozoic		SILURIAN	
			Sub-Cambrian landsurface, denudation
	Upper	Epi-Jotnian dislocations JOTNIAN	
			Sub-Jotnian landsurface, denudation, igneous rocks
Pre-Cambrian	Middle	Epi-Jatulian folding JATULIAN	
			Sub-Jatulian landsurface, denudation
	Lower	Serarchean granites ARCHEAN	

The Åmål-Kroppefjäll granite series (5) is usually classed among the Serarchean granites, together with the Filipstad and the Småland granites. The Åmål formation (4) which is correlated with the Småland porphyries, therefore, comes together with the järngneiss complex (1) into the Archean. Hence it follows that the Archean is divided by an unconformity, at least if it is true that the Åmål formation is older than the Åmål-Kroppefjäll granite series.

There is still some difference of opinion as to the age of the Dal formation (12), but most authors class it in the Jatulian. For a review of the literature regarding this question the reader is referred to the work of van Overeem (1948 p. 118—121). The Dal orogeny then belongs to the epi-Jatulian folding. If the Dal formation is of Jatulian age, then the Kappebo formation must be so too and in that case the Jatulian has to be divided in two by an unconformity. It is therefore advisable, in the case of Dalsland, to extend Högbom's chronological table, just as it was done for the Loos-Hamra area. The pre-Cambrian of the latter area is subdivided by von Eckermann (1936 p. 155) as follows:

JOTNIAN	Dala-Series	
		unconformity
PRE-JOTNIAN	Noppi-Series	
		unconformity
	Upper Loos-Series	
		unconformity
UPPER ARCHAEAN	Lower Loos-Series	
		unconformity
	Sub-Loosian-Series	
		unconformity
LOWER ARCHAEAN		

According to VON ECKERMANN (1936 p. 334), the Lower Loos-Series must be correlated with the Småland porphyries, and the granite that traverses the series (Risberg granite) with the Filipstad, Småland, and Åmål-Kroppefjäll granites (5). The synchronism of this series with the Åmål formation, however, remains questionable. VON ECKERMANN (1936, p. 335) suggests a correlation of the Noppi-Series with the Dal formation (12). If this be true, then the Kappebo formation (9) will have to be paralleled with the Upper Loos-Series. The long continued denudation (16) of the mountain system produced in Dalsland by the epi-Jatulian folding (= Dal orogeny), must have occurred in the Jotnian.

The geological map

The geological map has been surveyed on a 1:5.000 scale. An enlargement of the corresponding part of the topographic map (sheet Upperud N.V. 1926) has been used as base map. Small inaccuracies on it were corrected where met. The final map was drawn on a scale 1:10.000. Besides the spot elevations taken from the topographic map, those which figure on the geological map (sheet Upperud 1870) have been converted into meters and recorded on the map. The depth contours of the lakes have been reproduced from the book of A. SANDELL (1941). The northwestern part of the map of solid rock has been adopted from the map of P. HEYBROEK and H. J. ZWART (1949). In the use of topographic names I have followed the topographic map which in this respect differs rather often from the geological map. An exception has been made for the Kappebofjäll which is mentioned as Kappestigen on the topographic map. This has been done because the Kappebofjäll gave its name to the Kappebo formation and as such has become known in literature. The original map, on which the numbers of the collected hand specimens are recorded, is deposited with the collections of rock specimens and thin sections in the "Rijksmuseum van Geologie en Mineralogie" at Leiden, Holland.

The petrographic descriptions

In the petrographic descriptions a number of terms have been used, which need some explanation.

The term *porphyrocryst* (P. NIGGLI 1948, p. 147) is used instead of phenocryst, porphyroclast or porphyroblast, when the kind of origin of the big crystals in the groundmass is uncertain.

Aggregates of irregular shape and divergent composition, enclosed in the effusives and dyke rocks, are called *nodules*. Terms such as vesicles and amygdales have been avoided, because they imply a genetic interpretation of the origin of the aggregates.

An *amoeboidal* shape is said of such crystals with a „Gestalt mit vielen rundlichen Einbuchtungen und Fortsätzen, so dasz im Verband buchtiges Ineinandergreifen statt hat" (P. NIGGLI 1948, p. 179).

No discrimination has been made in the descriptions between structure, texture and fabric. The term *structure* stands for any one of these notions, which in my opinion is preferable to the use of several terms which, when used by different authors, have a different meaning.

The term *Na-potassium feldspar* is a collective term for such feldspars as orthoclase, sanidine, microline, perthite and micro-perthite, i. e., for potassium feldspars with a minor content of sodium (E. NIGGLI 1944, p. 64).

For the determination of the plagioclase by means of the Fedoroff universal stage the diagrams of REINHARD (1931) have been used. Occasionally X-ray refraction, according to the Debey-Sherrer method, was used for the determination of minerals. The X-ray diagrams of the powder to be examined were compared with the types of the standard collection of the "Rijksmuseum van Geologie en Mineralogie" at Leiden, which were made with the same apparatus and thus enabled a determination of the powder content. A short description of the method employed and further references to literature were given recently by E. DEN TEX (1949, p. 17—21).

CHAPTER I

THE GNEISS-GRANITES

Introduction

The oldest rocks of the Dalskog Dals-Rostock area are the gneiss-granites, exposed in the east, in the west and in the northwest. They form the basement of the Kappebo formation. The gneiss-granite in the northwest is overthrust.

When the geological map (sheet Upperud, 1870) was made, the following types of gneiss-granite were distinguished (TÖRNEBOHM 1870a, p. 10—18 and p. 59):
1. Jerbo gneiss, a gneiss of highly variable character; 2. Kroppefjäll gneiss, characterized by large eyes of red feldspar; 3. Teåker granite, consisting essentially of red feldspar and quartz. In accordance with the opinions of that time the gneisses were considered as sedimentary rocks. Only to the Teåker granite an igneous origin was ascribed. On the map referred to the overthrust gneiss-granite in the northwest is marked as Jerbo gneiss. The Teåker granite constitutes a small stock south of Teåkersjön, whilst the remainder was mapped as Kroppefjäll gneiss.

The gneiss-granite north of Teåkersjön has been remapped and described in greater detail by TÖRNEBOHM (1882—'83, p. 627—631). In doing so he recognized the intrusive character of these rocks, which subsequently suffered a strong transformation, ascribed by him to weathering. He distinguishes in this region: 1. Jerbo gneiss, a gneiss-granite; 2. diorite, genetically related to the Jerbo gneiss and 3. hornblende-granite, occurring as dykes and stocks in the former rocks. All three are traversed by 4. greenstone dykes, whilst the latest intrusion is 5. a red granite, because the latter traverses the first three and does not contain greenstone dykes. According to the description the red granite would nowadays be called aplite-granite.

In the brief description of the overthrust gneiss-granite near Ränn, VAN OVEREEM (1948, p. 6—7) draws about the same picture of the rocks encountered there. However, he believes that there exists no separate hornblende-granite, but that by a gradual increase of the hornblende content the rock passes into a hornblende-diorite, occurring as streaks in the gneiss-granite. I had no opportunity to study more closely the gneiss-granite in the northwest of my region. I may mention however that the rocks reported by TÖRNEBOHM (1882—'83) have indeed been found, but I paid no further attention to their distribution and mutual relations. On the map they are marked together under the name "Jerbo gneiss-granite".

Differences of opinion exist in literature concerning the formation of the Teåker aplite-granite, although, after the description of TÖRNEBOHM (1870a, p. 59) who considered it as intrusive in the sedimentary gneisses, no further description has been published. HOLMQUIST (1906, p. 226) takes it to be a relic of a less advanced degree of metamorphism in the Kroppefjäll gneiss-granite. SANDELL (1941, p. 24) regards it as a younger granite which cuts through the gneiss-granites and shows sharp contacts, but these are not always sharp against the Kroppefjäll granite which seems to be allied with it by palingenetic fusion. W. LARSSON (1947, p. 327) expressed the idea that the intrusion of the Teåker granite is related to the effusive activity in the Kappebo formation, and therefore must have been intruded at the same time. As will be set forth in the following discussion of the aplitic Teåker granite, the writer's investigations lead him to conclude that this rock is genetically related to the Kroppefjäll gneiss-granite, although it is slightly younger. It is, however, much older than the Kappebo formation.

For a long time the Kroppefjäll gneiss has not been regarded any more as of sedimentary origin, but as a gneiss-granite (E. SVEDMARK 1902, p. 18 and HOLMQUIST 1906, p. 224). For a better understanding of the position of these and other gneiss-granites we will have to review the regional literature on the subject over a wider district.

On inspection of the sheets of the geological map of Dalsland it appears that the Jerbo and Kroppefjäll gneisses have a widespread distribution in this province. They are to be considered as gneiss-granites, at least for the main part. This statement, however, leads me to join the ranks of many investigators who, ever since 1883 (NÄTHORST) and for various reasons, wanted a remapping of the area, since so many opinions have changed in geological science after 1870. From nothern Dalsland and southwestern Värmland granites and gneiss-granites which belong to the same gneiss-granite complex, were discribed later on by MAGNUSSON (1929a) and LARSSON (1947).

The outline map of Fennoscandia by GAVELIN and MAGNUSSON (1935) (see fig. 1) shows, as far as known, the extension of these gneiss-granites. They are considered to belong to the Småland and Filipstad granite group. They appear to have a sharp contact with the gneiss complex of southwestern Sweden, as is shown by the more recent publications (MAGNUSSON 1926 and 1929a, p. 53 and LARSSON 1947, p. 323). This contradicts the idea of HOLMQUIST (1906, p. 227), who supposes a gradual transition of the gneiss-granites of Dalsland into the gneiss complex of southwestern Sweden. According to this author there is only a difference in degree of their metamorphism. Since the gneisses of southwestern Sweden are not exposed in the Dalskog Dals-Rostock region no fresh arguments about this can be brought forward.

As said above, more recent studies of the gneiss-granites in northern Dalsland and southwestern Värmland have been published. K. WINGE (1900, p. 340—341) and P. J. HOLMQUIST (1906, p. 176) say that the Kroppefjäll gneiss, Gåsö gneiss and Tössö gneiss on the Åmål sheet (TÖRNEBOHM 1870c) are the more metamorphic equivalents of the Åmål and Bådane granites which are described as eruptive, and that these two granites are related genetically.

On the Säffle sheet (1929) and in the Stavnäs area MAGNUSSON (1929a) distinguishes two main types of granite viz. the more basic Åmål granite and the more acid Kroppefjäll granite. These two are genetically related, but the latter must be considered somewhat younger. It follows from the description that this Kroppefjäll granite, though not generally showing "augen" structure, has all the other properties in common with the Kroppefjäll gneiss-granite as it is found in the Dalskog Dals-Rostock area. The same rock has been recorded of the Vårvik area (W. LARSSON 1947) and on the sheet Fjellbacka (E. SVEDMARK, 1902), while the gneiss-granite between Ånimmen and Vänern must also be reckoned to belong to this type (VAN OVEREEM 1948, p. 10). Furthermore I found this rock in the field in an outcrop on the main road 3 km south of Åmål, in a locality southwest of Laxaby and in hand specimens of the collection of the "Rijksmuseum van Geologie en Mineralogie" at Leiden which had come from the Baldersnås peninsula. It is clear therefore that the Kroppefjäll gneiss-granite is widespread in the gneiss-granite area of Dalsland and southwestern Värmland.

It is more difficult to conclude whether the term Jerbo gneiss covers the same thing everywhere and is perhaps synonymous with the Åmål granite of the Säffle sheet (1929). The brief discussion of the Vårvik region (W. LARSSON 1947) looks promising in this respect. The granites are subdivided by LARSSON into four main types, viz. diorite, Åmål granite, intermediate granite and Kroppefjäll granite. All these types are linked together by transitions, but display a distinct individuality in their geological occurrence. They have been intruded in the mentioned order and are considered as differentiation products of one magma. It now becomes probable that the diorite, the hornblende-granite and the (Jerbo)gneiss-granite which TÖRNEBOHM (1882—'83) and VAN OVEREEM (1948) found north of Teåkersjön and west of Ränn correspond with the first two or three types of LARSSON. I think that with this subdivision LARSSON has furnished the key for the mapping of the entire area of gneiss-granites.

However, the investigations in the Dalskog Dals-Rostock region show that a fifth and final member has to be added to this consanguineous series, namely the Teåker granite, an extremely acid aplite-granite, the intrusion of which is of a later date than the Kroppefjäll gneiss-granite.

The statement of VAN OVEREEM (1948, p. 10—12) holding that the Kroppefjäll gneiss-granite between Ånimmen and Vänern differs from the hornblende-bearing granite to the west of Ränn only in metamorphism, becomes improbable if we keep in mind the above picture, drawn from literature.

Now the autochthonous gneiss-granite of the Dalskog Dals-Rostock area will be considered in detail.

The Kroppefjäll gneiss-granite

In the Dalskog Dals-Rostock region this rock, in its most typical form, is a coarse "augen-gneiss". However, it has far from always developed in this way, for quite often the "eyes" have been crushed, partly or completely, by strong dynamometamorphism, the rocks then being altered to granite-schists. Also in the acid parts of the Kroppefjäll gneiss-granite the eyed structure vanishes and the rock gets a more massive character.

Although there is a complete transitional series between the augen-gneiss and the granite-schist, since they mark only different stages of crushing of the original rock, they will for the sake of clarity be treated separately.

Augen-gneiss

Macroscopical characters:

Embedded in a bluish black very fine-grained matrix with a marked schistosity are large eyes of bright red feldspar and small eyes and streaks of milky blue quartz. The *feldspar eyes* (fig. 2 and 3) can reach a diam. up to 5 cm, but usually do not exceed 3 cm. Most often these large eyes alternate with a greater number of smaller eyes which do not measure over 1 cm. The number of feldspar eyes per unit area is highly variable. In places they may be lacking altogether, the rock then being composed entirely of the dark matrix with quartz eyes. The feldspar eyes are rarely bounded in part by regular crystal faces. Sometimes they are elliptically rounded and show a certain degree of cataclasm distinguishable on the hand specimen. Generally however the cataclastic process has been so strong as to stretch the red feldspar into irregular lenticular shapes, giving rise to numerous transverse cracks which partially have been filled up with quartz. If the cataclastic process involves the disintegration of the eyes, the structure of the rock disappears and a granite-schist results. We shall return to this presently. Several eyes of red feldspar show inclusions of matrix material.

When the eyes of milky blue *quartz* are rounded, they have an average size of 1 cm. A similar scale of progressive cataclasm may be observed in the quartz, ranging from more or less rounded grains to elongated eyes and trails of completely crashed quartzes parallel to the foliation. However, in the same rock, the quartz is always more distorted and crushed than the red feldspar.

The dark grey *matrix* mostly has a strong foliation. The foliation planes are bulging around the eyes and so give rise to a fluxion structure. Besides the occasional small grains of quartz and feldspar, occurring as streaks, the nature of the minerals of the matrix cannot be identified. The foliation plane has a micaceous lustre. Idiomorphic crystals of pyrite occur regularly though sparsely. Occasionally, fine scales of biotite may be observed in the matrix. North of Kabbo an augen-gneiss was found, containing biotite scales of a size up to 0.5 cm.

Acid parts occur regularly in the augen-gneiss. They are not clearly delimitated, but reveal themselves rather by the fact that the dark matrix is greatly reduced and then becomes only a minor constituent. This transition also causes the disappearance of the eyed structure. The rock then consists essentially of bluish white quartz, at times hypidiomorphic, and red xenomorphic feldspar and conveys the impression of a pegmatite. All these effects make for a close resemblance of the acid parts of the augen-gneiss to the aplite-granite, which will be treated subsequently. The grain size of the former (d. av. = 1 cm) however, is considerably larger.

Microscopical characters:

As may be expected the microscopical picture of this rock is dominated by dynamometamorphic structures. Highly cataclastic and undulatory quartz fragments surrounded by mortar quartz, together with substantially less cataclastic large Na-potassium feldspars are embedded in a matrix exhibiting a fluxion structure. The latter consists of streaks of sericite and fine granular quartz with little feldspar. In the streaks of sericite occasional scales of muscovite have formed. Small remnants of plagioclase with polysynthetic twinning are commonly observed between the sericite. Furthermore dark streaks occur in the matrix, consisting essentially of chlorite, finely disseminated ore and some epidote. Several of these streaks can be recognized as altered biotite by their form and the presence of small relics of biotite. As has been stated, only once coarse unaltered biotite was met in the augen-gneiss. Finely divided biotite rarely forms the main constituent of the dark streaks. Although the quantity of this dark material is variable, it always makes up a minor part of the matrix.

In searching for relic structures in this highly metamorphic matrix it strikes us that the sericite and the dark (chloritic) parts do not always form streaks, but are occasionally bounded by markedly straight outlines. In such cases the idiomorphic shape of the original minerals, namely plagioclase and biotite, is plainly recognizable. We have noticed already that remnants of these minerals have been encountered among the alteration products. From the pseudomorphic sericite aggregates it appears that the plagioclase crystallized in about the same size as the quartz.

It becomes also apparent from the less altered parts of the augen-gneiss that the initial xenomorphic character of the quartz crystals has been distinctly preserved, although broken up and surrounded by mortar quartz. They often fill up the interstices between the sericitized but idiomorphic plagioclases. In the more thoroughly metamorphosed parts of the augen-gneiss the quartz has partly been crushed into a granular mass, in which some scarce feldspar grains are found. The large irregular fragments of quartz enclosed in it, are remnants of the quartz crystals resulting from the crystallization. Irregular crystals of calcite have developed sometimes in the granular quartz-feldspar mass.

From macroscopical inspection we have already learned that the Na-potassium feldspar, forming the eyes, did not escape the general cataclastic deformation, as shown by the frequent cracks and the elongated forms of the eyes. This is also seen upon microscopical examination and it then appears that the cracks have been filled up by quartz and calcite. However, the cataclastic effects upon the Na-potassium feldspar are considerably smaller than upon the quartz and the sericitized plagioclase. Thus we have here a manifestation of the general rule that Na-potassium feldspar is one of the minerals most resistant to dynamometamorphism. This is also the reason why the acid pegmatite-like parts of the augen-gneiss look much more massive, for the amount of Na-potassium feldspar is much greater in those parts and the sericitized plagioclase is reduced. Exactly these aggregates of sericite, when under stress, are particularly susceptible to differential movements which cause the marked foliation and fluxion structure of the augen-gneiss. The thin sections of the acid, more massive parts show that this rock did not escape either the general effects of an intensive cataclasm, leaving not a single crystal unbroken. However, hardly any differential movements occurred.

The Na-potassium feldspar often has a perthitic structure. Occasionally a microcline lattice is produced, while complete or partial albitization is not uncommon, resulting in "chessboard" albite. That we are dealing with albite has also been proved by X-ray analysis [1]). Some of the Na-potassium feldspars have been kaolinized along a crowded network of cleavage and fracture lines. In thin section they are usually sieve-like in appearance owing to numerous small inclusions of quartz. Inclusions of idiomorphic, short prismatic laths of plagioclase, entirely altered to sericite, are never

[1]) I owe the interpretation of the X-ray powder diagram to Mr J. F. OSTEN, who investigated the possibilities of the determination of feldspars by X-ray analysis and who will soon publish on this subject.

missing. As far as it could be measured, they are of a much smaller size than the sericitized plagioclases of the matrix. The inclusions of plagioclase are rimmed by clear albite, secluding them from the surrounding Na-potassium feldspar. Biotite, either altered or in its original state and larger xenomorphic crystals of quartz also occur as inclusions. The boundary of the Na-potassium feldspar is seldom linear; but if linear, it suggests a hypidiomorphic shape of the crystal. Usually, however, the boundary is sinuous and makes it seem as though the Na-potassium feldspar penetrated into the matrix and in doing so half encircled idiomorphic plagioclases. Here again the latter are surrounded by a limpid rim of albite. The semi-enclosed plagioclases, forming as it were peninsulae in the Na-potassium feldspar, show plainly that the Na-potassium feldspar has been resistant against dynamometamorphism and has protected the inclusions and semi-inclusions within its boundaries from being crushed. Frequently, an irregular aggregate of sericite and quartz has developed just outside the Na-potassium feldspar, showing fluxion structure. This aggregate cuts off the half-enclosed plagioclases along the edge of the Na-potassium feldspar crystal.

In an advanced stage of dynamometamorphism finally the Na-potassium feldspar eyes also are crushed and nothing remains of relic structures, as we will see from the discussion of the granite-schists. Therefore this is the moment to consider the genesis of the Kroppefjäll gneiss-granite. It is obvious that the eyes originated before the dynamometamorphism in a matrix of coarse crystalline plagioclase, biotite and quartz. Thus, the order of crystallization has been:

1. plagioclase and biotite
2. quartz
3. Na-potassium feldspar.

As to the age of the sericitization of the plagioclase and the chloritization of the biotite nothing can be said with certainty.

The question arises whether the Na-potassium feldspar crystallized in the latest stage of the solidification of the magma or later on. In my opinion the first alternative is by far the most probable, for the following reasons:

1. In volume the Na-potassium feldspar is a very important constituent of the gneiss-granite (approximately $^1/_3$ to $^1/_4$) and is never lacking, though irregularly distributed. This is true not only for the Dalskog Dals-Rostock region, but also in the vast area where the Kroppefjäll gneiss-granite occurs. If the eyes resulted from injection after the rock's solidification, there should have been an enormous supply of potassium spread uniformly throughout the rock. This does not seem probable.

2. In the acid parts of the Kroppefjäll gneiss-granite without "augen" structure the Na-potassium feldspar has a fitfully xenomorphic character and an irregular distribution. The rock gives a strong impression of having been crystallized from a single fusion. In other areas where the "augen" structure is less conspicuous (MAGNUSSON 1929a) the Kroppefjäll gneiss-granite has a normal granitic character, though dynamometamorphic.

3. The idiomorphic plagioclases enclosed in the Na-potassium feldspar are usually smaller than the plagioclases of the matrix. This indicates that the crystallization of the plagioclase was not yet finished at the time the crystallization of the Na-potassium feldspar started.

4. From the southeastern part of the Gotthard Massif of the Alps a normal intrusive granite, the Medelser granite, is known which, apart from its lower degree of metamorphism and smaller grain size, shows a detailed resemblance to the Kroppefjäll gneiss-granite (H. M. HUBER 1943). There is every reason to consider the Medelser granite as a normal intrusive granite (cf. HUBER 1943, p. 104). In this granite also the Na-potassium feldspars are greater than the plagioclase and quartz crystals, thus a porphyric granite results.

Therefore, in my opinion, there is no reason to follow BARTH (1930, p. 113) when he considers the vast areas of augen-gneiss in Sweden and Finland as injection gneisses, in analogy with the origin of lenticular masses of augen-gneiss in the basement of southern Norway which have a maximum extension of 120 m only.

Granite-schist

This term applies to the highly dynamometamorphic parts of the Kroppefjäll gneiss-granite. They result from an advanced cataclastic transformation of the augen-gneiss by which the eyes disappear. Recrystallization occurs in places, but is of minor importance.

These highly cataclastic rocks can be termed mylonites, as HOLMQUIST (1906, p. 223) did, and we can use the terms granite-mylonite (s.l.) or gneiss-mylonite (s.l.) according to P. QUENSEL's nomenclature (1916, p. 99), as it is done by VAN OVEREEM (1948, p. 8) to designate the highly cataclastic granite, between Ånimmen and Vänern. LARSSON (1948, p. 635) has objections against this use of the term mylonite. Though this author gives no arguments upon which his opinion is based, there are reasons to support his view. Especially in Sweden, where the cataclastic rocks are of frequent occurrence, it would be useful to make a clear distinction between rocks having obtained their strongly cataclastic structure by regional dynamometamorphism and rocks cataclastically deformed by movements along great (thrust) faults, so that they originate in relatively small zones. Accordingly, the term mylonite should be restricted to the last case. This gives to the term a genetical (tectonical) value. For a critical review of literature in relation to all diversity in definitions of the term mylonite I may refer to the paper by WATERS and CAMPBELL (1935).

In the Dalskog Dals-Rostock region we have no indications that the granite-schists have been formed along great fault planes. The rock is not confined to a definite belt, but occurs rather in lenses of variable size in the augen-gneiss. As regards its regional distribution it must be stated that the granite-schist is to be found exclusively in the augen-gneiss west of the Kappebofjäll. It is more likely therefore, that the region as a whole has been subject to regional metamorphism of variable local intensity. To avoid confusion it does not seem appropriate in my opinion to use the term mylonite in this connection. If, nevertheless, it is used in this way, it is not allowed to take it for granted, as it has been done by VAN OVEREEM (1948, p. 19), that great (thrust) faulting has caused the crushing of the rocks. In this case it is even less admissible as VAN OVEREEM (1948, p. 8) states: "Whether the granite-mylonite and gneiss-granite have been deformed contemporaneously or not, could not be established with certainty", ..., "it is for the present preferable to leave this question undecided". In analogy to the Dalskog Dals-Rostock area there is a sound reason to assume that the "mylonitization" of the granite between Ånimmen and Vänern is to be ascribed to the same dynamometamorphism which brought about also the "gneissification" of the granite.

Evidently this remark does not imply that there has been no formation of granite-schist in small zones along the great faults, which must, according to the above suggested use of the term, then be called mylonites. A good example of this is to be found along the fault near Tonebyn. It must be noted in this connection that the great movement along the Kroppefjäll fault, which is probably syngenetical with the origin of the mylonite zones in central Värmland (see p. 176), did not form in this region a wide zone of mylonitization, as far as could be ascertained. The augen-gneiss of the Dalbo plain represents a comparatively little dynamometamorphic part of the Kroppefjäll gneiss-granite. However, it should be borne in mind that all outcrops are at a distance of 300 to 500 m from the fault. On the other hand a mylonite is found along this fault in the lower slate to the north of Viken at the eastern shore of Näsöln.

There is a complete series of transitions between the granite-schist and the augen-gneiss. The eyes of feldspar and quartz are entirely smashed to pieces which form long streaks. A still further crushing results in a fine-grained rock. The foliation, looking like a fluxion structure, is changed into a parallel foliation. The grain diameter at right angles to the schistosity is considerably smaller than that parallel to it. Some scattered larger fragments or eyes indicate the initial character of the rock. The microscopical examination shows the same progressive crushing of the eyes until only small protracted lenses of Na-potassium feldspar and streaks and lenses of utterly cataclastic quartz remain, embedded in a matrix which is composed of a parallel arrangement of streaks of sericite and of ore, chlorite and epidote. Here too, occasional remnants of plagioclase occur in the sericite streaks. Incipient recrystallization of the quartz and the formation of single larger muscovite crystals in the sericite has been observed.

On the Kroppefjäll plateau near Kroktjärn and to the west of it a granite-schist occurs in which recrystallization played a more important part. Its outward appearance has changed to a fine-grained grey mica-schist, in which muscovite flakes and incidental biotites are readily identified. Occasional scarce eyes of red feldspar betray here also the rock's origin. Microscopically it is seen that the quartz aggregates are almost completely recrystallized into a granoblastic pattern. Large muscovites and scarce biotites have grown from the sericite, which have a random orientation and do not conform to the schistosity: so called cross-muscovites and cross-biotites. In the Na-potassium feldspar a great number of small feldspar crystals have formed with the same refractive index as the original crystal, besides some small clear albites and a little calcite. They may replace the original crystal partly or totally. The granoblastic patches of quartz and feldspar have an average grain size of 0.1 mm. Relics of Na-potassium feldspars, which originated prior to the recrystallization, may be 30 times as large. Occasional relics of plagioclase, highly altered into sericite, indicate that in these rocks the dynamometamorphism has not been stronger than in the granite-schist, where recrystallization is lacking.

Little if any migration of material has attended the above described recrystallization. Due southeast of Kroktjärn the granite-schist has been changed into "hartschiefer" (nomenclature after QUENSEL 1916), apparently affected by local faulting. It has become a microcrystalline rock of quartz, feldspar and sericite, with a still distinct schistosity. The rock is pervaded by a great many veins of fine-grained quartz, in which some feldspars occur. These veins often have ill defined boundaries and then pass gradually into the rock. It is likely, therefore, that the vein fillings originated from the mobilized quartz and feldspar of the rock itself. There, migration of material did occur.

The fact that recrystallization took place in the granite-schists indicates that during the metamorphism of the rock temperatures have been relatively high. The circumstances then prevailing fall into the epidote-amphibolite facies of ESKOLA (1935) or occurred in the deeper parts of the epizone (GRUBENMANN and NIGGLI 1924).

That the presence of granite-schist in the Kroppefjäll is not merely local, is apparent from the description of the sheet Fjellbacka (SVEDMARK 1902, p. 19) in which a granite-schist is reported, derived from the Kroppefjäll granite in the parishes Rölanda and Lerdal. It is developed there as mica-schist and talc-schist.

Xenoliths in the Kroppefjäll gneiss-granite

In the augen-gneiss of the Dalbo plain, and to the south of Lapperud, inclusions have been found of mica-schists, mostly very rich in quartz. They have been encountered in sizes varying from 1 dm to several meters in diameter. The larger ones sometimes are pervaded by veins of augen-gneiss. On account of its petrographical likeness the mica-schist, which is exposed over a distance of 300 m along the track between Hult and Eriksbyn, should also be regarded as an inclusion or, more likely, as an assemblage of inclusions. This too is veined by augen-gneiss.

The mica-schist is a light to dark grey, sometimes nearly black, schistose rock. Upon microscopic examination the lighter types of this rock appear to be very rich in quartz, even to such a degree that one might call it a quartzite. In thin section of the less metamorphic inclusions the clastic origin of the rock is still plainly visible. Its consists essentially of rounded, evenly sized, quartz grains (d. = 0.25 mm) with sutured structure when in contact with each other. In addition there are a few albites of the same size. A matrix generally occurs between the grains. It is made up of sericite, small grains of quartz and feldspar and also of biotite, epidote and some ore, besides a little muscovite. The biotite, epidote and ore always occur together. In the stronger metamorphic types the quartz forms a mosaic structure and the mica is concentrated into streaks. Between these mica strings large muscovites have developed. The rounded form of the quartz grains is seldom preserved. The very dark varities consist essentially of black to light green pleochroic biotite and finely divided epidote, between which are inserted small grains of quartz. The inclusions must have been derived from sediments prior to the granites, namely from impure quartzites and slates. It may be logically assumed that these sediments belong to what is called by LARSSON (1947, p. 321) the "oldest supracrustal formation". This formation, which has been recognized by LARSSON in the gneiss complex of the Vårvik area, contains there also impure quartzite and slate, besides acid and basic volcanic rocks. It seems improbable that the inclusions have been derived from the Åmål formation, for the following resaons.

1. No slate has ever been reported from the Åmål formation.
2. The Åmål formation has not been encountered anywhere in southern Dalsland.
3. The pre-granitic age of the Åmål formation is questionable, as will be seen below (p. 136).

The Teåker aplite-granite

To the south of Teåkersjön an aplite-granite is exposed, which is named Teåker granite by TÖRNEBOHM (1870a). The boundary against the Kroppefjäll gneiss-granite is not clear-cut, but consists of a broad zone of composite rocks. This zone has been marked on the map by a hatching. Except to the south of Teåkersjön the aplite-granite occurs as small bodies and dykes in many other localities in the Kroppefjäll gneiss-granite, and is then usually more finely grained. The Teåker aplite-granite, therefore, is younger than the Kroppefjäll granite. The composite rocks of the two granites have also been encountered outside the zone mentioned above. Owing to their small dimensions they could not be plotted separately on the map.

The aplite-granite is a red, medium-grained rock, in which quartz and red feldspar (d. av. = 3 mm) can be distinguished, having little, if any, dark constituents. Although there is a distinct direction of schistosity, the rock has a remarkably massive outlook, as compared with the augen-gneiss and the granite-schist. Microscopically the rock proves to be composed essentially of Na-potassium feldspar, quartz and some plagioclase. In spite of its macroscopically massive appearance, the rock shows in thin section a definite cataclastic structure, to such an extent that no original structures can be recognized (fig. 4). The quartz has undulatory extinction and dentated edges; it is broken and surrounded by mortar quartz. The Na-potassium feldspar also is highly cataclastic and often shows a perthitic structure. Occasionally a microcline lattice is produced. Some feldspars have been slightly altered into kaolin. The plagioclase shows polysynthetic twinning and is less refringent than quartz, which indicates albite. The albite is slightly sericitized. Some epidote and ore occur as minor constituents. In the Na-potassium feldspar and in cracks there is a red pigment (hematite?).

The dykes of aplite in the Kroppefjäll gneiss-granite are essentially the same as the aplite-granite south of Teåkersjön, described above. However, the sericitization of the plagioclase has considerably increased, and in places, some muscovite has developed. A little tourmaline and leucoxene occur also. The Na-potassium feldspar of the aplite-granite has exactly the same colour as the eyes of the augen-gneiss.

Because of the strong cataclastic effects in the aplite-granite it is not necessary to assume that the forces causing the dynamometamorphism were smaller for the aplite-granite than for the Kroppefjäll gneiss-granite, as one might at first be inclined to suppose. The sericitized plagioclase in the aplite-granite forms a constituent of minor importance and this is precisely the most easily affected constituent by shearing stress, causing foliation. It is true that quartz and Na-potassium feldspar are broken up by the same stress, but the relative displacement of the fragments is very small, so the rock maintains its massive character. Accordingly, the schistosity of the more acid parts of the Kroppefjäll gneiss-granite diminishes in those places, where the content of sericitized plagioclase drops (p. 73).

In the smaller patches of aplite-granite, lying in the Kroppefjäll augen-gneiss, a few peculiar facts may be noted. The patches have an irregular outline. The aplite-granite is extremely fine-grained and encloses large crystals of milky blue quartz which in size, shape and colour are exactly similar to the quartz eyes of the surrounding augen-gneiss. Upon closer inspection one becomes aware of larger crystal faces of red feldspar in the aplite, which are not conspicuous because they are of the same colour as the aplite-granite. It looks as if the aplite were replacing only the dark matrix between the eyes of the augen-gneiss. The transitional zone between the aplitic parts and the augen-gneiss also points in the same direction. By a gradual increase of the dark matrix the eyes show up again very clearly and the rock passes into a normal augen-gneiss.

Microscopically, the aplitic parts turn out to be composed of a panxenomorphic, fine-grained (d. av. = 0.05 mm) aggregate of quartz and feldspar and finely divided sericite. Embedded are large crystals of quartz (up to 1 cm diam.) and partly recrystallized big feldspars with perthitic or "chess-board" structure. Recrystallization is particularly manifested by the formation of quartz crystals in the big feldspars, to such an extent that the original crystal is entirely broken up. It is frequently observed that a certain stage in this process leads to the formation of myrmekite. Besides the quartz and the partly preserved eyes of feldspar, occasional enclosed sericite assemblages have been encountered in the fine-grained aplitic rocks. These inclusions must be regarded as remnants of the initial sericitized plagioclases of the matrix. Apparently we are witnessing here a selective metasomatic replacement of the porphyric granite by the aplite-granite. First of all the sericitized plagioclase was replaced and next a great part of the Na-potassium feldspars, whilst the quartz eyes persisted.

Composite rocks

To the east of the Teåker aplite-granite stock south of Teåkersjön occurs a zone of aplite-granite and augen-gneiss, forming a chaotic mixture. This mixing up is mainly the result of transformation of the rocks themselves and in a lower degree due to the penetration of a great number of aplite veins into the augen-gneiss, constituting, as it were, a zone with many xenoliths of augen-gneiss in the aplite-granite. These transformations are brought about by injection of aplite-granite in the Kroppefjåll granite and by assimilation of the Kroppefjäll granite by the aplite-granite.

Only in this way can a plausible explanation be given of the fact that the Kroppefjäll gneiss-granite and the Teåker aplite-granite seldom have their normal character in this zone of composite rocks, but show instead all sorts of transitions lying pell mell, whilst not a meter of the rock has the same character. Along the new road between Sand and Kläppa a number of fresh outcrops originated in these composite rocks, which are of sufficient size to permit the study of the interrelations of the aplite and the augen-gneiss. Here one can see how the aplite-granite gradually passes into augen-gneiss by the appearance of dark streaks in which, as they become larger, eyes of red feldspar and quartz occur. In this transition zone part of the matrix of the augen-gneiss is usually aplitic. Aplite then may constitute the main part of the matrix, but more often it forms small nests and streaks in the latter. Apparently a metasomatic displacement of the matrix by aplite has been going on, just as we observed previously in the case of the small patches of aplite between the augen-gneiss. In this very irregular zone of composite rocks there occur also well defined patches and dykes of aplite-granite.

The nests and streaks of fine-grained quartz and Na-potassium feldspar in the matrix have also been found repeatedly in the augen-gneiss north of Kappebosjön, often in connection with veins of aplite. Apparently they

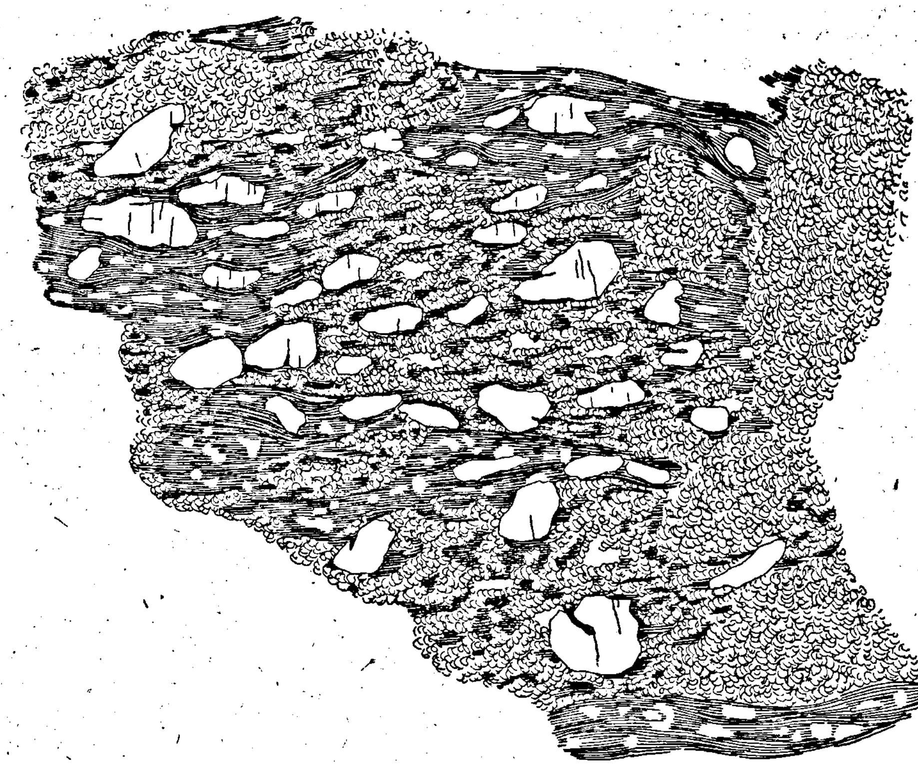

Fig. 2.

Composite rock consisting of augen-gneiss and aplite-granite. For explanation see text. ¼ × natural size. Due southeast of Gunnesbyn.

represent the utmost offsets of the zone of composite rocks, which indicates that the stock of Teåker aplite-granite must be present not far below the surface in this region too.. Just to the southeast of Gunnesbyn there is a small outcrop which provides a beautiful example of a composite rock of augen-gneiss and aplite-granite. Figure 2 gives a picture of a part of this exposure. It consists of coarse augen-gneiss with large eyes of red feldspar and quartz traversed by the aplite-granite with more or less straight contact lines (right of the figure), but also irregularly replacing the dark matrix. Aplite can take the place of the matrix completely (top left of the figure), or only partly, leaving dark spots in between (centre of the figure). Small crystals and nests of muscovite occur in this partly aplitic matrix.

Age relations

It is most likely that the Teåker aplite-granite originated from the same magma as the Kroppefjäll granite by differentiation and must be considered younger, though not much, than the latter. The acid pegmatite-like parts of the Kroppefjäll granite have exactly the same mineralogical composition as the aplite-granite. The Na-potassium feldspars of both granites are entirely identical in structure and in colour. The blurred contact between the Teåker aplite-granite and the Kroppefjäll granite also suggests a small time interval between the solidification of the two granites, since it is generally admitted that aplite-granite congeals at a low temperature, and therefore, as a rule, cannot assimilate other rocks (see p. 79). Be this as it may, both granites must be older than the Kappebo formation. The pebbles of aplite-granite form a large portion of the conglomerates of this formation. Besides, large pebbles of augen-gneiss have been found in them, whereas the graywackes of the Kappebo formation have been derived directly from debris of the Kroppefjäll gneiss-granite, as is shown by their content. Also the main stage of the dynamometamorphism of both granites must have occurred before the deposition of the Kappebo formation. The cataclastic effects in these rocks, from a regional point of view, are considerably stronger than those in the younger supracrustal formations. Recrystallization, as it is found in certain granite-schists, does not exist at all in the younger rocks. Moreover, in the Kappebo conglomerate east of Koljerudstjärn (200 m northeast of the northern point of the lake) a large outcrop has been found, where big, almost round pebbles of augen-gneiss, having a random distribution with respect to their foliation, form the main part of the conglomerate. In this connection an outcrop to the east of Tångebo deserves to be mentioned, where a strikingly little metamorphosed fine-grained pelite, rich in quartz, of the Kappebo formation, has been deposited on the rugged surface of the highly schistose augen-gneiss.

It is necessary to dwell for a moment upon the question of the age of the main stage of dynamometamorphism in the granites, because is has been the subject of a violent controversy between VAN OVEREEM (1948) and LARSSON (1947 and 1948).

LARSSON (1947) on the basis of his investigations in the Vårvik area presumes that the granites, to which belongs the Kroppefjäll granite, are of syntectonic or late-tectonic origin with respect to the folding of the Åmål formation and that their metamorphic structure was acquired during the folding of the Dal formation. VAN OVEREEM (1948) on the other hand has good reasons to believe that, as he puts it, the mylonitization and the gneissification of the granite occurred before the deposition of the Dal formation, a conclusion that is entirely confirmed by the geology of the Dalskog Dals-Rostock region. In reviewing VAN OVEREEM's work, LARSSON (1948) tried to refute the above conclusion in terms not strictly within the limits of argument.

Without going further into the matter at this moment, especially since some of the arguments will be discussed in other connections (p. 83, 136), I wish to put forward some observations which VAN OVEREEM reported to me in a personal communication and which show once more the pre-Dal age of the gneissification of the granite. He states: "In the vicinity of Ånimmen the mylonites are in a temperature-pressure facies different from the Dal formation. In the phyllites of the Dal formation, regionally speaking, no new biotite is formed, as it is in the granite. In the granite-mylonite the formation of microline is observed, but in the rocks of the Dal formation nothing like it has happened. In the Dal formation spilite which has been faulted into the basement, remnants of the original structure are readily seen by microscopic examination. The adjoining granite has not preserved a trace of its original structure, even though it must be a more resistant rock than the basic spilite".

Dykes, veins and filling deposits in the gneiss-granite

Besides the aplite dykes described above, the following types of dykes, veins and filling deposits have been encountered in the gneiss-granite:

1. Dykes of greenstone, albite-rhyolite and rhyolite, which are to be considered as the feeding channels of the volcanos which built up a large part of the Kappebo formation. They will be discussed at length in the next chapter.
2. Thin veins of epidote and quartz are common in the Kroppefjäll gneiss-granite (fig. 3). Sometimes they form a whole system of intersecting veins. In places they widen, so as to form nodules of quartz and epidote. VAN OVEREEM (1948, p. 18) reports the same veins in the gneiss-granite between Ånimmen and Vänern and observed that they are younger than the intrusions of quartz-porphyry in the gneiss-granite. As we will see below (p. 136) this quartz-porphyry is probably of the same age as the Kappebo formation, so the veins of quartz and epidote are to be considered as younger than the Kappebo formation.
3. Pegmatitic veins and fissure fillings have been found in the gneiss-granite. They may consist of: a. Na-potassium feldspar, quartz and chlorite-sand; b. Na-potassium feldspar and quartz; c. quartz and ore and d. Na-potassium feldspar, quartz and ore. All these types have been found also in the Kappebo formation, the last three, moreover, are reported from the Dal formation. LARSSON (1948, p. 367) found also veins filled with chlorite-sand in the Dal formation of northern Dalsland and uses this fact as a refutation of one of VAN OVEREEM's arguments in support of the pre-Dal age of the gneissification of the granite (VAN OVEREEM 1948, p. 21). Thus, all these veins are younger than the Dal formation and probably are related genetically.
4. In the augen-gneiss of the Dalbo plain a few fillings of large cavities have been found, composed of aggregates of large clear albite, together with an assemblage of albite and muscovite, usually finely divided.
5. In two localities in the Kroppefjäll gneiss-granite a diorite is encountered, viz. 850 m northeast of Tångebo and 500 m west of Öjerud. It is a medium-grained dark rock, in which pink to yellowish feldspars and dark green constituents are visible. Upon microscopic examination the rock appears to be composed essentially of highly saussuritic, hypidiomorphic plagioclase (65 % to 75 %), and of biotite (15 % to 25 %), almost entirely altered into chlorite, epidote and ore, beside a small amount of xenomorphic quartz (10 %). In the hand specimen as well as in thin section the rock looks remarkably massive, with scarcely any schistosity.

Although no direct field evidence is available to ascertain whether the diorite is syngenetic with the granite or of a later origin, its massive character suggests that it was formed afterwards. However, in the first locality the Kappebo formation is presumably overlying the diorite.

On the Upperud sheet (1870) of the geological map a diorite is recorded in the neighbourhood of St. Yxesjön, which has been described by TÖRNEBOHM (1870a, p. 60). Partly it occurs as a dyke in the Kroppefjäll gneiss-granite and the Teåker aplite-granite alike. It may be infered from these data that the diorite was intruded into the granite after the gneissification, but before the deposition of the Kappebo formation.

The direction of schistosity

It is useful to enter now into the matter of the direction of schistosity in the discussed region, because both VAN OVEREEM (1948) and LARSSON (1948) draw from it their arguments pro and contra in their discussion about the age of the gneissification of the gneiss-granite (see p. 81). As will be pointed out below, these arguments are in my opinion without value.

Though the measurements of the schistosity are not worked out statistically, it appears from the map that in the Dalskog Dals-Rostock region there is one single strike of schistosity which holds good for the gneiss-granites as well as for the Kappebo formation and for the rocks of the Dal formation which have any schistosity at all. TÖRNEBOHM (1870a, p. 31) noticed also this concurrence of the strike of schistosity. The strike fluctuates around north 30° east and the dip varies between 20° and 90° to the northwest. In general the attitudes are steeper in the north than in the south. All measurements of the schistosity have been recorded on the map of solid rock, except the majority of those taken in the Dalbo plain, while the mapped area of this plain is only partially shown on the map of solid formations. However, there, the strike of the schistosity is also about north 30° east with a dip usually between 50° and 60° to the west.

It has been reported also by LARSSON (1948, p. 636) of the region west of Könningen, near the southeastern point of Ånimmen, that the direction of the schistosity in the basement is the same as in the schistose beds of the Dal formation. According to his view, this indicates that the dynamometamorphism of the granite resulted from the folding of the Dal formation. We have seen earlier (p. 81) that this assumption cannot be correct, because the granite undoubtedly has a higher degree of dynamometamorphism than the Kappebo formation, and evidently higher than the Dal formation. This is clearly shown for instance, in the region between Ånimmen and Vänern, by the persistance of primary structures in the spilite of the Dal formation, whilst in the even more resistant granite, there is no trace of such structures (p. 82). Moreover, the pebbles of foliated augen-gneiss, distributed at random with respect to their foliation in the conglomerate of the Kappebo formation, give irrefutable evidence (p. 114). Thus apparently, we cannot determine the age of the dynamometamorphism of the granite from the direction of the schistosity.

It might be supposed that the earlier direction of the schistosity has been changed and that the present direction was determined by the Dal folding. But also this supposition offers difficulties, since the strike of the schistosity makes an angle of 30° with the north-south striking axes of the folds of the Dal formation in a large part of the region examined, whereas the general experience shows that schistosity, as a result of folding, trends parallel to the axes. Only in the southern part of the mapped area, the strike of the Dal formation changes in connection with the overturning, and thence becomes parallel to the schistosity.

In the Dal formation the foliation is confined to those beds in which it is apt to occur, such as the matrix of the basal and intercalated conglomerate, the slate bed, the calcareous slate beds and sometimes the spilite and particularly, in regions subjected to strong tectonic movements as in the overturned layers south of Helvetestjärn and in the southern part of the Prästbol syncline. The arkosic quartzitic sandstone has, in places, a schistose cleavage. As mentioned above, in large areas the strike of the schistosity makes an angle with the strike of the beds which, in the field, is best observed in the thin bed of intercalated conglomerate in the basal layers, and also in some outcrops of the slate bed. In the slate-quarry near Källsviken it may be observed that the schistosity is not influenced in the least by the secondary folds, but sets through the folded rock in one distinct direction, interrupted only by the quartzitic bed in this stratum or by thin non-foliated quartzitic intercalations. These facts were noted already by TÖRNEBOHM (1870a, p. 33 and 49) and described from various localities. We are witnessing here the well-known phenomenon that schistosity is truncated by hard beds and is continued with the same strike and dip on the other side of the hard bed. LARSSON's statement (1948, p. 636) that, west of Könningen, the direction of the schistosity is changed against the hard quartzitic sandstone, is quite contrary to the general experience, which is so clearly demonstrated in the Dalskog Dals-Rostock area. In my opinion therefore, LARSSON's observation should be considered with reserve. On the other hand no age relation can be established on the fact that a hard bed truncates the schistosity, as it was done by VAN OVEREEM (1948, p. 21). But it is clear that the schistosity of the Dal formaion has a quite different character than that of the gneiss-granite. Considering the strong crushing of the crystals of the latter, which can be compared with a mylonitization, important internal movements must have occurred there, whilst in the Dal formation this occurrence must have been very slight as the schistosity is truncated by hard layers.

It is as yet an unsolved question why the schistosity has the same direction in all rocks, whereas the gneiss-granite was metamorphosed earlier and more intensively than the other formations and that, moreover, the schistosity strikes obliquely the axes of the folds of the Dal formation. It is possible that the schistosity present in the basement played an important part in the folding of the Dal formation as a result of which the schistosity of this formation developed in the same direction.

CHAPTER II

THE KAPPEBO FORMATION

Introduction

Between the Kroppefjäll plateau in the west and the rocks of the Dal formation in the east and at some localities in the Dalbo plain a supracrustal formation is found consisting of graywackes, conglomerates, lavas, tuffs and agglomerates. This formation, being built up for the greater part by detritus of the gneiss-granite, is younger than the latter, while the still younger Dal formaion overlies it unconformably. Following A. SANDELL (1941, p. 184) we will call it the Kappebo formation.

On the Upperud sheet (1870) of the geological map and in the description pertaining to it (TÖRNEBOHM 1870a) a certain part of the rocks belonging to this formation, viz. the conglomerates and the so called Kappebo graywacke, have been included in the basal layers of the Dal formation. Other rocks belonging to this formation are incorporated in the "gneiss formation" and have been described under the names of helleflint, regenerated helleflint, older graywacke and slate. The helleflints are supposed to cut through the conglomerates and graywackes of the Dal formation. The regenerated helleflint is described by TÖRNEBOHM (1870a, p. 17) as a clastic rock composed of: "a dense helleflintic groundmass in which are embedded angular fragments of helleflint and little pebbles of granitic rock" (translated). We will see later that the helleflints are to be considered as lavas and tuffs and that the regenerated helleflint is an agglomerate. I have found no indications which might lead to a distinction of two different graywackes, as is done by TÖRNEBOHM (1870a, p. 18).

The rocks of the Kappebo formation extend to the south-southwest into the Sätersfjäll (sheet Rådanefors 1870). South of Rådanefors a supracrustal formation is also exposed which can be traced to the west on the sheet Fjellbacka (1902) and to the south on the sheet Wenersborg (1870). After the reading of the map descriptions one can take it that this supracrustal formation is a part of the Kappebo formation. In the description of the Rådanefors sheet V. KARLSSON and A. H. WAHLQUIST (1870, p. 23—24) express themselves in such a way that it becomes doubtful whether these rocks should be considered as belonging to the gneiss: "Here, as well as more to the north and to the south in Dalsland, they appear to constitute a separate group, sometimes with marked transitions to the real gneiss, sometimes entirely separated from it, in places by a more or less distinct conglomerate" (translated). The same is reported by ELIS SIDENBLADH (1870, p. 26—27) with respect to the "euritquartzite", "glimmerschist" and "helleflint" on the Wenersborg sheet, which "are sharply separated from the gneisses in this area" (translated). SIDENBLADH furthermore describes as conglomerate the same rock that has been termed "regenerated helleflint" by TÖRNEBOHM (1870a). GAVELIN (1909) mentions that the dykes which have been mapped on the sheet Wenersborg (1870) as diorites, belong to the leptite formation. E. SVEDMARK (1902, p. 21—22) recognized the rocks of this formation as porphyries and porphyric tuffs on the Fjellbacka sheet (1902).

In the Annual Report of the S. G. U. 1907 and later in a discussion on pre-Cambrian unconformities A. GAVELIN (1908, p. 9—10 and 1912, p. 557) asserts that all of the alleged basement-conglomerates on the sheets Rådanefors and Wenersborg are in reality "pseudo-conglomerates". "The terms "older graywacke" and "regenerated helleflint" in the Sätersfjall and in the Kappebofjäll represent exclusively strongly cataclastic rocks belonging partly to the Dal formation, partly to the basement" (translated). This statement appears to be incorrect.

The study of the Kappebo formation was resumed only in 1941 by A. SANDELL (1941, p. 183—188). This author sticks chiefly to the petrographic descriptions of TÖRNEBOHM (1870a). He comprises these rocks, however, under the name of "Kappebo series", which is sharply separated from the basement and "intimately related" to the Dal formation. He also draws up the following stratigraphy of the Kappebo series:

Kappebo graywacke
Granite-conglomerate
- - - - - hiatus - - - - -
Regenerated helleflint
Quartz-porphyry with graywacke

He considers the Kappebo formation as a geosynclinal formation and believes the regenerated helleflint to be a "direct parallel of the Alpine Flysch" (translated). As will be set forth later (p. 130 f) there does not exist a succession of different strata of a regional extent in the Kappebo formation. Rather, all the different kinds of rock which built up the formation, were deposited simultaneously from the beginning. Thus, a stratigraphical division as given by A. SANDELL is out of the question. The regenerated helleflint is but an agglomerate and has nothing in common with the Alpine Flysch.

W. LARSSON (1947, p. 326) refers briefly to the findings of A. SANDELL, but believes, with some reserve, that there is an unconformity between the Kappebo formation and the Dal formation; an opinion which is confirmed by the results of my field work. Quartzitic sandstone, recorded by LARSSON for the first time among the rocks of the Kappebo formation, has not however been met with.

The description of the rocks of the Kappebo formation will follow hereafter, while in a later chapter the formation as a whole will be considered more closely.

THE EFFUSIVES AND DYKES

Among the rocks of the Kappebo formation effusive and dyke rocks play an important, though by no means a predominant part. They occur as a great number of lenticular bodies amidst the clastic and pyroclastic rocks. The latter are quantitatively far more important. Generally it is quite difficult in the field to decide whether a given rock occurs as a dyke or as an outflow. Most often the contacts are hidden by a dense vegetation of wood or moss, making it almost impossible to follow a contact. A regular succession of strata does not exist in the Kappebo formation. The tectonic structure of this formation thus remaining inextricable, we cannot rely on it either for a clue to solve the question: "dyke or outflow". In some instances one may be fairly sure of the effusive character of the rocks, because they cover a considerable area, for example east of S. Damtjärn. On other occasions in the field they pass almost imperceptibly into tuffs, which also is an indication of their effusive origin. However difficult it may be to find out in each particular case whether a rock is an effusive or a dyke, the large amount of clastic and pyroclastic material leaves no doubt of the fact that outflows have played an important part. More so as the microscopic examination reveals perlitic and fluxion structures.

In the immediate surroundings of the Kappebo formation, where the Kroppefjäll gneiss-granite outcrops, the latter has been crossed in various places by a large number of dykes consisting of rocks having an identical character (both macroscopically and microscopically) as the effusives and dykes of the Kappebo formation. Apparently these are the feeding channels of the volcanos which once built up an integral part of the Kappebo formation and now are intersected by the actual level of erosion not far below the point of outflow. Dykes of this kind were found north of Tonebyn, near Lapperud and east of Åsen as well as at some other localities. They have been plotted schematically on the map with a hatching.

As dykes and effusives have been encountered:

1. albite-rhyolites
2. greenstones
3. rhyolites.

The first two are characterized by phenocrysts of albite, groundmass albite and dark constituents belonging to the epizone (GRUBENMANN - NIGGLI 1924).

If this paragenesis of minerals were not the result of allometamorphism, one would be entitled to speak of spilites and keratophyres. It will appear from the following discussion of these rocks that there is some evidence for a spilitic character, but I prefer to avoid any genetic implication in nomenclature and, if neccessary, employ only the term "spilitic" as an adjective. Consequently, we use the more general terms "albite-rhyolite" and "greenstone".

At a few places the rhyolite is encountered, probably as dykes, in the gneiss-granite and corresponds with the "quartz-porphyry" of northern Dalsland. This relation will be further discussed in the next chapter (p. 136).

Albite-rhyolite

The name "quartz-porphyry" has deliberately not been chosen to designate the acid effusives and dykes, since quartz phenocrysts are often lacking. Moreover, this name is out of date everywhere outside Sweden. TÖRNEBOHM used the term "hälleflinta" in mapping this area. According to his definition (see H. SANTESSON 1877, p. 49), both the albite-rhyolites and the fine-grained tuffs are comprised in this name. Though useful in the field, this term proves too comprehensive to be used for a petrographic description. The acid, aphanitic, mostly light coloured rocks, mapped as "hälleflinta", fall macroscopically into three groups:

1. without phenocrysts,
2. with phenocrysts of red feldspar,
3. with phenocrysts of red feldspar and quartz.

Under the microscope a part of the rocks, mapped as "hälleflinta", appear to be tuffs; the rest are effusives and dykes, which can be classed in four types, when besides the phenocrysts, the structure of the groundmass has been considered.

Type I: Felsitic rhyolite.
Type II: Albite-rhyolite with albite and quartz phenocrysts.
Type III: Albite-rhyolite with phenocrysts of albite only.
Type IV: Albite-rhyolite with a hypiodiomorphic to poikilitic groundmass.

We will proceed with the description of the different types.

Type I. Felsitic rhyolite

Under this heading come the aphanitic, acid rocks, without or poor in phenocrysts, of a grain size too small to permit a clear diagnosis of the mineralogical composition. When we are sure that the rock is a devitrified glass by the presence of relics of a perlitic structure, we may speak of "*felsitic pitchstone*". When perlitic, fluxion or brecciated structures are altogether absent, there is a possibility that it is not an effusive but a tuff. If there is no indication whatever of a pyroclastic origin, the rocks have been classed as felsitic rhyolites. These rocks are widespread in the whole area, where the Kappebo formation outcrops.

Macroscopical characters:

Aphanitic, brown-violet, chocolate brown, red-brown, light or dark grey rocks. Sometimes massive, mostly slightly foliated, seldom strongly foliated. Some samples show a brecciated structure, brought out by colour differences in the rock. Rare small crystals of pyrite occur commonly on a fresh fracture. The weathering crust is conspicuous by decoloration of the rock. The fracture is conchoidal in poorly or non-foliated felsites.

Microscopical characters:

Micro- to cryptocrystalline rocks with as main constituents: quartz, feldspar and sericite. Some sparse phenocrysts of albite, sometimes also spherulites occur. If sericite is predominant, the rocks are fibrous to parallel fibrous and thence foliated, if not, they are of a granular habit. Fluxion, brecciated, perlitic and lenticular structures are found. Fluxion and perlitic structures especially are best observed in ordinary light and with moderate magnification.

The *perlitic structure* is characterized by spheroidal or ellipsoidal concentric bands, visible by their yellowish to brownish colour (fig. 5). This colour is produced by a brown-red pigment (hematite?), particularly apparent in reflected light. The mass between the coloured strips is colourless to cloudy grey. The last is due to minute inclusions in the feldspars, probably

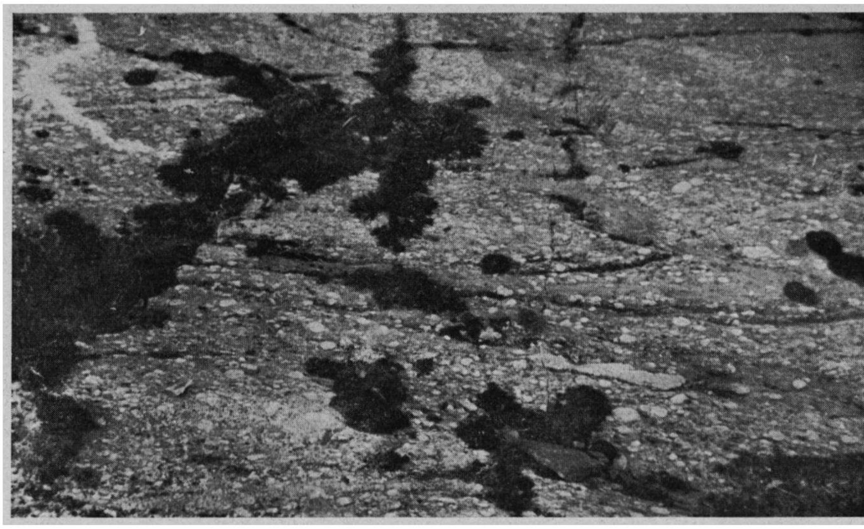

Fig. 3.

Augen-gneiss with large rounded eyes of feldspar and veins of quartz-epidote. The largest eyes have a size of 5 cm. Southwest of Ericksbyn.

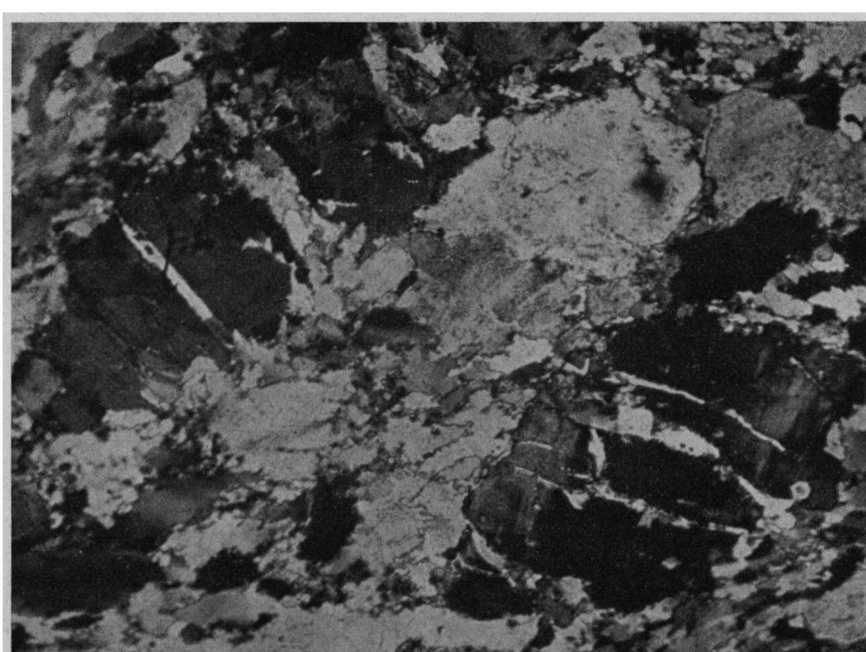

Fig. 4.

Photomicrograph of the Teåker aplite-granite with distinct cataclastic structure of the quartz (light) and the Na-potassium feldspar (dark). Crossed nicols. 35 X lin.

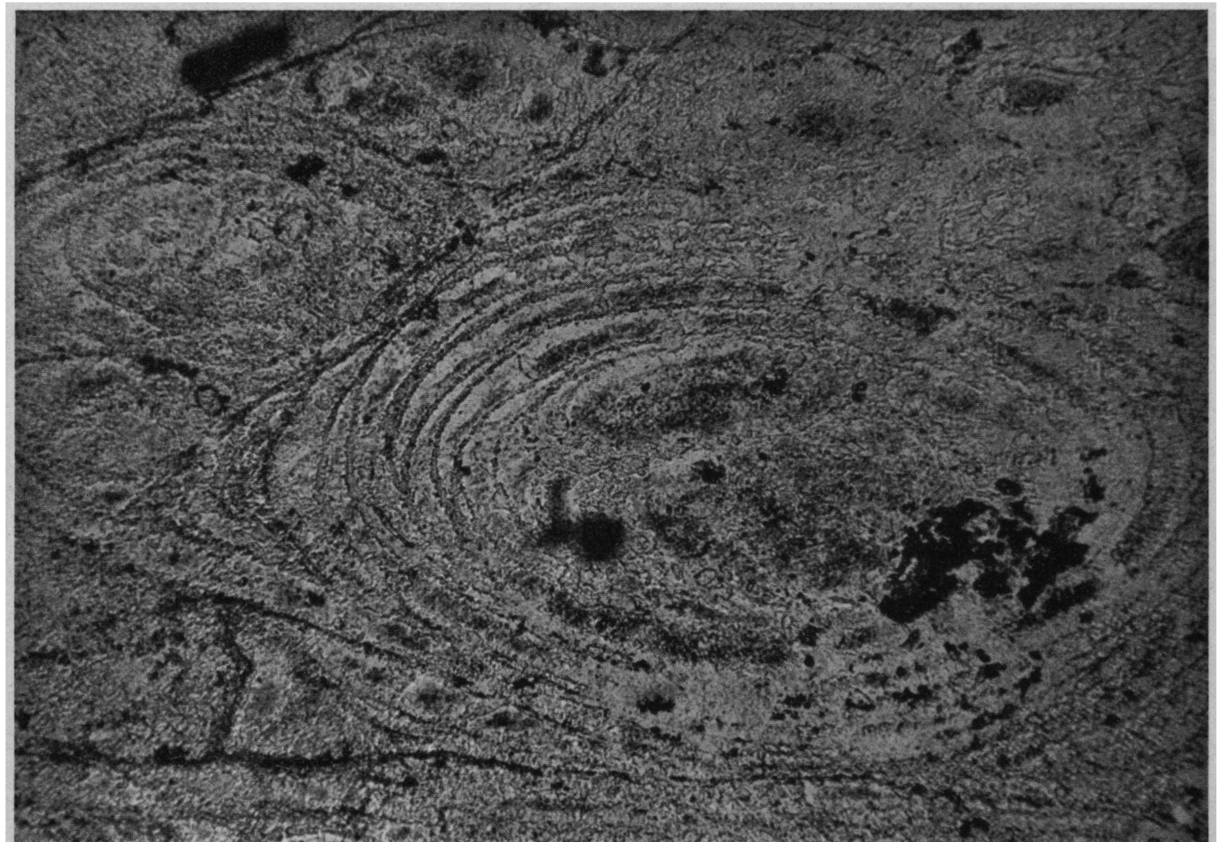

Fig. 5.

Photomicrograph of perlitic structure in a felsitic rhyolite from the eastern shore of N. Strutsåstjärn. 53 × lin.

consisting of sericite scales. With nicols crossed the perlitic structure is shown by the fact that the concentric circular or elliptical bands are cryptocrystalline, whilst the interlying mass is microcrystalline. The grain size of the granular microcrystalline parts may vary strongly (from d. = 0.01 mm to d. = 0.1 mm).

The *fluxion structure* too, is brought out by a difference of colour produced by the red-brown (hematite?) pigment which is concentrated mainly in the cryptocrystalline parts. When this pigment is lacking, the distribution of micro- and cryptocrystalline material may show a fluxion pattern under crossed nicols. Sometimes there is also a marked fluxion structure, exhibited by streaks of sericite. In some cases it becomes difficult to distinguish real fluxion structure from "pseudo-fluxion structure" caused by dynamometamorphism. Generally speaking, the former distinguish themselves from pseudo-structures by the sinuous, non-parallel trend of the sericite streaks.

The *brecciated structure* again, is brought out by difference of colour caused by the red-brown pigment (hematite?) or by the distribution of crypto- and microcrystalline parts.

The *lenticular structure* may result from cryptocrystalline lenses lying between granular strings or inversely by granular lenses being surrounded by cryptocrystalline strings. This structure too, must be due to the motion of the lava flow.

Spherulites have been found sometimes. They do not have well rounded forms but occur only as segments or in drawn-out shapes.

The above structures may be lacking altogether, the rock being then of equigranular structure.

Veins of quartz and albite, impregnating the rock, are of frequent occurrence. The crystallization is hypidiomorphic. The feldspars are, in contradistinction to those of the rock itself, perfectly clear. Sometimes it was found that they have not been affected by the dynamometamorphism to which the rocks have been submitted. In these cases the quartz crystals of the veins do not show the undulatory extinction of the other quartzes of the rock.

Minerals:

Porphyrocrysts [1]) are represented by: very sparse phenocrysts of *albite* (5 % An) with polysynthetic twinning. Sometimes they are small (d. av. = 0.15 mm, circa 10 × the grain size of the groundmass). They are seldom idiomorphic in that case, but most often amoeboidal [1]) so as to give the impression of having been corroded. Incidentally larger crystals occur, which are idiomorphic. These have occasionally been broken up (cataclastically, may be protoclastically). More often than not the cracks have been filled up with quartz. Rare large crystals of *pyrite* occur (d. up to 1 mm). Often the edge is weathered to brown and only the centre shows the bright yellow reflection colour.

The *groundmass: Quartz* and *feldspar* can only rarely be identified in the micro- to cryptocrystalline mass. Coarser crystallized quartz often occurs in lenses and strings. *Sericite* is dispersed in minute scales between the quartz and feldspar in varying quantities. Moreover, it appears in bands and streaks. The sericite is light apple-green pleochroic, but not always to the same degree (X = colourless to light lemon yellow, Y = Z = light green to green).

[1]) Cf. p. 67.

The scales are too small to allow further optical measurements. It is often associated with a mineral resembling leucoxene, especially in the streaks.

A mineral resembling *leucoxene* is of common occurrence, mostly in small dots or in long strings, often in the streaks of sericite, less often in larger patches, without crystal faces. In normally transmitted light it looks opaque, but in reflected light it shows the white reflection of highly refringent minerals in finely divided state.

Incidental small crystals of *epidote*, found exclusively within this opaque mass, show the same white reflection, which suggests that at least a part of this leucoxene-like mineral consists of fine epidote. More convincing indications in support of this view are brought in by the later study of the greenstones (p. 103). Another part must be accredited to residual matter left after the solution of carbonates, which too gives a white reflection. Since three out of four chemical analyses of the albite-rhyolites show a small content of Ti, true leucoxene may well be present.

Carbonate (in various rocks the presence of *calcite* can be demonstrated) occurs as larger crystals in the coarser parts; sometimes in the venules of quartz and albite and as the filling mass of small fissures, where it has a xenomorphic character. Individual small idiomorphic rhombohedra (d. up to 0.1 mm) occur scattered throughout the rock. The carbonate often has a dark rim. When it has gone into solution by surface water, there is a leucoxene-like pigment left in the pores. The carbonate therefore contains iron.

In many rocks a red-brown pigment occurs, readily observed under reflected light, which is probably finely divided *hematite*. This pigment is only found in those rocks which show a red to red-brown colour on the sample. Apparently, this colour is due to the pigment. Scarce small grains of ore, showing steel-blue reflection, are present. Sometimes idiomorphic regular crystal habit is to be observed, which might be an indication of *magnetite*. On rare occasions small amounts of *chlorite* (pennine) have been found in the sericite-leucoxene veinlets.

In some of the samples occur *cross-biotites*, scattered through the rock, as thin scales of varying size (0.25—1 mm). They are brown to greenish-brown pleochroic and seem to have originated from a green or brown pigment of irregular distribution. In cross-section one sees that the ends of the basal plane are ravelled and that the frays are diverging a little.

Analysis:

A chemical analysis of a felsitic rhyolite with perlitic relic structure showed the following result:

Analysis 1. Felsitic rhyolite.

Locality: Eastern shore of N. Strutsåstjärn, 190 m from the northern point.

Petrochemical Laboratory, Leiden (analyst: B. HAGEMAN).

Computed after P. NIGGLI's method[1]).

Weight %		Niggli values		Basis		Standard katanorm		Standard [1]) epinorm	
SiO_2	76.41	si	455	Kp	7.3	Mt	1.0	Hm	2.3
Al_2O_3	11.16	al	39	Ne	26.9	Hm	1.7	Ab	44.7
Fe_2O_3	3.29	fm	18	Cal	1.5	An	2.5	Cc	0.2
FeO	0.40	c	8	Cs	0.9	Or	12.1	Ms	3.5
MnO	0.04	alk	35	Fo	0.2	Ab	44.7	Or	9.6
MgO	0.10	h	10	Fa	0.5	Wo	1.2	Gram	0.4
CaO	1.10	co_2	2	Fs	3.5	En	0.3	Fe-Akt	1.0
Na_2O	4.78	k	0.21	Q	59.2	Q	36.5	Xon	1.3
K_2O	2.02	mg	0.05		100.0		100.0	Q	37.0
CO_2	0.22	w	0.88						100.0
H_2O_+	0.29								
H_2O_-	0.06								
	99.87								

Q 59.2	π 0.04	qz 215	Ab_{75} An_4 Or_{21}
L 35.7	γ 0.18	Magma type:	natronengadinitic
M 5.1	μ 0.04	Main group:	trondhjemitic

(P. NIGGLI 1936)

[1]) All analyses have been calculated according to P. NIGGLI's method, as set forth in C. BURRI and P. NIGGLI (1945, p. 24 f. and p. 579 f.).

In computing the standard epinorm, however, we followed the scheme of E. NIGGLI (1944, p. 239), in which first Cal, Sp and Hz under addition of Q are split up in C and Cs, Fo, Fa resp.. The scheme of C. BURRI and P. NIGGLI is somewhat different, in so far as the above mentioned segregation does not take place first, but instead begins with the formation of Cc, Mgs and Sid out of Cs, Fo and Fa resp., dependent on the amount of CO_2, by which process Q is set free. If sufficient CO_2 is available Cc, Mgs and Sid are subsequently formed out of Cal, Sp and Hz resp. and C is set free. This implies that whenever Cs is too small to bind all CO_2 as Cc, first Mgs and Sid are formed, before more Cc can be built on the base of Cal. This does not seem to me to be a natural course of events and seems therefore less correct.

The small grain size makes computation of the mode impossible; the mineral content, however, shows clearly that the standard epinorm comes much closer to the truth than the standard katanorm. The residual Fo, Fa and Cs which are used in building Gram, Fe-Akt and Xon in the standard epinorm will, in reality, have served partly for the formation of Ms (Fa and Fo) and partly for the formation of epidote (Fa and Cs). A little Fa, in combination with Fs, will have gone into Mt (at the expense of Hm).

Type II. Albite-rhyolite with albite and quartz phenocrysts

This type is characterized by distinct phenocrysts of quartz and albite. Just like the felsitic rhyolites they occur scattered throughout the Kappebo formation without any traceable regularity.

Macroscopical characters:

Red, red-brown or light grey, aphanitic rocks, sometimes massive, mostly foliated. Phenocrysts of quartz and red feldspar are at times very distinct; they can reach diameters up to 4 mm, most often, however, they are smaller and sometimes they are poorly, or not at all distinguishable. When the rocks are foliated, the foliation plane is dotted with little knots consisting of phenocrysts. Large idiomorphic crystals of pyrite (d. = 3 mm) occur rarely. The weathering crust is characterized by a fading out of the colour of the rock, eventually leading to a smudgy white colour at the surface.

Microscopical characters:

The groundmass is identical with that of the felsitic rhyolites, consisting of a micro- to cryptocrystalline granular to fibrous aggregate of quartz, feldspar and sericite. Here too, fluxion and brecciated structures as well as spherulites are found, but perlitic structure has not been observed. For a detailed description I may refer to what has been said about the felsitic rhyolites.

All transitions have been found from rocks with abundant, large phenocrysts of quartz and albite (d. up to 4 mm), which make about $^1/_4$ of the content, to those with very small and scarce phenocrysts (d. up to 0.3 mm). The albite phenocrysts are often gathered in small bunches.

Veins of quartz and albite with the same properties as in the felsitic rhyolites have also been found in these rocks.

Minerals:

For a discussion of the mineral contents of the *groundmass* we may refer again to what has been said in connection with the minerals of the groundmass of the felsitic rhyolites.

As *porphyrocrysts* [1]) occur: 1. phenocrysts of *quartz*, idiomorphic or hypidiomorphic by corrosion, mostly with undulatory extinction, sometimes incipient cataclastic. 2. phenocrysts of *albite*, often idiomorphic to hypidiomorphic,

[1]) Cf. p. 67.

frequently gathered in small groups so as to form eyes in the rock. If they are small the albites generally are hypidiomorphic to amoeboidal[1]) with, at times, definitely concave boundaries. The larger crystals most often are idiomorphic, but may have rounded corners. The amoeboidal shape as well as the rounded corners must be ascribed to corrosion. In the thin sections of some of the samples, in addition to the normally prismatic sections (b:l = 1:2 or 1:3), very long-drawn albites occur in which the proportion of length to breadth may be as high as 9:1. The albite phenocrysts are sometimes bent and broken up. Polysynthetic twinning is common; in the larger crystals "chessboard structure" is frequently observed.

3. Occasionally, *pyrite* is found in large idiomorphic crystals (d. up to 3 mm).

Analysis:

A chemical analysis of an albite-rhyolite of type II gave the following result:

Analysis 2. Albite-rhyolite with phenocrysts of quartz and albite.
Locality: 400 m south of Rösetjärn.
Petrochemical Laboratory, Leiden (analyst: B. Hageman).
Computed after P. Niggli's method[2]).

Weight %		Niggli values		Basis		Standard katanorm		Standard[2]) epinorm	
SiO_2	75.54	si	446	Ru	0.2	Ru	0.2	Cc	0.9
Al_2O_3	12.11	al	42	Kp	8.2	Mt	0.8	Ru	0.2
Fe_2O_3	2.25	fm	14	Ne	25.9	Hm	1.1	Hm	1.6
FeO	0.37	c	8.5	Cal	3.2	An	5.4	Ab	43.1
MnO	sp	alk	35.5	Cs	0.4	Or	13.6	Ms	7.3
MgO	0.25	ti	1.1	Fo	0.5	Ab	43.1	Or	8.5
CaO	1.34	co_2	5.8	Fa	0.4	Wo	0.5	Gram	1.0
Na_2O	4.72	h	9.6	Fs	2.4	En	0.7	Fe-Akt	0.9
K_2O	2.26	k	0.24	Q	58.8	Q	34.6	Xon	0.4
TiO_2	0.26	mg	0.15		100.0		100.0	Q	36.1
P_2O_5	sp	w	0.85						100.0
CO_2	0.74								
H_2O+	0.49								
H_2O-	0.15								
	100.48								

Q 58.8 0.09 qz 208 Ab_{69} An_9 Or_{22}
L 37.3 0.10 Magma type: natronengadinitic
M 3.9 0.13 Main group: trondhjemitic

(P. Niggli 1936)

[1]) Cf. p. 67.
[2]) See note analysis no. 1, p. 91.

The small grain size prevented the determination of the mode. All the same it becomes apparent from the mineral content that the standard epinorm gives a better approximation of the truth than does the standard katanorm.

Here too, however, the base-compounds, which build 5 % of the accessory constituents in the standard epinorm, will have to be distributed in a different manner, in order to constitute the observed minerals which are: leucoxene (epidote), hematite (pigment), calcite and chlorite.

Type III. Albite-rhyolyte with phenocrysts of albite only

The rocks belonging to this type are distinghuished from the foregoing type by the absence of quartz phenocrysts, but, as far as the groundmass is concerned, they are essentially identical with the types already mentioned.

Macroscopical characters:

Grey, red-brown, light red or yellow-brown, aphanitic rocks, with small phenocrysts of feldspar.

Microscopical characters:

The structure of these rocks forms a link between type II and type IV. The *groundmass* shows a close relationship to that of type I and II. Here again we find a crypto- to microcrystalline, granular to fibrous groundmass, consisting mainly of feldspar, quartz and sericite. *Skeletoncrystals of feldspar* have frequently been observed in the groundmass (l. = 0.15 mm), which are the equivalents of the idiomorphic albites of the second generation found in type IV. The accessory constituents are the same as described for type I.

As *phenocrysts*, only *albites* occur, which show polysynthetic twinning. Sometimes they are cataclastic and cloudy grey because of finely dispersed inclusions.

Fluxion structures, as well as *spherulites* have been encountered. The latter show marked transitions to a cryptogranophyric intergrowth of feldspar and quartz, which is particularly characteristic of type IV and will be described there.

Restricted to type III, however, are *nodules* with a maximum size equal to the albite phenocrysts. They consist of aggregates of granular quartz and are of limited but of rather regular frequency.

Type IV. Albite-rhyolite with a hypidiomorphic to poikilitic groundmass

Macroscopical characters:

Aphanitic, red-brown, violet-brown, light or dark grey rocks. In most cases phenocrysts of red feldspar, measuring a few mm, can be observed. The fracture, as in other types, is conchoidal. Occasionally, however, they present an extremely fine-grained surface. Sometimes small crystals of pyrite may be identified on a fresh fracture. Magnetite too may be abundant in the rock as tiny idiomorphic crystals.

Microscopical characters:

A few phenocrysts of albite are embedded in a very irregular groundmass, in which lie small laths of albite of a second generation, surrounded by a mass of feldspar, quartz and sericite with a variable amount of dark constituents, often in microcrystalline state.

The quartz and feldspar, at first sight finely divided, frequently appear to form larger aggregates with simultaneous, though undulatory extinction. They turn out to be feldspars of larger size, intergrown with numerous cryptocrystalline quartzes, which lend them a mottled appearance (fig. 7, p. 102). Transitions have been found to a distinct micrographic structure. Therefore we may use the term *crypto-granophyric* structure. Similar structures have been described by O. NORDENSKJÖLD (1892—'93, p. 207—208) in effusives of Småland and by N. SUNDIUS (1922, p. 105 and p. 123) in the phenocrysts of effusives of the uppermost horizon of helleflints of Grythyttefält. There are transitions from the undulatory extinction of these aggregates to the extinction typical for the spherulites. These transitions have been noticed already in the albite-rhyolites of type III. This might be suggestive of the existence of a continuous series. The granophyric crystal complexes can become so extensive as to enclose completely some albite laths of the second generation, eventually resulting in a *poikilitic structure.*

Sericite may occur in such profusion of minute scales that, with exception of the albite prisms, the groundmass becomes devoid of any structural elements, except for a pronounced foliation.

Veins of quartz and of quartz-calcite-epidote have been encountered.

Minerals:

Porphyrocrysts [1]) only occur as phenocrysts of *albite* (0—3 % An); marked polysynthetic twinning is common. The albites are idiomorphic or hypidiomorphic by corrosion, in which case they have rounded shapes. They are frequently cataclastic and often form clusters; their size varies considerably (l. = 0.5 to 4 mm). In some samples the usual prismatic albites are accompanied by very long-drawn, often xenomorphic, individuals (l : b = 8 : 1). The measured specimens of this type showed pericline-twins exclusively. The phenocrysts have many saussuritic inclusions.

In the *groundmass* the lath-shaped *albites* of the second generation are particularly conspicuous (l. av. = 0.1 mm; l. max. = 0.2 mm).

Identification of the xenomorphic *feldspars* with *granophyric* intergrowth of quartz, is extremely difficult because of their undulatory extinction. With the aid of the Fedoroff stage a single determination was obtained, namely $2V = -86°$, which would indicate albite-oligoclase. This result should be considered with every circumspection.

Quartz occurs, except in granophyric arrangement with feldspar, also in small xenomorphic crystals which occasionally may reach the size of the albite laths of the second generation. A few small veins and microcrystalline aggregates of quartz have been found.

Sericite occurs in scales; sometimes it forms streaks or nests. It may be apple-green or colourless as is usual. Colouring is most common in sericite when it is associated with a leucoxene-like mineral, e.g. in the streaks. Conse-

[1]) See p. 67.

quently, the green colour is probably due to the presence of iron. The amount of sericite may vary within wide limits, but forms always an important constituent.

In addition we find as non quartz-feldspar minerals substantially the same as found in the preceding types. However, in the various samples their relative frequency and association differ strongly. If this be due to some extent to local variations, for the most part it must be the result of differences in the amount of dark constituents in the rocks belonging to this type.

The most common of the minor constituents is a *leucoxene-like* mineral, as a rule evenly distributed throughout the rock, but sometimes concentrated in the sericite streaks. Sometimes this mineral has grains of ore in its centre and must be therefore real leucoxene. For the most part, however, it will be made up by finely divided *epidote* (see p. 90 and p. 103).

The red hued rocks are characterized in thin section by a red-brown pigment, probably consisting of extremely fine *hematite.*

Grains of ore, giving bluish-black reflections, frequently exhibiting idiomorphic regular crystal boundaries, are of common occurrence and increase steadily in quantity and size (d. from very small to 0.15 mm) as the samples become less acid. From a powder of such a less acid sample the ore was obtained by mechanical separation and the X-ray analysis of the concentrate showed *magnetite.* In some of the less acid samples a part of the magnetite is gathered to form strings.

Rare crystals of idiomorphic *pyrite,* weathered brown at the edges, have been found repeatedly.

Carbonate occurs in xenomorphic crystals of varying size and irregular distribution. More often than not it has a sieve-like appearance caused by the orderly arrangement of inclusions. Minute rhombohedra of carbonate (d. = 0.06 mm) also have been developed, at times with opaque edges. Occasionally a HCl test on the sample showed the presence of calcite. The amount of carbonate is subject to strong variation, independent of the acidity of the rock.

Epidote is present in more than negligible amounts only in the less acid members of this type. It may be finely dispersed throughout the rock and/or concentrated in trails and veins.

Chlorite (pennine) too is mainly confined to the less acid members and is then often associated with titanite and epidote, frequently forming clusters.

In a few instances *titanite* has been found, mostly of xenomorphic habit, in nests of chlorite (pennine). This mineral does not occur in any other type of the albite-rhyolites.

Scarce needles of *apatite* have been observed.

Analysis:

Two rocks of this type have been submitted to chemical analysis, one poor, the other rich in non quartz-feldspar minerals.

Analysis 3. Albite-rhyolite with phenocrysts of albite and a second generation of idiomorphic albite in the groundmass; poor in non quartz-feldspar minerals.

Locality: 150 m north of Tonebyn.

Petrochemical Laboratory, Leiden (analyst: B. HAGEMAN).

Computed after P. NIGGLI's method [1]).

Weight %		Niggli values		Basis		Standard katanorm		Standard [1]) epinorm	
SiO_2	74.89	si	435.5	Ru	0.2	Ru	0.2	Ru	0.2
Al_2O_3	12.24	al	42	Kp	14.3	Mt	0.7	Hm	1.9
Fe_2O_3	2.59	fm	14	Ne	19.8	Hm	1.4	Ab	33.0
EeO	0.24	c	9	Cal	3.4	An	5.6	Cc	0.5
MnO	0.03	alk	35	Cs	0.5	Or	23.9	Ms	8.1
MgO	0.16	ti	1.4	Fo	0.35	Ab	33.0	Or	18.0
CaO	1.45	co_2	3.1	Fa	0.35	Wo	0.7	Gram	0.7
Na_2O	3.57	h	10.5	Fs	2.8	En	0.5	Fe-Akt	0.7
K_2O	3.96	k	0.42	Q	58.3	Q	34.0	Xon	1.5
TiO_2	0.32	mg	0.10		100.0		100.0	Q	35.4
CO_2	0.39	w	0.89						100.0
H_2O+	0.54								
H_2O-	0.08								
	100.46								

Q 58.3	π 0.09	qz 195	Ab_{53} An_9 Or_{38}
L 37.5	γ 0.13	Magma type:	engadinite-granitic
M 4.2	μ 0.09	Main group:	leucogranitic

(P. NIGGLI 1936)

In this analysis too the standard epinorm gives a better approach to the reality than the standard katanorm. However, the accessory minerals according to the standard epinorm (5.5 %) are not in accordance with the facts. In this sample were found: epidote, possibly leucoxene, calcite and a little hematite pigment.

[1]) See note to analysis no. 1, 91.

Analysis 4. Albite-rhyolite with phenocrysts of albite and a second generation of idiomorphic albite in the groundmass; comparatively rich in non quartz-feldspar minerals.

Locality: 50 m to the east of the southern point of N. Strutsåstjärn.

Petrochemical Laboratory, Leiden (analyst: B. HAGEMAN).

Computed after P. NIGGLI's method [1]).

Weight %			Niggli values		Basis		Standard katanorm		Standard [1]) epinorm		Variant epinorm	
SiO_2	66.88		si	297	Ru	0.3	Cp	0.4	Cp	0.4	Cp	0.4
Al_2O_3	14.29		al	37	Cp	0.4	Ru	0.3	Ru	0.3	Ru	0.3
Fe_2O_3	4.09		fm	25.5	Kp	14.0	Mt	3.4	Hm	2.9	Ab	39.4
FeO	1.35		c	9	Ne	23.6	Hm	0.7	Ab	39.4	Cc	0.7
MnO	0.09		alk	28.5	Cal	4.9	An	8.1	Cc	0.7	Ep	3.5
MgO	0.99		ti	1.33	Sp	0.3	Or	23.4	Ms	12.3	Mt	3.4
CaO	1.85		p	0.45	Fo	2.0	Ab	39.4	Or	14.5	Hm	0.3
Na_2O	4.28	4.20[2])	co_2	3	Fa	1.7	Cord	0.5	Gram	4.1	Ms	9.1
K_2O	3.86	3.95[2])	h	18	Fs	4.4	En	2.7	Fe-Akt	2.5	Or	16.9
TiO_2	0.43		k	0.37	Q	48.4	Q	21.1	Fe-Ant	0.5	Ant	2.3
P_2O_5	0.25		mg	0.25					Q	22.4	Q	23.7
CO_2	0.53		w	0.36		100.0		100.0				
H_2O+	1.23									100.0		100.0
H_2O-	0.03											
	100.15											

Q 48.4	π 0.12	qz 83	Ab_{56} An_{11} Or_{33}
L 42.0	γ 0.00	Magma type:	tasnagranitic
M 8.7	μ 0.25	Main group:	granitic

(P. NIGGLI 1936)

The determination of the mode again is impossible, owing to the small grain size, in particular of the dark constituents. The non quartz-feldspar minerals of the analysed sample consist of: sericite, calcite, magnetite, epidote, possibly some leucoxene, a little chlorite and a little apatite. In reality therefore, the Gram and Fe-Akt of the standard epinorm are not formed. Since the non quartz-feldspar minerals are not too scarce here, there might be some sense in contriving a variant epinorm for this analysis and those of the greenstones, which will be treated subsequently. In this variation magnetite and epidote are formed, but, as no volumetric analysis can be made, its value is only a relative one. Nonetheless it would appear that this modified epinorm comes closer to reality.

In the calculation of this variation the following scheme is adopted:

[1]) See note analysis no. 1, p. 91.

[2]) Second determination.

1. Cp and Ru enter into account as such.
2. Ne and Q give Ab.
3. Cal, Sp, and Hz are segregated, under addition of Q, in C and Cs, Fo and Fa respectively.
4. Cc is formed from Cs corresponding to the amount of CO_2.
5. Formation of Ep (30 % Pi) according to the equation:

$$160\,Ep = 60\,Cs + 27\,Fs + 42\,C + 31\,Q$$

6. Formation of Mt out of Fs and Fa.
7. If Fs is in excess, formation of Hm.
8. After this the usual calculation of the standard epinorm is continued, starting with the formation of Ms.

Some peculiar rocks associated with type IV

The rocky hills in the vicinity of Lapperud, the western flanks of which rise steeply from Kappebosjön, fall outside the area of clastic and pyroclastic deposits of the Kappebo formation and are constituted chiefly by Kroppefjäll gneiss-granite. The latter has been penetrated by a great number of dykes of albite-rhyolite belonging to type IV. In this system three dykes have been found, whose outlook and mineral composition are diffirent.

A short description of these aberrant types of dyke rocks follows:

1. On the spot of the former Lapperud farm, a dark grey to black, dense rock with numerous veins of calcite has been found. Especially on a fresh fracture a great number of octahedra of magnetite are to be observed, with a maximum size of 0.5 mm. On microscopic examination this rock appears to be closely related to the others of type IV. The original structure has been largely blurred by the prolific formation of sericite.

A number of highly altered albite laths in the groundmass exhibit, in places, distinct relic structures of hypidiomorphic granular character. The quartz has escaped the general sericitization and lies as clear xenomorphic grains between the altered feldspar (one half of which has been converted into sericite). Much xenomorphic calcite occurs in the groundmass. A leucoxene-like mineral (epidote?), in a finely divided state, has been found abundantly.

In the groundmass just described occur large porphyrocrysts[1]) of magnetite (d. max. = 0.5 mm and d. av. = 0.15 mm).

They often form skeleton crystals enclosing parts of the groundmass. This leads to the conclusion that they are porphyroblasts. Small quantities of chlorite (pennine) are commonly found near these porphyroblasts of magnetite.

2. At the cape in Kappebosjön to the north of Lapperud, a peculiar kind of rock has been found close to the water-level (fig. 6). It is a red-brown, aphanitic rock, in which are embedded irregular long-drawn, angular patches of a dark greyish-green rock. These patches have a length of 5 to 10 cm and lie at random. Alongside some of the foliation planes, occurring at

[1]) Cf. p. 67.

large intervals, thin bands of that dark greyish-green material have developed. The entire surface of the rock is sown with small rhombohedral holes (d. max. = 2 mm). The distribution and size of these holes are irrespective of the greyish-green and red-brown parts of the rock. On a fresh fracture, the original filling of the holes is still intact and appears to be made up of a white carbonate, which gives no reaction with diluted HCl.

On microscopic examination, the red parts of the rock are found to consist of a microcrystalline aggregate of quartz, feldspar and sericite, with finely dispersed leucoxene, showing in places relics of a hypidiomorphic granular structure. The darker parts of the rock are characterized in thin section by the abundance of chlorite (pennine), which for the rest are the same as the lighter parts. This circumstance gives them a striking resemblance to some of the basic effusives (greenstones) which, in this area, are associated with the albite-rhyolites and will be discussed subsequently.

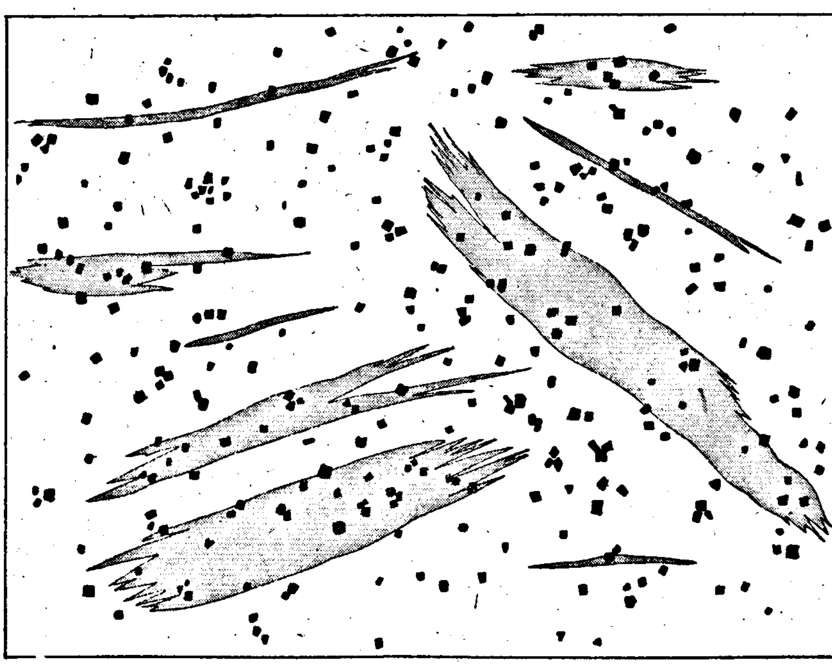

Fig. 6.

Surface of the albite-rhyolite, on the shore of Kappebosjön north of Lapperud, with patches of greenstone and holes due to weathering of dolomite porphyrocrysts. App. ½ nat. size. The foliation is horizontal.

In this groundmass, both in the lighter and the darker parts, lie big hypidiomorphic porphyrocrysts[1]) of carbonate (d. av. = 1.5 mm). Wavy extinction, bent crystal faces and lamellae give evidence of the dynamometamorphism to which they have been subjected. Often they are sieve-like in appearance owing to the numerous inclusions of groundmass material.

[1]) Cf. p. 67.

This might be a reason to consider these porphyrocrysts as porphyroblasts. The X-ray analysis of the powder of the carbonate showed the porphyroblasts to consist of dolomite.

3. On the spit of land between Öjerud and Lapperud a third peculiar kind of rock appears as a dyke in the Kroppefjäll gneiss-granite. Macrospically it is a foliated ore with veins of calcite. Under the microscope the rock is found to be essentially made up of sericite and ore. They are intermingled in contorted trails reflecting a fluxion structure. In this sericite-ore assemblage lie grains, looking like leucoxene, but presumably consisting of epidote. From a powder sample the ore has been separated mechanically and an X-ray diagram proved it to be hematite. A number of quartz-feldspar lenses and veins are lying in the groundmass of sericite and ore. In the veins, calcite and chlorite occur. Calcite may become predominant, the small veins linking up into larger ones. These again are the veins of calcite which macroscopically can be seen and may attain a width of 2 cm.

Older analyses of helleflints

Apart from the new analyses given above, four analyses of helleflints are known from the area under discussion, whereas one more has been made of a helleflint just north of the northwestern boundary of the present map.

All five of them may be found in the list of analyses compiled by H. Santesson (1877) as well as in W. Larsson's (1932—'33).

Of these analyses three have been prepared from a helleflint situated between the slates of the Dal formation and the overthrust gneiss-granite, in the slate-quarry of Halängen (Santesson 1877, p. 60, no. 47, 50 and 51; Larsson 1932—'33, p. 146, no 835; p. 162, no 962; and p. 164, no 967). The helleflint providing the analyses is by no means related to the outflows of the Kappebo formations, but must be considered as a friction mylonite, formed at the thrust plane of the great horizontal Lysesjö thrust mass which will be dealt with in a later chapter. The same remark holds for the analysis of the helleflint near the croft of Dalen, south of Vägtjärn and due north of the northwestern margin of the map. (Santesson 1877, p. 60, no 37; Larsson 1932—'33, p. 144 no 821).

One further analysis is known of a helleflint which is to be included in the Kappebo formation and has been found south of Damtjärn in the Kappebofjäll. (Santesson 1877, p. 58, no 4; Larsson 1932—'33, p. 144 no. 818). The Niggli values of this analysis are (Larsson 1932—'33) si = 331, al = 11, fm = 47, c = 4.5, alk = 37.5, k = 0.34 and mg = 0.14. Hence it becomes apparent that this analysis does not fit at all within the picture of the new analyses with its extraordinary high *fm* and strongly negative *al — alk*. As neither the sample nor the thin section of this helleflint was available and as there is a fair possibility that the rock is of pyroclastic origin, this analysis has been neglected.

Greenstone

Associated with the albite-rhyolites, flows and dykes of greenstone are of common occurrence. This association is particularly conspicuous and convincing where both types of rocks form dykes in the Kroppefjäll gneiss-granité, as they do north of Tonebyn and near Åsen. The albite-rhyolites, however, outnumber the greenstones. This is especially the case in the neighbourhood of Lapperud. Less striking, but still apparent, is the association in the region of the Kappebo formation. Here, too, both rocks occur together in many localities. There is no evidence for the assumption of a time interval between the extrusions of albite-rhyolite and greenstone, because the pyroclastic members of both of them occur intermingled.

The less acid members of the albite-rhyolites therefore must, in my opinion, be regarded as intermediaries to greenstones. In this connection we may call attention to the peculiar type of rock encountered at the shore of Kappebosjön, to the north of Lapperud (p. 99). The green spots of this rock show close affinities to some of the greenstones, whilst the brown-red bulk must be classed among the albite-rhyolites. A further close relationship of the greenstones to the albite-rhyolites is brought out by the nature of the dark constituents.

The macroscopical character of the greenstones as a whole is rather monotonous. They are essentially dark green, aphanitic to fine-grained rocks with a few reddish phenocrysts of feldspar. Microscopically, however, various types may be distinguished in relation to their structures. Viz.:

Type I: Crypto- to microcrystalline greenstone, with occasional fluxion structure.

Type II: Greenstone with relics of a microcrystalline hypidiomorphic to intersertal structure.

Type III: Greenstone with relics of meso- to macrocrystalline intersertal structure.

Transitions are found between types I and II. Type III is represented only by two samples. Greenstones, just like albite-rhyolites, exhibit a varying degree of schistosity. This is due to differences of intensity of metamorphism, as well as to the variable amount of fibrous minerals (sericite and chlorite) which react readily upon dynamometamorphism.

Type I. Crypto- to microcrystalline greenstone

In these fine-grained rocks fluxion structures have been found whose tortuosities must be traced back to the flowing of lava, thus being suggestive of the fact that at least a part of these rocks is of effusive origin. When these structures are lacking, then the rocks may have been formed, in part, as tuffs, since hypidiomorphic or intersertal structure has not been met with. In those cases where there is no indication whatever of a tuffaceous origin the rocks have been included in this group.

Macroscopical characters:

Mostly dark green, aphanitic rocks which carry light green veins and spots of epidote. These veins may widen to form nodules. In the larger veins and nodules calcite can be often macroscopically identified. Many of the small veins have been folded and broken up. Frequently the green rock has light pink flames. The flames range from sizes of a few cm up to several m. The larger ones contain often pale green strings of epidote, which are probably fluxion structures.

Fig. 7.

Photomicrograph of an albite-rhyolite of type IV showing poikilitic structure with idiomorphic albites of the second generation lying in cryptogranophyric crystal compexes. Crossed nicols. 135 × lin.

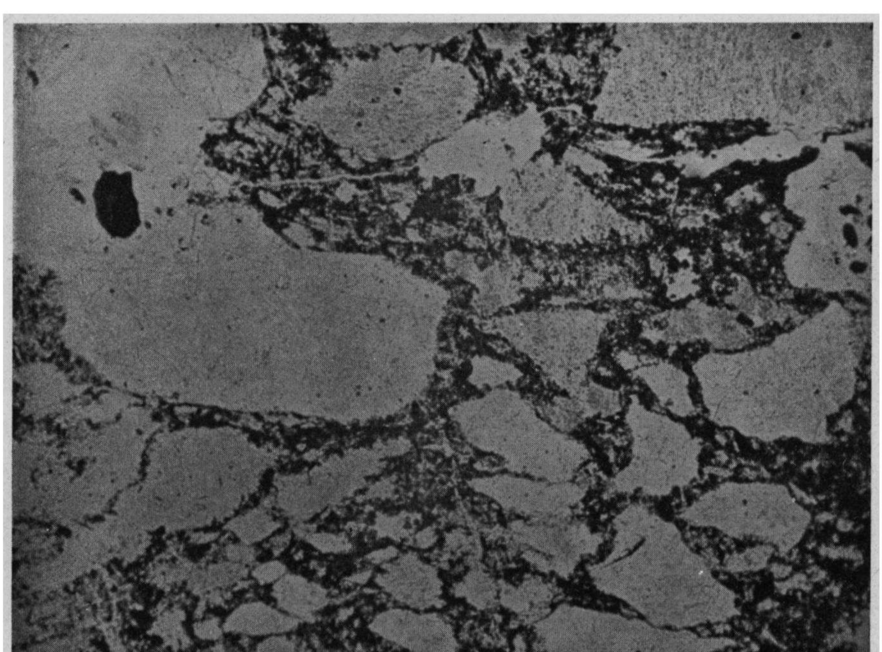

Fig. 8.

Photomicrograph of a graywacke of the Kappebo formation showing the size difference and the angular form of the grains. The feldspars are slightly cloudy owing to alteration products. Ordinary light. 33 × lin.

Fig. 9.

Ripple marks in the arkosic quartzitic sandstone of the basal layers of the Dal formation. Northeast of Helvetestjärn.

Fig. 10.

Intercalated conglomerate with quartzite pebbles in the basal layers of the Dal formation. Northeast of Helvetestjärn.

Microscopical characters:

Under the microscope the rocks appear to be crypto- to microcrystalline equigranular, with the exception of a few albite phenocrysts. They may be of decussate structure, but mostly they have a marked foliation which may include all minerals, except epidote. The lighter constituents are represented by phenocrysts of albite, while quartz and feldspar occur in the groundmass. The *albite phenocrysts* (av. 2 % An; d. av. = 0.5 mm) are scarce, but found regularly.

The small grain size makes it difficult, if not impossible, to estimate the *quartz/feldspar* ratio in the groundmass, but both minerals are always present, though in varying quantities. The determination of a few feldspar grains of the groundmass succeeded by means of the Fedoroff universal stage. They proved to be albite (2 % An).

The non quartz-feldspar minerals are represented mainly by epidote along with chlorite, sericite, magnetite and calcite, in strongly varying proportions. The amount of non quartz-feldspar minerals is far from being constant, thus causing the light rose and dark green patches which often lend the rocks a flamed appearance.

Epidote occurs in small grains scattered throughout the rock; incidental porphyrocrysts (l. up to 0.5 mm) have also been found. The small grains of epidote are not always distributed evenly throughout the rock, but are concentrated in given tracts. In the extreme case a continuous field of epidote results in which accessory small amounts of quartz and chlorite appear. As a consequence these concentrations are not sharply divided from the surrounding rock, as are the well defined nodules which also occur. The latter consist again of epidote, epidote and quartz or epidote, quartz and calcite, chlorite being always present in small quantities. When quartz appears in the nodules, it often occupies the centre and epidote the edges. Both kinds of epidote or epidote-quartz assemblages can vary considerably in size (d. = 0.3 to 2 mm). Macroscopically they appear as light green spots in the dark green rock. Some nodules have long-drawn shapes in thin section, so that they look like sausages. Possibly they are sections of folded veins.

Together with epidote, a leucoxene-like opaque looking mineral is met everywhere; the same as is frequently found in the albite-rhyolites. However, in the greenstones it is encountered together with epidote in larger quantities and, with fair certainty, we may ascribe this leucoxene-like mass to finely dispersed epidote. For this view the following reasons can be brought forward:

1. The strict association with epidote.
2. The leucoxene-like mass encloses small crystals of epidote, which give the same white reflection in reflected light.
3. Even distinctly crystallized epidote gives white reflections on the edges.
4. With strong magnification and condensor the mass proves to be not wholly opaque.

Idiomorphic *magnetite* is frequent as an accessory mineral; sometimes it is abundant. The *sericite* found in the greenstones, usually is not apple green, as it is in many albite-rhyolites, but of the normal colourless type. In these cryptocrystalline greenstones it is abundant as scales and fibres. *Chlorite* is common, sometimes abundant and is finely dispersed in the rock. It is deep green to pale yellow pleochroic and shows an anomalous interference colour (brown), distinctive for pennine. Less common is a colourless to light brown-green pleochroic *biotite*, altered partially or mainly into chlorite. The light colour must be due to discoloration. *Calcite* appears in greenstones of this type only in nodules and fissures.

Frequently, distinct *fissure veins* are also found, which are filled with epidote, quartz, calcite and sericite, in various combinations. On one occasion a fissure vein of albite (2 % An), showing needle-like inclusions, was found.

A few greenstones have been found, whose foliation planes have a brown hue caused by small crystals of biotite. In thin section they appear to be light brown to brown-black pleochroic biotites, developed in fine scales and orientated irrespective of the schistosity. These *cross-biotites* are not confined to the foliation planes, but may occur at random in the massive parts of the rock. They are in everything comparable to those of the albite-rhyolites. Here too, there is a peculiar ravelling at the ends of the basal plane.

In several greenstones of this kind *fluxion structures* have been found, exhibited in different ways:

1. By sinuous strings of an increased epidote or epidote-sericite concentration.
2. By sinuous strings with abundant crystals of magnetite.

Between these strings, lenses of coarser, granular quartz occur, in which calcite, feldspar and chlorite may be present. There are also eyes which consist, or have consisted, of one single crystal of quartz, attaining sizes up to 1 mm. These porphyrocrysts have been broken up, but the simultaneous extinction shows that the fragments originally formed one single crystal.

A few cryptocrystalline greenstones have been found in which actinolite is the most important mineral. They are dark green, almost black rocks, showing only the smallest amount of light constituents in thin section. These occur as scarce *feldspars*, partly altered into saussurite, and *quartz* porphyrocrysts. In addition there is some feldspar in the groundmass.

The bulk of the groundmass, however, is made up by thin needles of colourless to light green pleochroic *actinolite* with a maximum extinction angle of 20°. The needles have a length of 0.07 mm and a breadth of 0.006 mm. The light colour, therefore, is the result of their small thickness. Possibly *sericite* is present in fine scales between the actinolite needles, but it is impossible to be certain because of the minute dimensions of the crystals. Fine-grained *epidote* is also found in considerable quantity and a few small grains of *ore* with dark brown reflecting edges. *Chlorite* (pennine) and green *biotite* are minor constituents.

The most remarkable feature of this rock is the occurrence of scarce porphyrocrysts of quartz, reaching up to 0.5 mm in size. Incidentally, these crystals have been invaded from the edges by needles of actinolite. In one individual the original idiomorphic hexagonal habit of the quartz is still clearly exhibited, though the centre of the crystal has been converted completely into epidote and actinolite. Although the remaining edge has been split up, the fragments still have simultaneous extinction.

A greenstone of somewhat different habit, found 450 m east of Kleven, also belongs to this type. It consists of microcrystalline quartz and feldspar and fibrous sericite, biotite and chlorite in irregular distribution. Numerous hypidiomorphic crystals of magnetite of variable size (d. max. = 0.2 mm) lie scattered throughout the rock. The peculiarity of the rock, however, lies in the long-drawn bent nodules, filled with calcite in the centre and chlorite towards the rim. At the interface of these two minerals epidote has been found, which penetrates into the calcite as idiomorphic stalks. The more or less linear arrangement of these nodules and their shape suggest the idea of a fissure-vein, that has been folded and broken up. Calcite also occurs outside these nodules in large xenomorphic crystals with frequent inclusions. A few small rhombohedra of calcite have been observed.

Type II. Greenstone with relics of microcrystalline hypidiomorphic to intersertal structure

Macroscopical characters:

In the field the greenstones of this type cannot be distinguished from the foregoing type. They are greyish green, dark green or almost black aphanitic to fine-grained, more or less foliated rocks. On a fresh fracture one can often discover light green or white spots, due to epidote and calcite resp.; occasional reddish phenocrysts of feldspar have been met. Folded fissure-veins also occur in this type.

Microscopical characters:

Often microcrystalline, essentially equigranular rocks, in which *albite* is important, occurring as grains throughout the rock, but at times relics are found of lath-shaped crystals with crenate edges caused by alteration. In this case the albite laths indicate the original hypidiomorphic or intersertal structure. Of this structure only a few scattered remains can be found; apparently the rock has gone through many alterations since the formation of the albite laths. Albite (3 % An) appears also as phenocrysts (d. = 0.2 mm to 1.2 mm), showing polysynthetic twinning and containing numerous inclusions of saussurite (= zoisite, epidote, calcite, chlorite).

Free *quartz* is always present in small amounts and in xenomorphic shapes, sometimes with inclusions of *rutile* needles.

We find again the same non quartz-feldspar minerals as in the greenstones of type I. A short description therefore may suffice. *Epidote* is often predominant. *Calcite* is common in larger or smaller quantities, irregularly distributed. It is xenomorphic, always considerably larger than the other minerals and has a sieve-like look as a result of numerous inclusions. One sample yielded calcite in large quantities (d. av. = 0.25 mm) with a great number of flaky brown inclusions, too small to be identified. *Magnetite* in idiomorphic crystals of varying size (d. av. = 0.08 mm, d. max = 0.2 mm) is encountered regularly. *Sericite* in this type of greenstones is generally of minor importance, whilst *pennine* is mostly abundant. In a few dark coloured stones *tremolite* and *actinolite* occur as fine needles. Tremolite is entirely colourless; actinolite colourless to light green pleochroic. The needles are distributed irrespective of the relics of igneous structure and have grown criss-cross through the feldspars, occurring also in trails. This indicates a late-magmatic or a secondary origin of the minerals. In some greenstones of this type *biotite* has been found. It has a pleochroism from colourless to light brown-green or from pale yellow-green to intense green and is partially altered into chlorite. Rarely, small *cross-biotites* have been encountered. In some of these greenstones occur trails of cryptocrystalline matter between the parts with igneous structure, thus forming transitions to the foregoing type.

The quantities as well as the mutual proportions and association of the dark constituents are far from being constant. This may be conveniently illustrated by the tabulation below, in which are listed the dark constituents and their relative frequencies of eight samples:

Sample	I	II	III	IV	V	VI	VII	VIII
epidote	XX	XX	X	XXX	XXX	XXX	XX	XX
chlorite	sp	XXX	X	X		X	X	X
magnetite	X	X	XX	XX			X	X
calcite		X	XXX	XX	XX	X	X	
sericite	XXX	X	X		sp			X
actinolite						XX	XXX	
tremolite					XX		X	
biotite							sp	XXX

XXX abundant among the dark constituents
XX important constituent
X minor constituent
sp accessory constituent

The order of the samples indicates an increasing quantity of non quartz-feldspar minerals, made up by estimation.

The variable association of the minerals is clearly shown by the table. Epidote appears to be by far the most common mineral whilst tremolite and actinolite and even more so biotite are restricted to the more basic samples.

Analyses:

Of the greenstones of type II two analyses have been made. One (anal. no. 5) represents a greenstone relatively poor in dark constituents (sample II of the above table) of which chlorite is the most important and another (anal. no. 6) of a dark greenstone with much actinolite (sample VII of the above table).

Analysis 5. Greenstone.

Locality: 150 m to the north of Tonebyn.

Petrochemical Laboratory, Leiden (analyst: B. HAGEMAN).

Computed after P. NIGGLI's method [1]).

Weight %		Niggli values		Basis		Standard katanorm		Standard [1]) epinorm		Variant [2]) epinorm	
SiO_2	53.72	si	161	Ru	1.2	Ru	1.2	Ru	1.2	Ru	1.2
Al_2O_3	17.00	al	30	Cp	0.3	Cp	0.3	Cp	0.3	Cp	0.3
Fe_2O_3	5.85	fm	40.5	Kp	6.8	Mt	6.2	Hm	4.1	Ab	58.7
FeO	3.78	c	7	Ne	35.3	An	11.0	Ab	58.7	Cc	0.5
MnO	0.04	alk	22.5	Cal	6.6	Or	11.4	Cc	0.5	Ep	6.7
MgO	4.01	ti	3.8	Sp	0.7	Ab	58.7	Ms	16.0	Mt	5.1
CaO	2.28	p	0.3	Fo	8.1	Cord	1.3	Zo	1.1	Ms	11.0
Na_2O	6.46	co_2	1.6	Fa	4.5	Hy	1.6	Gram	10.5	Or	2.7
K_2O	1.87	h	2.7	Fs	6.2	Fa	0.2	Fa	4.5	Ant	9.3
TiO_2	1.67	k	0.16	Q	30.3	Fo	8.1	Fo	3.1	Fe-Ant	2.2
P_2O_5	0.27	mg	0.44		100.0		100.0		100.0	Q	2.3
CO_2	0.39	w	0.58								100.0
H_2O+	2.76	o	0.32								
H_2O-	0.10										
	100.20										

Q 30.3 π 0.14 qz — 29 Ab_{72} An_{14} Or_{14}

L 48.7 γ 0.00 Magma type: Melanatronsyenitic

M 20.7 μ 0.43 Main group: Natronsyenitic

(P. NIGGLI 1936)

The small grain size did not allow the calculation of a mode, but the variant epinorm corresponds best with the actual mineral contents. However, in reality, amesite is present in the chlorite (pennine). The amount of C required for the formation of amesite can be obtained by decreasing the percentage of Ms, as this appears in excess of the actual amount. From the Kp, which is set free hereby, Or is formed. Since numerical values are lacking, this computation has not been carried out.

[1]) See note to analysis no. 1, p. 91.

[2]) For the calculation of this variant epinorm see analysis no. 4, p. 98.

Analysis 6. Actinolite-bearing greenstone.
Locality: 800 m to the west of Dansbo.
Petrochemical Laboratory, Leiden (analyst: B. HAGEMAN).
Computed after P. NIGGLI's method [1]).

Weight %		Niggli values		Basis		Standard katanorm		Standard [1]) epinorm	
SiO_2	50.70	si	130	Ru	1.2	Ru	1.2	Ru	1.2
Al_2O_3	16.09	al	24	Kp	7.8	Mt	4.9	Hm	3.3
Fe_2O_3	4.62	fm	43	Ne	24.0	An	18.1	Ab	40.0
FeO	4.52	c	18	Cal	10.9	Or	13.0	Cc	0.5
MnO	0.19	alk	15	Cs	4.6	Ab	40.0	Ms	18.2
MgO	6.44	ti	3.3	Fo	13.5	Wo	6.1	Zo	5.6
CaO	6.63	co_2	1.5	Fa	5.5	Hy	0.4	Gram	17.0
Na_2O	4.38	h	17.9	Fs	4.9	Fa	2.8	Cs	3.7
K_2O	2.17	k	0.25	Q	27.6	Fo	13.5	Fo	5.0
TiO_2	1.74	mg	0.57		100.0		100.0	Fa	5.5
CO_2	0.42	w	0.24						100.0
H_2O+	2.10								
H_2O-	0.09								
	100.09								

Q 27.6	π 0.27	qz − 30	Ab_{56} An_{26} Or_{18}
L 42.7	γ 0.16	Magma type:	Mugearitic
M 29.7	μ 0.45	Main group:	Natrongabbroitic

(P. NIGGLI 1936)

There is no point in computing a modified epinorm here, as a satisfactory picture of the mineral contents appears only to be obtainable after the mode has been determined and this is prevented by the smallness of the grain size.

Type III. Greenstone with relics of meso- to macrocrystalline intersertal structure

Although, on microscopic examination, only two samples prove to possess these coarse relic intersertal structures, this constitutes a character so different from what we have seen previously, that it is justifiable to consider them as a type apart.

One of these two greenstones has been found 30 m from the eastern shore of Kappebosjön, 850 m north of Höljen. Macroscopically, the rock is dark green in appearance, fine-grained, slightly foliated and with numer-

[1]) See note to analysis no. 1, p. 91.

ous grey spots which appear to be calcite. Thus, in the field, it looks in no way different from other greenstones.

The microscopic examination reveals relics of igneous structure, which indicate that the rock has a meso- to macrocrystalline magmatic origin, e.g. strongly altered and corroded remains of long prismatic plagioclase have been found (measuring 0.7 mm). When several such crystals lie together a relic intersertal structure may be recognized, the interstices being filled up with epidote.

The bulk of the rock, however, is made up of a fine but unevenly grained aggregate without definite structure, consisting of *feldspar, quartz, epidote, chlorite,* coarse grains of *ore* rimmed with *leucoxene* (d. av. = 0.25 mm) and irregularly distributed xenomorphic *calcite.* Schistosity is practically non-existent. Xenomorphic calcite may occur in large crystals and is most often crowded with inclusions of light constituents. Moreover, inclusions of epidote and chlorite have been found occasionally.

In the aggregate just described occur clusters of epidote formed by a number of crystals with different optical orientation and separated by small veins of chlorite. These clusters may show conspicuously straight boundaries. The rectilinear boundaries are determined partially by the remaining big laths of feldspar which bring about an intersertal structure. Even in those places where the feldspars have been entirely altered and disintegrated, the relics of intersertal structure are recognizable from the acute angular shape and straight edges of the epidote clusters. There are also aggregates of epidote whose outlines indicate the form of idiomorphic pyroxene (edges of the faces 100, 110 and 010 or 110 and 111). Apparently these are pseudomorphs of epidote with some chlorite after pyroxene.

A vein has been observed in the rock filled chiefly with stalky quartz and chlorite. The stalks are at right angles to the wall of the vein, epidote and ore being interstitial between them. In this vein lie a few large crystals of quartz, which enclose idiomorphic needles of epidote and apatite, along with some chlorite.

In this connection a greenstone which outcrops on the Kroppefjäll plateau due southeast of Vägtjärn may be mentioned, which consists of an irregular aggregate of chlorite and epidote, calcite, sericite, magnetite, some quartz and feldspar without any traceable structure. In this rock remains of amphiboles largely altered into chlorite have been discovered.

The second greenstone belonging to this type is found 250 m south of Ängen, among the Kroppefjäll gneiss-granite. Owing to a dense overgrowth of wood and mosses it is impossible to decide whether it is a dyke penetrating the gneiss-granite. Under the microscope, however, the rock shows a well preserved primary structure, in contrast to the strongly cataclastic gneiss-granite itself, in which all traces of any primary structure have disappeared. It may be logical therefore to include this greenstone in the Kappebo formation.

Its outward appearance is that of a dense dark rock, sown with innumerable white spots (d. = 2 mm) which appear to consist of calcite; they are weathered out at the surface, as a result of which it has become pecked with small holes.

Microscopically it proves to be a porphyric rock made up of feldspar and ore, with phenocrysts (d. = 1.5 mm) consisting of feldspar, completely altered into sericite. The groundmass shows a distinct intersertal structure of small feldspar laths (l. = 0.25 mm) altered, to a large extent, into sericite and saussurite. The idiomorphic shape is blurred by the abundant alteration products, but the polysynthetic twinning may be still recognizable. The interstices are filled by a granular ore, while the whole is traversed by long thin needles of ore. Conjointly with the ore occurs some chlorite. The calcite porphyrocrysts [1]) (d. up to 2 mm) lie in this groundmass usually without crystallographic form and contain numerous inclusions of ore. It is worth noting that the ore, both in the calcite and in the groundmass, is identical in form and composition. This indicates that the calcite is of later, and consequently of porphyroblastic, origin. Free quartz has not been encountered.

Rhyolite

In the Kroppefjäll gneiss-granite to the west of Svartetjärn and to the south of Blixerud, some rhyolites have been found which are entirely similar to the quartz-porphyries occurring in the Kroppefjäll gneiss-granite "mylonite" between Ånimmen and Vänern, described by van Overeem (1948, p. 12—17). Because of the dense vegetation it is not possible to decide by direct observations in the studied area, whether they occur as dykes in the gneiss-granite. But this can be concluded from the striking difference in dynamometamorphism with the gneiss-granite which in this region is altered into a granite-schist.

Microscopically the rhyolites are distinctly different in character from the albite-rhyolites. The quartz phenocrysts (d. up to 3 mm) are hypidiomorphic, corroded and somewhat undulatory. The feldspar phenocrysts (d. = 2 to 3 mm) are most frequently altered entirely, or nearly so, into sericite and a finely divided dark pigment, in which occur a few grains of epidote. The shape of these altered aggregates is often that of idiomorphic feldspar. Occasionally remnants of feldspar have been found in the alteration products. Next to these almost completely altered feldspars, there are also phenocrysts of feldspar, only slightly sericitized, which appear to consist of albite since they show polysynthetic twinning und are less refringent than quartz. These phenocrysts are idiomorphic or amoeboidal by corrosion. Also epidote-chlorite clusters have been found, which, on account of their form, must be considered as altered biotite. The groundmass, consisting essentially of granular quartz and feldspar in microcrystalline distribution, is somewhat coarser (d. av. = 0.08 mm) than that of the albite-rhyolites, but differs especially by reason of the frequent grains of epidote and chlorite, while sericite is decidedly less numerous. Sparse grains of ore and small rhombohedra of calcite have been encountered.

The camera lucida drawing given by van Overeem (1948, p. 14, fig. 4) of a thin section of his quartz-porphyry is very characteristic for these rocks and might as well have been prepared from one of my own thin sections. However, van Overeem reports large phenocrysts of Na-potassium feldspars in this rock, whereas in the rhyolites of my area only albite phenocrysts have been found besides the entirely or almost entirely altered feldspars, which cannot be determined with more precision. As a matter of fact, upon inspection of the thin sections of van Overeem's collection, I have been able

[1]) Cf. p. 67.

to recognize a few phenocrysts of Na-potassium feldspar in the quartz-porphyry and the porphyrite associated with it. Yet, here too, by far the best part of the feldspar phenocrysts has been altered to such a degree that the determination of their character is no longer possible. I found also that the quartz-porphyries and the porphyrites look strikingly similar in structure and differ only by the presence or absence of quartz phenocrysts and by the amount of biotite phenocrysts and of quartz in the groundmass, which in my opinion, has not been sufficiently stated in VAN OVEREEM's descritption. Moreover, the quartz-porphyries and the porphyrites are linked to one another by transitional forms. Hence it is evident that they are related genetically, originating from one and the same magma.

The Niggli values of the analyses of these rocks given by VAN OVEREEM (1948, p. 15 and 16) are the following:

	si	al	fm	c	alk	k	mg
quartz-porphyry	363.5	45.5	14	8.5	32	0.45	0.56
porphyrite	221	38	18.5	21	22.5	0.31	0.55

The analysis of the quartz-porphyry shows a striking similarity with the analyses from the albite-rhyolites of the Kappebo formation, given above (analyses 1, 2 and 3).

THE SEDIMENTS

The sedimentary rocks of the Kappebo formation are composed of psephites, psammites and pelites, which genetically fall into two groups viz. epiclastic and pyroclastic sediments. The first group is represented by slates, graywackes, breccias and conglomerates, the second by tuffs and agglomerates. These two groups are linked by transitional forms. For one thing this is due to the fact that the sedimentation of the epiclasts mainly coincided with the volcanic activity. Therefore, the material transported over the surface of the land contains products of volcanic origin as well as disintegration products from the underlying gneiss-granite. On the other hand, the pyroclastic sediments may contain many fragments and crystals derived from the gneiss-granite, broken up within volcanic vents (so called accidental fragments). These were ejected in the explosive stage of eruption together with volcanic ash, lapilli and bombs. The ratio of products of volcanic origin to those derived from the gneiss-granite gives an indication as to the genesis of the sediment. The structures in both types of sediments also exhibit differences, but here again transitions are present.

Epiclastic rocks

The most important epiclastic rock of the Kappebo formationn is a coarse graywacke of 4 mm to 0.5 mm grain size, the coarser parts of which (4 mm — 2 mm diam.) could be termed granule breccia. However, neither the distribution, the structure nor the composition of the rocks give any reason for considering the coarser parts separately. Far less common are those epiclasts whose grain size is less than 0.5 mm but which, down to a grain size, of 0.06 mm lie within the sand grade and hence should also be called graywackes. Siltstones and slates with a grain size less than 0.06 mm have been found only in limited quantities. In places, larger pebbles and rock fragments occur in the coarse graywacke, the rock being then a conglomerate or a breccia. The proportion of pebbles and fragments varies widely; sometimes they appear only scattered through the graywacke, sometimes they constitute the main part of the rock.

All the epiclastic rocks of the Kappebo formation are ill sorted with respect to grain size and contents; the clastic fragments are poorly or not rounded (fig. 8, p. 102).

Coarse graywacke

This rock, quantitatively the most important of the epiclastic sediments, is almost uninterruptedly exposed over large distances on both sides of the eastern arm of Kappebosjön and to the south of Svartetjärn. It occurs also as a great number of lenticular bodies among the other rocks of the Kappebo formation.

It is composed essentially of grains of milky blue quartz and red feldspar in a light to dark grey matrix. Scales of slate, fragments of felsite and granite particles occur in small quantities. They are very poorly sorted; in thin sections it appears that the diameter of the largest grains is about twenty times that of the smallest. The proportion of quartz grains to feldspar grains varies greatly. In the coarser types a proportion 1:1 is not rare. More often than not and especially in the finer graywackes quartz is considerably more abundant than feldspar. There are rarely any signs of stratification by differences of colour in the matrix, by the position of the commonly elongated fragments of slate and felsite or by differences in grain size. Most of the graywackes show a certain degree of schistosity.

The usually angular or sliver-like *quartz* fragments all show undulatory extinction, at times very strongly so, even in the less schistose types of graywacke. Only in the markedly schistose graywacke are they cataclastic. The strongly undulatory extinction has apparently been formed in the quartz before the deposition of the sediment.

All types of *feldspars* which figure in the gneiss-granite are met with again as angular fragments in the graywacke, such as Na-potassium feldspar with perthitic structure or with a microcline pattern and "chessboard" albite. They have often inclusions of quartz and idiomorphic plagioclase, as in the eyes of the augen-gneiss. Plagioclases altered into sericite and such forms as result from the replacement of the Kroppefjäll granite by the aplite-granite viz.: myrmekite and feldspar with quartz drops, are also found. Moreover, grains consisting of a quartz-feldspar assemblage have been found, which must be considered as fragments of gneiss-granite.

In addition the following constituents have been found in subordinate quantities:

1. Fragments of *slate*, which upon microscopic examination appear to consist of a cryptocrystalline aggregate of sericitic matter, sometimes mixed with a little ore (magnetite?).
2. Angular fragments of *felsite*, which microscopically show to be composed of an aggregate of cryptocrystalline light constituents and sericite.
3. Clastic crystals of *muscovite*, probably derived from the recrystallized granite-schist.
4. Lenses with abundant *epidote*,
5. *Albite* crystals,

The fragments of felsite, the lenses of epidote and the crystals of albite must originate from the volcanic rocks of the Kappebo formation.

All these angular fragments of crystals and rocks are embedded in a matrix which consists mainly of microcrystalline quartz, feldspar and sericite. The matrix may occupy more than one half of the rock or may be almost lacking. The proportion of sericite to quartz and feldspar may vary greatly. Other minerals which have been found in the matrix are: chlorite, epidote and ore (sometimes looking red = hematite?). These minerals occur mainly as minor constituents, but occasionally a combination of them, or a single one, may become important. Rare idiomorphic crystals of pyrite, up to 1 mm in size, have been encountered in the matrix.

Everything leads to the assumption that the essential part of the constituents of the graywacke are direct clastic derivatives of the gneiss-granite. The colour of the quartz and the colour and structure of the feldspar support this view. Moreover, such an origin is undoubtedly borne out by the contents of the conglomerate and breccia associated with the graywacke, as will be explained subsequently. The clastic crystals of muscovite are probably derived from the recrystallized granite-schist.

The remaining smaller part of the constituents descends apparently from the Kappebo formation itself. The helleflints are entirely similar to the acid effusives and tuffs of this formation, which must also be the source of the albite crystals. The lenses with abundant epidote may be looked upon as fragments of greenstone or greenstone tuff. The dark minerals of the matrix too, are likely to be derived from the last mentioned rocks. Finally, deposits of slates are known also in the Kappebo formation which, as will be described later, must have furnished the slate fragments.

Fine graywacke, siltstone and slate

Less common than coarse graywacke are fine graywacke and siltstone, the grain size being 0.5 to 0.06 mm diam. and less than 0.06 mm diam. respectively. These rocks of light to dark grey tone are usually found to form lenses in the coarse graywacke, but occur also among the other rocks of the Kappebo formation. The small grain size of the siltstone gives it the appearance of a schistose felsite. The only essential difference of the siltstone and fine graywacke from the coarse graywacke lies in the grain size, their structure and composition being much alike. Obviously rock fragments are absent in these fine-grained rocks. The quartz/feldspar ratio seems to have shifted towards a higher quartz content. The amount of matrix material is generally larger than in the coarser graywacke. The green shade of some of the rocks is due to a matrix rich in chlorite and epidote. The clastic fragments of crystals are not rounded in these rocks either. Scarce idiomorphic crystals of pyrite occur, up to 2 mm in size.

In some places slates have been found in the Kappebo formation. The main occurrences are near Blixerud and on the Kappebofjäll 450 m northeast of the northeastern point of Kappebosjön. In the first locality the slates have a perfect transverse cleavage and were quarried in a small way for roofing purposes. A distinct bedding is observable approximately at right angles to the schistosity. Microscopically it is seen that the slate bears a close relationship to the siltstone and may be considered as a variety of it. The sliver-like quartz (and feldspar?) fragments are less frequent and the increased matrix is richer in sericite-like minerals.

Conglomerate and breccia

If on the one hand the coarse graywackes are related to more fine-grained deposits, on the other there exist all transitions to very coarse sediments, for occasionally rock fragments and pebbles with a diameter considerably above 4 mm are found in the coarse graywacke. Whenever there is a certain concentration of these large rock fragments, one is entitled to speak of a conglomerate or a breccia which is lenticular and has no definite boundaries in the graywacke. The matrix of the graywacke, however, still forms an important part in these conglomerates and breccias, so the fragments and pebbles generally do not touch one another. On the western side of the Kappebofjäll, from Bengterud to Höljen, many of these lenticular conglomerates and breccias occur but, contrary to the statement of SANDELL (1941, p. 184 and pl. II p. 160), they do not form a continuous layer. He mapped it as such and even imputed it to an unconformity. However, near Blixerud and to the east of Koljerudstjärn the conglomerate and the breccia are exposed without a break, the amount of matrix being considerably reduced.

The conglomerates and breccias of the Kappebo formation are polymict; the constituents vary enormously in size, ranging from 4 to 250 mm diam., and are generally angular or poorly rounded. In the conglomerate and the breccia near Blixerud and Koljerudstjärn, however, parts of well rounded pebbles occur. The constituents of the conglomerate and the breccia can be divided into two groups, still more clearly than in the case of the graywacke. One group is made up of the disintegration products of the gneiss-granite, the other apparently consist of earlier deposits of the Kappebo formation itself. The detritus of the gneiss-granite is quantitatively by far the most important constituent. As such have been found:

1. grains of milky blue *quartz* (d. up to 8 mm).
2. red *feldspar* (d. up to 20 mm).

These are the slightly reduced eyes of the augen-gneiss, which in the breccia give rise anew to an "eye" structure when they occur frequently, and at first sight give it the appearance of an augen-gneiss. They differ from the latter by the occurrence of fragments of other rocks and by the clastic nature of the matrix.

3. *Aplite-granite* forms large angular fragments and pebbles which may attain 250 mm in diameter and which comprise the most important constituent of the conglomerate or breccia.
4. Less common are pebbles of *augen-gneiss* with diameters up to 25 cm. Their low frequency must be ascribed to the fact that this rock, as a result of weathering and transport, is readily disintegrated into its components. However, in the conglomerate to the east of Koljerudstjärn they are of common occurrence and in places even constitute the major element. As has been pointed out (p. 81) this conglomerate is of special importance, as the pebbles of the augen-gneiss are scattered indiscriminately with respect to the direction of their schistosity. This shows that the augen-gneiss must have got its schistosity before the deposition of the Kappebo formation.
5. A few fragments of the *acid parts of the Kroppefjäll gneiss-granite* have been encountered.

The components of the conglomerate and the breccia, derived from previous deposits of the Kappebo formation itself, form a small minority and are represented by the following rocks:

1. *Slate*, as in the graywacke, occurs commonly as long-drawn fragments

which may attain lengths up to 8 cm. In places, these fragments of slate may become an important constituent of the conglomerate and the breccia.
2. Mostly angular and elongated fragments of *felsite* occur commonly, but not in great quantities.
3. Rare fragments of *greenstone* and *greenstone tuff* have been found.
4. A few pebbles and fragments of *siltstone* also occur.

The matrix of the conglomerate is always formed of graywacke which is similar in every respect to the graywacke described above.

Pyroclastic rocks

The pyroclasts are the most common rocks of the Kappebo formation. Especially in the central part of the Kappebofjäll and more to the south between Höljesjön and Sketjärn they have an overwhelming predominance, but they occur also frequently elsewhere. Obvious volcanic ejecta are found both from the eruptions of albite-rhyolites and of greenstones. The pyroclastic rocks produced by them could be divided according to that dual origin if it were not that these different elements occur intermingled. We must admit therefore that much of the ejected material is transported by running water into a secondary position and became mixed during the process with ejecta of other eruptions. Also epiclastic elements can be added in this way to the pyroclastic sediment.

Tuff

The tuffs are fine-grained to felsitic rocks of widely variable colour and character. Light, almost white, red, chocolate brown, light brownish or dark grey, almost black or light to dark green shades occur. Sometimes they are laminated or flamed. Schistosity is medium to strong. Quartz or feldspar grains may be embedded in the aphanitic mass, which give the rock a porphyritic structure. Again, small lenses of differently coloured felsite may be visible in the rock. On account of these characteristics it is often impossible upon macroscopic examination to distinguish the tuffs from the effusives and the pelitic epiclasts. The larger crystals of quartz and of red feldspar could be mistaken for phenocrysts. In most cases, however, microscopical differences of structure can be observed between tuffs and other rocks which are macroscopically similar.

Upon microscopic examination almost all of the tuffs show a porphyritic structure, the groundmass being generally cryptocrystalline with embedded idiomorphic crystals of albite, which are sometimes broken up (protoclastically?). Also hypidiomorphic epidote and even occasional augite crystals, partly altered into actinolite, are met with in some greenstone tuffs. These crystals are likely to be of intratelluric origin and were ejected together with the ash during the explosive stage of the eruption. They are, therefore, entirely comparable with the phenocrysts of the effusives.

Most of the enclosed crystals consist of sliver-like or angular fragments of quartz and Na-potassium feldspar. They are derived from the underlying gneiss-granite and must be considered, at least partly, as "accidental" material. Among other features the shape and the nature of these inclusions clearly separate the tuffs from the effusives. From the epiclasts the tuffs are mainly distinguished by the great quantity of cryptocrystalline matrix.

However, as the quantity of sliver-like crystal fragments increases the structure of the graywackes is approached. In this particular case much epiclastic detritus must have been washed into the tuff during the secondary transport, whereby the distinction between tuff and graywacke becomes less defined.

The most important minerals of the cryptocrystalline matrix are quartz, feldspar and sericite. To these constituents other minerals are often added, such as epidote, magnetite, chlorite, calcite and a red pigment (hematite?). If the latter constituents are only accessory, the tuff may be considered as an albite-rhyolite tuff. If, however, they form an important part, then the rock is a greenstone tuff or a mixed greenstone albite-rhyolite tuff.

The structure of the matrix of the tuffs is only rarely massive. Most often the non quartz-feldspar minerals are arranged in streaks or bands producing a linear structure of the rock. On the one hand this may be due to schistosity induced by dynamometamorphism, but on the other hand it is often recognized as a sedimentary stratification. The bands and streaks, like flow structures, enclose larger crystals and crystal fragments. In a similar way they close round lenses consisting of micro- to cryptocristalline aggregates, which may not differ at all from the rest of the matrix, but may also show slight differences in mineral content, grain size or structure. These are the coarser parts of the ash or fine lapilli which settled together with the fine ash particles. Macroscopically these coarser parts of ash may show up as lenses of a different colour in the rock. Pronounced linear structure is not always present in the tuffs and in this case the coarser ash particles are drop-like or even angular, giving rise to a brecciated structure. In a few mainly acid tuffs such coarse ash particles with much epidote and chlorite have been found; these must therefore have originated from an eruption of greenstone. Conversely, acid ash particles have been found in mainly basic tuffs. This phenomenon demonstrates more than anything else the secondary mixing of two different kinds of tuffs. As has been stated, the coarser ash particles are in general entirely comparable to the different types of matrices and therefore need no further description. An exception to this rule are lenses and sometimes streaks of quartz and feldspar of considerable larger grain size, in which much calcite occurs. The origin of these lenses is not known to the writer. It may be mentioned further that some enclosed ash particles show a perlitic structure and that others have been found which contain an idiomorphic crystal of albite.

Infrequent tuffs occur which consist almost exclusively of sericite, while inclusions are lacking entirely or appear only as extremely small scales of crystals. These rocks could be termed slates but for their light yellow to greenish colour. Sometimes lamination is perfect as a black banding which is caused by finely divided ore.

As to the question whether these tuffs were originally deposited mainly as vitric, crystal or lithic tuffs, we may state that the cryptocrystalline matrix predominates to such an extent that it is impossible to call them lithic tuffs. It is likely that the matrix must for the greater part be considered to be devitrified on account of its uniform and extremely fine-grained character and therefore they are probably vitric tuffs. This is suggested also by the perlitic structure of some of the coarser ash particles. Of very rare occurrence are such forms as remind us of the typical curved glass fragments which are so characteristic of many vitric tuffs. They appear as parts of coarser crystallization, in the matrix. In view of the complete crystallinity it is not sure that they constitute relics of original glass particles.

As in the albite-rhyolites and the greenstones, occasional small cross-biotites also occur in the tuffs.

Lapilli-tuff and agglomerate

As distinct from the tuffs, the lapilli-tuffs and agglomerates are usually easily identified in the field. They are composed of fragments of aphanitic felsite and greenstone, which are distinguished in the rock by differences of colour. Just as in some tuffs one may notice coarser ash particles on macroscopic examination. These rocks have been termed regenerated helleflints by TÖRNEBOHM (1870a, p. 17) and SANDELL (1941, p. 184). In addition to the fragments of felsite large crystals of quartz occur, as well as red feldspar and incidental fragments of aplite-granite and even augen-gneiss. The size of the fragments varies greatly. By definition the average diameter of the particles in lapilli-tuff is greater than 4 mm. From this size upwards all dimensions have been found including coarse agglomerates with many fragments of 12 cm diameter.

As well in the lapilli-tuffs, as in the tuffs, the amount of fragmental quartz and feldspar crystals together with the accidental rock material may increase strongly, marking transitions to fine-grained conglomerates and breccias. Transitional types between agglomerates and the coarser conglomerates and breccias have been hardly, if ever, observed. The big lumps of the agglomerates were apparently less amenable to secondary transportation than the finer ejected material and consequently are seldom observed in mixed deposits.

It is not necessary to dwell on detailed petrographic descriptions of the lapilli-tuffs and agglomerates, since they are, except for the size of the fragments, identical with the tuffs. The only point that need be mentioned is that the cryptocrystalline matrix is considerably more reduced than in the tuffs and sometimes only fills up the interstices between the lapilli and the blocks. In these coarse pyroclasts too, fragments of felsite with perlitic structure or with albite phenocrysts have been found. It is somewhat surprising that albite-rhyolites of the types II, III and IV have never been found as lapilli or blocks in the coarser pyroclasts.

CHAPTER III

THE KAPPEBO FORMATION (continued)

Petrology of the effusives and dykes

In order to facilitate the comparison of the results, the most important data of the six analyses of effusive and dyke rocks of the Kappebo formation are listed below. As we have said (p. 101), conclusive field geological and mineralogical evidence shows that the albite-rhyolites and the greenstones are related genetically.

Alb.-rh.	Anl.	Q	L	M	π	k	γ	mg	si	al	fm	c	alk
Type I	1	59.2	35.7	5.1	0.04	0.21	0.18	0.05	455	39	18	8	35
Type II	2	58.8	37.3	3.9	0.09	0.24	0.10	0.15	446	42	14	8.5	35.5
Type IV	3	58.3	37.5	4.2	0.09	0.42	0.43	0.10	435.5	42	14	9	35
Type IV	4	48.3	42.5	8.7	0.12	0.37	0.00	0.25	297	37	25.5	9	28.5
Greenst.													
Type II	5	30.3	48.7	20.0	0.14	0.16	0.00	0.44	161	30	40.5	7	22.5
Type II	6	72.6	42.7	22.7	0.27	0.25	0.16	0.57	130	24	43	18	15

	Magma type
Anl. 1	natronengadinitic
Anl. 2	natronengadinitic
Anl. 3	engadinite-granitic
Anl. 4	tasnagranitic
Anl. 5	melanatronsyenitic
Anl. 6	mugearitic

The potassium content

The first three analyses are almost identical, except for the potassium/sodium ratio (*k* figure) which in analysis no. 3 shows a marked increase of potassium on account of which it has been classed in a different magma type. Analysis no. 4 is provided by a less acid sample of albite-rhyolite which structurally, however, is identical with the sample represented by analysis no. 3. The *k* ratio of analysis no. 4 is also higher than that of the first two, whereas the figures for the greenstones (no. 5 and 6) in this respect correspond to the first two analyses. No doubt the rocks of these six samples have originated from the same magma. Apparently the *k* ratio has not a determinative quality in these rocks.

The number of analyses is too small to justify the assertion, however attractive it be, that a high *k* ratio is only found in type IV. It would lead us into still further speculation to assume that there exists a causal relation between the high *k* ratio and the slower cooling of the albite-rhyolites of type IV.

The strange feature about the high *k* ratio in the analyses 3 and 4 is that it is not appreciably reflected in the mineral contents of the pertaining albite-rhyolites of type IV. The phenocrysts as well as the idiomorphic feldspars of the second generation are albite. The measurement of the granophyric feldspars of the groundmass on the Federoff universal stage is greatly hampered by their undulatory extinction and their intergrowth with quartz. The few apparently reliable measurements obtained, however, also seem to indicate albite. On account of this discrepancy the alkali determinations of analysis no. 4 have been repeated. The results agree with the first determination, within the normal limits of error. Even if we suppose the maximum amount of potassium to be contained in the sericite, as computed in the standard epinorm (Ms), there still remains a quantity of potassium not to be neglected for the formation of potassium feldspar (Or). For convenient inspection the values of Or and Ab of the standard epinorm of the analyses no. 1—4 incl. are listed:

Analysis 1:	Or = 9.6 %	Ab = 44.7 %
Analysis 2:	Or = 8.5 %	Ab = 43.1 %
Analysis 3:	Or = 18 %	Ab = 33 %
Analysis 4:	Or = 14.5 %	Ab = 39.4 %

As generally admitted, up to 10 % of orthoclase may be included in the albite lattice. The above figures point to a ratio Or/Ab two to five times as high. Thus, the bond of K-ions in the crystal lattice of albite equally fails to account in a satisfactory manner for the apparent lack of potassium-bearing feldspar.

Of the composition of sericite little is known, but the few existing analyses seem to indicate a decrease of the K content rather than a increase as compared to muscovite (Ms), but at the same time it is shown that in the constitution of the lattice Mg + Fe together stand for ± 20 % (HUBER 1943, p. 175). This fact does not find expression in the calculation of Ms, since the amount of Ms is determined solely by the excess of aluminium. We must reckon, therefore, with the possibility that an appreciably larger amount of sericite may have been formed than the epinorm would suggest, so that, in absolute quantity, more potassium is contained in the sericite. Taking these considerations together we might be able to understand, as far as the rocks of the analyses no. 1 and 2 are concerned, that no K-bearing feldspar is present, but the rocks of the analyses 3 and 4 must contain K-bearing feldspars in the groundmass, respectively potassium in the granophyric feldspars.

Structural forms

As far as we can see in these very fine-grained rocks, the mineral content of the samples of the analyses no. 1, 2 and 3 is the same. Only the structural development differs. These differences must, therefore, be entirely attributed to the physical conditions prevailing when the magma solidified.

Analysis no. 1 was taken of a devitrified glass with perlitic structure and extremely few phenocrysts of albite (type I). Consequently in the rocks of this type, a small degree of intratelluric crystallization can be traced (albite phenocrysts), upon which extrusion and rapid cooling followed. There has been little or no crystallization from the fluid lava.

Analysis no. 2 was made of an albite-rhyolite of type II. In this type the number of albite and quartz phenocrysts often becomes considerable, lying in a micro- to cryptocrystalline groundmass which at times exhibits fluxion structures. Thus, there has been a rather important intratelluric crystallization interrupted by extrusion and rapid cooling. It may be of interest to observe that in those places, where an original glass base has been demonstrated by perlitic structure, the grain size of the quartz and feldspar, as cristallized in the solid state, varies strongly from crypto- to microcrystalline. We need not, therefore, consider only the cryptocrystalline parts of type I and II as devitrified glass, but we may safely assume that a great portion of the groundmass of these types has also been solidified in the form of glass.

Analysis no. 3 has been taken from a sample which, structurally, belongs to type IV. This type is characterized by the presence of albite phenocrysts, while phenocrysts of quartz are lacking. The groundmass often shows a distinct igneous structure. Here the cooling, after the intratelluric crystallization, has been slower, allowing a proper crystallization from the fluid phase. The only satisfactory explanation for the absence of quartz phenocrysts seems to be that these crystals of intratelluric origin have been destroyed by resorption. Two observations support this assumption:

1. In type III, which also for other reasons, holds an intermediate position between type II and type IV, more or less rounded granular aggregates of quartz are found. They are possibly partially resorbed and disintegrated phenocrysts.
2. In the acid members of type IV (analysis 3) xenomorphic quartz grains have been found, which are considerably larger than the grain size of the groundmass. These might be relic phenocrysts.

The albite-rhyolites of type IV are most abundant in those spots where no sediments or pyroclastic rocks of the Kappebo formation occur i.e. where the albite-rhyolites are found as dykes in the Kroppefjäll gneiss-granite. This fact is consistent with the conclusion on the structural development, arrived at from the above considerations, viz. that the albite-rhyolites of this type have cooled comparatively slowly.

In the same way the differences of grain size and crystallization, as displayed by the three types of greenstones, are likely to mark the stages of a continually slower cooling.

The mineral paragenesis

The mineral paragenesis of the albite-rhyolites and the greenstones indicates the formation or transformation of these rocks at a low temperature. They are rocks of the epizone (Grubenmann—Niggli 1924) and in the facies classification of Eskola (1939) occupy a border-line position between the epidote-amphibolite facies and the greenstone facies.

The question arises whether this association of minerals is a product of the solidification process or of a subsequent metamorphism. To decide this, within the limits of the possible, we must first of all consider to what kind of metamorphism the Kappebo formation has been subjected. Nothing is known about a later magmatic activity in the Kappebo formation, except for the rare thin veins of pegmatite as occur also in the gneiss-granite (p. 82). The mineral association, therefore, cannot be the result of contact metamorphism. Metasomatism from the outside is not to be considered either, since nothing of the kind has been found in the adjacent rocks. However, the Kappebo formation has been subject to dynamometamorphism, causing some degree of schistosity in all the rocks involved. It is well-known that such a metamorphism may give rise to sericitization, saussuritization and chloritization, especially when it occurs at some depth below the surface. One would be inclined, therefore, to attribute the mineral association of the Kappebo lava flows to dynamometamorphic alterations, were it not for the fact that some observations seem to contradict this assumption.

The pseudomorphism, in the greenstone of type III (p. 109), of epidote and some chlorite after pyroxene does not indicate a dynamometamorphic action, especially as the rock is hardly foliated. It is also difficult to ascribe the formation of hypidiomorphic epidote in some greenstone tuffs (p. 115) to dynamometamorphic processes. They lie together with idiomorphic albite, sliver-like quartz and Na-potassium feldspars in the fine matrix of the tuffs and it is logical to assume that they were all ejected at the same time by the volcano.

In view of these observations one may ask whether the mineral association of epizonal character did not, rather, come into existence at the time of the solidification of the rock. In that case, the process of solidification must chiefly have developed at a low temperature, or over a wide temperature interval with low final temperature. This implies a magma rich in volatiles. In other words, we may ask whether the Kappebo outflows have not to be considered as a spilitic series (spilites and keratophyres). In order to get any certainty about this, the chemical composition of the albite-rhyolites and greenstones under discussion will be compared with that of a great number of spilitic rocks.

Before we proceed to do so, it may be useful to note that, since Dewey and Flett (1911) first called the attention to the problems connected with a spilitic series, a voluminous literature on the subject ensued. It is not my intention to dwell upon these problems, nor on the numerous contradictory opinions about them. A critical review of the literature, until 1935, was published by J. Gilluly (1935). Since that time, the flow of new publications on this subject has not diminished. I wish to mention only the most recent papers by Vuagnat (1946) and van Overeem (1948), in which are given further references. Notwithstanding many contradictory opinions, we can say that a definition of what is to be considered as a spilite, becomes more and more clear and can be summarized as follows:

A basic effusive (or dyke) rock, with dark constituents characteristic for the epizone and with albite and feldspar, while this mineral paragenesis is not the result of allometamorphism. The spilites are often associated with acid lava flows, keratophyres, the feldspar of which is mainly plagioclase with a high Na content. This association of rocks is termed a spilitic series.

Van Overeem (1948, p. 90—97), in his detailed discussion of the spilites of the Dal formation, has compiled most of the modern analyses (total 137) of spilitic rocks and represented them in diagrams. This provides the means to form an objective opinion about the chemical characteristics of these rocks. It is striking to note the perfect congruence between the chemistry of the Kappebo outflows and these characteristics of the spilitic rocks. For a closer comparison, the analyses of the Kappebo formation have been plotted in the same diagrams as used by VAN OVEREEM (1948).

First we have a variation diagram (fig. 11) which shows the curves of the Niggli values with a very regular trend. This is, in part, due to the small number of analyses. If these curves are compared with the average variation curves of the Atlantic, Pacific and Mediterranean provinces, as given by C. BURRI and P. NIGGLI (1945, p. 61), it appears that the *al* curve, for high values of *si*, follows the Pacific *al* curve pretty closely, but exceeds it for smaller *si*. It is found that *fm* is always higher than the average *fm* curves of the Atlantic, Pacific or Mediterranean provinces, while the *c* curve for high values of si, agrees best with that of the Atlantic province, but is appreciably less for low *si*. The *alk*, for higher values of *si*, matches best the corresponding curve of the Pacific province, but, for low values of *si*, remains slightly above it. On account of this comparison the effusives of the Kappebo formation cannot be classed in one of the above provinces.

It might be argued that the curves, as given by C. BURRI and P. NIGGLI (1945) are average curves and that, when a small number of analyses is available, important divergences are likely to occur. However, the divergences observed here are of special significance, as they are in accordance with those stated by VAN OVEREEM (1948, p. 95—96) on a great number of analyses of spilitic rocks. His findings may be summarized as follows:

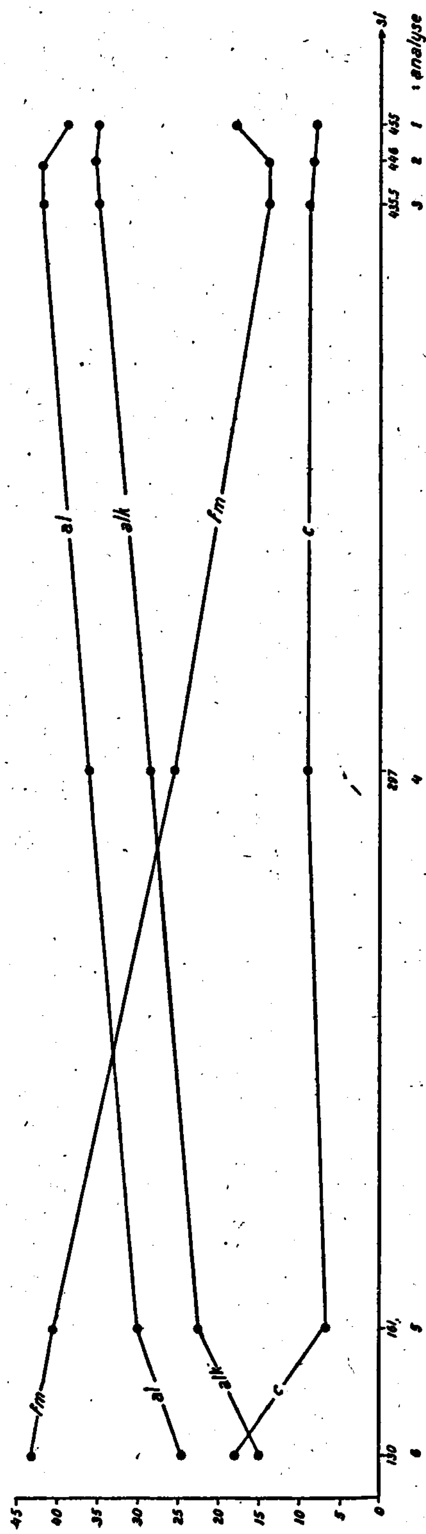

Fig. 11. Variation diagram of the effusives and dykes of the Kappebo formation

The values of *al* and *alk* correspond roughly with the curves of the Pacific province, the *fm* values being greater and the *c*'s smaller than the curves of any one of the three provinces. The *c* values give still the best match with the *c* curve of the Atlantic province.

There could hardly be a more perfect correspondence between this general tendency of spilitic rocks and that observed in the effusives of the Kappebo formation.

For a comparison of the ratios *mg* and *k* it is better to refer to other diagrams, viz. the $\pi - k$ and the $k - mg$ diagram and in addition to the $(FeO + Fe_2O_3) - MgO - CaO$ diagram and the $fm - mg$ diagram. For this purpose these diagrams (fig. 12, 13, 14, and 15) have been plotted for the effusives of the Kappebo formation. Also marked are the fields within which fall most of the spilite analyses compiled by VAN OVEREEM. In all diagrams the Kappebo effusives lie within these fields.

Hence it follows that the chemical composition of the albite-rhyolites and greenstones discussed here, corresponds to that of spilitic rocks. This, added to the considerations mentioned above, makes it probable that the Kappebo outflows represent a spilitic series.

It is desirable to consider in detail the chemical characteristics of the spilitic rocks, as compared to the Atlantic, Mediterranean and Pacific provinces, because this is done very briefly by VAN OVEREEM (1948, p. 96). For this purpose the delimitations of the main areas of distribution of the above provinces have been traced in the diagrams used.

In the $k - \pi$ triangle (fig. 12) the spilitic rocks occupy a large area about the Na(Ne) corner. The greatest concentration is found immediately in the Na(Ne) corner and from there along the π axis until $\pi = 0.5$. This has been a reason for many investigators to consider the high Na(Ne) content as determinative of spilitic rocks and to ascribe the formation of albite to this cause. Although the high Na(Ne) content is a frequent phenomenon and albite, no matter from what sort of origin, is of common occurrence, the $k - \pi$ diagram of the spilitic rocks (VAN OVEREEM 1948, fig. 19) shows that one half of the spilites deviate from this narrow zone, poor in potassium, and occur in the remainder of the area indicated. The principal areas of distribution of the Atlantic and Pacific rock provinces lie almost entirely within the field of the spilitic rocks. Neither of these areas, however, encloses the great concentration along the π axis near the Na(Ne) corner. The principal area of distribution of the Mediterranean province falls for the most part outside the field of the spilitic rocks.

A similar wide scattering of points is found in the $k - mg$ diagram (fig. 13). However, a concentration of points is found in the field between $mg = 0.3$ to 0.7 and $k = 0.0$ to 0.15. Here again, the field of the spilitic rocks very nearly encloses the fields of the Pacific and Atlantic provinces, but in addition extends to the ordinate axis, also for low values of *mg*. The field of the Mediterranean province falls almost completely outside the area of spilitic rocks. In this diagram the acid spilitic rocks (keratophyres) are found mainly at low *mg*, the basic ones (spilites) at high *mg*. Consequently it is possible to distinguish a spilite area and a keratophyre area, connected by a transition zone. Such transition zones have been marked, in this and the following diagrams, by a dotted line.

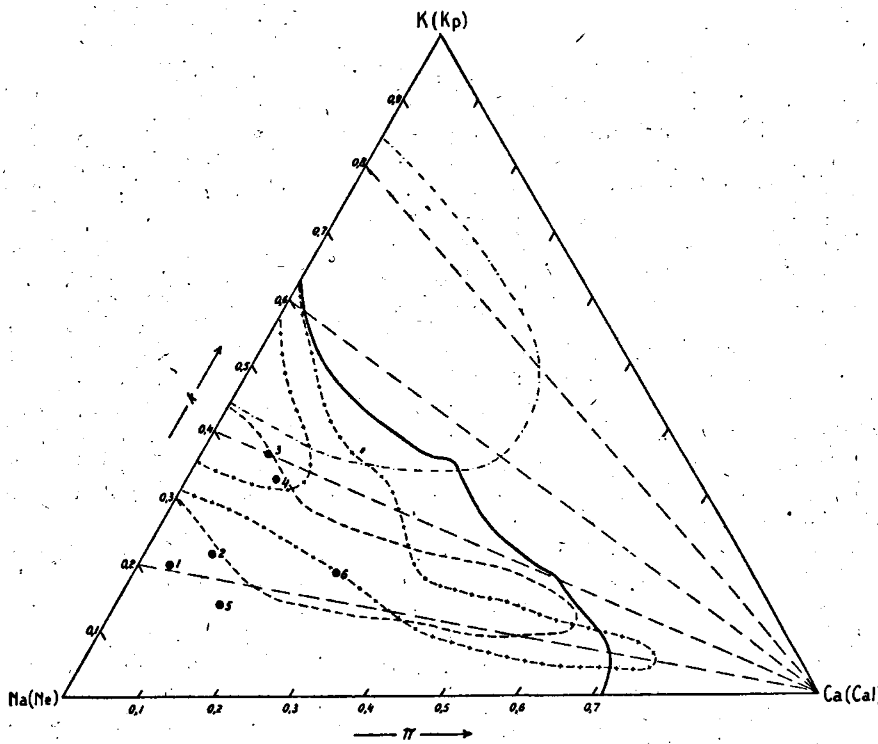

Fig. 12.
$k - \pi$ diagram of the effusives and dykes of the Kappebo formation
For explanation of the symbols see fig. 16.

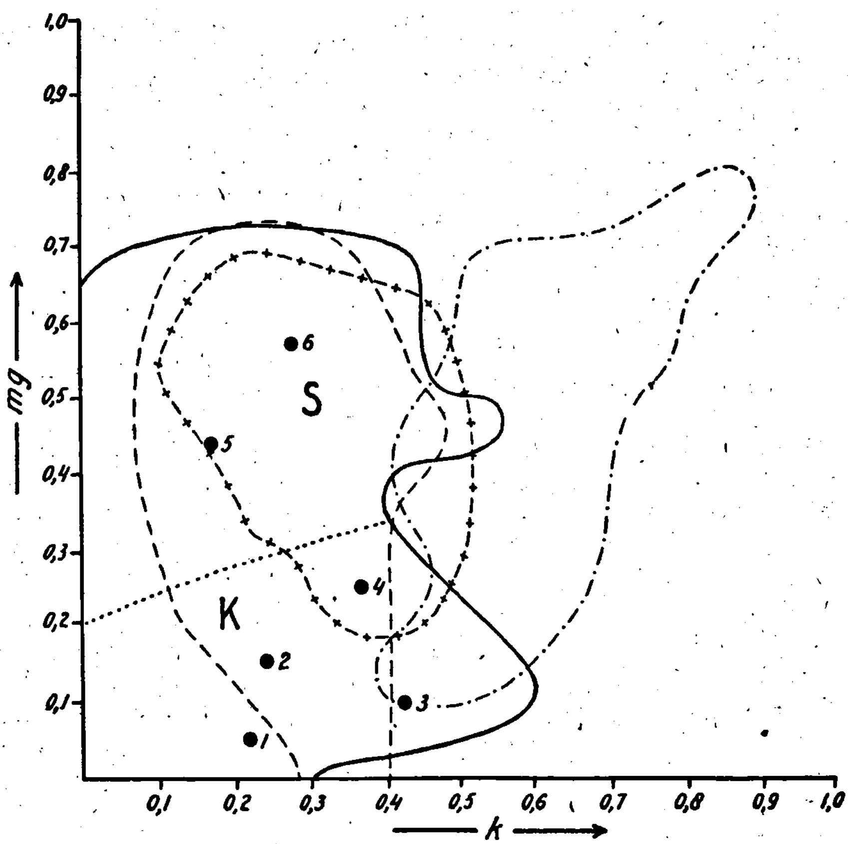

Fig. 13.
$k - mg$ diagram of the effusives and dykes of the Kappebo formation.
For explanation of the symbols see fig. 16.

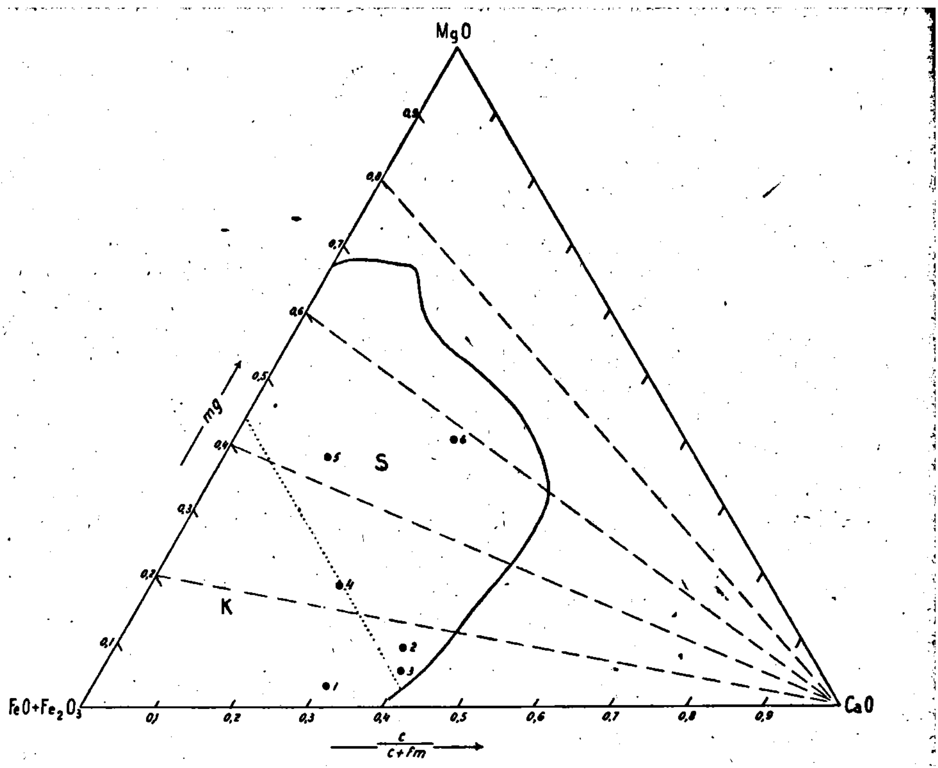

Fig. 14.
(FeO + Fe_2O_3) — MgO — CaO diagram of the effusives and dykes of the Kappebo formation.
For explanation of the symbols see fig. 16.

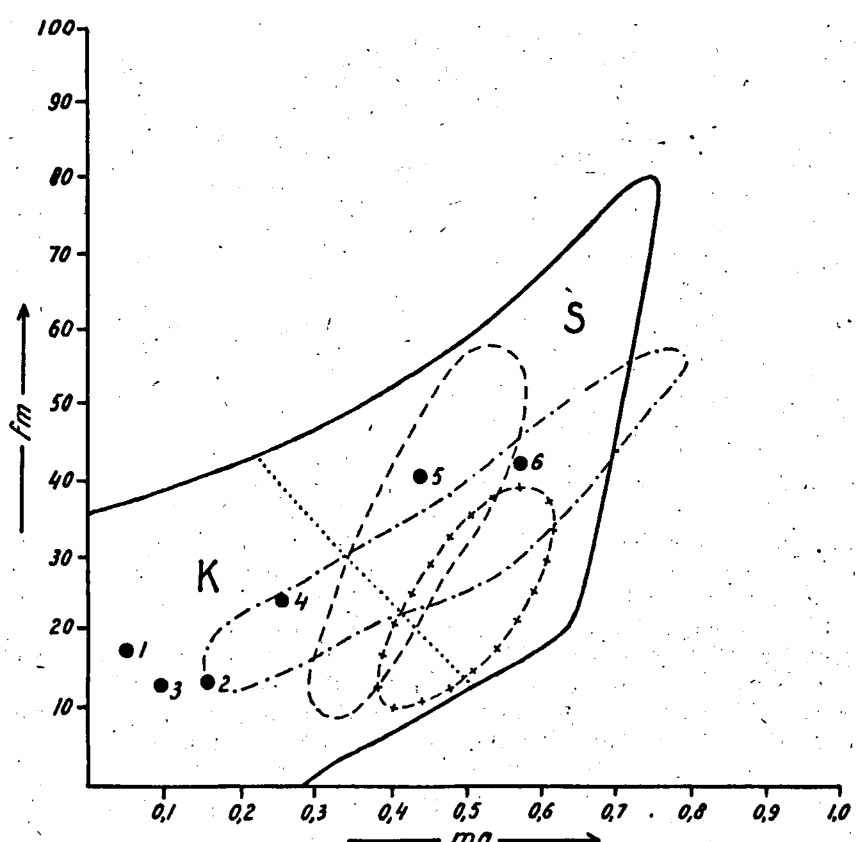

Fig. 15.
fm — *mg* diagram of the effusives and dykes of the Kappebo formation.
For explanation of the symbols see fig. 16.

As in a $\gamma - mg$ diagram many points would fall outside the triangle (because of an excess of al in many spilitic rocks no Cs is formed), I have followed VAN OVEREEM (1948) in plotting a ($FeO + Fe_2O_3$) — MgO — CaO diagram (fig. 14). The field of spilitic rocks lies about the ($FeO + Fe_2O_3$) corner. It has its boundary at 40 % to 50 % CaO and passes only slightly beyond the $mg = 0.7$ line. Here too, a clear distinction between a spilite area and a keratophyre area can be made. The keratophyres have a higher relative ($FeO + Fe_2O_3$) content, whereas the spilites contain relatively more MgO or CaO.

The $fm - mg$ diagram (fig. 15) shows an exceedingly extensive area symmetrical with respect to a slightly curved line which roughly coincides with

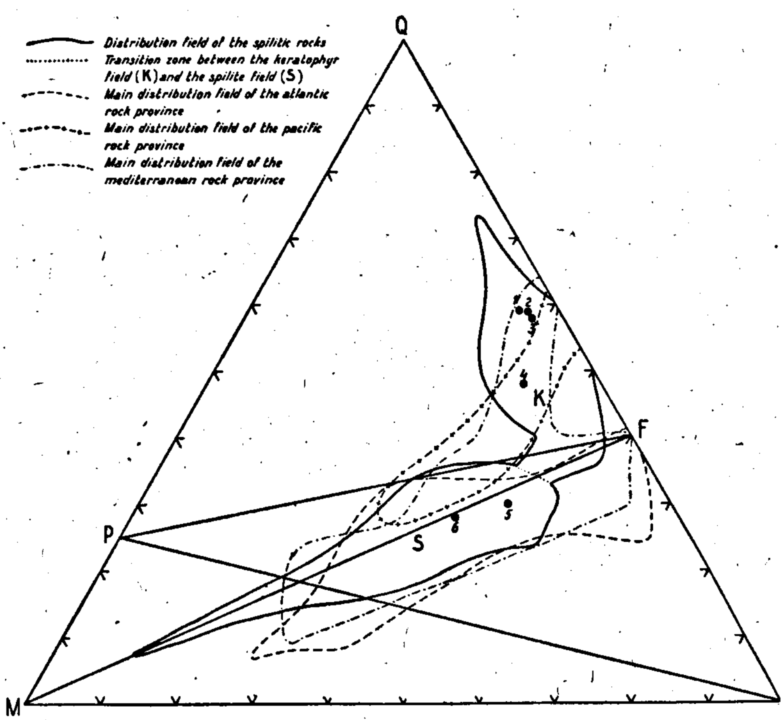

Fig. 16.

Q — L — M diagram of the effusives and dykes of the Kappebo formation.

the diagonal of the diagram. An important concentration of points is found around the diagonal at $mg = 0.4$ to 0.6 ($fm = 40$ to 60). A division between a keratophyre area and a spilite area is also distinct here and can be indicated by a line perpendicular to the diagonal. This extensive field of the spilitic rocks encloses the principal fields of the Pacific, Atlantic and Mediterranean provinces alike. The principal field of the Pacific province, however, lies below the diagonal ($mg = 0.2$ to 0.3 larger than fm), whereas the principal field of the Atlantic province lies about a line running from $mg = 0.2$, $fm =$ 10 to $mg = 0.5$, $fm = 60$. This line is steeper than the diagonal and intersects it. The principal field of the Mediterranean province, on the contrary, is to be found along a less inclined line which begins on the diagonal at $mg = 0.15$, $fm = 15$ and trends towards $mg = 0.8$, $fm = 60$.

In contradistinction to the preceding diagrams, the field of the spilites in the Q — L — M diagram (fig. 16) is very constricted, and follows, as far as the basic rocks are concerned, the M — F line. The field of distribution of the keratophyres joins it near F and thence, in a sharp turn, finds its continuation just within the line L — Q, in the direction of Q. Thus, it appears to be a typical field which is not altogether comparable with the fields of the three major rock provinces. If anything, the main field of the Mediterranean province shows the closest resemblance to the spilitic field.

VAN OVEREEM (1948, p. 94) concludes, referring to the wide spreading of the spilitic rocks in the various diagrams "that in principle "every" gabbroid magma (Pacific, Atlantic or Mediterranean) is capable to supply a spilitic rock, provided that a high content of volatiles is present". In this respect, however, I should like to restrict this statement where the Mediterrenean province is concerned which, through its high potassium content, occupies fields essentially different from the spilitic rocks, in the $k - \pi$ diagram and the $k - mg$ diagram.

Thus, if the spilitic rocks have originated from normal Pacific or Atlantic (may be Mediterranean) magmas, then the processes which produced them have influenced the chemical composition of the final rock in the following ways:

1. Decrease of the Ca content throughout the whole series.
2. Increase of the combined Fe and Mg content (fm) in the whole series.
3. Increase of the Fe content of the joint amount of Fe and Mg in the acid members (keratophyres) of the series, i.e. decrease of mg in the keratophyres.
4. A strong tendency to increase the Na content and to decrease the K content.

By these changes the joint alkali content and the alumina content remained practically unaffected.

The Kappebo effusives as a spilitic series

As it has become probable that the Kappebo effusives represent a spilitic series, they will be given a more detailed examination from this point of view.

It is admitted by the majority of the authors that the low temperature of solidification of a spilitic magma is due to its richness in volatile constituents. Opinions on the speculative question of the origin of such a magma and the source of the volatiles are widely divergent. Besides, the origin need not have been always the same. An account of the opinions on this subject is given by VAN OVEREEM (1948, p. 101 and 102). As far as the rocks treated here are concerned, there is no reason to make a choice out of these ideas or to add a new one, and accordingly, I will leave the matter alone and only try to elucidate the process of crystallization, as inferred from the observations. To do this, we had best turn our attention to the greenstones first.

In the greenstone of type III there are indications that pyroxene and plagioclase crystallized first in intersertal structure. The pyroxene suffered pseudomorphic alteration into epidote and some chlorite, and the originally idiomorphic plagioclase has been corroded and disintegrated by alteration products in such a way, that little remains of the initial igneous structure. In the fine-grained greenstone no pseudomorphism has been found. The granular epidote is scattered through the rock, but the same mineral also occurs in patches of a greater concentration, which have no sharp boundaries dividing them from the bulk of the rock. In addition, there are nodules in these rocks, consisting of epidote and quartz, in which sometimes calcite and chlorite occur. It seems to me that this marks a gradual transition from a hydromagmatic to a hydrothermal process of crystallization, in the course of which pyroxene, if present, was changed into epidote and chlorite or in other cases, at lower temperatures, epidote formed directly. A residual solution, rich in epidote, however, remained active in the rock, which finally crystallized in spots and nodules.

As final crystallization products of the hydrothermal stage veins occur, usually filled with epidote, quartz and calcite, also sericite and rarely chlorite, found in various combinations. Apparently, at the time, the consolidation of the rock had gone so far that the residual solution could only circulate and crystallize in veins. As a peculiarity of these pegmatitic veins must be mentioned the occasional idiomorphic crystals of epidote, sometimes entirely enclosed by a quartz crystal. They are probably comparable to the quartz-epidote pegmatite, described by BESKOW (1927, p. 176—179) from Södra Storfjället.

In the course of the processes, outlined above, the original plagioclase was altered into almost pure albite with saussuritic inclusions.

In some greenstones a green biotite was found, altered partly or mainly into chlorite. Possibly the biotite is an early product of crystallization, but there are no observations which might confirm this. A few dark greenstones are characterized by the presence of actinolite (and tremolite). These minerals evidently have formed after the first crystallization, as the orientation of the needles is irrespective of the relics of igneous structure and they even penetrate into some of the quartz porphyrocrysts. The amphibole, however, is distributed uniformly throughout the rock and has not been found in veins or nodules. Therefore, it is probably a newly formed product of the hydromagmatic stage. The formation of actinolite (and tremolite), along with epidote, might be explained by assuming a slightly different chemical composition of the fusion, e.g. a higher Mg content.

A peculiar unexplained feature is the presence of quartz porphyrocrysts which, in a few instances, even occur in the dark actinolite-bearing greenstones. They possibly derive from an intratelluric stage of crystallization. How they could have formed there, in such a basic fusion, is not clear.

If we come now to a more detailed consideration of the acid albite-rhyolites, we find essentially the same dark constituents as in the greenstones, but in smaller quantities. Nodules are not encountered; at most a somewhat irregular distribution of the dark constituents with, partly, a greater concentration in clusters and strings. Only the veins of quartz and albite, sometimes with a little calcite, are to be considered as hydrothermal products. To all appearance the fusion of the acid albite-rhyolite remained fluid down to a low temperature, and afterwards it crystallized or solidified as a glass in a comparatively small temperature interval. In other words, the hydromagmatic stage has been more important and the hydrothermal stage less important than in the greenstones.

It is likely that the sericite in the albite-rhyolites was formed, for the most part, at the time of the crystallization of the rock. In the course of the solidification the potassium feldspar, apparently, became instable and was partly replaced by sericite which gave the albites of the albite-rhyolite a "cloudy" appearance, because of the inclusions of extremely fine sericite scales. Another part of the sericite will have crystallized directly from the fusion at a lower temperature.

In analogy to the acid albite-rhyolites, we may attribute the sericite of the greenstones and the less acid albite-rhyolites to the same kind of origin. Thus, sericite is among the latest crystallization products, which is shown also by the occurrence of sericite in some of the veins in the greenstones.

Idiomorphic magnetite, which is common, must have crystallized in a relatively early stage, for this mineral is not found in the veins and nodules. Moreover, the grain size of the magnetite increases with the increasing of the primary grain size of the rock. The large porphyroblasts of magnetite in the albite-rhyolite near Lapperud (p. 99) are an exception to this rule. Calcite, on the other hand, with its large, irregular and unevenly distributed crystals, which are most often sieve-like because of the numerous inclusions, is to be regarded as a mineral essentially connected with the hydrothermal stage. The common occurrence of calcite in veins also points in that direction. On account of these considerations and of what has been said about the formation of the sericite, the vein of sericite, calcite and hematite on the spit of land to the northeast of Öjerud (p. 101) must be considered as a vein, in which a residual solution has crystallized. In the albite-rhyolites, however, small rhombohedra of calcite are also found in the groundmass. Possibly, the hydromagmatic stage lasted sufficiently long to permit the crystallization of idiomorphic calcite in the groundmass. The large idiomorphic crystals of dolomite, with abundant inclusions, in the albite-rhyolite on the shore of Kappebosjön, to the north of Lapperud, and the large hypidiomorphic calcite which has been found in one of the greenstones, do not fit in the picture outlined above. Possibly, they are connected with some other still unexplained phenomenon.

In the greenstones it is manifest how great a difference there can be in the association and the mutual relations of the dark constituents, even to such an extent that scarcely two of the samples examined under the microscope show the same details of contexture. This is illustrated by the table (p. 106), in which are listed the mineral associations of eight greenstones of type II. VUAGNAT (1946) and VAN OVEREEM (1948) emphasize an important phenomenon in the solidification of a spilitic magma, viz. the circulation of aqueous solutions which may give rise to a differentation in situ. This gives also a natural explanation of the great differences in the Kappebo greenstones.

As a matter of fact, after crystallization has started in a magma rich in gas, the opportunity to migrate for the solutions and volatile constituents remains open, since a large temperature interval has to be traversed before the solidification is completed. In this case not only an opportunity, but also ample reasons for such a migration exist. The magma remained in motion practically up to the moment of solidification in the outflows and the feeding channels, and as a consequence, pressure conditions varied continuously, the more so as we are led to assume that tectonic movements of the crust accompanied the outflows (p. 133). These circumstances provide ample scope for local differentiation (so called differentiation in situ), since the conditions varied strongly from one place to another. In the mobile fusion crystallization, resorption and migration of material, therefore, had a different course from one point to another, which resulted in rocks of a different association of minerals.

Thus, it appears that we must assume a somewhat different development of crystallization for the greenstones and the acid albite-rhyolites. This difference is probably determined by the chemical constitution of the original fusions, rather than by the outward physical conditions. In this connection it must be noted that the basic albite-rhyolites of type IV form, in many respects, a transition to the greenstones. The albite phenocrysts, here, show saussuritic inclusions and epidote is also present in the veins, while the quantity of dark constituents is greater and strongly varied in association and mutual proportions.

I am aware of the fact that the deductions with respect to the course of the crystallization, outlined above, bear a hypothetical character, but they provided the opportunity to summarize the most important characteristics of the effusives and dykes of the Kappebo formation.

The sedimentation

The most striking feature of the Kappebo formation is the irregular distribution of the different kinds of rock, there being no definite strata of homogeneous sediments over any considerable distance. Accordingly, it is impracticable to make a stratigraphical subdivision on petrographical grounds and SANDELL's (1941, p. 183 f.) fanciful attempt to do so, is considered to be completely incorrect. In the course of sedimentation marked differences in supply and in grain size of clastic material must have occurred from one place to another and, geologically speaking, from one moment to another. Moreover, complications due to volcanic activity occurring simultaneously with the deposition of the epiclasts, resulted in an irregular interstratification of lava flows and pyroclasts between the epiclastic sediments. As a consequence. the Kappebo formation is made up of smaller and larger lenses and irregular bodies of all rocks present in almost infinite variety, including many transitions between the various epiclastic and pyroclastic members. Only very roughly may some regions be outlined in which one particular rock predominates, as has been indicated in the description of the different types.

That conditions were variable from one place to another from the outset of the sedimentation, is shown by the deposits which have been found at or near the normal contact of the Kappebo formation and the gneiss-granite and which, therefore, must have formed about at the same time. They are mainly epiclastic rocks of all grain sizes viz.: slate, siltstone, graywacke and conglomerate, but also albite-rhyolite, greenstone, agglomerate and tuff. In order to find out whether the highly irregular position of the rocks in those parts of the Kappebo formation which were deposited later, may be due to a complicated tectonic structure, the northern part of the Kappebofjäll as far as Skogstjärn has been investigated in detail along a great many traverses. It was found that continuous layers of individual rocks do not exist and that the alternation of the rocks is sometimes so frequent that they cannot be recorded completely, even on a 1:5.000 scale map. For this reason, the different rocks of the Kappebo formation have not been marked by separate shades or symbols on the map. The inevitable simplifications in this manner of recording would tend to a misrepresentation of the way in which the Kappebo formation is built up, while in those regions where the network of traverses is less closely spaced, considerable inaccuracy would result.

As we have seen, the sediments of the Kappebo formation are characterized by a high content of meta-stable minerals (feldspar and muscovite). It follows that the sedimentary material has not been subjected to continued chemical weathering. Moreover, it is poorly sorted as to grain size and poorly rounded. Also previous sediments of the Kappebo formation have been partly broken up, as is shown by the fragments of slate, siltstone and felsite in the conglomerate and breccia. These characteristics of the sediments and the irregular way of deposition indicate a short, speedy, though irregular transport and a rapid deposition. This was probably effected by a number of rivers and streams with a steep slope. It implies that at the time of deposition considerable differences of level must have existed and so, in terms of morphology, young land forms must have been present or, in tectonical terminology, the earth's crust has been disturbed by movements just before or during the sedimentation. The presence in the Kappebo formation of its own disintegration products may be taken as evidence of continental deposition.

That the Kappebo formation was deposited upon a surface of strong relief may be seen from the westward bend of the boundary of the Kappebo sediments into the basement near Blixerud. Here is one of the few places where the strike and dip of the formation could be determined with certainty and it was found that the direction of strike was nearly normal to the curve. On a small scale the irregularity of the surface of deposition is plainly visible in an outcrop of gneiss-granite in the Kappebofjäll just to the east of Tångebo. Here, remnants were found of pelitic sediments of the Kappebo formation, which must have been deposited in a great number of gullies and excavations. Also small remnants of Kappebo sediments on the gneiss-granite, at great distances from the line of continuous exposure of the Kappebo formation, indicate considerable differences of elevation of the surface mentioned. These facts are entirely consistent with the environments inferred above.

The sediments of the Kappebo formation are as yet known with certainty only from Bengterud in the north down to V. Blekan, in the off-sets of the Sätersfjäll, 17 km to the south (sheet Rådanefors 1870). In an east-west direction isolated remnants of these sediments have been found between Svingsjön (sheet Upperud 1870) and Kålungen, i.e. over a distance of 13 km. From the descriptions of the sheet Rådanefors (KARLSSON and WAHLQUIST 1870) and Wenersborg (SIDENBLADH 1870), however, it may be assumed that the belt of supracrustal rocks, extending from Rådanefors 20 km to the south, belongs equally to the Kappebo formation. If this be true then the Kappebo formation is exposed in a narrow zone over a distance of 50 km, although with some interruptions. Only a small part of this has been studied in detail and we may assume safely that only a small part of the original formation was left by erosion, which precludes comments upon the shape and the development of the basin of sedimentation as a whole. In this respect further insight may be expected if additional investigations were to reveal the identity of the Åmål formation with the Kappebo formation. We will find (p. 136) that this possibility is not to be excluded.

The combination of the epiclastic sediments of the Kappebo formation is a typical example of what, in recent years, has become known in geologic literature as the "graywacke suite" (JONES 1938, p. lxiii). They are "Einschüttsedimente" (G. FISCHER 1933, p. 341) which, as stated, have been formed in a topography with great differences in level, therefore in a region of young tectonic movements. It is evident that on the coasts of rapidly subsiding geosynclines similar conditions appear and that the rocks of the graywacke suite will be frequently deposited there. It is for this reason that various authors (PETTIJOHN 1943 and 1949, TYRREL 1933) consider these poured-in sediments as a typical geosynclinal formation and, in accordance with this idea, the Kappebo formation is mentioned by SANDELL (1941, p. 186) as a typical geosynclinal deposit.

In my opinion, this view implies an unfounded restriction of the origin of strong differences of topographic level and it means also a depreciation of the widely divergent circumstances which, as regards sedimentation, may exist in a geosyncline. The enormous thicknesses of Mesozoic limestones, slates, and marls of the Thetis geosyncline as well as the Devonian limestones and sands and the Carboniferous shales and coal measures of the Variscan syncline are to the same or even to a greater degree typical geosynclinal deposits, whereas the conditions of deposition must have been entirely different from those of the graywacke suite. For a rather complete enumeration of the possible circumstances attending sedimentation in a geosyncline, I may refer to the paper of O. T. JONES (1938, p. lxii—p. lxvi).

In this connection, it must be pointed out that the poured-in type of sediments of the graywacke suite is often compared with the Flysch of the Alpine geosyncline (FISCHER 1933 and PETTIJOHN 1943 and 1949) and also with the Molasse deposits of the Alps. In fact, these rocks were also formed under conditions of great differences of elevation, shortly after or contemporaneous with tectonic movements in the earth's crust. The Flysch and Molasse, however, are defined in relation to space and time in the development of the Alpine mountain system, which is reflected in their petrography. Therefore, it is not correct to assimilate on petrographical grounds only, a deposit of the graywacke suite with the Flysch or the Molasse.

It might be suggested that the still preserved portion of the Kappebo formation has been deposited in the coastal zone of a geosyncline. However, neither the shape of the sedimentation basin, nor any remnants of sediments of the central zone of the geosyncline being known, such an assumption would be premature. Whenever the condition of a considerable difference of level over a small distance exists, the deposition of the Kappebo formation might have occurred as well in a fault block valley, alongside a fault zone or in a trough of restricted dimensions.

Tectonic movements

It appears to be impossible to give a detailed account of the tectonic development of the region before the deposition of the Dal formation. Only a few aspects can be considered.

So far we have found that, after the intrusion and solidification of the Kroppefjäll granite and the Teåker aplite-granite, intense orogenic forces must have operated, which gave a definite dynamometamorphic character to these rocks. The resultant mountain range must have been eroded deep into the gneiss-granite before the deposition of the Kappebo formation because, as far as is known, it rests everywhere upon the gneiss-granite. This implies that, if we admit for the thickness of the original roof of the batholith the small figure of 600 m (DALY 1933, p. 126), the thickness of the eroded pack of rocks must be estimated at least at 1000 m. It is not likely that after such a deep erosion, considerable differences of elevation still existed in the landscape, which nevertheless are required for the deposition of the Kappebo formation. It seems probable, therefore, that just before and during the deposition renewed tectonic movements have taken place.

The question arises whether we have to consider these movements as folding, as block faulting or as subsidence of a geosyncline or trough. In this respect one may state only that there are indications that movements along faults played a part before or during the sedimentation of the Kappebo formation, but this does not answer the question, because faults may occur in all three kinds of tectonic movements mentioned. In the discussion of the tectonics of the Dal folding, it will be pointed out that before the deposition of the Dal formation, movements took place along the fault which runs from Rörnäs to the south across Högesjön, via Blixerud down to Abborrtjärn (p. 164). It appears also (p. 166) that the position of the boundary between the gneiss-granite and the Kappebo formation near Åsen indicates a tectonic movement (probably a fault) which must be older than the deposition of the Dal formation. It is striking that from Kroktjärn to the south, a greenstone dyke is found which follows the fault first mentioned and that in the vicinity of Åsen, numerous dykes of albite-rhyolite and greenstone occur in the gneiss-granite. Towards the border of the Kappebo formation the frequency of the dykes is greatly increased, even to such an extent that they form small stocks of greenstone and rhyolite. This association of greenstone and albite-rhyolite dykes with the old faults suggests that movements occurred along these faults even before or during the deposition of the Kappebo formation, which as weaker spots in the earth's crust, lent passage to the uprising magma.

The unconformity between the Kappebo formation and the younger Dal formation is apparent from the reading of the map. Both to the east and to the west, the Dal formation rests directly upon the gneiss-granite over great distances, whilst the remnants of the Kappebo formation between Svingsjön (sheet Upperud 1870) and Kålungen indicate that this formation must have been present all over the region. It is, therefore, fairly sure that the Kappebo formation has been subjected to orogenic movements followed by deep erosion before the Dal formation was deposited. This erosion is indicated also by the presence of rounded pebbles and rock fragments of the Kappebo formation, which have been found in great quantities in the basal conglomerate of the Dal formation.

About the structure and the extension of the Kappebo orogeny we are completely in the dark. As we have seen, a definite succession of strata of a regional extent is not present in the Kappebo formation, so that the principal clue for unravelling its tectonics is lacking. Dip measurements could only be made exceptionally and these scanty data do not, by any means, allow the determination of what part of the deformation is due to the Kappebo orogeny and what to the Dal folding. A certain amount of information concerning the direction of the Kappebo orogeny may be found in the line of outcrop of the normal contact plane of the Kappebo formation and the gneiss-granite between Bengterud and Svartetjärn. Roughly speaking, it has a direction from the north-northeast to the south-southwest, even in those places where the strike of the Dal formation runs plainly north-south.

The schistosity

The dynamometamorphic changes in the rocks of the Kappebo formation are expressed by a general schistosity. The degree of schistosity varies widely: highly sheared schists are found, but also rocks which possess only a schistose cleavage, the foliation planes being relatively far apart. These extremes have only been met with locally, the majority of the rocks of the Kappebo formation holding an intermediate position.

The Kappebo formation is less schistose than the gneiss-granite, as is shown for instance, by the graywacke which is often conspicuously slightly foliated, though mineralogically it comes very close to the highly schistose gneiss-granite. More reliable means of comparison are provided by microscopic examination. In the gneiss-granite, a strong cataclasm of the quartz is common, even in those places in which the original structure is still visible. In the Kappebo formation it is restricted to the most schistose types. Finally, the frequent occurrence of such primary structures as are easily destroyed by dynamometamorphism in the greenstone, albite-rhyolite and tuff, indicate that these rocks were less metamorphosed than the gneiss-granite.

The schistosity of the Kappebo formation differs from that of the Dal formation. In the latter, only the appropriate beds in the lower part of the series are clearly schistose, whereas in the Kappebo formation this is the case for all the rocks. False cleavage is often met with in the rocks of the Kappebo formation, especially between Åsen and Stammen.

Comparisons

As appears from the literature study of the spilite problem by GILLULY (1935) most spilitic lava flows are of subaqueous origin. However, the exceptions show that "this locus is not essential for their development". The spilitic lava flows of the Kappebo formation are, in all probability, subaerial formations and thus form such an exception. It is of interest to examine whether the Kappebo formation as a whole can be compared to other series of rocks in which extrusions of spilitic lavas occurred above the sea-level. Although a thorough study of the literature on this subject has not been made, three instances of series of rocks may be cited, which show a great similarity with the Kappebo formation. They will be briefly discussed below.

1. In southwestern Wales (Pembrokeshire) a volcanic series of spilitic character occurs, the "Skomer Volcanic Series" of lower-Palaeozoic age (H. H. THOMAS 1911). The strongly differentiated series which, at least in part, has been deposited on land, consists of eight different kinds of lava ranging from soda-rhyolites ($SiO_2 = 79.64\%$) to olivine-dolerites ($SiO_2 \pm 46\%$). The sediments consist of coarse rhyolitic breccias, conglomerates, feldspatic grits and quartzites. The finer sediments are made up of subangular quartz, angular soda-rich feldspar and a few fragments of trachytic

and other acid volcanic rocks. Red clays also occur (THOMAS 1911, p. 208). Over small distances the sediments vary considerably in thickness. The Skomer Volcanic Series lies at the southern border of the Caledonian geosyncline of Great Britain. The contrast of level, required for the accumulation of such coarse sediments, was probably brought about by the subsidence of this geosyncline (O. T. JONES 1938).

2. In the geosyncline of Novaya Zemlya likewise, during the upper-Devonian, conditions prevailed which are summarized by BACKLUND (1930, p. 39) as follows:

"Die angeführten Beobachtungen und Daten geben mit recht groszer Sicherheit an, dasz die "spilitischen" Erguszgesteine nicht als Tiefseeergüsse gefördert wurden, sondern im Gegenteil nahezu als Landvulkanite zu bezeichnen sind. Die lebhafte Gesteinswechsel der sie begleitenden Sedimente, sowie der Wechsel von Erosion und Sedimentation zeugt von lebhaften Bewegungen der Erdrinde im Gebiet des oberdevonischen Vulkanismus, ungefähr wie sie für eine Flyschsedimentation innerhalb einer Geosynklinale charakteristisch zu sein pflegen. .... Der lebhaften Erdkrustenbewegung entspricht ein ebenfalls verhältnismäszig lebhaften Gesteinswechsel innerhalb der Vulkanitserie: ihre Differentiationsamplitude liegt zwischen einem Quartzkeratophyr (= "Spilit" im engeren Sinne) einerzeits und einem Pyroxenit ("Spilitaugitit") andererseits, mit einem Schwerpunkt vielleicht innerhalb der "Spilitdiabase" oder "spilitischen Basalte"."

3. A third instance may be found in the Permo-Carboniferous of the region between Andermatt and Disentis, at the northern margin of the Gotthard Massif, Switzerland, which has been described by E. NIGGLI (1944, see particularly p. 285—286). The Permo-Carboniferous has been deposited here in a trough, formed between two mountain ranges which later (during the Alpine orogeny) were to become uplifted as the Aar Massif and as the central and southern parts of the Gotthard Massif. On the margin of the trough coarse, clastic sediments have accumulated which vary considerably in thickness. In the centre of the trough poorly sorted argillaceous sediments were mainly deposited. During this sedimentation, which was continental, an intense volcanic activity prevailed in the trough, yielding quartz-porphyries, quartz-keratophyres and spilitic basic effusives. The volcanites, immediately after their deposition, were partly eroded, and supplied material for the psammitic sediments in the trough.

In all three series of rocks, mentioned above, the sediments, as in the Kappebo formation, are coarse fragmental, poorly sorted, made up of ill rounded material, extremely heterogeneous and greatly variable in thickness. This indicates tectonic activity immediately before or during the sedimentation. In the case of the Palaeozoic of Pembrokeshire and the Permo-Carboniferous at the northern margin of the Gotthard Massif, it is evident that these movements are related to the subsidence of a geosyncline or trough and not to mountain building. As to the Kappebo formation and the upper-Devonian of Novaya Zemlya, not much can be said as yet about the nature of these movements. BACKLUND (1930, p. 56) believes that the tectonic movements of the latter region are connected with the first oscillation, in positive direction, of the geosyncline, but, as far as I understood, this belief is based solely on the coarse clastic structure of the sediments, which suggests a Flysch facies (BACKLUND 1930, p. 39). However, far from all coarse, poorly sorted, clastic sediments have been formed in a Flysch facies.

A further point of correspondence between the Kappebo formation and the three series cited, lies in the strong differentiation of the spilitic outflows. Acid, basic and intermediate members of the series are always present.

It seems justified, therefore, to conclude that both the sedimentation and the volcanism of the Kappebo formation are the result of very definite geological conditions which, also in other regions, at different times induced similar sedimentation and volcanism.

These conditions seem to occur in regions of subsidence, but it would be premature to consider such an environment as the only possible one.

The stratigraphical position of the Åmål formation.

In the gneiss-granite a rhyolite has been found, as described on page 110, which is distinctly less dynamometamorphous than the granite-schist in which it occurs. It corresponds perfectly with the quartz-porphyry of the area between Ånimmen and Vänern described by VAN OVEREEM (1948, p. 12—17). On page 13 this author reports that the quartz-porphyries and the associated porphyrites occur as dykes in the "granite-mylonite", the lower degree of dynamometamorphism showing that they are younger than the "mylonitization". Leaving aside whether the term "mylonite" is justified for the highly schistose gneiss-granite in those places (see p. 75), the latter statement agrees perfectly with what I have found west of Svartetjärn and south of Blixerud. We may, therefore, safely assume that this rhyolite (= quartz-porphyry) and porphyrite are younger than the gneissification. Hence it becomes probable that they are to be included in the effusives of the Kappebo formation.

In the review of VAN OVEREEM's work by LARSSON (1948, p. 634—638) the reviewer denies the dyke-like occurrence of the quartz-porphyry and porphyrite in the gneiss-granite, but considers them as xenoliths in the granite, where, according to his view, they have been gneissified together. In that case, even for "an observer familiar with pre-Cambrian geology" it is difficult to explain why these xenoliths should have the property "to split off small apophyses which run further parallel to the mother dyke", as VAN OVEREEM (1948, p. 13) states and as I have been able to verify in the field, and to account for the observation which VAN OVEREEM mentioned to me in a personal letter as follows:

"To the west of Storön and Könningen lies a long strip of quartz-porphyry which can be traced for several kilometers until it disappears under the Dal formation. This strip has been marked separately on my map; it is but narrow, some 20 to 60 m. In places, there are a few thin intercalated layers of granite-mylonite, but in other places they are absent. This appearance of quartz-porphyry does not resemble a xenolith" (translated).

As regards the relation of these dykes to the gneissification VAN OVEREEM writes in the same letter:

"One may imagine that the quartz-porphyry, being a more resistant rock, partly escaped the effects of dynamometamorphism, but I refuse to accept that the same should be true of the porphyrite. Yet, in the porphyrite too the original structure is often preserved, which in this granite-mylonite zone is never observed in the case of the granite" (translated).

I am perfectly aware of the fact that VAN OVEREEM's conclusion (1948) to the effect that these porphyritic rocks occur as dykes in the granite, which in my opinion is correct, may be of revolutionary significance with respect to the stratigraphical position of the Åmål formation. I want to consider this more closely, without forgetting that there is insufficient information at hand to yield a definite conclusion.

The porphyritic rocks between Ånimmen and Vänern, described by VAN OVEREEM (1948), are interrelated, at least on the map, with the porphyritic and tuffaceous rocks of the Åmål sheet (1870). The latter constitute, in conjunction with the clastic sediments (feldspar-quartzite and conglomerates), the Åmål formation. That this is not only an apparent interrelationship is evident from the thin sections of the samples of Åmål rocks, which I had the opportunity to collect southeast of Bjäkebol in the parish of Tösse. They show a striking similarity to those of the quartz-porphyries and porphyrites between Ånimmen and Vänern. Moreover, this may be concluded also from the description of rocks of the Åmål formation to the south of Åmål as given by MAGNUSSON (1929a, p. 8—11).

This would imply that the Åmål formation is younger than the gneiss-granite, which fact seems to be confirmed also by the conditions near Örnäs just to the south of Åmål. During a short trip to this locality I found the feldspar-quartzite of the Åmål formation lying normally on the Kroppefjäll gneiss-granite with a dip of 45° to the northeast. Although the contact plane is nowhere exposed, no indications have been found of a fault. It is important that the feldspars of the feldspar-quartzite are entirely similar as to colour, composition and structure (perthitic Na-potassium feldspar), to those of the underlying Kroppefjäll gneiss-granite, while fragments of the granite itself are enclosed in the quartzite. The exposure just mentioned gives conclusive evidence of the age relations of the present rocks. Away from the contact this quartzite is no longer feldspar-bearing, but shows a flamboyant green colour. As appears from microscopic examination this is caused by the irregular distribution of epidote, sometimes abundant, between the quartz grains. On account of this petrographical peculiarity it does not seem likely that these rocks belong to the basal layers of the Dal formation. However, the possibility exists and should be taken into consideration. This is all the more imperative since investigators of renown (MAGNUSSON 1926, 1929a, 1929b and LARSSON 1947) consider the Åmål formation to be older than the gneiss-granite. The arguments in literature brought forward in favour of this view cannot dispel the doubt as to its correctness. They will be briefly discussed here:

MAGNUSSON (1926, p. 116 and 1929b, p. 28) refers mainly to the investigations of WINGE (1900, p. 34) and HOLMQUIST (1906, p. 117). WINGE, in a short account of a lecture, says that the granites break through the Åmål formation. On the page quoted of HOLMQUIST's work, nothing is to be found about the relative age of the granite of western Värmland and Dalsland with respect to the Åmål formation. However, he classes it among a group of granites of which he says (HOLMQUIST, 1906, p. 148): "Wenn diese Granite deutliche Kontakte gegen die Porphyr-Hälleflint-Gneiss-gesteine zeigen, verhalten sie sich geologisch zu diesen als jüngere Eruptivgesteine".

Fragments of porphyries in the granite are reported by WINGE (1900, p. 340), MAGNUSSON (1929a, p. 20 and 1929b, p. 32) and LARSSON (1947, p. 324—325). The last author gives a brief description of a local granitification of the Åmål formation by the intruding granite. The enclosures of fragments of volcanic rocks in the granite seem to afford a strong argument in favour of the younger age of the granite. It is, however, weakened to a certain extent since we know from LARSSON's paper (1947, p. 323) that an older supracrustal formation has been recognized in the oldest gneiss-complex.

The central position of the granite in the Gillberga basin on the Åmål formation may also be due to big overthrusts, as is pointed out by MAGNUSSON (1929a, p. 52). This conception becomes more probable since large scale overthrusts must be assumed in middle Dalsland, as will be described in the chapter on the Dal orogeny.

The doubt as to the younger age of the granite is still increased by the fact that granite pebbles have been repeatedly reported in the conglomerate of the Åmål formation (TÖRNEBOHM 1870c, p. 18; HUMMEL and ERDMANN 1870, p. 25, fig. 2 and HOLMQUIST 1912, p. 399).

It is not my intention to draw a conclusion from the enumeration of these contradictory arguments; I only wished to emphasize that the information as yet available from literature on the subject must be considered insufficient to decide conclusively the relative age of the Åmål formation with respect to the granite.

However, if it becomes apparent that the Åmål formation is younger than the granite, it is still possible that the former is not synchronous with the Kappebo formation. As previously stated (p. 133), the deposition of the Kappebo formation in the Dalskog Dals-Rostock region is separated from the post-granitic orogeny by a long period of denudation. During this period the Åmål formation could have been deposited elsewhere.

CHAPTER IV

THE DAL FORMATION

Introduction

Since Dalsland was first mapped by the Swedish Geological Survey, in the sixties of the last century, the folded sedimentary Dal formation has been known. This formation rests unconformably on the Kappebo formation and the still older gneiss-granite. The original mapping of the Dal formation in the Dalskog Dals-Rostock region (sheet Upperud, 1870) is remarkably correct, if we take into account its scale and the early date of its origin, as well as are the annexed descriptions (TÖRNEBOHM 1870a). However, little attention was paid to the tectonics of the formation at that time. Yet, the schematic sections published by TÖRNEBOHM (1870a, p. 21 and 45 and 1870b) give a better idea of the structure of the Dal formation than do the later tectonic interpretations by HAUSEN (1931) and by SANDELL (1941). The tectonics of the Dal formation in the Dalskog Dals-Rostock region are dealt with in a separate chapter, in which it will also be compared with that of the adjoining areas.

An extensive stratigraphic, petrologic and tectonic study of the Dal formation, north and northeast of the investigated area, has been published recently by A. J. A. VAN OVEREEM (1948). Owing to the close resemblance between the rocks of the Dalskog Dals-Rostock area and the region to the north of it, frequent references to VAN OVEREEM's work will be made in the following discussion of the Dal formation. The many authors who have indirectly been concerned with the Dal formation are not cited here, as they may be found in the above-mentioned treatise in which a general account is also given of the literature on the much discussed question of the age of the Dal formation. The investigated area does not provide further evidence regarding this question. The Dal formation is considered to be of Jatulian (=pre-Jotnian) age by LARSSON (1947) and VAN OVEREEM (1948).

In the Dalskog Dals-Rostock region the Dal formation has the following development:

Liane layers	subgraywacke	400 m
	conglomerate	10 m
	subgraywacke with slate lenses	110 m
Quartzite layers	quartzite	280 m
Upper slate layers	quartzose slate	200 m
Spilite-bearing layers	quartzitic beds alternating with spilitic outflows	200—300 m
Lower slate layers	quartzose calcareous slate	250 m
	slate	100 m
Basal layers	arkosic quartzitic sandstone	250 m
	intercalated conglomerate	10 m
	arkosic quartzitic sandstone	110 m
	basal conglomerate and breccia	locally
		1920—2020 m

The whole series, although disturbed by faults, occurs between the Dalbo plain and the Kappebofjäll. More to the west, only the basal layers and the lower slate layers reach an important extension. The intercalated conglomerate thins out in a westward direction and the slate bed of the lower slate layers becomes thinner in the same direction.

To the lower slate layers, the spilite-bearing layers and the upper slate layers together the term "slate layers" was applied by TÖRNEBOHM (1870a, p. 23).

The basal layers

The basal layers of the Dal formation are mainly composed of arkosic quartzitic sandstone, locally passing into arkose. In addition, a basal conglomerate and breccia often mark the transgression plane and in the middle of the series an intercalated layer of conglomerate has been found. In places occur lenses of a fine-grained conglomerate as well as thin beds of graywacke of small extension.

Of the rocks of the Dal formation the basal layers are the most widely exposed. Besides in the Dals-Rostock monocline [1]) they have been encountered in small outcrops in the Dalbo plain and cover large areas around Dalskog and Teåkersjön. In the Dalskog Dals-Rostock region the thickness of the basal layers may be evaluated in general at 370 m, but in places it may be far greater. This must be ascribed to the irregularity of the surface on which the Dal formation was deposited. This irregularity is particularly apparent from the trend of the normal contact plane e.g. to the south of Teåkersjön and the western arm of Kappebosjön, to the southeast of Gunnesbyn, to the north of Kabbo, to the south of Römäs and near Helvetestjärn. In several of these localities it may be observed that the contact plane makes an angle with the bedding plane. More often than not the depressions of the transgression surface are filled up with basal conglomerate and breccia. The construction of the cross-sections shows the differences of elevation of the transgression plane amounting to 100 m (section VI). Such an irregularity of the pre-Dal surface could be determined to the south of Marsjön by P. HEYBROEK and H. J. ZWART (1949, p. 429). The basal layers which outcrop in the overthrust mass to the west of Landsbol are thinner (about 250 m), although probably the whole of the section is exposed here (p. 158). This seemingly abrupt change in thickness becomes clear if we realize that these beds were carried along by the thrust mass and were deposited at least 14 km more to the west (p. 172).

There is often a marked difference in metamorphism between the basal layers and the rocks which they overly unconformably, especially when the latter are gneiss-granites. The angular fragments in the basal conglomerate and breccia have a random orientation and bear no sign of deformation. Neither do the well rounded pebbles of the intercalated conglomerate, the smaller ones of which are often globular. However, from Helvetestjärn to the south the metamorphism increases strongly and east of Åsen even the hardest quartzite pebbles are transformed into protracted lenses. This deformation is also seen in the microscopical structure of the more fine-grained layers. This must be related to the intensive tectonic movements which are

[1]) By the Dals-Rostock monocline is meant the eastward dipping portion of the Dal formation, exposed between the Kappebofjäll and the Dalbo plain.

responsible for the overturned position of the layers. Owing to the same cause and eventually to the movements along the Kroppefjäll fault, the basal conglomerate in the Dalbo plain is highly metamorphosed. In the vicinity of other faults the rocks sometimes have been brecciated and cemented again by quartz veins. This has happened intensely due west of Dalskog.

Arkosic quartzitic sandstone

This rock has been termed quartzitic sandstone by TÖRNEBOHM (1870a, p. 24). VAN OVEREEM (1948, p. 25) calls it an arkose. Although, in places, the feldspar content of the arkosic quartzitic sandstone is such that we may speak of an arkose, it is usually between 10 and 25 % so that it is preferable not to call it an arkose, but to use arkosic or feldspatic as an adjective. Since the rock is cemented by quartz and is well indurated it should be classified as a quartzite, according to most authors (i.a. GROUT 1932, p. 366). However, there are authors (cf. RICE 1945, p. 333) who want to restrict the term quartzite to recrystallized sandstone and use the terms siliceous quartz sandstone or quartzitic sandstone for silica-cemented sandstones. In order not to depart unnecessary from the original name given by TÖRNEBOHM (1870a), which has become familiar in literature, the latter term has been chosen.

The arkosic quartzitic sandstone is usually of a light red-brown colour, sometimes also grey or light lilac. With the naked eye a granular structure is often discernible. This is especially distinct by the grains of feldspar which are always present in smaller or greater quantities. On a fresh fracture the feldspars are red, but at the weathered surface they are white or even dissolved entirely; the rock then shows small holes. At first sight the rock looks massive, but at a closer inspection it proves, in most cases, to be distinctly stratified by fine bands, alternately richer and poorer in feldspar. Ripple marks have often been observed (fig. 9, p. 103). Near Rörnäs and Kabbo and to the southeast of Korsgården cross-bedding is developed close to the contact plane, the dimensions of sets of the same inclination amounting to several meters. In the last two localities mentioned it is readily seen that the cross-bedded deposits have formed in depressions of the transgression surface.

Under the microscope the grains appear to be rounded to well rounded. The sorting is poor and the size of most of the grains varies between $^1/_8$ and $^1/_2$ mm. The variation diagram of the grain size given by VAN OVEREEM (1948, p. 26 fig. 8) has been made from an extraordinary fine sample, but still gives a fair idea of the sorting.

The quartz content varies strongly, but is usually between 75 % and 90 %. The quartz grains are oval. Besides quartz, grains of feldspar have been found, consisting chiefly of Na-potassium feldspar, but also of plagioclase. The Na-potassium feldspar exhibits a perthitic structure or has a microline lattice. The feldpars are commonly coloured by a red pigment (hematite?). Grains of a felsitic rock, often rich in sericite, occur also in the quartzitic sandstone.

Frequently, a small amount of interstitial material is present, consisting of very finely divided quartz together with sericite and ore-like material. In most cases however the rock is cemented by quartz in optical continuity with the detrital grains, which gives rise to a mosaic structure. Without analyzer, the granular structure is still clearly visible. A small quantity of ore (leucoxene and magnetite?) is present as little grains.

In many cases some recrystallization is observed in the rock, brought out by a suture structure at the boundary of the secondary enlarged quartz grains. Near faults and in the overturned beds south of Helvetestjärn a mortar structure is developed within the rocks. With increase of the amount of feldspar the quartzitic sandstone passes locally into an arkose. This is notably the case close to the transgression plane, where it is sometimes calcareous, and in the vicinity of the intercalated conglomerate. In the south of the investigated area, near Örsjön, the entire section between the transgression plane and the intercalated conglomerate is composed of arkose. Northeast of Dalskog it is above the intercalated conglomerate that much arkoses are found, which often become coarse-grained and pass into a fine conglomerate. The arkose and the fine conglomerate are not confined to a definite horizon, but occur as lenses scattered throughout the beds. Lenses of fine conglomerate of this kind have also been found in the overthrust arkosic quartzitic sandstone near Landsbol as well as at the shore of Kappebosjön, south of Tonebyn, and at the shore of Teåkersjön, south of Södra Halängen. The well rounded pebbles and grains of this conglomerate are rarely more than $^1/_2$ cm in size and consist for a great part of quartz and red feldspar, joined by small pebbles of aplite-granite and fragments of felsite. Infrequent small pebbles of the violet quartzite, which play such an important part in the intercalated conglomerate, have also been found. These conglomerates are often porous, because there is too little interstitial material to fill all the pores between the grains and the pebbles. Consequently these rocks are readily accessible to circulating surface water and, as a result, are mostly strongly weathered.

In the great outcrop of quartzitic sandstone along the road west of Dansbo, a number of thin beds of slate and graywacke, about 20 cm in thickness, have been found. The graywacke consists of poorly sorted quartz and red feldspar, together with fragments of slate, about 2 cm long, cemented by a green matrix.

Basal conglomerate and breccia

Not infrequently a basal conglomerate, highly variable both in grain size and composition, has developed on the contact plane of the Dal formation and the underlying rock. Its composition is mainly determined by the subjacent rocks, the fragments being for the most part little or not rounded. This indicates transport over a short distance. The matrix consists of graywacke. Upon these characteristics alone it is often difficult to distinguish the basal conglomerate of the Dal formation from the conglomerates and breccias of the Kappebo formation. However, there are a few criteria which may serve as a guide in cases of doubt. In the basal conglomerate of the Dal formation the matrix is not always graywacke, but may be arkose or quartzitic sandstone which may also occur as laminae and lenses in the conglomerate. Moreover, well rounded pebbles of light violet quartzite which play such an important part in the conglomerate intercalated in the quartzitic sandstone higher in the section, are rarely found. These components are not encountered in the Kappebo conglomerate.

Nearly all the rocks described above occur as fragments or pebbles in the conglomerate. To the south of Teåkersjön and the western arm of

Kappebosjön, fragments of the Teåker aplite-granite, up to 8 cm in size, form the most important constituent. Of the augen-gneiss only a few large pebbles have been found, but the big red feldspars and the milky blue quartzes in the conglomerate must be derived from this rock. South of Rörnäs the conglomerate has abundant inclusions of graywacke and slate, while to the east of the Kappebofjäll inclusions of felsite and porphyry predominate. Under the microscope it is possible to recognize the different types of rock of the Kappebo formation which we have seen before. In the Dalbo plain the fragments and pebbles of aplite-granite are again predominant. The thickness of the basal conglomerate and breccia may vary strongly. To the south of Teåkersjön and the western arm of Kappebosjön vast quantities of it have been deposited, just as to the south of Rörnäs, near Helvetestjärn and in the Dalbo plain, south of the railroad. In other localities it is thin or altogether absent.

Intercalated conglomerate

All along the Kappebofjäll and to the north and northwest of it, a very typical polimictic conglomerate is intercalated in the arkosic quartzitic sandstone. This bed of conglomerate is situated 250 m below the top of the series. Its distance to the base of the formation is not uniform since the basal plane is irregular, but is usually between 100 and 150 m. The bed thickness varies from 4 to 20 m. To the east and northeast of Dalskog the intercalated conglomerate forms an excellent key horizon to unravel the tectonic relations. It marks the course of three anticlines within the monotonous quartzitic sandstone area. The most characteristic feature of this conglomerate is the abundance of well rounded pebbles of a violet to light lilac quartzite, which may attain 18 cm in size (fig. 10, p. 103). In addition, we find well rounded pebbles of aplite-granite and also, rarely, of graywacke. The pebbles of felsite and porphyry which were encountered, are more angular and oblong. This is probably due to the original shape of the pebbles before they were transported. The matrix of this conglomerate is made up chiefly of light-toned arkose. The pebbles of quartzite, as described also by VAN OVEREEM (1948, p. 27), are composed of well rounded and well sorted quartz grains of about $^1/_2$ mm diam. Suture and mortar structures are common. There are practically no other minerals in these pebbles. Up to due east of Dalskog the conglomerate can be traced with certainty. Possibly, it occurs again in a small outcrop on the edge of the marginal terrace west of Korsgården, but there the pebbles are much smaller. More to the west the intercalated conglomerate appears to be pinched out. At the shore of Kappebosjön, south of Tonebyn, and at the shore of Teåkersjön, south of Södra Halängen, conglomerates have still been found among the quartzitic sandstone, but their content and grain size show that they are to be paralleled to the lenses of fine-grained conglomerate which, northeast of Dalskog, are of such a frequent occurrence in the quartzitic sandstone and do not belong to the conglomerate bed with quartzitic pebbles, as it is supposed by TÖRNEBOHM (1870a, p. 27). TÖRNEBOHM (1870a, note on p. 26 and 27) points out that this conglomerate with quartzite pebbles in Dalsland "has a small extension in its east-west direction as compared to its north-south direction. If its occurrence, from the northernmost findspot (in the Laxaby parish, sheet Baldersnäs) is traced to the south, it is found that it is mainly confined to a narrow zone which makes a curve, bending slightly eastward, and extends via Tanesjön and Råvarp and west of Örsjön down to lake Långhalmen (sheet Rådanefors), at the southernmost end of which it is found for the last time" (translated).

As far as may be ascertained from literature data, this conglomerate is also an intercalated conglomerate in other parts of Dalsland (HUMMEL and ERDMANN 1870, p. 48—52 and VAN OVEREEM 1948, p. 27). Near Billingsfors (Laxaby parish, sheet Baldersnäs 1870) one might think at first that it is developed there as a basal conglomerate. However, I have been able to ascertain that there a major fault, striking north-south, conceals the normal contact of the conglomerate with the underlying rock.

From its composition it is clear that this polimictic conglomerate is not to be considered as an intraformational conglomerate and it is necessary, therefore, to provide an explanation for the strange fact that, apparently in the middle of the sedimentary basin, a narrow belt of conglomerate has been intercalated. So, we may well ask whether it is to be related to the north-south trending facies boundary between the eastern and western part of the basin of sedimentation which was established by VAN OVEREEM (1948, p. 105) alongside lake Åklången. This boundary must have been a shore line at the time of deposition of the intercalated conglomerate, while the region lying to the east must have been subject to denudation. It now appears very probable that the conglomerate has been deposited along that coast and thence it becomes clear that the conglomerate has a wide extension in a north-south direction, but is very limited in an east-west direction. If this view is correct, then we must admit a transport of material from the east, which discounts the generality of VAN OVEREEM's statement (1948, p. 105) that the supply of material for the deposition of the Dal formation came from the west.

The problem of the origin of the quartzite pebbles, raised already by TÖRNBOHM (1870a, note p. 27), is as yet unsolved. At any rate it is made clear by the presence of these pebbles that at least a part of the arkosic quartzitic sandstone is constituted by second cycle material.

The lower slate layers

Next to the basal layers comes a sequence in which slates play an important part. In contradistinction to the slates which occur higher up in the Dal formation, they will be called lower slate layers. At the base they consist of a comparatively thin bed of slate, overlain by a great thickness of quartzose calcareous slate, composed of a variety of rocks different in character and constitution. The thickness of the lower slate layers can be estimated at 350 m in the Dals-Rostock monocline. West of it, the top of the series is not found. A section of the incomplete series, exposed in the Bäcke valley, still amounts to 230 m (HEYBROEK and ZWART 1949, p. 430).

Slate

Immediately above the basal layers lies a comparatively thin layer of dark slate, usually with a good cleavage. The transition to the basal layers is a gradual one, just as it is more to the west (HEYBROEK and ZWART 1949, p. 430) and to the north (VAN OVEREEM 1948, p. 28) of Dalsland, owing to the development of thin bands of slate which gradually increase in number and thickness. The thickness of this layer is about 100 m in the monocline

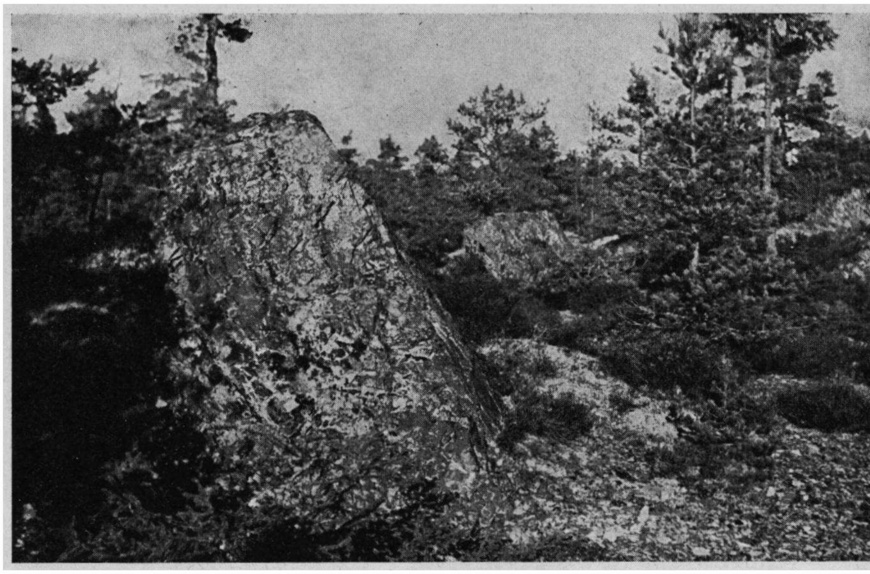

Fig. 17.
Hogbacks of massive beds of quartzite of the quartzite layers of the Dal formation. West of Råbäck.

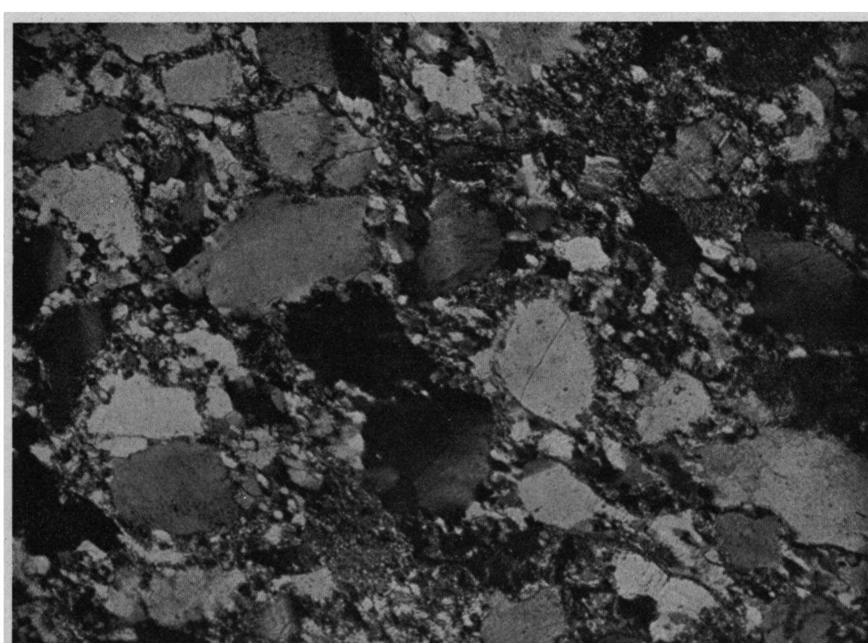

Fig. 18.
Photomicrograph of the subgraywacke of the Liane layers of the Dal formation. Crossed nicols. 42 × lin.

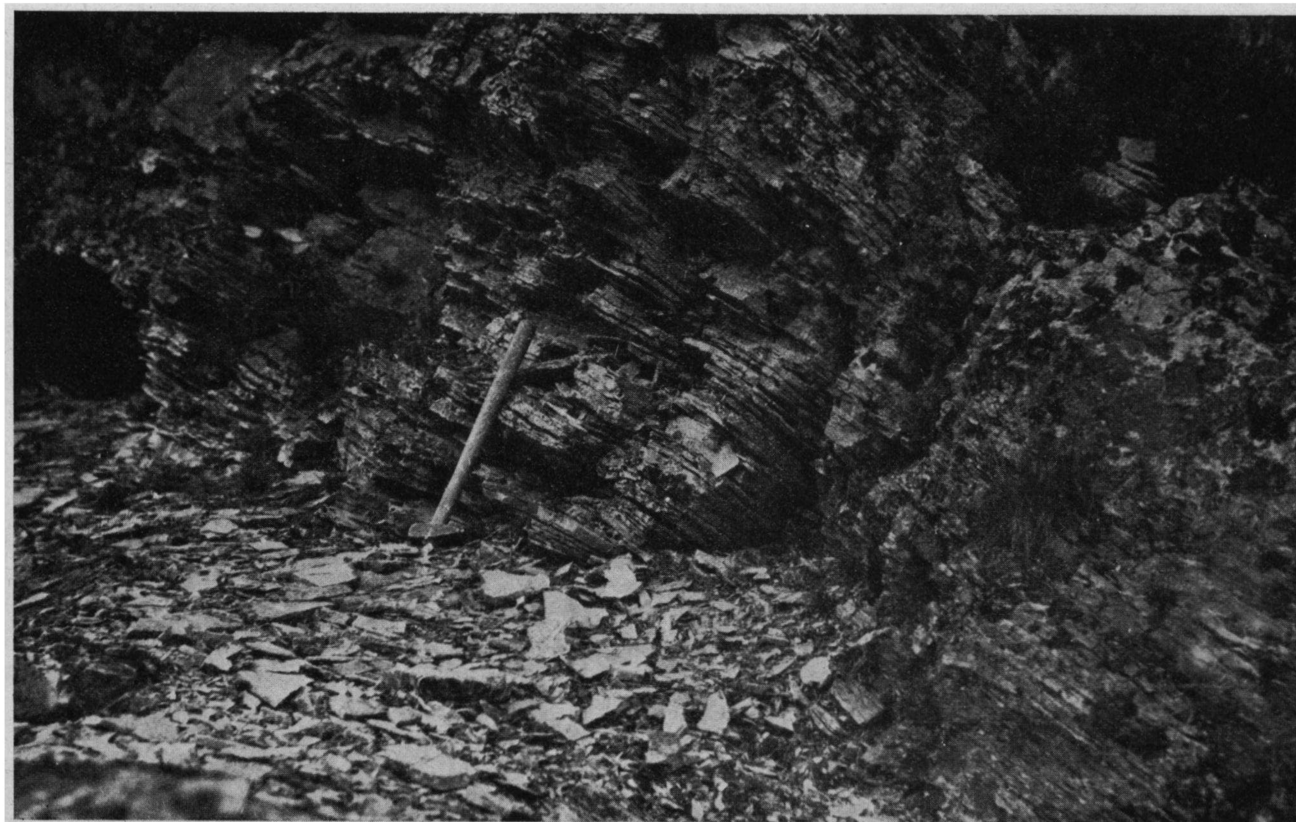

Fig. 19.

Very fine laminated quartzite of the quartzite layers of the Dal formation. North of Råbäck.

of Dals-Rostock, but thins down in a western direction and is 30 m thick north of Teåkersjön (HEYBROEK and ZWART 1949, p. 430).

Upon microscopic examination the rock appears to be almost entirely composed of a sericite-like mineral in a state of extremely fine subdivision. Quartz occurs as infrequent grains of larger size (0.08 mm), scattered through the rock. In places, idiomorphic crystals of pyrite (2 mm) and calcite (1 mm) were found.

As stated by TÖRNEBOHM (1870a, p. 31) the cleavage of the slate is due to a transverse foliation, especially prominent where the tectonic movements have been intense. Its good cleavage led to an extensive exploitation of slate-quarries in past years for roofing purposes. In the region under discussion, this was done mainly in three big quarries, viz. near Halängen, near Källsviken and along the eastern shore of Långtjärn, besides in a whole series of smaller ones. In the middle of the slate layer is an intercalated quartzitic sandstone bed, varying in thickness from a few meters to 10 m.

Quartzose calcareous slate

The great variety of rocks overlying the slate are termed calcareous slate by TÖRNEBOHM (1870a, p. 36) and marl by VAN OVEREEM (1948, p. 28). Actually, the rocks are all well consolidated and slightly metamorphosed, quartz being always present in a considerable quantity. Accordingly, the term quartzose calcareous slate (GROUT 1932) has been chosen here, also because of its close conformity with the original name.

The transition to the underlying slate is a gradual one and is effected by an increase of the amount of calcite and quartz in the rock. The most common rock type in this series is a quartzose slate with abundant thin lenses and irregular knots of light pink calcite. Calcite is frequently concentrated in irregular veinlets across the stratification which give the rock a brecciated appearance. This can be ascribed, as pointed out by VAN OVEREEM (1948, p. 29), to the fact that the calcium-carbonate becomes mobile at a low degree of metamorphism and is deposited subsequently in joints of restricted dimensions, perpendicular and parallel to the bedding. Owing to the irregular distribution of quartz and calcite, the stratification planes and cleavage planes have been irregularly corrugated during the folding of the beds.

Microscopically the rock is found to consist of sericite, ore-like material and a variable though important quantity of ill rounded quartz grains of 0.05 mm average diameter. It is crossed by the irregular veins and bands of calcite.

In various places in the Dals-Rostock monocline, not far above the base of the series, the lime content is increased to such an extent that layers of white quartzose limestone are intercalated. North of Prästbol, not far from the base, a red-brown rock has been found ranging from quartzose limestone to calcareous quartzite. In the railway cutting to the west of Dalskog, it forms a continuous bed which is also found more to the north, west of Berga. The red-brown colour is due to finely disseminated ore. The bed is often well stratified by alternately coarser and finer laminae and shows, sometimes, marked incongruous little folds of small dimension, probably caused by slumping.

In the upper portion of the series, in particular, rocks occur, distributed arbitrarily, with an important increased quartz content. They are mostly red-brown rocks with a brown crust due to weathering, the surface of which is studded with numerous rhombohedral holes, caused by the solution of calcite. They are termed "iron-bearing argillaceous sandstone" by VAN OVEREEM (1948, p. 30). Often they have a brecciated structure with larger and smaller fragments of the same rock of slightly different shade, embedded in a matrix of about the same composition. This is a typical example of an intraformational breccia or, as CAYEUX (1935, p. 144) calls it, a "brèche sédimentaire s.str." which must have been formed by wave action on a newly deposited sediment.

Under the microscope these rocks appear to be composed essentially of a mixture of microscrystalline quartz, calcite and a dark pigment, made up mainly of ore (d. av. = 0.01 mm). In some, slightly coarser, parts a granular structure may be recognized. The proportions of the minerals may vary strongly and thus determine the differences of colour in the intraformational breccias. In this fine- grained mass occur numerous, large, idiomorphic crystals of calcite, often bordered by a rim of ore. They may attain several mm in size (cf. VAN OVEREEM 1948, Pl. I, fig. 4). Big quartzes, of variable size, also lie scattered in the microcrystalline matrix. They are often full of inclusions and have an irregular shape. Irregular streaks of coarsely crystallized quartz also occur. The large crystals of calcite and quartz must be considered as porphyroblasts.

I did not find any indication in the Dalskog Dals-Rostock region that these rocks contain tuffaceous matter or that the porphyroblasts result from volcanic activity, as it is suggested by VAN OVEREEM (1948, p. 31—32).

But I fully agree with VAN OVEREEM (1948, p. 29—30), when he denies the existence of a younger, unconformable breccia, overlying the Dal formation and built up by its disintegration products, as has been assumed by TÖRNEBOHM (1870a, p. 56—58). This alleged breccia must be considered mainly as the intraformational breccia (the so called "sandstone breccia" of TÖRNEBOHM) and as the quartzose slate which, through the veins of calcite, has obtained a brecciated character (the so called "slate breccia" of TÖRNEBOHM), both forming part of the slate layers, discussed above. Moreover tectonic breccias, such as occur below the thrust masses in the northwestern part of the region, have been mapped by TÖRNEBOHM as younger, unconformable breccias (cf. HEYBROEK and ZWART 1949, p. 433).

The spilite-bearing layers

The lower slate layers are followed by a succession of beds, characterized by the presence of outflows of spilitic greenstone. The latter alternate with quartzose sediments and quartzites, occasionally mixed with some tuff. These intercalated sediments are strongly variable in character and content, but since quartzite predominates, they will be called quartzitic beds. The thickness of the spilite-bearing layers varies between 200 and 300 m.

Spilites

After the very detailed description and the petrologic study of the spilites of the Dal formation published by VAN OVEREEM (1948, p. 43—104) it is not necessary to treat these volcanic rocks again. The distribution of the spilitic effusives in the monocline of Dals-Rostock, however, is much more irregular than more to the north in Dalsland where, almost continuously, three layers are exposed which are called by VAN OVEREEM (1948, p. 45) first, second and third spilite. The first and the second spilite layer are met with repeatedly in the region under discussion, but they have an irregular shape and often fail altogether. Also irregular but almost continuous is the third spilite layer, its variable thickness may attain 100 m. North of the Kroppefjäll sanatorium it is exposed over a great width. This may be ascribed largely to tectonic causes (p. 163), though here the position of the layer is not clear. South of the sanatorium practically no spilite has been found, though the third spilite might be expected there. It appears to have been pinched out altogether, just as it has east of Dansbo.

This irregularity is difficult to explain. Probably, we are here at the limit of the area of distribution of the spilitic effusives, where the lava flows were less fluid. Owing to the large viscosity they did not form a horizontal surface. In addition, small differences of level of the surface of deposition have had a great influence on the course of such a viscous lava flow. In the third spilite thin beds and lenses of quartzite have been intercalated in various places, indicating that this layer is not to be considered as the consolidation product of one single flow of lava, but must be the result of several such flows.

The quartzitic beds

As stated above, these beds are made up of rocks varying both in character and content. The lower beds are not very different from the rocks of the quartzose calcareous slate layer; rather all the sediments described from this layer are also represented in the lower parts of the quartzitic beds. The increase of the quartz content towards the top, which could be observed in the quartzose calcareous slate, continues in these beds, as a result of which the calcite-veined slate, low in quartz, is rarely found here, whereas the rocks with a higher quartz content, including the intraformational breccia, become predominant. With further increase of quartz towards the top of the section, the rock passes into a red quartzite which usually forms a thick bed underlying the third spilite. In the lower portion of the quartzitic beds, occur thin laminae of purple shale, besides the rocks already mentioned. Under the microscope they appear to be composed of small quartz grains, a dark red pigment, calcite (sometimes abundant) and ore. In these thin beds mud cracks have often been found. The red quartzite higher up in the section frequently shows an extremely fine lamination owing to thin dark bands produced by a dark pigment, occasionally also by fine bands rich in calcite. VAN OVEREEM (1948, p. 34) found indications that these sediments have been mixed with tuffaceous material. West of Hult, a few thin beds of graywacke are intercalated between the quartzites of this series.

The upper slate layers

In the monocline of Dals-Rostock the upper slate layers consist of 200 meters of fairly homogeneous quartzose slate. Only near the top occur one or more intercalated beds of quartzite, up to a few meters in thickness. The quartzose slate is always more or less calcareous. Microscopically it appears to consist mainly of a great quantity of a sericite-like mineral, together with finely divided quartz and some calcite. The sericite-like material is sometimes partly replaced by finely disseminated ore, often associated with a brown pigment. Occasionally, ripple marks have been found on the bedding planes. On the road just north of Tegen this is very well exposed.

The quartzite layers

Contrary to what has been found to the north (between Råvàrp and St. Ärven, VAN OVEREEM 1948, p. 37) the quartzite layers in the Dals-Rostock monocline are constituted by fairly homogeneous quartzites of 280 m thickness. Only near the top of this very pure quartzite a few lenses of violet slate are intercalated. They probably form the counterpart of the "purple slaty sandstone", ranging in thickness from 80—160 m, which was found more to the north (VAN OVEREEM 1948, p. 37).

As we have seen above, near the top of the upper slate layers occur one or more beds of quartzite, marking the transition to the sedimentation of the homogeneous quartzite. This rock is pale yellow, nearly white, or light violet in colour. Upon microscopic examination it appears to consist almost exclusively of quartz associated with some sericite and ore, in very small quantities. The quartz grains are surrounded by mortar quartz or connected by a suture structure. A few lenses of breccias occur, composed of angular fragments of quartzite (up to 1 cm diam.) cemented by quartz or calcite.

Stratification of the quartzite is usually distinct, but a good parting along the bedding planes is not nearly so common. In the field, therefore, this rock is apt to occur in massive beds of several meters thickness. Due west and north of Råbäck, however, a thin lamination is very conspicuous in places, as it is underlined by micaceous bands. Parting along the lamination planes, less than one half cm apart, is therefore perfect (fig. 19). Due west of Råbäck alternations of similar easily parting beds with beds of massive rock have brought about small "hogbacks". The fissile beds have been worn away more rapidly by erosion, as a result of which the massive beds stand out as pillars in the landscape (fig. 17).

The Liane layers

The uppermost portion of the Dal formation has been called Liane-schist layer (TÖRNEBOHM 1870a, p. 42), after the Lianefjäll where it is exposed over a large area. These layers will be referred to here as the Liane layers since, in the investigated area, there is no trace of schistosity.

The Liane layers outcrop exclusively in the Årbolfjäll and are composed of the following beds:

Subgraywacke	> 400 m
Conglomerate	10 m
Subgraywacke with slate lenses	110 m
	> 520 m

The contact of the Liane layers with the underlying quartzite is exposed only in a few small outcrops and appear as a sharp boundary. Close to the contact plane the subgraywacke contains a few small scattered pebbles of quartzite and felsite. On the map there is no sign of any unconformity between these adjacent layers.

Subgraywacke

This rock has been called graywacke by TÖRNEBOHM (1870a, p. 42; 1882—'83, p. 632) and SANDELL (1941, p. 31), whereas LJUNGNER (1927—'30, p. 30) and VAN OVEREEM (1948, p. 41) refer to it as sparagmite, which is more or less synonymous with graywacke. Indeed, the most common rock type of the Liane layers is much like a graywacke on account of the nature and the amount of its matrix (fig. 18). The feldspar content, however, is low and the grains of quartz are all rounded and well sorted, so the rock represents an intermediate type between a typical graywacke and a quartzite. Therefore, the term subgraywacke, according to the definition given by PETTIJOHN (1949, p. 255), is justified here.

The subgraywacke is of a grey, dark grey or purplish colour. It is well stratified, usually in thick massive beds. Near Myrerna it is quarried for building-stone. The rock is chiefly made up of quartz grains (d. = 0.5 — 0.25 mm) which are well rounded and well sorted. In addition occur scarce grains of microcline, perthite and albite. The large amount of matrix, which may constitute as much as $^1/_3$ of the rock, consists for the most part of sericite and extremely fine-grained quartz (and feldspar?). Ore and some epidote are minor constituents. Epidote is also found in the matrix in a few crystals of larger size (d = up to 0.5 mm); possibly these are porphyroblasts. At the bifurcation of the road, to the east of Dansbo, outcrops a very dark subgraywacke. Here, the matrix appears to consist principally of ore. A similar rock has been described also of the lower beds of the Liane layers at the southeastern point of Lilla Ärven (VAN OVEREEM 1948, p. 40).

Slate and conglomerate

In the lower 110 m of the series frequently occur lenses of a dark grey slate, which have a very fine-grained and uniform structure. Owing to their small dimensions the minerals are hard to identify, but it is likely that the main part is made up of sericite and quartz (and feldspar?). These slate lenses form the counterpart to the 70 m of "bituminous slate" which occur more to the north, not far from the base of the Liane series (VAN OVEREEM 1948, p. 40).

In the subgraywacke, at 110 m above the base of the Liane series, occur numerous pebbles of quartzite, ranging up to 5 cm in size and forming a horizon of oligomictic conglomerate of, at most, 10 m. This conglomerate can be traced further to the north all along the Lianefjäll syncline (TÖRNEBOHM 1870a, p. 43 and VAN OVEREEM 1948, p. 42). Upon this conglomerate bed a monotonous series of subgraywacke has accumulated of which still 400 m subsists in the Årbolfjäll.

Basin of sedimentation

The sediments of the Dal formation are comparatively well sorted and the constituent material is generally well rounded. Hence, we may assume a long transport of the material and considerable chemical weathering.

The presence of ripple marks, intraformational breccias, mud-cracks and oligomictic conglomerate are indications of the deposition of the whole series of Dal sediments about sea-level. VAN OVEREEM (1948, p. 105 and p. 45) supposes the arkosic quartzitic sandstone and the Liane layers, from the intercalated conglomerate onward, to be continental deposits. In my opinion this possibility is not excluded but, on the other hand, there is no convincing evidence for such an assumption. The uniform character of deposition in both sets of layers, which attain thicknesses of 350 m and over 500 m respectively, seems to indicate the contrary. Also the thin bed of conglomerate, intercalated in the basal layers, which can be traced over 55 km in a north-south direction, suggests a uniform distribution by a sea.

As has been stated, the basin of sedimentation of the Dal formation is clearly divided in two parts. In middle Dalsland the north-south trending facies boundary passes alongside Åklången (VAN OVEREEM 1948, p. 105—107). In the western part of the basin of sedimentation the lower layers in particular are thicker and the series as a whole shows more petrographical changes. The investigated area falls entirely in the western part of the basin of sedimentation, but the thickness and development of the layers here still show marked differences as compared with the strata lying to the north in the same part of the basin, as may be seen from the table below:

	basin of sedimentation		
	western part		eastern part
	Dals-Rostock	South of Råvarp	
Liane layers: above the conglomerate	> 400 m	> 500 m	> 65 m
Liane layers: up to and including the conglomerate	120 m	175 m	
Quartzite layers	280 m	365 m	275 m
Upper slate layers	200 m	165 m	290 m
Spilite-bearing layers	200-300 m	200-270 m	55 m
Lower slate layers	350 m	240 m	00 m
Basal layers	370 m	220 m	45 m
	> 1920-2020 m	> 1865-1935 m	> 730 m

This table shows that in the Dalskog Dals-Rostock region the older layers are much thicker, the younger layers thinner than in the region south of Råvarp. The latter bear more resemblance to the deposits of the eastern part of the basin of sedimentation, also in petrography. Of the strong variation of rocks in the boundary beds of the quartzite and the Liane layers to the south of Råvarp nothing but a faint reflexion is found in the Dals-Rostock monocline. In the eastern part of the basin it is lacking altogether.

From the paper by HEYBROEK and ZWART (1949, p. 429) it appears that the western part of the basin of sedimentation, at least as far as the lower layers are concerned, extends to the west far beyond Åklången. Not only are the basal layers and the lower slate layers present in great thickness in the Bäcke valley (just as they are near Dingelvik, north of Iväg; see

p. 171), but the same succession of strata is found in the crown of the Bäcke valley overthrust, though in minor thicknesses. These strata must have been deposited at least 10 km more to the west.

In southern Dalsland there is no evidence to suggest that the folding of the Dal formation started at the same time as the effusion of the spilites, as is supposed by LARSSON (1947, p. 328) in connection with his investigations in the Vårvik region. Rather, the constant subsidence of the region still continued for a long time after the extrusion of the spilites.

Along with VAN OVEREEM (1948, p. 109) and SUNDIUS (1947, p. 368) I have not been able to discover any sign of an angular unconformity between the quartzite and the Liane layers, as has been frequently put forward.

CHAPTER V

DAL OROGENY AND LATER TECTONIC MOVEMENTS

The structures of the Dalskog Dals-Rostock area

Since comparatively large tracts of the mapped area are covered by quaternary deposits, a map of solid formations has also been prepared on which the structural features stand out more clearly. This necessarily involved interpolations and interpretations regarding those parts which are not open to direct observation. In some instances, an approximate solution was all that could be arrived at. In drawing the cross-sections, no discrimination has been made between interpretations and directly observed facts; only occasional references to this point will be found in the descriptive text of the tectonics. The 1:15.000 map gives as true a picture as possible of the observed facts. Faults and overthrusts have only been drawn in full where they have been seen in the field and where they appear directly from the succession of strata. Everywhere else, as they become apparent from the construction of the sections or must be assumed for some other reason, they have been represented by a dashed line. Furthermore, all measured dips and strikes have been recorded on this map. The reader will therefore be able by simple comparison of the two maps to distinguish interpretation from actual observation.

In order to elucidate the position of the great horizontal overthrust in the northwest of the region and the tectonics of the Teåkersjö anticline a part of the map by P. Heybroek and H. J. Zwart (1949, p. 426) has been reproduced on the map of solid formations and incorporated in the cross-sections.

A. Sandell (1941, p. 163—188), in his study of the tectonics and morphology of Dalsland, paid special attention to the tectonics of the area under discussion. A comparison of his tectonical map (p. 168, Pl. I) with the present one, shows the profound difference which exists between Sandell's views and my own. In the following treatment of the tectonics I therefore refrained from confronting my results in detail with those of Sandell. A more general discussion of Sandell's conceptions will be reserved for a later chapter on the morphology. Neither did I find any connection with Hausen's tectonical interpretation of the Dal folding (1931), therefore I shall not make further reference to it.

In a discussion of the tectonics, the region divides itself in a natural way into a number of structures, which, though not disconnected, constitute cartographically separate units. As such we may distinguish:

First a number, of north—south trending anticlines and synclines with a pitch of the axis towards the north. From west to east we find in succession:

1. The Teåkersjö anticline
2. The Prästbol syncline
3. The four anticlines between Dalskog and Dansbo,

separated by narrow synclines.

Moreover, to the northwest a part of the great horizontal overthrust is exposed, hence:

4. The Lysesjö overthrust.

The eastern flank of the easternmost anticline is formed by

5. The Dals-Rostock monocline,

exposing a complete section through the sedimentary series of the Dal formation, which may be followed from north to south over the map.

In the southwest we have:

6. The region between the Kroppefjäll plateau and the Dals-Rostock monocline,

where the Dal formation has been almost completely eroded, as a result of which there is little known about its structure.

The Dals-Rostock monocline is bounded to the east by:

7. The Kroppefjäll fault and the upthrown block of the Dalbo plain.

Finally we will discuss:

8. The renewed movement along the Kroppefjäll fault,

which has broken up the pre-Cambrian peneplain.

The Teåkersjö anticline

The Teåkersjö anticline (sections II—VI) has developed almost entirely in the arkosic quartzitic sandstone of the Dal formation. The transgression plane of this formation is partly exposed near the south shore of Teåkersjön and forms the southernmost vestiges of the gently arched anticline. North of Teåkersjön, the bend of the anticline is indicated by the trend of the slate bed near the farmhouse of Dalen (P. Heybroek and H. J. Zwart 1949, p. 430). This proves clearly the plunge of the axis towards the north; moreover we may assume that the bottom of Teåkersjön consists for the greater part of arkosic quartzitic sandstone.

The western flank of this anticline is bounded by a fault running from the south point of Damtjärn, past Teåker, along the entire west shore of Teåkersjön. The western wall of the fault is formed by an upthrown block of the basement. At the southwest point of Teåkersjön is an outcrop of the lower beds of the basal layers, dipping gently to the north and butting against this fault. The dip of the fault is towards the west, as is shown by the narrow belt of quartzitic sandstone which is deflected north of Teåker and assumes a steeply overturned position to the west (sections II and III). P. Heybroek and H. J. Zwart (1949, p. 427, sections I and V) did not notice this phenomenon and supposed a normal position of the fault. The stratigraphical throw of the fault in its northern part may be estimated at some 250 to 300 m. This is shown by the sections of P. Heybroek and H. J. Zwart (p. 427, section I), and by the small outcrops of quartzitic sandstone on the Kroppefjäll plateau west of Teåker (sheet Upperud 1870) which show that the base of the Dal formation here was situated not far above the present surface.

In the northeastern flank of the anticline occurs a fault, dipping slowly to the northeast, which cuts out the slate bed (P. HEYBROEK and H. J. ZWART 1949, p. 431). Near Stora Halängen, the slate reappears in several secondary north—south trending folds, the flank-fault, just mentioned, dying away rapidly. Further southwards, south of the east—west directed fault passing by Idala and Södra Halängen, the flank of the anticline is also broken. I will refer to this again presently. More detailed information about this slightly arched anticline is hidden on the bottom of Teåkersjön. One may assume however, with SANDELL (1941, p. 190), that the east—west striking fault via Idala and Södra Halängen is continued below the lake and is possibly connected with an east—west trending fault, which one may suppose to exist west of Teåkersjön by the presence of the markedly east—west trending valley through the Kroppefjäll plateau, along the lakes L. and St. Yxesjön and N. Damtjärn. The existence of such a fault might also be suggested by the geological map (Upperud sheet, 1870).

It seems logical to assume a similar slightly curved fault of an east—west direction south of Teåkersjön in the typical narrow valley along the lakes St. Örlevattnet and L. Råvattnet (SANDELL, 1941, p. 190). Geological evidence in support of this assumption has not been found in the studied area. Moreover, it is not likely to be easily found in this gneiss-granite region.

The Prästbol syncline

The Prästbol syncline (sections I—VIII) limits, to the east and northeast, the Teåkersjö anticline and is divided by the east—west striking fault near Idala into two parts having different structures.

The northern part is a flat syncline (sections I—IV) in which mainly the lower slate layers are exposed. At the west side it passes, via the little secondary folds mentioned above, into the east flank of the Teåkersjö anticline. These small secondary folds extend for some distance towards the centre of the syncline, as is shown in the railroad cutting, due north of the east—west trending fault past Idala, by the sandy bed which has developed here near the base of the calcareous slate beds. TÖRNEBOHM (1882—1883, p. 650, fig. 17) gives a sketch of these small secondary folds in the northern wall of the railroad cutting, with which I fully agree. Its east flank, which has been disturbed by faults, can be followed up to the northern boundary of the map near N. Båsane. North of Bergatjärn the centre of it is covered entirely by the Lysesjö thrust mass, which will be discussed presently. In this connection it must be noted that the inliers in this overthrust prove the underlying strata to consist everywhere of the calcareous slate beds (P. HEYBROEK and H. J. ZWART 1949, p. 432), thus the covered part must also be a flat syncline.

In the southern part of the Prästbol syncline, a number of blocks have moved along essentially north—south striking faults, thus we may speak of a fault trough rather than of a syncline. Apparently this part has been subjected to a greater compression. It has been thrown down along the east—west trending fault of Idala and has been displaced towards the west, as a result of which higher beds of the lower slate layers are exposed south of the fault.

Again this part is subdivided in two by the slightly bent fault striking east—west, which occurs northwest of Stammen, since on either side of this line the movements of the blocks in the western flank have been somewhat different. In the southern part, which for the most part coincides with the Källsviken peninsula and can be traced further to the south beyond Kappebosjön, the effects of the compression are clearly exhibited.

At the west side of the peninsula there is an outcrop of a reverse fault in the arkosic quartzitic sandstone along the road north of Källsviken (sections VI and VII). On the fault plane, one finds a thin band of slate of 2 or 3 m in thickness, which must have acted as a lubricant. This band of slate has a perpendicular position west of the slate-quarry north of Källsviken, but at Källsviken itself, the dip is about 60° to the west. Simultaneously with this change of dip, the beds of quartzitic sandstone on the west side of the fault shift from the normal eastward dip to an overturned position. Most probably the continuation to the south of this reverse fault is to be found in the valley down to Rotjärn, cut out in the Kroppefjäll plateau. The proof of a fault, traversing this valley, is given by a small outcrop of arkosic quartzitic sandstone in the southeast wall of the valley. The partly overturned block is bounded to the west by a fault which passes along the shore of Teåkersjön and forms the limit between the Teåkersjö anticline and the fault trough-like syncline of Prästbol. The dip of this fault is unknown.

In the deepest part of the fault trough, to the east of the structure discussed above, a number of small secondary folds have developed which are clearly exposed in the large slate-quarry north of Källsviken (section VI), with the quartzitic bank in the slate forming a profitable key bed.

These folds are accompanied by a series of minor faults which have an inclination to the west, except for the most eastern one which has probably a perpendicular position. From the east point of the slate-quarry two minor reverse faults run in southeastern direction towards Kappebosjön. Here too, slate has been found covering the fault planes. The inclination of the small faults is 35° to the northeast.

This central part of the fault trough is limited to the east by the block on the upthrown side of the fault (sections VI and VII). Of the inclination and the amount of throw, nothing can be said with certainty. With some probability, the latter may be evaluated at 250 m. It is quite conceivable that this fault, the vertical displacement of which is at any rate considerable, continues to the south beyond Kappebosjön and then in the notch on the edge of the Kroppefjäll plateau just west of Öjerud. In that case, it would at the same time mark the eastern limit of the basal layers of the Prästbol syncline on the south shore of Kappebosjön.

As stated above, the part of the syncline north of the east—west trending fault to the northwest of Stammen has a somewhat different development. Although here too, block faulting has taken place by stronger compression, it proved impossible to retrace these movements in detail. To the south of Södra Halängen, in the west flank of the syncline, runs a strike fault along which an eastern block has slipped down and as a result the basal layers, with an eastward inclination, have come into contact with the calcareous slate beds dipping to the north. As has been said, this fault can be followed along the east shore of Teåkersjön. Its inclination is not known. At some distance to the south, as opposed to what we have just seen, an eastern block has been upthrown, causing the quartzitic sandstone beds to be exposed at the surface also on the east side of the fault. This block is parted by a north—south trending band of slate, both the slate and the quartzitic sandstone dipping to the east. The question, how all this has been brought about, must be left undecided.

In this part, too, the east side of the fault trough is bounded by faults which cause older rocks to be exposed in two steps. Structurally the eastern flank is the continuation of the conditions found southwards, to the east of Källsviken.

The four anticlines between Dalskog and Dansbo

Four anticlines, separated by narrow synclines, follow in eastern direction the syncline of Prästbol. Near the northern boundary of the region only a single broad anticlinal structure has developed, which will be called the S. Båsane anticline. Both in its eastern and its western flank, this anticline shows faults along which the centre part has been raised. The inclination of these flank-faults is not known, but they are considered here as normal faults. For the easternmost fault this is borne out by the fact that the strata to the east give a strong impression of having lagged behind in the upheaval of the anticline and also by the apparent eastward bend of the fault south of Dansbo. This easternmost flank-fault is the most important and its throw is increasing rapidly from north to south, so that the entire lower slate layers and the greatest part of the spilite-bearing layers are soon cut out. A second fault occurring east of the flank-fault and joining it near Bockhålstjärn reëstablishes the contact between the lower spilite-bearing strata and the quartzitic sandstone to the south of this lakelet.

The fault in the western flank is of less importance. In the north, it starts near Landsbol and towards the south cuts out the slate bed gradually. This is accompanied by a crumpling of the bed. More to the south, the position of this fault is not clear, since quaternary deposits and the great thrust mass in the west cover considerable areas of the surrounding rocks. Northwest of Dalskog, the fault passes into a fault zone in which several blocks have been involved.

In the western flank of the S. Båsane anticline two anticlines have originated which become visible in the vicinity of Dalskog. Particularly the western one, the Gunnesbyn anticline, is hidden from the eye by quaternary deposits to such an extent that we possess only the scantiest information about this structure. A belt of quartzitic sandstone dipping to the west, overlying the basement with normal contact south of Tonebyn, is the only evidence of the western flank. This flank is bordered to the west by a fault passing along Tonebyn and the Dalskog road. The western block of quartzitic sandstone, strongly veined with quartz and irregularly dipping to the north, is thus brought into contact along the road with the centre of the anticline consisting of gneiss-granite. Likewise, the eastern flank of the anticline is only indicated by small remnants of quartzitic sandstone lying on the core of basement rock east of Gunnesbyn and beyond Kappebosjön near Rörnäs. To the west of Korsgården on the edge of the terrace there is a small outcrop of conglomerate which I am inclined to class among the intercalated conglomerate. If this be true, we may infer to have found the beginning of the anticlinal bend of the intercalated conglomerate, which must have been folded far south into the adjoining syncline.

A second anticline, the Kabbo anticline, which has arisen in the western flank of the S. Båsane anticline, is somewhat better exposed. The anticlinal bend of the intercalated conglomerate can be traced in a series of small outcrops east of Dalskog, the same bed being a reliable index horizon to determinate the eastern flank with more precision. The western flank is outlined by remnants of quartzitic sandstone with a western dip south of Korsgården. It appears from this that the valley of Korsgården is a narrow syncline situated between the Gunnesbyn and Kabbo anticlines. A similar narrow syncline is also found east of the Kabbo anticline. Owing to the small outcrop of intercalated conglomerate north of Kabbo we obtain an understanding of the structure of this strongly squeezed syncline. The remarkably irregular boundary between the quartzitic sandstone and the basement, especially in this anticline, must be attributed to the rugged surface on which the Dal formation was deposited (see p. 140).

The way in which these two anticlines have originated from the western flank of the S. Båsane anticline cannot be ascertained, since in the neighbourhood of Dalskog large areas are covered by quaternary deposits. However, the origin of the anticline that rises in the eastern flank of the S. Båsane anticline due south of Myrevarv is shown very clearly. Towards the south, this anticline is increasing rapidly in significance, the core being formed here by the Kappebofjäll. At the uplift of the Kappebofjäll anticline the flank-fault of the S. Båsane anticline was bent to the east and consequently has a deflection between Myrevarv and Bockhålstjärn. This indicates that the Kappebofjäll is younger than the flank-fault of the S. Båsane anticline. At the same time it becomes apparent that the eastern flank of the latter has been taken over by the Kappebofjäll anticline. The steepening of the strata and the possibility of renewed motion along the flank-fault gave ample scope for its upheaval. Furthermore the younger age of the Kappebofjäll anticline as compared to the S. Båsane anticline is shown by their delimitation in the valley of Tångebo. The narrow syncline, separating these two anticlines in the north, soon passes towards the south into a fault that cuts obliquely through the anticline of S. Båsane and also further south through the narrow syncline bordering it to the west. South of this fault, the S. Båsane anticline has disappeared completely. The fault is not exposed here but becomes apparent, when the cross-sections and the map are constructed. To the south it passes across the east arm of Kappebosjön, on the south shore of which, a zone of mylonitization is found dipping to the west. On this account a westward dip of the fault in its northern part has been assumed.

We are able to understand the structure of these four anticlines by assuming that the three flank-anticlines are of a later origin than the S. Båsane anticline, as we found already to be true in the case of the Kappebofjäll anticline. The anticline of S. Båsane thus splits up in the south into three anticlines, each with a separate northward pitching end, whereas the pitching end of the S. Båsane anticline itself, situated more to the north, encloses the others. This anticline is not continued towards the south. The space formerly occupied by it is divided between the three more recent structures, the Kappebofjäll anticline getting the largest share. Only in this way can it be seen that the core of the S. Båsane anticline terminates abruptly against the fault passing through the valley of Tångebo. Likewise the cylindrical shape, assumed by the core of the anticline soon after its beginning, exhibited by the trend of the intercalated conglomerate, must be ascribed to compression between the flank-anticlines.

The Lysesjö overthrust

North of Teåkersjön and north of the Bäcke valley P. HEYBROEK and H. J. ZWART (1949) found the remnants of large horizontal thrust masses of gneiss-granite upon the sediments of the Dal formation. The broader aspects of this phenomenon will be dealt with later (see p. 168). It must be mentioned here, however, that the continuation of this overthrust, termed Lysesjö overthrust by the authors in question, outcrops in the area under discussion. The line of outcrop of the thrust plane enters the map near the south shore of Bergatjärn, then passes with a sharp turn along the east shore and can be followed from there in a north-northeasterly direction to the northern margin of the map near N. Båsane. The thrust mass is formed by gneiss-granite and quartzitic sandstone and is overlying autochthonous sediments of the Dal formation, viz. almost exclusively the lower slate layers of the Prästbol syncline. An outlier of this thrust mass is found as a thin slab of granite lying on the lower slate layers west of Prästbol.

The edge of the thrust mass has been broken by two west—east trending faults, which lowered the thrust plane in three steps from south to north. Coincident with the upper limit of the map near Damängen there is a third west—east striking fault. North of it, the thrust plane has been raised again. Possibly, renewed movements along the east—west trending fault passing near Södra Halängen and Idala, have taken place after the overthrusting, and the outlier near Prästbol, lying about 20 m below the thrust plane near Bergatjärn, may be regarded as a sunken part of the thrust mass. Likewise, the somewhat crooked trend of the east—west trending fault may be related to renewed movements along the fault plane.

A belt of quartzitic sandstone stretching from Bergatjärn to N. Båsane has been overthrust with the gneiss-granite. That it belongs to the thrust mass, follows from its normal contact with the gneiss-granite which is often accompanied by lenses of basal conglomerate.

The thrust plane, almost horizontal near Prästbol and around Bergatjärn, has an inclination of about 20° to the west near Högebotjärn and about 60° to the west near Landsbol. Concurrently the quartzitic sandstone in the thrust mass increases. At the eastern shore of Bergatjärn we find only a thin bed of quartzitic sandstone under the granite and randomly implicated with it. Near Landsbol and N. Båsane, on the contrary, we have probably a complete series plunging under the granite. Remnants of the overlying slate have namely been found, in highly moulded state, on the thrust plane. The steep position of the thrust plane and the thick beds of quartzitic sandstone plunging under the gneiss-granite indicate that we have here the crown of the thrust mass. The lowering of the thrust plane from south to north, conditioned by the faults, thus provides in this direction sections at succeedingly higher levels through the crown of the thrust mass. This crown proves once more (P. HEYBROEK and H. J. ZWART 1949, p. 432) that the Lysesjö overthrust must have come from the west and also, that it never extended farther to the east over the sediments of the Dal formation.

In a steep scarp southeast of Högebotjärn a second thrust plane outcrops along which calcareous slate has been thrust upon the quartzitic sandstone, the thrust plane bearing a thin layer of crumpled slate. This is probably a detached patch of calcareous slate which has been pushed for some distance ahead of the crown of the thrust mass. Southeast of Bergatjärn, where this overthrust calcareous slate lies upon autochthonous calcareous slate, it becomes difficult to locate the thrust plane. Here too, however, its course can be traced by a few outcrops in the zone of mylonitization. It is quite possible that there is such an advanced mass of calcareous slate east of the outlier near Prästbol, as the gneiss-granite is bounded to the east and northeast by hills of calcareous slate which have a greater elevation than the gneiss-granite, whereas in particular on the northwestern side of the outlier, outcrops are found in which the superposition of the gneiss-granite upon the calcareous slate is clear.

As shown on the map, the outlier is divided into three parts by two north—south trending bands of calcareous slate. The western band was generated by the erosion of the very thin cover of gneiss-granite, as a result of which the calcareous slate came to the surface. In the eastern band circumstances are different as the calcareous slate has been encountered in lateral contact with the gneiss-granite on the eastern side. Consequently, faulting must have occurred here after the overthrusting.

The Dals-Rostock monocline

The eastern flank of the central series of anticlines consists of a large monocline dipping to the east, in which the complete set of strata of the Dal formation is exposed. This monocline had been intensively faulted and imbricated in the north. For a better understanding of the sequence of events here, we must consider the area in its broader aspects (see fig. 22 and sheet Upperud, 1870).

From this it appears that the anticline of S. Båsane near the farm of this name, rises rather steeply from the western flank of the Lianefjäll syncline and is for the first time clearly seen by the anticlinal bend in the quartzitic sandstone. It is surprising to note how little influence this anticlinal arching has had on the strata to the east of this bend. The whole movement of this upwarping seems to have been absorbed by the eastern flank-fault. The complete slate series (i. e. the lower slate layers, the spilite-bearing layers, the upper slate layers) and the quartzite layers can be followed almost in a straight line to the south from Stora Ärven where they form the western flank of the undisturbed Lianefjäll syncline, to where the lower slate layers and the spilite-bearing layers end bluntly against the flank-fault near Myrevarv and Bockhålstjärn. The upper slate layers and the quartzite layers terminate abruptly against another fault running from the southeast shore of Stora Ärven in a southern direction and joining the flank-fault south of Bockhålstjärn.

Now the curious thing is that the Liane series, situated more to the east between Stora Ärven and Näsöln, is bent as though to make room for the emerging anticline. This difference in behaviour between the neighbouring unbent strata and the more remote bent strata resulted in a gap, which is filled up by a roughly wedge-shaped block, the western side of which has been tilted along faults, causing a repetition of the spilite-bearing layers, the lower slate layers and the quartzite layers. The fault, bounding it to the west, runs from Stora Ärven to south of Bockhålstjärn. In the southwest its delimitation is formed by a system of faults which runs approximately from Tegstjärn to Sjögoanäs on the Näsöln beach. These statements do not imply that the wedge has come into existence as an effect of the bend of the Liane series. Presently, I will try to state the genetic relations between the structures described.

The western part of the wedge is rather complicated. Especially around Tegstjärn there have been movements of different blocks causing, among other things, the spilite-bearing layers to come three times to the surface and the jamming of a quartzite wedge between older strata. The bend of the Liane layers is partly neutralized by its shifting to the west along the fault zone of Sjögoanäs. South of this zone of dislocation the normal succession of strata in the monocline has been restored as far as the flank-fault of the S. Båsane and the Kappebo anticlines.

The inclination is not known of any of the faults in this highly faulted area around Tegstjärn. It does not seem possible, therefore, to grasp the meaning of all movements along these faults. Roughly speaking, the order of events must have been as follows: The anticline which bordered the syncline of the Lianefjäll on the western side has been raised by further compression between two flank-faults, thus forming the S. Båsane anticline, the strata of the eastern flank lagging behind. The lag of these strata, however, has been compensated by a repetition of the strata to the east of a second fault running from the southeast shore of Stora Ärven to south of Bockhålstjärn. The throw of this fault decreases to the north and dies out somewhere along the southeast shore of Stora Ärven. Thus the Liane layers have not been affected by it and follow the bend without a fault. Afterwards the anticline of the Kappebofjäll arose on the flank of the S. Båsane anticline which, near Dansbo and north of it, has been neutralized by a steepening of the strata on the flank and possibly by renewed movements along the flank-fault. In the region around Tegstjärn, however, where the anticline of the Kappebofjäll emerges, we find a more or less sharp transition from an undisturbed position of the flank in the north to an eastward shifted flank in the south. It looks probable that the fault zone between Tegstjärn and Sjögoanäs absorbed the movements at this transition point.

Towards the south, near Dansbo, the flank-fault bends to the east causing the lower slate layers to outcrop west of the fault. From there the fault breaks through the spilite-bearing layers and the upper slate layers, its throw being rapidly diminished, to disappear completely south of Gulkanerud. Its function is taken over partially by a north—south striking fault passing west of Årbol along Gulkanerud, which cuts out part of the upper slate layers and the quartzite layers. The only field evidence for this fault is found in its southern extremity in the shape of crumpled layers; for the rest it is covered by quaternary deposits and appears only from the construction of the cross-sections.

East of Dansbo an east—west trending fault arises, which is readily observed east of the road junction. South of this fault the upper layers of the Dal formation are shifted a little to the west and at the same time the strike of the Liane layers changes from north—south to north—northwest—south-southeast. By so doing, it gives way to the underlying Dal formation, which here is almost complete owing to the dying out of the flank-fault. The two minor faults crossing the basal layers obliquely on either side of Hultet are probably due to a difference of compression at the uplift of the Kappebofjäll anticline, which further to the east is absorbed by the flank-fault. On this account they have been given a westward inclination in the sections.

The monocline, which is practically undisturbed between Skogtjärn and Råbäck, shows further to the south many disturbances. These are related to an overturning of the sedimentary series by which the normal eastward dip is inverted to a westward dip. This inversion takes place in a zone between N. Strutsåstjärn and Dals-Rostock. It is important to note that the overturning of the sediments goes on without major transverse faults, the layers assuming a south-southwest strike thus deviating to the west. With the same strike and dip the basal strata of the series proceed for 20 km in a southwesterly direction to Bollungssjön (sheet Rådanefors) as a narrow strip of Dal formation in its basement. It has to be pointed out here that the important Kroppefjäll fault, which traverses the monocline in the Dals-Rostock area at a sharp angle so that, near Örsjön, successively all the layers except the basal layers are cut out, is not to be considered as the cause of the over-

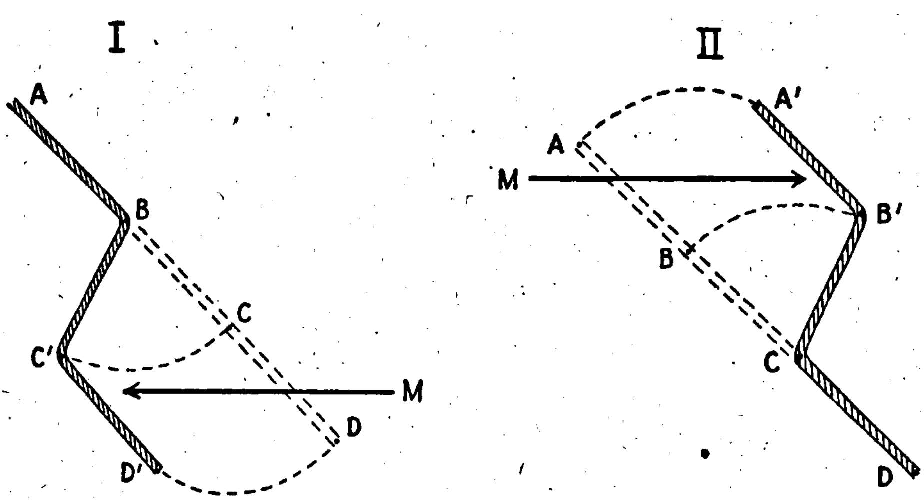

Fig. 20.

The deformation of a stratum AD in case of underthrusting (I) and overthrusting (II). Section.

turning, but must have originated at a latter date. This fault will be treated in detail presently and need not be discussed in the following analysis of the overthrow.

The mode of occurrence of the overthrow gives us a clue as to the direction of the movement, a rare advantage in a tectonical analysis. In other words, we are able to distinguish between an active and a passive block, viz. an east-southeast active block moved towards a west-northwest passive block, the layers of the monocline in between being overturned, which overturning is to be regarded as the initial stage of an underthrust.

This statement may be made clear by means of a schematic section (fig. 20).

A given stratum AD uniformly dipping to the east is moved into an overturned position over a distance BC under the impact of an active block. Whether the active block is situated to the east and the relative movement (M) was directed westward (case I, i.e. underthrusting) or is situated to the west with a relative movement (M) directed to the east (case II, overthrusting) cannot be decided from the resulting shape of the strata. However, there is a difference of position with respect to the original position in both cases.

In the case of underthrusting (I) the upper part of the stratum (AB) has kept its original position, whereas in the case of overthrusting (II) the upper portion is transferred to A′B′. The inverse holds in both cases for the lower portions of the stratum (CD). If it can be demonstrated that the part AB of the stratum either has persisted in its original position or has been moved eastward, then we are able to conclude the existence of an underthrusting or an overthrusting respectively. The inverse is true for the section CD of the stratum. A similar trend of thought has been put forward by T. S. LOVERING (1932, p. 651).

In the overturned sediments near Dals-Rostock we are confronted with the upper bend of the overthrow (point B of figure 20). There is no sign of an eastward shifting of the sediments above this bend. In view of the fact that the pitch of the axis in this part of the Dal formation has a general direction to the north, (see also TÖRNEBOHM 1870a, p. 45, fig. 11) such an eastward shifting of the sediments should be expected north of the overthrow near Dals-Rostock, conveyed for instance, by a number of east—west trending faults or by a gradual deflection of the sediments to the east, the most easterly position being just there, where the overthrow occurs. Nothing of the kind has been recorded. Though dissected by faults the general direction of the monocline does not depart from the north—south direction of the Dal formation, until the strata are overturned and bend to the south-southwest. Thus, the layers have been dragged to the west by the overthrow and we may infer an underthrusting.

If we assume that the strike of the Dal folding in southern Dalsland had also originally been directed north—south, then it follows that the amount of underthrusting near Bollungssjön has to be evaluated at 10 km. The overturned strata, therefore, must probably have been broken over a large distance along an underthrust fault. However, this could not be demonstrated in the studied area with certainty, though there is every reason to suspect the fault, which has formed a typical mylonite in the lower slate layers north of Örsjön, to be an underthrust fault. On this account the fault has been drawn in the sections with a slight dip to the west. Its continuation to the south is likely to be truncated in Örsjön by the Kroppefjäll fault and covered laterally by the eastern fault block.

The behaviour of the basal layers in the overturning is to a high degree independent of the other layers. Between Skogstjärn and S. Strutsåstjärn a structure has developed whih may be best compared to an anticlinal nose, causing a slight bend of the intercalated conglomerate to the east. The nose structure is even better exposed by the trend of the boundary between the basal conglomerate and the overlying quartzitic sandstone. The basal conglomerate is massed south of Helvetestjärn in a depression of the transgression plane. This, together with the considerable differences of topographical elevation, is responsible for the aberrant course of the line of outcrop of this plane here and has, therefore, no structural significance.

The nose is bounded on its eastern and southeastern side by an S-shaped fault. Further to the east the basal layers assume an increasingly steep position in a direction from north to south, then, after a small transverse fault, passes over a little distance into a low overturned dip (25°) to the north-northwest and finally with a sharp turn, comes to a steeper inclination (45°) to the west-northwest. The existence of the nose and the low dip of the overturned layers just below, seems to be the result of the overturning. In the process of overturning the strata pass through a position in which a deficit of space results in the vertical direction (maximum in the perpendicular position), at least when a shortening and widening of the layers does not occur by internal movements within the stratum. As for the hard quartzitic sandstone, the internal movements must have been slight, causing the inflexion point to be displaced upwards. It looks very much as though the S-shaped fault northeast of S. Strutsåstjärn has neutralized part of the movement of the sharp overthrow, as is shown by the strongly brecciated rocks of the Kappebo series on this spot. The final displacement along this fault, however, is small. To what extent this fault forms the boundary between the Dal and the Kappebo formations in a southerly direction, cannot be ascertained, owing to the small displacement. For morphological reasons it has been drawn on the map and sections as far as Sketjärn.

It is surprising that no trace of the above-mentioned structures is to be found in the strata lying more to the east, but there is a gradually progressing overturning to the south. This must be related to the fault, which from Hjulsängen to south of Långtjärn, forms the boundary between the lower slate layers and the basal layers, and cuts out a portion of the latter and either equalized the differences in movement on both sides or brought different forms into lateral contact by subsequent vertical movements. The inclination of this fault, which reveals itself in the field by crumplings of the slate beds, could not be determined. South of Långtjärn, it is probably continued by a fault cutting obliquely through the basal layers in the direction of Sketjärn. A. E. Törnebohm (1870a, p. 32, fig. 3) gives a section through the southern extremity of Långtjärn, in which the slate beds show small folds. On the spot I found only irregular crumplings of the quartzitic sandstone and the slate beds. From Hjulsängen to the south, the belt of the lower slate layers becomes gradually broader until beyond the overthrow, where it is decreasing rapidly in width. The contrary is true for the spilite-bearing layers. Whether this is due to a primary difference of thickness of the layers, or comes from tectonic causes, cannot be ascertained. The dip measurements in the very incompetent calcareous slate should be treated with caution. This circumstance and the dense vegetation of the region render it very difficult to detect faults. A slight indication of faulting might be found in the spilite, west of Dals-Rostock station, dipping at low angles to the west, and bearing traces of tectonic movements. The extent and the significance of these movements, however, could not be determined. Since the lower slate layers and spilite-bearing layers together, from Råbäck to the overthrow, may be represented without trouble in uniform thickness in the sections, it is likely that the primary difference of thickness is the cause of the peculiar trend of the boundary between the two layers.

The same question may be put about the strongly widening belt of spilite between Dals-Rostock station and Mörtjärn. Here, however, there are indications of a tectonic origin, for near Dals-Rostock station the upper slate layers dip to the east in small folds, whereas, west of this spot, a western dip prevails. The bending of the sandy intercalations in the spilite north of the Kroppefjäll sanatorium also seems to be due to tectonic causes. The structural representation of this area, as laid down in the sections, is based on the assumption that the fault in the overturned lower slate layers north of Örsjön is a thrust fault, the foot wall having been pushed forward to the west. It has been assumed, furthermore, that this feature is continued to Dals-Rostock station and that, by the westward movement of the foot wall, the hanging wall has been lifted up, driving the spilite-bearing layers and the upper slate layers partly into a horizontal position. A fault trends from Taperud to the south-southwest, with the upthrown block on the east side.

The region between the Kroppefjäll plateau and the Dals-Rostock monocline

In the southeast corner of the mapped area the vestiges of the Dal folding have been blurred by erosion. On the Kroppefjäll plateau no traces

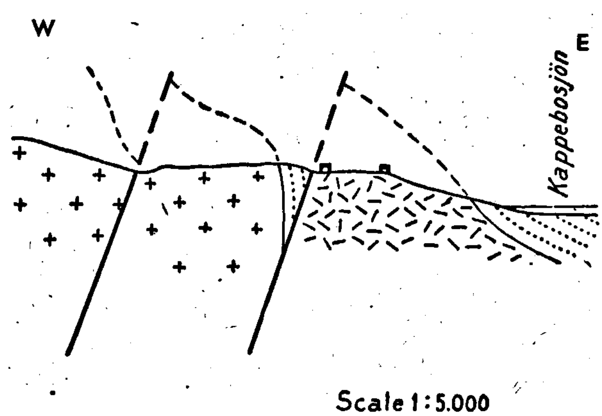

Fig. 21.

East—west section 450 m south of Rörnäs.
For explanation of the symbols see section sheet.

have been left, but between this plateau and the Dals-Rostock monocline only small outcrops of the basal beds of the Dal formation as well as some faults and fault indications, which can be followed across and along the Kappebo formation, give a reflection of the more north- and eastward situated structures.

The eastern flank of the Gunnesbyn anticline can be traced across Kappebosjön, past Rörnäs to near Högesjön, a small strip of quartzitic sandstone and basal conglomerate serving as a guide. South of Rörnäs it has been raised in two steps, going from west to east, along two north-northeast—south-southwest striking faults. The easternmost fault can be traced, across Högesjön past Blixerud and Smedtjärn to Abborrtjärn, by the presence of fault breccias and abnormal contacts between the Kappebo formation and the gneiss-granite.

This fault must have existed before the period of denudation preceding the deposition of the Dal formation, while renewed movements must have attended the Dal orogeny, for on the west side, the basal layers (fig. 21).

with a basal conglomerate, overlie the gneiss-granite in normal superposition, though the beds are steeply inclined or overturned by subsequent movements. East of the fault, in the raised block, rocks of the Kappebo formation outcrop, overlain in normal superposition by the lower beds of the basal layers, a little more to the east. Relative to the Kappebo formation the eastern fault block is a down-thrown block, so the initial movement of the fault must have been reverse to and greater than the Dal movement.

A remnant of arkosic quartzitic sandstone lying along the west shore of the eastern arm of Kappebosjön, is the southward continuation of the eastern flank of the Kabbo anticline. The basal layers of the Dal formation, near the farmhouse of Rörnäsön and to the south-southwest of it, may have been placed there by a fault occurring to the east. However, further indications for this supposition are lacking. It is assumed, therefore, that the position of these outcrops has been determined by primary unevennessess of the transgression plane of the Dal formation.

Through the eastern arm of Kappebosjön runs the fault against which, more to the north, the S. Båsane anticline terminates abruptly. Between Kappebosjön and Lillesjön this fault is indicated by a friction mylonite which has an inclination of 57° to the west, whereas south of Lillesjön it is exposed as a regular thrust fault with a low dip (20°) to the west. Above the thrust plane, we find essentially Kroppefjäll granite; below, there is the same granite and rocks of the Kappebo series. On the thrust plane itself, crushed rocks of the Kappebo formation have been found, even there where gneiss-granite covers gneiss-granite.

The two faults, just mentioned, bend somewhat to the southwest off Högesjön. The location and position of this bend indicate a direct relation with the overthrow and underthrust which we encountered south of Dals-Rostock. Of the eastern fault, we know that after the bend, its inclination to the west flattens; so here again, we may infer an underthrusting from the east. It appears from this that the shortening attending the underthrust movement from the east, not only took place in the zone between Dals-Rostock and Bollungen but equally, in the adjoining western zone. In this connection it may be pointed out that west of Fjällsäter the gneiss-granite overlies the Kappebo formation with a normal contact, which must be ascribed to overturning, whereas on the opposite side of the fault, near Blixerud, an eastern dip was observed in an intercalation of slate in the Kappebo series. Here, in consequence, the Kappebo sediments cover the gneiss-granite in normal superposition.

Besides these two major faults, a number of minor ones are drawn on the map. These have been traced with some degree of probability on the evidence of friction breccias and morphological features. By drawing one of these faults, I have connected the brecciated rocks occurring on the east shore of Smedtjärn with the long narrow valley between Abborrtjärn and Svartetjärn, which here separates the Kappebo formation from the Kroppefjäll granite. Perhaps all the faults in this region have passed into thrust faults. However, field evidence confirming this point is lacking. Another fault must have separated the little granite mass at the west shore of Lillesjön from the rocks of the Kappebo formation and further south must have caused a dyke-shaped furrow in the upthrown granite mass, while a third fault at the southeast shore of Lillesjön has raised the gneiss-granite. More to the south this third fault can be followed as a trail of brecciated rocks. The conspicuously straight valley, invading the Kappebofjäll from Åsen to N. Strutsåstjärn, which is continued with an S-shaped bend through this lakelet into South and North Damtjärn, is probably connected with a fault, especially as brecciated rocks are found along this valley, particularly to the west of Åsen. The S-shaped bend of N. Strutsåstjärn is clearly reflecting the structure to the east of it.

As marked on the map, near Åsen, the Kroppefjäll gneiss-granite is exposed between the overturned Dal formation in the east and the Kappebo formation in the west. The gneiss-granite is abundantly veined by rhyolites and greenstones of Kappebo age. If it is imagined that the horizontal position of the Dal transgression plane is restored, it looks as if the Kappebo formation were situated below the gneiss-granite at the time of deposition of the Dal formation. This indicates a complicated fold, or a fault prior to the period of denudation preceding the deposition of the Dal formation. Probably we have a fault here. In the field, this ancient fault cannot be recognized. This is due to the fact that in a western direction the veins of rhyolites and greenstones become increasingly abundant, even leading to the complete elimination of the gneiss-granite through which a belt exclusively greenstone and rhyolite (outflow or dyke?) results, separating the gneiss-granite from an area where also the clastic rocks of the Kappebo formation occur. The fault, as drawn on the map, indicates the limit beyond which no gneiss-granite has been found. The great number of dykes and outflows in and about the gneiss-granite give reason to think that the fault originated during the deposition of the Kappebo formation and acted as a conduit for the effusives.

The Kroppefjäll fault and the upthrown block of the Dalbo plain

The Dals-Rostock monocline is bounded to the east by the Kroppefjäll fault which passes across Näsöln, via Dals-Rostock to Örsjön and cuts the monocline obliquely. The eastern block of the Dalbo plain has been raised with a considerable throw along this fault.

The Kroppefjäll fault can be followed outside the mapped area over a very great distance viz. from Råvarpen in the north to the vicinity of Munkedal in the south. We will return to this later (p. 173).

In the neighbourhood of Näsöln, it is no longer a simple fault but rather a fault zone with three more or less parallel step faults. The western one passes along the west shore of Näsöln. Several spits projecting from this shore belong to the raised block. On such a spit, south of Sjögoanäs, outcrops have been found of the spilite-bearing layers, which are brought by the fault into lateral contact with rocks of the Liane layers. On the Lövnäs spit, just north of the mapped area, quartzite outcrops with a steep eastern dip (sheet Upperud 1870). The fault cuts off the fault zone between Tegstjärn and Sjögoanäs, as well as the bend of the Liane layers south of Stora Ärven (sheet Upperud, 1870). Hence, it is younger than these structures. The central one of the three step faults passes through Näsöln, as a result of which quartzose calcareous slate beds, strongly mylonitized in places and with a western dip, are exposed on the east shore. Along the easternmost step fault the block of the Dalbo plain has been raised, in which the outcrops are mainly basement; only a small belt, lining the fault, is covered by the basal beds of the Dal formation, dipping to the west.

South of Näsöln, no traces of step faults have been found. Here the fault zone is narrowed into one single fault which, going south, cuts obliquely the layers of the monocline which have a normal inclination first, but are overturned south of Dals-Rostock station. Consequently, the Kroppefjäll fault is also younger than the overturning. This becomes obvious too, if we consider the fact that the block of the Dalbo plain has been raised along this fault right on the spot where previously the lowered block of the underthrust must have been situated. The amount of throw of the fault zone near Näsöln approximately equals the whole thickness of the Dal formation, and may therefore be evaluated at 2000 m. This has already been pointed out by Törnebohm (1870a, p. 62), but Ljungner (1927—'30, p. 227) wrongly had doubts about it.

At first sight, it seems as if the throw near Örsjön were considerably less, because of the presence of the lower slate layers both east and west of the fault. The latter are exposed on the block of the Dalbo plain near St. Rud at the east shore of Örsjön (sheet Upperud, 1870 and Sandell, 1941, p. 182—183). However, that here too the vertical displacement must be considerable, becomes evident if we realize that to the west of the fault the dip of the overturned lower slate layers is directed away from the fault and must return to normal position somewhere in depth. These layers in normal position have been raised by the fault. In other words, very distant parts of the lower slate layers have been brought into lateral contact by the fault at the north shore of Örsjön.

Since the fault plane is never exposed in the studied area, its dip could not be determined. Near Buterud, however, a dip of 70° to the west has been recorded by van Overeem (1948, p. 116 and sections). In my sections a westward dip of 70° has been assumed, in accordance with the inclination of the fault near Buterud.

Just outside the eastern limit of the map a north—south striking fault runs through the Dalbo plain along which an eastern block is sunken and a narrow strip of basal beds of the Dal formation escaped erosion.

Renewed movement along the Kroppefjäll fault

Following the Dal orogeny, a long period of erosion wore down the existing mountains to such a degree that a definite peneplain resulted. The formation of a sub-Cambrian peneplain after the Dal folding is not a phenomenon locally restricted to Dalsland and surroundings, but is observed almost all over Sweden. Among many others Högbom (1910) in particular pointed out that essentially this peneplain already existed in sub-Jotnian times, and since then in southern and middle Sweden experienced only minor fluctuations, while in broad lines the peneplain still persists in the actual surface. For Bohuslän and Dalsland this peneplain has been described at length by E. Ljungner (1927—'30, p. 231—243).

The elevations of hill tops and high plateaus show that from Skagerack to the Kroppefjäll fault, the peneplain slopes gently upwards to about 230 m, then passes with a steep scarp to the Dalbo plain, which from 80 m near the fault, drops slowly to Vänern (44 m). This proves that much later a renewed movement occurred along the Kroppefjäll fault which caused the block of the Dalbo plain to sink about 150 m. Thus the renewed movement is contrary to the initial one along this fault.

North of the discussed area, the old surface rises above the Dalbo plain, which passes into hilly country. In other words, the later movement diminishes in a northerly direction and soon can no longer be recognized from morphology.

There is no evidence in the studied region for a closer dating of the age of the peneplain. I therefore refer to E. LJUNGNER (1927—'30, p. 239) who, from the position and the thickness of the Cambro-Silurian sediments at the south shore of Vänern, comes to the following conclusion (p. 239 and p. 242):

"Der Peneplan muss in dem Faltungsgebiet der Dalformation zwischen der Zeit dieser Faltung und der kambrischen Transgression ausgebildet worden sein."

"Der Peneplan ist also in West Schweden erst (während und) nach der kambrosilurischen Sedimentierung durch Dislokationen zerstückelt worden. Man muss darum, besonders angesichts der vorzüglichen Ausebnung der jungpräkambrischen Faltung, annehmen, dass die Ausbildung des Peneplans noch bis ins Kambrium fortgedauert hat."

It is clear, therefore, that the renewed movement along the Kroppefjäll fault must be younger than the Cambrian. Direct observations which would enable a closer dating of the age, are not known. However, E. LJUNGNER (1927—'30, p. 249 and 1937, p. 56) supposes the young fault movements in the Vänern area (see S. DE GEER, 1910) to be connected with the Caledonian folding epoch at the end of the Silurian. In the Oslo region the compression was directed about northwest—southeast by that time, causing renewed movements of the perpendicular faults in the Vänern area. It must be noted, however, that there are also other assumptions as to the age of the young fault movements.

Overthrusts in middle Dalsland

As stated, the overthrust from Bergatjärn to the north, as well as the thrust outlier near Prästbol, form the crown of the Lysesjö overthrust, which has been described by P. HEYBROEK and H. J. ZWART (1949, p. 431). For a detailed description of the thrust zone between Teåkersjön and Marsjön I may refer to their paper, while epitomizing here the results obtained by them.

In the northeastern wall of the Bäcke valley, the thrust plane of the Bäcke valley overthrust is exposed. Gneiss-granite of the basement here overlies the lower slate layers of the Dal formation. On the accompanying map (fig. 22) I have traced, tentatively, the line of outcrop of the thrust plane with a curve around Bäckefors to Ekeliden. For, according to A. SANDELL (1941, p. 153, fig. 93), the gneiss-granite rests here on the basal quartzitic sandstone. The direction of the Bäcke valley thrust has been from west to east. The crown of the thrust mass is exposed at Damtjärn, north of Teåkersjön. A second thrust sheet, the Lysesjö overthrust, covers the area between Marsjön, Bergatjärn and Båsetjärn. This thrust mass rests in part on the Bäcke valley thrust mass and partly on autochthonous sediments of the Dal formation. The authors are led to the conclusion that these two overthrusts cannot be separated with respect to their genesis and time of origin, but are rather to be considered as one thrust mass, which has been broken at a given moment of its initial movement. The upper part (the Lysesjö overthrust) continued the drive along a new thrust plane, while the underlying part (the Bäcke valley overthrust) stayed behind.

In the Dalskog area, the crown of the Lysesjö overthrust has been found, evidence again of a push from the west and at the same time proving that

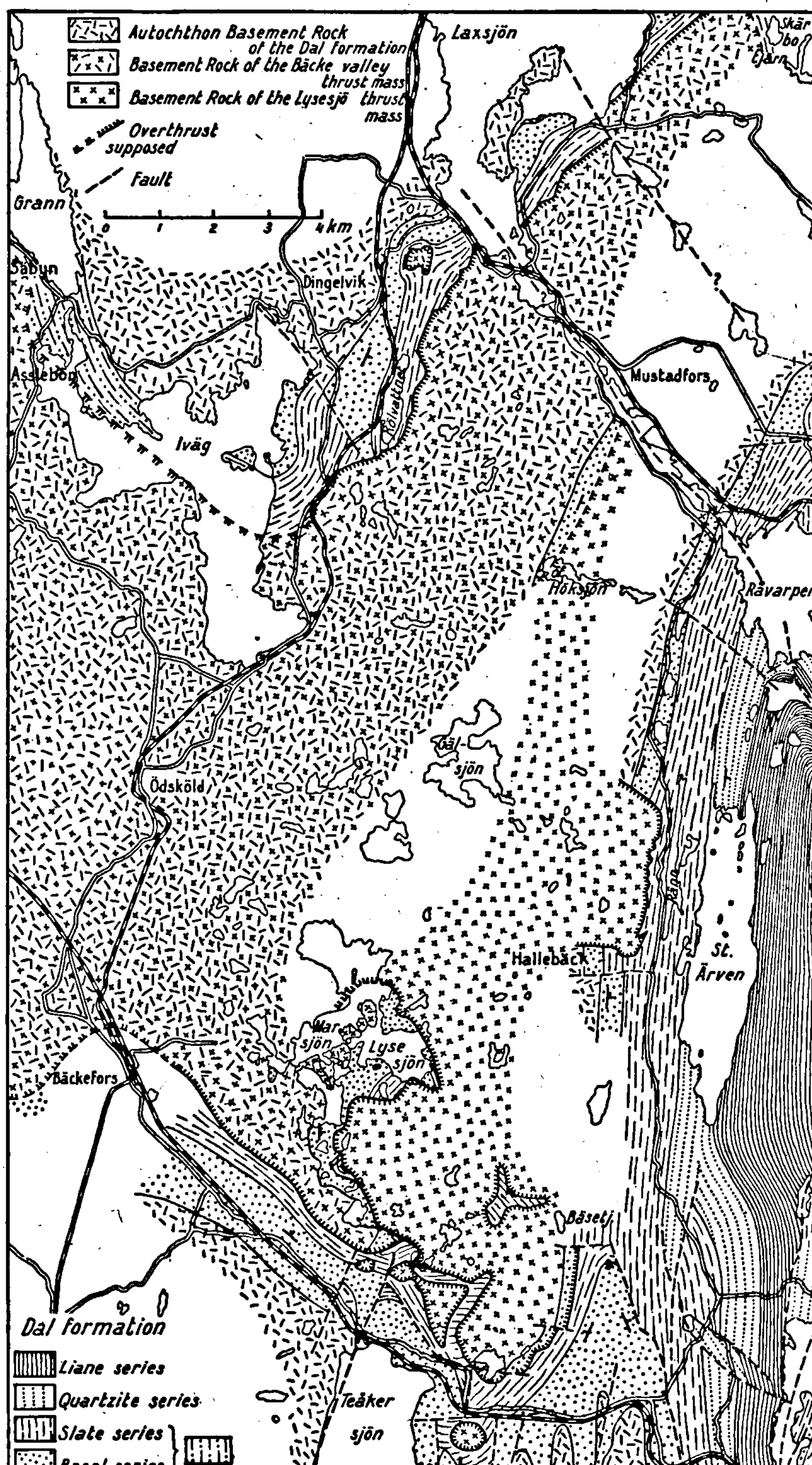

Fig. 22.
Overthrusts in middle Dalsland.

the Lysesjö thrust sheet never extended farther to the east. This crown lies at a distance of 3 km from the crown of the Bäcke valley thrust mass; consequently the Lysesjö overthrust has pushed ahead over a distance of 3 km after the break-up. The detached parts of the Lysesjö thrust mass in this area still have a width of 4 km, measured in the direction of the thrust; hence, we calculate a minimum slip of horizontal thrust of $4 + 3 = 7$ km.

As we are facing here a thrust effect, too extensive to be of local character only, I have examined whether the various data about overthrusts in middle Dalsland fit into the picture, drawn by P. HEYBROEK and H. J. ZWART (1949). I have experienced no difficulties in doing so. The results have been brought together in the adjoining map (fig. 22). The data being as yet too fragmentary, the conclusions are of a hypothetical character.

The data have been secured, for the greater part, from detailed mapping, carried out in 1945, 1946 and 1948 under the auspices of Leiden University, part of which have not been published before. The results of the explorations of A. SANDELL (1941) have been quoted only occasionally and under reserve, as it appeared to me that his diagnosis for the presence of an overthrust must often be deemed unsatisfactory.

On the western flank of the Liane syncline which extends from Tisselskog into the Dals-Rostock area, we find a normal contact of the Dal formation with the autochthonous gneiss-granite over large distances. This contact can be traced with certainty from the vicinity of Tisselskog as far as Ljungberget, south of Hallebäck. Opposite lake Ränn this normal contact is overridden by a thrust mass, apparently coming from the west. This thrust mass outcrops in a down-thrown block, delimited by two faults. The southernmost of the two faults especially has a distinct throw and an east—west strike (VAN OVEREEM, 1948, p. 113, fig. 25). This makes it quite comparable with the post-thrust faults which, in the Dalskog area, caused step faulting of the thrust plane. The relations between Ljungberget and Båsetjärn are uncertain, as no detailed geological mapping has been done here. To the south and the west of Båsetjärn, however, all gneiss-granite is overthrust and forms part of the Lysesjö thrust mass. There is every reason to believe that the overthrust near Ränn is also a part of the Lysesjö overthrust. VAN OVEREEM (1948, p. 114) considers the overthrust near Ränn as an extremely local phenomenon since there is only a very poor mylonitization on the thrust plane. However, even if this thrust mass is taken as a part of the Lysesjö overthrust, it need not be surprising, for in that case it must be considered as the frontal part of the Lysesjö overthrust which has been pushed only a small distance over the lower limb.

An invitation from the "Svenska Institutet" led to the mapping of a small area between Laxsjön and Iväg in October 1945 by H. VAN HEES, J. MEKEL and L. V. D. HARST. They found there an overthrust, by which gneiss-granite has been superimposed, almost horizontally, on the lower slate layers of the Dal formation. The line of outcrop has been traced from the south point of Laxsjön in a south-southwestern direction, along the east shore of Kölvattnet, south of which it takes a more southwestern direction, up to the main road, where the mapping stops. This overthrust and its continuation in northeastern direction (see below) has also been described by A. SANDELL (1941, p. 141). East of Dingelvik station we find a thrust outlier lying on the lower slate layers. Between the overthrust gneiss-granite in the east and the autochthonous gneiss-granite in the west lies a narrow strip of sediments of the Dal formation, more or less parallel to the line of outcrop of the thrust plane. They consist of the basal layers and a portion of the lower slate layers. Apart from a little, secondary anticline, developed mainly in the lower slate layers, they possess a monoclinal dip to the east. The contact between the basal layers and the gneiss-granite to the west is not exposed. The angle of dip of the basal layers and their thickness (about 300 m) do not raise any doubts as to their normal contact with the underlying basement. This structure is cut at the south side by a northwest—southeast trending fault, which forms the prolongation of the northeast shore of lake Iväg and probably determined the origin of this shore. South of this fault the strata of the Dal formation are offset to the west, with at first the same dip to the east. In the neighbourhood of Skuggetorpsön, however, they form an anticlinal bend and, apparently, continue as the strip of sediments of the Dal formation, which are marked on the geological map (sheet Baldersnäs 1870) along the northwest shore of lake Iväg, and along lake Grann and Torrsjön. The latter area has not been studied in detail. From the geological map one may conclude with fair certainty that the eastern boundary of the strip of Dal formation sediments indicates a normal contact with the basement rocks, because a basal conglomerate has been recorded at this contact. From the geological map it becomes apparent also that the sediments are not simply shaped in monoclinal fashion, though a westward dip is predominant.

A. Sandell (1941, p. 147—148, fig. 91) gives a little section of an overthrust near the southwest shore of lake Grann and adds the following description:

"The west side (of lake Grann) is flanked by an overthrust which stands out clearly at the granite mass due east of Säbyn, where the calcareous slate dips below the former in flat position. This overthrust is also found at the west shore of lake Iväg near Asslebyn and at the Krummenäs peninsula" (translated).

The mapping did not reveal the direction of movement of the thrust east of Kölvattnet, but the proximity of the great west—east trending thrust phenomenon makes it impossible, in my opinion, to assume another direction. The following facts must be kept in mind in this connection: 1) that only the southern part of the Bäcke valley and Lysesjö overthrusts are known, 2) that the known front of the Lysesjö thrust from Prästbol to Ränn has already a length of 12 km and 3) that the distance in a north—south direction from the northernmost point of the latter thrust mass to the Kölvattnet thrust mass is not more than 5 km. These considerations make it probable that the thrust masses to the east and west of lake Iväg form part of the Bäcke valley Lysesjö thrust mass. On this account I have tentatively connected the lines of outcrop of the thrust planes east and west of lake Iväg on the map.

The outcrop of the thrust sheet near Kölvattnet can be followed in a northeastern direction across Laxsjön over a distance of about 6 km. This area was mapped by H. Coutinho, D. D. Bannink and H. Odé, also in the autumn of 1945. We find an identical disposition here as near Dingelvik, at the other side of Laxsjön. In the northwest lies the autochthonous gneiss-granite of the basement, which is overlain normally by the sediments of the Dal formation. The sedimentation starts most often with the deposition of a basal conglomerate. The strata have a general dip to the southeast, though minor disturbances of this position have been recorded. Here too, an overthrust gneiss-granite rests on the lower slate layers southeast of the strip of sediments. The line of outcrop of the thrust plane runs from the southern point of Laxsjön in a northeastern direction up to south of Hansebotjärn (Åsnebotjärn). It looks as though two northwest—southeast trending faults have displaced the thrust plane on either side of the Baldersnäs peninsula. We cannot, however, make certain about it on this spot. If this should prove to be correct, it would indicate that the northwest—southeast trending faults in northern Dalsland (e. g. the Tanesjö fault) are younger than the overthrust.

South of Hansebotjärn the outcrop of the thrust plane probably turns to the east, along the south shore of Långvattnet and thence to the south along the west shore of Skärbotjärn, where the lower slate layers again appear below the overthrust. Just to the south of Skärbotjärn, HUMMEL and ERDMANN (1870, p. 58, fig. 17) found a thrust outlier of fine-grained granite lying on the lower slate, but did not recognize it as such. In the east of Hansebotjärn, extends a large outcrop of conglomerate and arkose. On the geological map (sheet Baldersnäs, 1870), it is marked as granite. Possibly this area of arkose belongs to the thrust mass, but the field observations do not suffice to unriddle this region definitely. However, there is good reason to look for the northern extension of the overthrust in this area; I wish, therefore, to call the attention of future investigators to this region.

The long strip of quartzitic sandstone in the basement, situated between Mustadfors and Öf. Höksjön, holds a similar key position for a closer definition of the overthrust phenomenon. A. SANDELL (1941, p. 139—141) gives a section through this sandstone band, which lies in a valley. It shows clearly, that the basal layers rest on the west side with a normal contact on the basement. A basal conglomerate has developed at the plane of contact. One would be inclined, to judge from the section, to regard the eastern limit of the quartzitic sandstone as a fault. In the drawing, however, a different scale has been adopted for horizontal and vertical direction. If we redraw the cross-section in equal proportions, the position of the supposed fault plane becomes much less inclined viz. from 65° to 45°. I do not know how far the drawn inclination of the contact plane is based on actual observation, more so since A. SANDELL (1941, p. 140) himself speaks of an overthrust. The possibility remains, that we have here a nearly flat lying overthrust. If this be true, we would have found here the continuation of the Lysesjö overthrust lying partly upon the crown of the thrust mass, which overlies the lower slate layers to the east of Kölvattnet. This would mean a perfect analogy with what we have found in the area between Teåkersjön and Marsjön and in that case the overthrust east of Kölvattnet ought certainly to be included in the Bäcke valley overthrust.

Hence all overthrusts yet known in middle Dalsland have been fitted into the frame drafted by P. HEYBROEK and H. J. ZWART (1949) for the region north of Teåkersjön. Further detailed exploration will be required, however, to confirm this conception. A few consequences of the conception outlined above are:

a) The presence in middle Dalsland of a broken overthrust, the front of which has a length of 28 km and which has a minimum amplitude of 14 km.

b) The line Bäckefors—Iväg marks a culmination in the thrust mass, which causes autochthonous rocks to appear at the surface north of Iväg and southeast of Bäckefors.

c) Since it is known that the Lysesjö overthrust, being the top part of the Bäcke valley overthrust, moved onward only for about 3 km, it can never have stretched beyond the line Dalskog—St. Ärven—Tisselskog.

It is striking, that in many outcrops, the thrust plane lies in the lower slate layers of the Dal formation. The thrust plane apparently found its way by preference through these slaty rocks, which are bounded at bottom and top by rigid strata. It is possible, therefore, that the culmination of the thrust mass between Bäckefors and Iväg owes its origin to a pre-existent slight folding of the Dal formation.

It is tempting to look for the continuation of the Bäcke valley overthrust south of Bäckefors in the Kroppefjäll plateau east of Högsäter. The three elongated north—south trending strips of quartzitic sandstones in the basement give the impression of being determined by overthrusting (sheet Rådanefors cf. KARLSSON and WAHLQUIST 1870, p. 25—27). Alternatively, one may suppose that they mark a set of minor thrust faults, ensuing from a more intensive compression in southern Dalsland. There is too little known about this region to substantiate these speculations.

The existence of large overthrusts in middle Dalsland corroborates the assumption, which so far has only cautiously been suggested, that the granites in the centre of the Gillberga basin were located there by overthrusts. This basin, the border of which is lying at a distance of 20 km north-northeast from the overthrusts here considered, extends from there 40 km to the north and has been studied especially by N. H. MAGNUSSON (1926, 1929a and 1929b). The contact of the granite with the underlying rocks of the Åmål formation is formed nearly everywhere by a strongly foliated or mylonitic zone. MAGNUSSON believes that the granite was injected in place, the zones of mylonitization being caused by comparatively small movements during the folding. He gives this view however with proper reserve, as is readily seen from the report of the discussion following his lecture (MAGNUSSON 1926, p. 125): "Speaker emphasizes......, that we must consider the possibility of the place of intrusion lying outside the actual basin." (translated)

The Kroppefjäll fault outside the Dalskog Dals-Rostock area

As has been said, the Kroppefjäll fault can be followed in Dalsland over large ditances.

North of Näsöln, the continuation of the central and the eastern faults of the fault zone cannot be read from the geological map (sheet Upperud, 1870) without a closer investigation. We find, however, the continuation of the westernmost fault along Lövnäs and Heden, where the eastward dipping quartzite has been brought into lateral contact with the Liane layers (cf. TÖRNEBOHM 1870b). North of Heden the fault passes on via Vandringstjärn and Lillesjön and vanishes in the fault zone of Buterud (VAN OVEREEM 1948, sections). It is situated here in the eastern flank of the Lianefjäll syncline and consequently must have traversed the axis of the syncline. In this area the eastern step faults have disappeared, but to the west begins a step fault which can be followed in a northern direction on to Råvarpen. The dip of the fault plane near Buterud is 70° to the west (VAN OVEREEM p. 116 and sections).

The total amount of throw of the fault zone increases rapidly from north to south; south of Buterud it amounts to 1300 m already and as previously stated, to 2000 m near Näsöln.

That the main movement along the Kroppefjäll fault is younger than all structures intersected, is nowhere more convincingly shown than in the Dalskog Dals-Rostock area. This prevents me from following SANDELL (1941) and VAN OVEREEM (1948, p. 109) when they say that the first origin of the Kroppefjäll fault is older than the Dal formation. The fact that in the Råvarpen and Åklangen area the Kroppefjäll fault zone more or less coincides with the north—south trending facies boundary in the Dal formation has led the authors mentioned to this conception. As it appears now that more to the south the Kroppefjäll fault obliquely intersects the western part of the basin of sedimentation and the structures occurring in it, it is easy to distinguish from VAN OVEREEM's sections (1948) three very distinct features:

1. The north—south trending facies boundary, approximately conterminous with Åklangen and almost undisturbed south of Lillesjön.
2. A fault zone starting from the line near Buterud and proceeding to the north. It is related to the emergence of the western syncline, whereas the eastern syncline continues to plunge to the north. It is clearly shown by the sections that this fault zone is contemporaneous with the orogeny.
3. The Kroppefjäll fault zone, at a short distance west of the above elements, which bears no signs of having been affected by the folding. This fault zone is probably late- or post-orogenic.

In my opinion, therefore, there is no reason to regard the fault zone from Buterud along Tisselskog to Tanesjön as the continuation of the Kroppefjäll fault, even though one of the faults of the Kroppefjäll fault zone vanishes into the former zone.

South of Örsjön the Kroppefjäll fault follows the overturned basal layers of the Dal formation up to Bollungssjön (sheet Rådanefors, 1870) where these layers also, terminate against the fault. Further to the south, the fault can only be traced on morphological grounds; it runs alongside Rådanesjön, whence, according to LJUNGNER (1927—'30, p. 238) it curves to the west-southwest past Östersjön in the direction of Munkedal. A friction breccia found near Östersjön has a dip of 45° to the north-northwest. Although there is no reason to doubt that the post-Cambrian movement occurred along this line, it is quite possible that this movement followed a different fault plane west of Rådanefors, whilst the continuation of the Kroppefjäll fault may have a more southerly trend.

The stages of the Dal orogeny and later movements

If we review the autochthonous structures of the Dalskog Dals-Rostock area, it appears that the compression increased continually from north to south. In the north, we find two gently arched anticlines separated by a flat syncline. To the south, the Prästbol syncline has suffered further compression along faults and, in the S. Båsane anticline, three superposed anticlines came into existence, separated by narrow synclines. Still more to the south, the shortening is still stronger, involving overthrow and thrust faults, in two zones at least. As we saw, the latter has to be ascribed to underthrusting from the east.

In the discussion of these structures, we had the opportunity to say a few words about their relative ages and it was stated that the Lianefjäll syncline, north of the mapped area, is older than the structures to the south of it. On the other hand, this syncline must have been formed during the principal stage of the Dal orogeny, for its structure is entirely similar to that of the main part of the Dal formation. During this first stage of folding, broad synclines and anticlines of symmetrical character were formed (VAN OVEREEM 1948, sections). Probably the Teåkersjö anticline also originated in this stage.

Of later date than the formation of the Lianefjäll syncline, the anticline bordering it to the southwest has been further arched, by which movement the S. Båsane anticline came into existence. On the flanks of the latter the Kappebofjäll anticline arose, together with the Kabbo and Gunnesbyn anticlines, the southern part of the Prästbol syncline being further compressed. That the overturning and underthrusting are younger than these structures might be infered from the flattening in southern direction of the fault plane, which limits the Kappebofjäll anticline to the west. This fault was apparently forced into a flat position by the underthrust block and thereby became a thrust fault. Thus, from north to south, we find progressively younger structures. The autochthonous rocks of southern Dalsland, to all appearance, have been submitted to an extra compression after the principal stage of folding. The resulting new structures in the Dalskog Dals-Rostock area are younger the further south they occur.

It is not necessary to assume a new stage of folding for each of those structures; rather they arose successively from a slow progressive compression, which was particularly effective in the south. Proportionally to the yielding of the crust to compression, the deformation reached a shorter distance to the north.

One may suppose that the pressure, which achieved the greater compression of the autochthonous rocks in southern Dalsland, caused the formation of great flat lying overthrusts more to the north. Though not impossible, this assumption is premature, as the eventual continuation of the overthrust to the south of Bäckefors has not been investigated; besides, the actual thrust mass has to be taken as a little erosion remnant of the primary mass. Add to this that the thrust plane, although often following the lower slate layers, intersects many autochthonous structures (fig. 22), a phenomenon which for example becomes clear north of Teåkersjön, near Ränn and near Dingelvik. It is likely, though not proved, that the great overthrusts in middle Dalsland are younger than the additional compression in southern Dalsland, because the thrust outlier near Prästbol gives the impression of having cut off the underlying structures, resulting from the extra compression.

The main movement along the Kroppefjäll fault we found to be decidedly later than the folding and additional compression in the south. However, a direct time relation to the great overthrusts cannot be estabished and, accordingly, all we can say about its age is that it originated either in the ultimate stage of the Dal orogeny or later, but in any case before the forming of the peneplain. In view of its strike, which is roughly parallel to the Dal folding, although it intersects the structures, we may assume that the fault belongs to the latest stage of the Dal orogeny.

Of post-orogenic origin are the transcurrent faults, striking northwest—southeast, in northern Dalsland, of which the Tanesjö fault is the most important. They are younger than the overthrusts in middle Dalsland, which they cut (near Baldersnäs), but prior to the formation of the peneplain. Only the renewed action along the Kroppefjäll fault, which can be understood as a Caledonian movement, is younger than the peneplanation.

Summarizing we can distinguish the following stages of movement:

Post peneplain: (Caledonian?)		6. Renewed movement along the Kroppefjäll fault.
Post Dal orogeny:	}	5. Transcurrent faulting in northern Dalsland.
		4.? Main movement along the Kroppefjäll fault.
Dal orogeny:	}	3. Great overthrusts from the west.
		2. Extra compression in southern Dalsland.
		1. Simple folding.

The Kroppefjäll fault in its broader aspects

In relation to the Kroppefjäll fault N. SUNDIUS' paper "Orogene Erscheinungen im südwestlichen Schweden" (1944) is of outstanding importance. In this paper are treated a number of north—south striking fault or overthrust lines with a westward inclination, which the author relates to the folding of the Dal and Almesåkra formations.

In recent years, it has become more and more clear that southwestern Sweden has been affected in pre-Cambrian times by more than one orogenic period. Thus, W. LARSSON (1947) came to distinguish four orogenic periods for the Vårvik region. Almost simultaneously and independently VAN OVEREEM (1948) was led to the same conclusion. Three of those four periods are strikingly prominent in the Dalskog Dals-Rostock area. Outside the area where unconformities are observed in the field, E. LJUNGNER (1927—'30, p. 217—226 and 1937) found several mylonites with eastern and northeastern dip in Bohuslän, indicating very large movements, which must be older than that of the Bohus granite.

The westward dipping faults treated by N. SUNDIUS (1944) therefore are to be considered only as a single effect of a certain orogenic period. These faults consist mainly of three fault or thrust lines viz. from west to east: the Svinesund-Kosterfjord thrust, the Kållandsö line, continued by the "central Värmland mylonite zones", and finally the boundary dividing the geologically different western and eastern Sweden, which particularly north of Vänern, is conspicuously determined by faulting. For further literature on the subject reference may be made to SUNDIUS' paper.

The rocks to the east and west of these lines, especially of the last two, indicate an important throw of the respective faults or thrusts. According to SUNDIUS (1944, p. 291)the parallel trend, in places very conspicuous, and the equal dip of the faults favours a common origin in the same period by the same pressure [1]).

A strike parallel to the Kållandsö line and an inclination to the west is also exhibited by the Kroppefjäll fault; moreover we found that along this fault, too, a movement with a 2000 m throw occurred in pre-Cambrian times. Apparently the Kroppefjäll fault belongs to the same fault system.

An upper age limit of this system was given by MAGNUSSON (1937) on

[1]) SUNDIUS speaks of a force acting from the west. I will not enter further into the difficult question of the direction of the force involved in tectonic deformation.

account of the fact that the central Värmland mylonite zones disappear under the sparagmite formation (Eocambrium) near Mjösen (Norway). A lower age limit is now provided by the Kroppefjäll fault since we saw that it originated late in the Dal orogeny or even after.

Considering the number of orogenic periods southwestern Sweden has known and the non-parallel position of the mylonites near the Almesåkra formation, which are included also in this system by SUNDIUS, a closer specification of the age, based on our knowledge of the latter formation, seems premature to me (SUNDIUS 1944, p. 292—294).

Now, if we look at the map of S. DE GEER (1910), it becomes apparent that the post-Cambrian movement of the Kroppefjäll fault likewise forms part of a whole system of faults, dissecting the sub-Cambrian peneplain. At the same time, it is noted that not only the Kroppefjäll fault displayed renewed action, but the Kållandsö line and the border-line between western and eastern Sweden equally have again been in action in post-Cambrian times. The vertical displacement attending these renewed movements in the Vänern area is small (max. 200 m) as compared with the previous movements.

On the stability of the earth's crust in central Fennoscandia

In a paper under this heading, BACKLUND (1928) advocates the theory that the sub-Cambrian peneplain is not only roughly identical with the present surface of the land and with the sub-Jotnian surface (HÖGBÖM 1910), but does not differ sensibly from the pre-Dal and even from the pre-Åmål surface. His theory is based on the fact that remnants of the sediments pertaining to the various surfaces, are found at approximately the same elevation in central Fennoscandia.

That the remnants of the horizontal Cambrian and Jotnian strata justify such a conclusion has been put forward repeatedly (HÖGBÖM 1910). In the case of the strongly folded formations, such as the Dal and Åmål formations, the surface of deposition, according to BACKLUND (1928, p. 5 and 9) might have been restored, broadly speaking, in its original position, by means of "magmatic and isostatic readjustment".

This opinion was not shared by GAVELIN (1928, p. 467), who pointed out that there is much and weighty evidence for a strong erosion after the Dal folding. HAUSEN (1931, p. 37) too, calls attention to the marked unconformity between the pre-Dal surface and the sub-Cambrian peneplain. BACKLUND (1931) defended himself against this criticism in virtue of the ill-founded tectonical picture of the Dal formation, as given by HAUSEN (1931).

After the detailed studies by VAN OVEREEM (1948) and the writer, it is appropriate to take up the argument and once again condemn BACKLUND's theory. For we know now that the pre-Dal surface within short distances shows differences of altitude of 2000 m, as for instance in the Lianefjäll syncline (VAN OVEREEM, 1948 sections) and along the Kroppefjäll fault. In view of the considerably larger extension which the Dal formation must have had in former times, proved by the remnants of sediments of this formation in the adjoining autochthonous and overriding basement, one is enabled with some assurance to estimate this difference of level of the pre-Dal surface at 3000 to 4000 m.

Nevertheless the arguments of BACKLUND are not without interest. The fact that remnants of sedimentary series of different ages are found nearly at the same level in central Fennoscandia, proves that the elevation in the earth's crust of widely different surfaces of denudation is approximately on one level. This indicates an equalization of the thickness of the crust by processes of isostasy and erosion after each folding. Such equalization does not imply that a given surface, transformed by the folding, will reassume roughly its former position in the crust.

CHAPTER VI

MORPHOLOGY OF THE BEDROCK.

For a full understanding of the main features of the morphology it was often found to be necessary to go beyond the limits of the investigated area. For the various locations I have to refer to the sheet Upperud N. V. of the topographic map.

The southeastern part of the mapped area belongs to the Dalbo plain. This plain, here about 80 m above sea-level, slopes gently down to the shore of Vänern (44 m above sea.l.). In the west and in the north it is bounded by the hilly country of Dalsland. Its northern limit is constituted by an area of stronger relief, which from Näsöln and southern När to the north, gradually rises up. The western limit, on the other hand, is formed by a sharp and steep escarpment, about 80 m in height (fig. 25), which runs along the west shore of Näsöln, via Dals-Rostock to the west shore of Örsjön and thence trends in a southwest direction. This western boundary of the plain is at the same time the eastern boundary of a hilly country and a tableland called the Kroppefjäll.

In the southwestern corner of the region and near Bergatjärn we find parts of a plateau that extends over a large distance of middle- and south-Dalsland. It lies 180—200 m above sea-level, the altitude of the tops averaging 230 m. This plateau occupies a north—south strip on the Upperud sheet, between Stora Ärven and Bodanesjön in the east and Järbo and Ödsköld in the west. Practically all of the rocks exposed are gneiss-granites. By the depression of the Bäcke valley, trending northwest—southeast and terminating to the southeast in the Teåkersjö basin, this plateau is divided in a northern and a southern part. The southern part is named the Kroppefjäll plateau.

Between the Dalbo plain and the Kroppefjäll plateau extend the hills of the Kroppefjäll consisting of a belt of long ridges striking more or less north—south, separated by lakes and valleys. This is the prolongation of the hilly country of eastern Dalsland, of strong relief but moderate elevations, which enters into the area on the northeastern side and to the south occupies only a small strip. The highest parts of these hills are at approximately the same elevation as the Kroppefjäll plateau and generally are flat topped and wide. Most often steep scarps separate the hills from the shallow valleys, which have a flat bottom. These hills consist mainly of rocks of the Dal and Kappebo formations.

In the Dalskog Dals-Rostock area four morphologic units (fig. 23) may thus be distinguished:

1. The Dalbo plain in the east.

2. The table land in the west.

3. The hilly area between the first two units, with north—south to northnortheast—southsouthwest trending ridges.
4. The Teåkersjö basin.

The modelling of these different units is due to four processes, far apart in time, viz.:

1. The formation of the pre-Cambrian peneplain.
2. The breaking up of this peneplain by a dislocation along the Kroppefjäll fault.
3. The pre-glacial and glacial erosion.
4. The deposition of quaternary sediments.

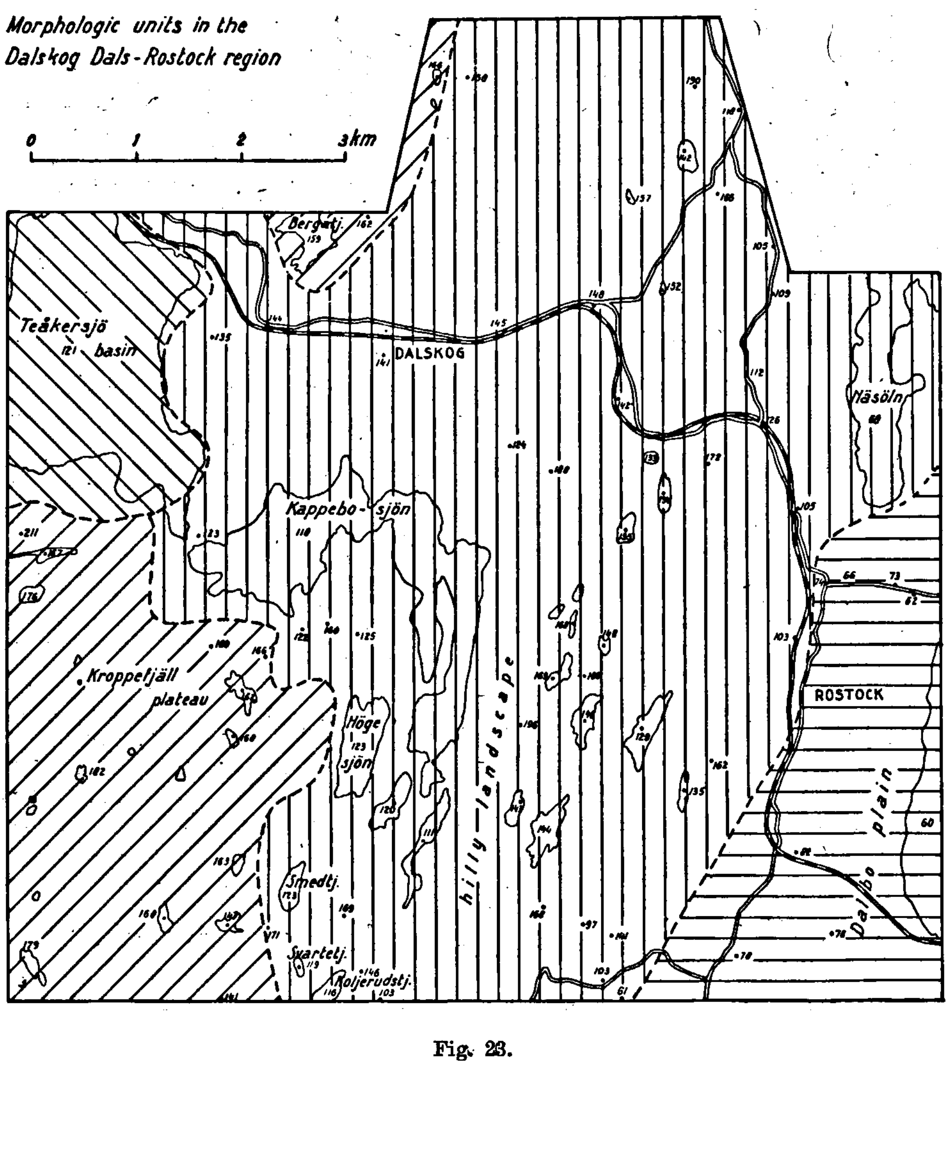

Fig. 23.

The formation of the pre-Cambrian peneplain and its breaking up by the Kroppefjäll dislocation has been dealt with in the chapter on tectonics. West of the Kroppefjäll fault, the peneplain may be recognized as a "Gipfelfluhr" in the hills and the plateau land at an altitude of 200 to 230 m. The block east of the Kroppefjäll fault slipped down to 80 m above sea-level and formed the Dalbo plain. The extremely smooth character of this plain (fig. 24) however, is due to the quaternary deposits, by which practically all unevennesses and valleys in the bedrock were filled up to the level of the "Gipfelfluhr", i. e. of the original peneplain. Consequently a great number of generally small hummocks of bedrock stand only a few meters above the plain which is formed by loose sediments. The existence of some large and deep lakes (Kålungen and Örsjön), which somehow avoided being filled up, proves that considerable differences of elevation must have been levelled in this way (see p. 192).

In the higher area to the north and west of the plain the morphological importance of the quaternary deposits is much smaller. Here, the morphology is determined essentially by the shape of the bedrock, as pointed out already by TÖRNEBOHM (1870a, p. 3). This implies that where the pre-Cambrian peneplain is situated above the level of the Dalbo plain, the morphology is determined chiefly by the pre-glacial and glacial erosion of the peneplain. As may be expected, a close relationship exists between the erosional forms and the geological constitution of the region. Where the gneiss-granite is exposed, there has been a small and uniform degrading of the peneplain and a plateau land resulted (fig. 26). The greater differences of altitude in the hilly country must be ascribed to the variety of the exposed rocks belonging to the Dal- and Kappebo formation, which were differently affected by erosion. The brecciated zones bordering the faults also appear to offer less resistance to erosion than the adjacent rocks.

As regards the nature of the erosive forces, it is evident that the glacial erosion during the Pleistocene has had a preponderant part in the modelling of the bedrock. This is proved by the great number of lakes and lakelets, the "sheepbacks" and glacial striae. However, we may safely assume that the inland ice met a fairly dissected peneplain. It cannot be ascertained in which period or periods after the Cambro-Silurian the pre-glacial erosion has been active, since any trace of later transgressions is lacking for miles around. Perhaps Dalsland has been above sea-level ever since the Silurian and the erosion has been active all the time with varying, though always moderate, intensity.

We will enter now into the details of the erosional features of the bedrock and their relation to the geology. First of all we should make mention of A. SANDELL's treatise (1941) of the morphology and the tectonics of the Dal formation. This author uses the connection between the geological and morphological features to provide a tectonic analysis of Dalsland on morphological grounds. He paid special attention to the area around Dalskog, which roughly coincides with the area studied by the writer (SANDELL 1941, pl. I, p. 169). His method led to the assumption of quite a number of overthrusts and a somewhat smaller number of faults. From the outset of my field work I have endeavoured to check SANDELL's tectonic conceptions, but I often found his deductions and therefore also his method to be erroneous. The reason is to be found in his neglect of a number of factors which played an important part in shaping the morphology. For one thing, SANDELL did not, or not sufficiently, take into account that not only faults and overthrusts, but also stratification planes constitute planes of discontinuity in the crust, which can be brought out by erosion. Furthermore there are many factors operative in the formation of the landscape, the effect of which cannot be fully evaluated. In this region, which has been finally transformed by the erosive action of vast glaciers of the Pleistocene period, one may think for instance of the direction of glacial flow, the possibilities of swerving, the lateral pressure of the ice and also of the preexisting configuration of the surface. Moreover the position and the degree of schistosity, the stratification and the cleavage of the rocks affect the rate of erosion. All this makes it very risky to draw geological conclusions purely from morphological data. Such conclusions always need geological proof, which SANDELL (1941) most often neglected to provide. In the field the morphology of the landscape can be very useful, as it may signal the presence of tectonic disturbances or petrographical changes, but it very seldom offers conclusive evidence.

It is striking that the long valleys, lakes and ridges in the hilly country trend parallel to the strike of the Dal formation. This is comprehensible if we recall that here the direction of movement of the ice very nearly coincided with the direction of strike. However, even in those places in Dalsland where this is not the case, the strike is determinative for the topographic forms (VAN OVEREEM 1948, p. 122). We must assume therefore that certain rocks were more resistant to glacial erosion and formed ridges, while others, less resistant, were carved down to valleys. Though, broadly speaking, this may be true, there are numerous exceptions to this rule. The exceptions may be due to the brecciation of the adjacent rocks by faults. The resulting brecciated zones are an easy victim to the erosive forces. Again this rule, though generally true, has too many exceptions to be applied inconsiderately in the tectonical interpretation of the area. Apparently, a multitude of other factors are involved.

The most resistant rock proves to be the gneiss-granite, which has formed the high plateau in the west. In this respect it does not make any difference whether the gneiss-granite has been overthrust as a thin slab upon younger rocks (as it is north of the Bäcke valley) or constitutes an autochthonous mass (as e. g. the Kroppefjäll gneiss-granite).

The effect of erosion on the rocks of the Kappebo formation is very irregular. On the Kappebofjäll with its broad, smooth ridge, they reach about up to the level of the old peneplain. To the south the Kappebofjäll turns into a wide depression near Åsen, the same rocks being still exposed at the surface. Also west of the Kappebofjäll, the erosional forms in the Kappebo formation are very irregular, which must be due partly to the influence of fault-breccias.

The hard basal strata of the Dal formation seem to have been readily attacked by glacial erosion. This is shown remarkably well in the Teåkersjö basin, which has been scoured out entirely in these rocks and is surrounded on three sides by gneiss-granite (fig. 27). In the other localities too, as for instance between Dansbo and Helvetestjärn, near Dalskog, near Tångebo and near Rörnäs these rocks occur in the depressions. Northwest of the mapped area, the Bäcke valley and the depression of Marsjön and Lysesjön are of similar origin as Teåkersjön. Yet, on the other hand these rocks occasionally

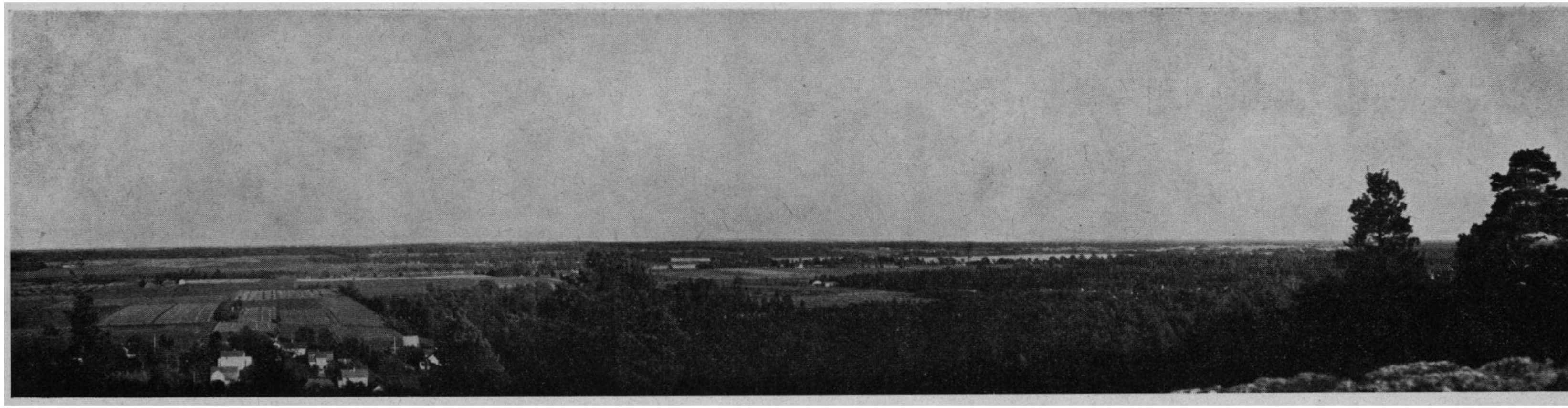

Fig. 24.

View of the Dalbo plain from the Kroppefjäll escarpment near Dals-Rostock. Extreme left Gunnarsnäs Kyrka. In the middle Kålungen. In the background left of centre Mellerud.

Fig. 25.

The Kroppefjäll escarpment near Dals-Rostock seen from the Dalbo plain.

Fig. 26.

View from Kappebofjäll to the west. In the foreground (centre and left) Kappebosjön and right the plain of the Tångebo valley with the river terrace north of Kabbo. In the background the Kroppefjäll plateau.

Fig. 27.

Teåkersjön viewed from the northeast (from Stora Halängen). The horizon is formed by the escarpment of the Kroppefjäll plateau.

stand far above their surroundings, as north of Bengterud and southwards of Sketjärn, where from a high cliff of quartzitic sandstone one may enjoy the magnificent views of the Dalbo plain. The strong erosion of this otherwise very resistant rock must be imputed to the numerous joints commonly occurring, which promote its rapid disintegration. This is proved by the contents of the fluvio-glacial gravel deposits, which consist for a high percentage of ill rounded quartzitic fragments (see p. 189—190).

Strangely enough, the lower slate layers form a prominent ridge over large distances west of Dals-Rostock, generally outranging even the spilite-bearing layers, which show themselves resistant towards erosion throughout. The elevation of the calcareous slate in these parts is in sharp contrast to the region east of Upperud, where a string of lakes has been formed in this rock (VAN OVEREEM 1948, p. 122).

The behaviour of the quartzoze slate and the quartzite towards erosion is also extremely variable. Near Råbäck and Dals-Rostock, the elevation is appreciable, but to the north near Årbol and Tegen they occupy a broad valley which is the continuation of the depression of Stora Ärven. The variability of behaviour of the hard quartzite may be due to differences in the number of joints, as was pointed out by VAN OVEREEM (1948, p. 123). However with respect to this rock, we must also bear in mind that locally the very thin stratification greatly facilitated erosion. In this connection we noted (p. 148 fig. 17) the existence of small "hogbacks" west of Råbäck.

The rocks of the Liane layers, here, as practically all over Dalsland, prove to be very resistant against erosion. West of Näsöln they constitute the Årbolfjäll, a solitary ridge. In the narrow valley, between the Årbolfjäll and the ridge of quartzite west of it, there is no fault (SANDELL 1941, pl. I, p. 169). The normal sequence of the Dal formation can be traced throughout the valley. The slaty intercalations in the boundary beds between the quartzite and the Liane layers appear to have facilitated erosion to a large extent. Moreover this is the only valley through which the ice could have flowed, with a steep gradient (about 60 m pro 1.5 km), from the elevated valley near Årbol to the Dalbo plain.

The boundaries of the various strata are often characterized by more or less high scarps in the landscape. They usually arise from petrographical variations in the underground and not from faults or overthrusts, as SANDELL believes. In this connection I want to stress especially the fact that the high steep western escarpment of the Kappebofjäll has nothing to do with a fault or overthrust as has been pointed out as well by SANDELL (1941), as by HAUSEN (1931, p. 26).

The influence of faults on the morphology of the bedrock proves to be even less regular than the influence of the petrographical habit of the rock. If a fault shows up morphologically, then its manifestation follows roughly this rule: If the same rock is exposed on either side of the fault its trace is characterized by a valley or a gully in the bedrock. If, on the other hand, different rocks are brought into contact, laterally, by a fault, then a fault scarp occurs. The first case, as may be logically expected, is encountered mainly in the Kappebo formation and in the Kroppefjäll granite, for instance west of Svartetjärn and near Rotjärn, as well as in the southeast corner of Teåkersjön and the eastern arm of Kappebosjön. In the second case, besides the influence of the fault, the petrographical changes of the rock are also of interest in determining the land form. The resulting fault scarps are observed e.g. between Håltjärn and Långtjärn, south of Dalskog and north of Högesjön. The Kroppefjäll fault is such a particularly prominent feature in the landscape, because it forms the limit between the Dalbo plain and the hills. As we have seen previously, this is not owing to erosion. However, the glacial erosion did accentuate this line and formed the lakes Näsöln and Örsjön.

CHAPTER VII

QUATERNARY DEPOSITS AND MORPHOLOGY

Although it has not been my purpose to study the quaternary deposits of the Dalskog Dals-Rostock area, a few of the most striking phenomena, connected with them, will be briefly discussed. I want to stress that no systematical observations have been carried out in the quaternary deposits, but rather incidentally a number of mainly morphological data were collected. Hand-drilled holes and levellings, indispensable for a full understanding, have not been carried out. On the map the Quaternary has been left uncoloured, the subdivisions being marked by different symbols, in those places where they are encountered and distinguished.

In the region surveyed determinations of the highest marine limit (M. G.) are not known. On the outline map of late-glacial deposits of southern Sweden by G. DE GEER (1910) the marine limit has been indicated at 150 m, 1 km north of the edge of the map, east of Alhagen. Taking this figure as a reference, the distribution of land and sea at the time of the maximum water-level of the Yoldia Sea has been plotted on a chart (fig. 28). Necessarily this is only a rough approximation of the actual situation, as contour lines are lacking on the topographic map and a sufficient number of altitude measurements is not available.

The bare bedrock

The quaternary deposits in this area, just as in western Dalsland and north Bohuslän, are confined to the depressions in the landscape, whereas the salient rock masses and rocky plateaus are fairly clear of them, no matter whether they occurred under or above the water-level immediately after the glaciers' withdrawal from the region. G. DE GEER (1902, p. 41) noticed this phenomenon in north Bohuslän and once again called the attention to it in his classic description of the glacial deposits of Dals-Ed (1909, p. 7) in the following terms:

"On the mountain plateau the bed rocks are often almost quite bare, with the exception of the concavities which are occupied by small lakes or peat bogs. Scattered boulders and scanty deposits of morainic material are indeed to be found here and there, but still it is very remarkable how little material was generally left by the receding land ice. This is the more striking, since great tracts of the region are situated above the highest marine limit, where thus no later degradation of moraine matter by the sea-waves can have occurred. Furthermore it is difficult to assume that sub-glacial rivers could have swept the rocks bare to such a degree, if they had been at first moraine-covered to a larger extent. In such a bare tract the great and almost continuous ridges of the terminal moraines form a very striking feature in the landscape, and show that the ice-border, to afford the time necessary for deposition, must have been stationary during quite a number of years or, as now known, about a century."

The assertion that practically no material was left by the land ice between the typical ridges of the terminal moraines, does not hold for the

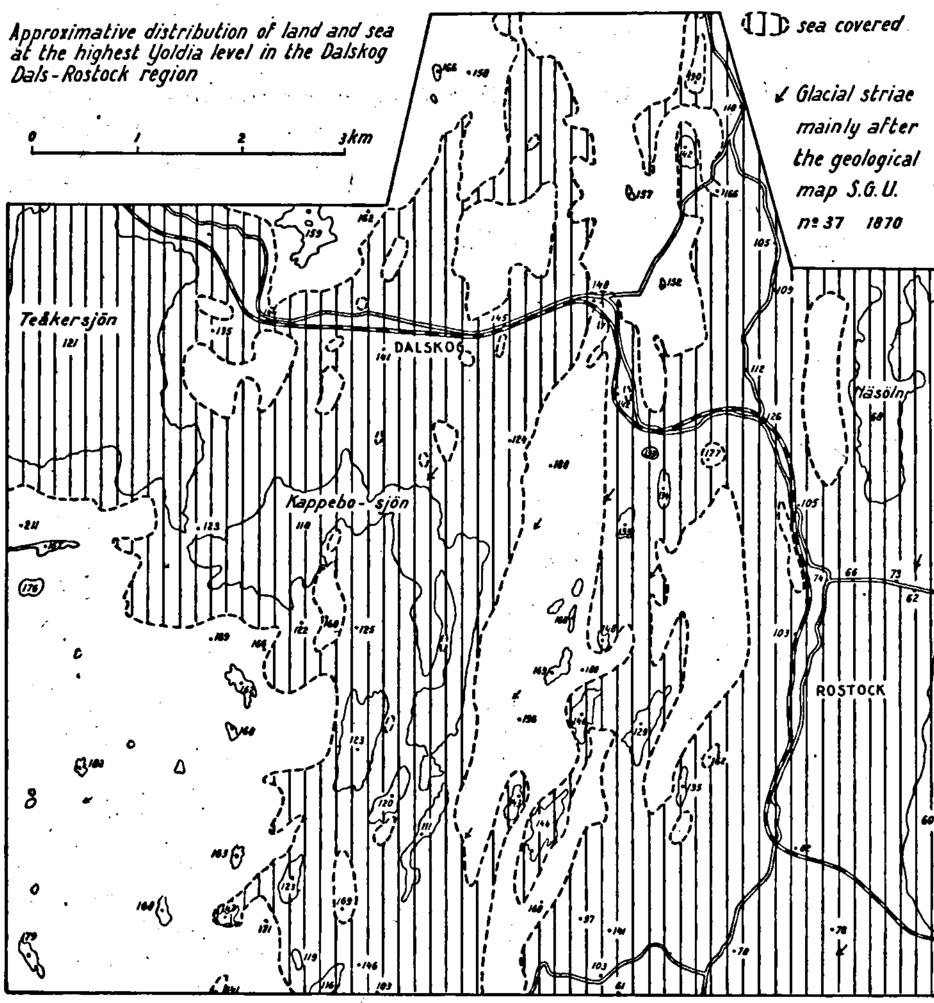

Fig. 28.

Dalskog Dals-Rostock region. Deposits, at times, of huge thickness of fluvioglacial material, even in such places where they cannot possibly have been formed during a long stationary stage of the ice, are in contradiction with DE GEER's views. IVAR HESSLAND does not agree either with DE GEER's contention. In his work on north Bohuslän (1943, p. 17—23) he supposes that large quantities of moraine material have been swept off the plateau by subglacial meltwater streams and have been deposited in the valleys, whereupon the cleaning is completed by the wave action of the steadily falling water-level of the sea. The high level of the meltwater in the melting glacier provides a hydrostatic pressure, causing rapid subglacial streams, which may extend even below sea-level. Eventually the glacier could be lifted by the meltwater, if this stretches far enough upward, thus providing room for a fast flowing subglacial sheet of water (E. LJUNGNER 1927—'30, p. 396—403). A strong scouring and eroding action must be its result. It seems to me that this gives a reasonable explanation also for the bare bedrocks in the Dalskog Dals-Rostock region. Since a great part of the mountainous region was situated well above sea-level the degradation by wave action need not be taken into consideration.

Deposits of pebbles on the east side of steep crags

To the east side of the steep face of the Kroppefjäll there has been, in places, deposition of gravel with very large boulders up to 1 m diameter, reaching high up the scarp. It is poorly sorted material of essentially coarse gravel, with numerous lumps scattered throughout. A deposit of this kind has been formed near Råbäck at the lower end of the valley which separates the Årbolfjäll from the Kroppefjäll scarp.

Another is found to the south, in the triangle between Dals-Rostock station, Mörtjärn and west of Stenbacken, where the scarp of the Kroppefjäll curves inward. Further south yet, similar deposits occur north of Örsjön, again in a concavity of the Kroppefjäll edge. The somewhat jagged edge of the Kroppejfäll has been smoothed by these deposits so as to form a regular slightly bent curve. In other localities also these deposits, profuse of large blocks, can be found, but always on the east side of a steep crag. So they have been encountered along the scarp, formed by quartzitic sandstone and conglomerate, north of Örsjön and especially along the track to Åsen, where there is a slight incurvation of the wall. Furthermore they can be found between Skogstjärn and the steep face of the Kappebofjäll and along the steep crag, composed of graywacke, southwest of Koljerudstjärn. For the origin of these deposits I want to suggest two possibilities:

1. Their position, more or less parallel to the glacial striation (fig. 28) and their component material, suggests that they are esker sediments, deposited in subglacial channels by meltwater streams of very high strength of current (G. DE GEER 1897). Their association with special topographical features of the bedrock might be accounted for by the assumption that crevasses in the glacier existed on the east side of the long north—south trending scarps.

2. Another possible mode of origin may be found by comparing these deposits with the "praecrags" found west of the lake Stora Le.

G. DE GEER (1895, p. 212) supposes that they originated by deposition of moraine material on the stoss-side of rock masses:

"In Dalsland these vast subglacial deposits are far more conspicuous (than in America), since they occur in a region where for the rest the bedrock is lying bare at the surface over large distances. It is of special importance here that, contrary to the general rule, a definite and local cause can be held responsable for the formation of these hills. For their distribution is markedly connected to the west shore of lake Stora Le and most probably they have been formed here chiefly by the loose material which the land ice picked up from the bottom, in sliding past from the northeast, crossing the valley of the lake at an angle and dropped upon the encounter of the cliffs" (translated).

There is a striking analogy between the topographical character of the deposits under discussion from the Dals-Rostock area and these from Stora Le. All of them have been deposited on the east side of a rocky plateau or a ridge, traversed obliquely by the ice. At Stora Le, as in the Dals-Rostock district, moraines have been deposited at the stoss-side. The fact that these deposits have been found mainly in excavations of the crags fits well in this theory.

The important deposits of gravel and pebbles, accumulated against the scarp near Dals-Rostock, have been exposed to the wave action of the sea which covered the Dalbo plain. Consequently the finer material was washed out and, in places, beach lines have been formed. This instance induced A. ERDMANN (1868) to consider all eskers as shore formations. TÖRNEBOHM (1870a, p. 65), though denying the generalisation of this rule, equally believes in the presence of a shore formation near Dals-Rostock. It is clear, however, that at the present state of our knowledge about the glacial and marine deposits, this view does not hold.

Terminal moraines

The region surveyed is situated between two main ridges of terminal moraines. One of them passes south of the mapped area from Brålanda to Råggärd, thus crossing the Kroppefjäll and is the result of the stationary stage G [1]). The other one is a part of the composite Fennoscandian terminal moraine, resulting from stage H. This stage forms the boundary between the Gotiglacial and the Finiglacial phase.

In the northwest of the mapped area a small part of these terminal moraines is to be seen. In the area between Mellerud and Näsöln possibly three substages H_1, H_2, H_3 might be distinguished in three closely gathered ridges of moraines. If this supposition be true, it is the terminal moraine H_1 which passes over Mellerud, Gunnarsnäs and Backa, where it crosses the deep valley of the western branch of Näsöln up to the Årbolfjäll, constituting a barrage across Näsöln.

To the west not a trace of this terminal moraine is left, unless the little patch of morainic material north of Dansbo is a remainder of a terminal moraine, which is not probable. A more plausible assumption is that stage H_1 comes to coincide with stage H_2, which has resulted in a terminal moraine 1,5 km to the north of H_1, reaching Näsöln near Bårviken. This view is supported by the presence of hillocks between the morainic ridges of Gunnarsnäs and Bårviken, on the east side of Näsöln. Another 1,5 km to the north of H_2 we find the terminal moraine of stage H_3, which however falls entirely outside the map. The terminal moraine that is damming up Näsöln in the south has not subsisted in its original shape. The moraine ridge has been degraded and erosional gullies have been formed. A shallower one close to Årbolfjäll and a broader and deeper one near Backa. It is not clear to what cause we have to ascribe these gullies. A stream of meltwater during the stationary stage in which this moraine was built, would have caused a marginal delta. Just the contrary can be observed. At the distal side of the moraine ridge there is an excavation in the Dalbo plain. A glacial stream

[1]) Notation according to H. MUNTHE (1940, p. 16—19).

debouching into the deep valley of Näsöln during a stage in which the ice sheet covered the northern parts of Näsöln, would have flowed out into a sea of about 100 m depth and consequently could not have had any influence upon the moraine ridge at some distance from its mouth. By the time the sea had retreated so far that the terminal moraine emerged, Näsöln had been delivered from the ice long ago and could use directly its present outlet in the north.

Northwest of Tegen and northeast of Amerika we find a morainic area, through which runs a definite ridge of a terminal moraine, trending southeast—northwest. To judge from the scarce spot elevations of the topographic map this moraine must have been deposited above sea-level. This becomes the more probable since in the southern flank of the ridge gullies have been scoured out by streams of meltwater. The position of this ridge is just in a straight line with the one of Bårviken and must be considered as its prolongation.

The morainic area south of Teåkersjön near Kläppa, lying against the scarp of the Kroppefjäll, can, in my opinion, only be a terminal moraine. Huge quantities of large boulders cover, up to a considerable height, the steep northern wall of the Kroppefjäll. The east—west direction of the axis of these deposits moreover makes it difficult to consider them as an esker.

Marginal deltas

The farm Amerika is situated close to the face of an escarpment which separates an upper morainic platform, covered with large blocks from a lower plain, consisting of sand and gravel. Close to the scarp on the plain one may still find some large blocks, but their number is rapidly decreasing towards the southwest. The escarpment corresponds very nearly to the marine limit, thus marking the spot where the moraine has accumulated above sea-level. As stated above, across this moraine area stretches a ridge of a terminal moraine, due to an important stationary stage. The subglacial streams, breaking away from underneath the stationary ice cap, flowed over this morainic area at the distal end of the terminal moraine and deposited their coarser material on it. As soon as they reached the sea-level, while flowing over the moraine, the strength of the current dropped and by and by the finer material subsided and circumstances were favourable to the formation of a large marginal delta. These fluvio-glacial deposits filled up the lower parts between the quartzitic cliffs about up to sea-level. Subsequently the glacial streams sought their way over this surface and through the narrow passages between the rocks to the lowlands near Dalskog and Tångebo, where enormous masses of sand and gravel have been deposited.

Enclosed on both sides by rising rock beds and shut off at its proximal side by a terminal moraine, this marginal delta presents a typical example of a marginal terrace, according to the terminology of H. Nelson (1909). This gently sloping marginal terrace extends down to the northern shore of Kappebosjön, covering an area of 4 km². Similar marginal deltas, due to the same stationary stage, have been formed west of the region under discussion, although under somewhat different conditions, at the Ödskölds moar and near Dals-Ed (G. de Geer, 1909).

The deposits of the marginal terrace near Dalskog consist of sand, coarse sand and loamy sand in which occur many boulders. We deliberately do not call them pebbles, because the main part of the stony contents is made up of poorly or not at all rounded quartzitic sandstones and quartzites of the Dal formation. The well rounded pebbles of basement rock are only a small minority.

The most southern altitude measurement on the marginal terrace is close to the church of Dalskog at 141 m. Supposing a height of 150 m for the sea-level at the time of deposition and thence for the proximal part of the terrace the resulting gradient is 1:300, which is in accordance with similar deposits in Sweden (H. NELSON 1910, p. 48). The surface of the terrace is remarkably smooth, although slight differences of level occur in narrow passages between the cliffs. This is obvious from the fact, that these obstacles must have affected the rate of flow and the supply of material. This marginal terrace has been strongly cut down by later river erosion, leaving intact only a tiny portion of the steep distal slope, which is so characteristic for subaqueous deltas. It is still plainly visible, however, to the south and southeast of Tonebyn.

The stratification dipping at an angle of 20°—30° to the south, observed in the gully of the brook 750 m south of Korsgården, is indicative of the fact that we are in the presence of the frontal beds of the delta. The high angle of inclination of the beds at the distal side of a submarine delta is caused by the material transported over the flat top of the terrace and falling down the edge under an angle of deposition, determined by the nature of the material and the surrounding medium. The delta is not growing from the bottom to the top, but in off-shore direction.

Although the distal slope is lacking south of Korsgården, as a result of post-glacial erosion, we are entitled to assume, on account of the typical dip of the frontal beds, that it has been present here also near the shore of Kappebosjön and stood out above it for some 15 m. This conclusion is justified as well by the occurrence of terrace deposits in the little rock-bound valley of Kabbo. These deposits have been preserved from post-glacial erosion by the rivulets that have worn away the terraces south of Korsgården and south of Tångebo. It is true, here too that post-glacial erosion has done its work, but by the very minuteness of the reception basin of the streamlets this was bound to be no more than a fraction, compared to that of the surrounding valleys. For this reason the fluvio-glacial delta deposits at the northern side of this valley lie almost at the level of the original terrace, while at the shore of Kappebosjön the distal slope has partly escaped erosion. The fluvio-glacial deposits to be found at the northeast corner of Kappebosjön, lying against the Kappebofjäll, high up above the valley floor, may be evidence that here too the end of the terrace was lying near Kappebosjön.

400 m south of Korsgården I found in the bed of the brook a plastic grey clay below the terrace deposits. This indicates that the marginal terrace has grown in off-shore direction over the clay already deposited on the sea bottom. This also is a frequent phenomenon in marginal deltas. It was noted already by TÖRNEBOHM (1870a, p. 68) in the marginal delta of Ödskölds moar, 10 km northwest of Dalskog.

As stated above, there has been a considerable post-glacial erosion of the terrace around Dalskog. Streamlets, with a rate of flow many times as great as the present ones, have cut in the marginal terrace the normal pattern of fluvial erosion found on plateaus with steep edges. On the terrace itself there is not much of incision, the gradient being too small. Near the edge, however, a narrow incision is formed which downstream becomes gradually deeper and widens near the base of the terrace. In front of the terrace a fan talus was built by deposition of the eroded material. In the Dalskog terrace this process continued for a rather long time and the terrace edge has been shifted by retrogressive erosion up to 1,5 km to the north (near Tångebo). In this way we get the present image of a terrace with a capriciously dissected edge, joined by a gently sloping flat valley floor. Two valley plains thus came into existence, one south of Tångebo, another south of Korsgården. Especially the one south of Korsgården shows an extraordinarily regular and smooth character. This is to be ascribed to the location of the cliffs east and west of the valley, obliging the meandering stream to grade a relatively small area in proportion to its water-supply. The valley floor south of Tångebo, though equally bounded by rock masses, is less regular, e. g. north of Kabbo a river terrace has subsisted.

The erosive action of the present brooks is very small, which is shown by their meandering course, even in the incisions covered with vegetation, while other gullies are perfectly dry. This phenomenon is very well displayed near the Dalskog church.

Kappebosjön must have had, at a time when the water-level was a few dm higher, a slight abrasive effect on its shores. This is evidenced by the occurrence, in places, of coastal plains. They have a clay bottom, which proves that the fluvio-glacial deposits do not extend below the level of Kappebosjön.

South of Gunnesbyn an erosion mark occurs on the distal slope of the marginal terrace, which cannot be regarded as the result of fluvial erosion. In the terrace edge there have been carved three tongue-shaped little valleys surrounded on three sides by steep walls. In the north they terminate by a steep wall against the undissected terrace plain. These tongue-shaped valleys have, almost immediately at their beginning, a flat bottom running down to Kappebosjön. The slope of these valleys is much steeper than those of the river-formed valley plains. We are here witnessing the effects of soil slip. Possibly the distal slope of the terrace, after the retreat of the sea, was too steep so that sudden tongue-shaped earth-slides resulted. Perhaps the formation of quicksands in the extremely wet sediment has been of importance.

The irregular depth-profile of the northern branch of Kappebosjön must be imputed chiefly to post-glacial dumping of material, derived from the marginal terrace.

In the narrow valley, running from Långtjärn in southern direction to Örsjön, there has been an accumulation of large quantities of fluvio-glacial deposits. This can be observed especially in the gravel pit on the road north of Örsjön. The coarse sand and gravel, with pebbles up to 20 cm diam., show a stratification dipping to the south. This, together with the steep topographic slope of the deposit south of the road, descending to Örsjön, indicate that we have here a marginal delta. This marginal delta is not connected at its proximal side with a terminal moraine. We are led to the assumption that it has been formed during a period of rapid recession of the ice sheet. Though there was an ample supply of material, the lapse of time was too restricted to permit the formation of a marginal terrace reaching up to sea-level. Törnebohm (1870a, p. 69) describes an occurrence of marine clay with fossils, just south of Långtjärn, lying directly on the bedrock. It seems likely to me that in the high parts of the valley no fluvio-glacial sediments have been deposited, so that the subsequent marine sediments here lie on the bedrock, but more to the south they have been deposited on the delta beds. However, sufficient field data to confirm this conception are lacking.

The Dalbo plain

The exceptionally smooth bottom of the Dalbo plain must be due to the levelling by wave action of the retreating sea. It is not conceivable that deposition of morainic and fluvio-glacial material could have taken place so regularly over such wide areas. A striking feature in this plain are the large concavities, filled up partly by the lakes of Kålungen and Örsjön. The difference of level between these depressions and the average height of the surrounding deposits was too great to be blurred out by wave action. The main level of the plain is about 15 m higher than the water-level of the lakes. The maximum depth of Örsjön is 23 m (Sahlström 1929), so the "hole" in the plain is 38 m deep. The depth of Kålungen is not known. Consequently there resulted a very flat plain, in which depressions with comparatively steep walls have been left.

North of Örsjön and along the shore of Kålungen the post-glacial, fluvial erosion has been important and has scoured out similar forms as at the edge of the Dalskog marginal terrace. There too we had a plain separated by a scarp from the Kappebosjö depression.

In the Dalbo plain the erosion by the present rivulets is extremely small. They display the same characteristics as on the terrace of Dalskog.

A plate-shaped depression, which must have been filled with water until recent times occurs south of Näsöln, which itself is barred by a terminal moraine. The only drainage of this little basin takes place through a very narrow gully near N. Bäckebol. Considering the narrow V-shape of this gully it is improbable that it should have originated at a time of strong fluvial erosion, after the sea and ice had retreated. We should rather assume therefore, that the little basin has been tapped by the brook that is running even now through this gully. Its present flow seems to be sufficient for that.

Peat

In numerous smaller and larger shallow excavations in the bedrock there has been formation of peat. Also in places where the quaternary deposits have a nearly horizontal surface it appears that, owing to the poor drainage, circumstances are rapidly leading to the formation of peat. Thus peat has been formed on the marginal terrace of Dalskog, on the coastal plain east of Famsed, north of Skogstjärn, on the Dalbo plain near Dals-Rostock and in other localities. Several of the peat-bogs, which we encountered, are dealt with in the description of the Upperud sheet of the "Torvmarkskartor" (1921).

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