

SVERIGES GEOLOGISKA UNDERSÖKNING

SERIE C NR 619 AVHANDLINGAR OCH UPPSATSER ÅRSBOK 61 NR 4

PER GEIJER

THE PRECAMBRIAN QUARTZITE
IN THE NORBERG DISTRICT,
CENTRAL SWEDEN,
AND ITS IRON-SAND BED

WITH SOME ASPECTS OF THE EVOLUTION OF
THE FENNOSCANDIAN SUPRACRUSTALS IN CENTRAL SWEDEN



STOCKHOLM 1967

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ABSTRACT

The special interest associated with the Norberg quartzite derives from the fact that it forms in these parts the basal member of the Larsbo (-Mälar) Series of detrital sediments, which — with a slight erosional unconformity — follows on top of the volcanic leptites that make up the lower series of the supracrustal complex, referred to the Svecofennian cycle. The accumulation of quartz sand is regarded as proving that the detrital material came from a land area of mainly granitic nature, and thus from outside the area of volcanics. The local iron-sand bed, with magnetite as the chief constituent, shows the characteristic features of a beach deposit of that nature. Secondary traits are that the detrital ilmenite grains have been greatly altered during diagenesis and that later metamorphism has caused the magnetite to recrystallize and its content of titanium to separate as ilmenite. The quartz was also texturally affected. The zircon grains have retained their detrital shapes unaffected. The zircon indicates granitic rocks as the probable source of the sediments, but the magnetite and ilmenite are presumably derived from basic plutonics. The formation of the bed is ascribed to storm waves, probably acting on dune sands above the ordinary high-water mark, and it is found probable that the open sea lay in an easterly direction.

The last section of the paper is devoted to a discussion of the relations between the two upper, sedimentary series, and the lower, predominantly volcanic series of the supracrustals, and of the first phases of the Svecofennian folding.

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INTRODUCTION

In a monograph on the Precambrian geology and ore deposits of the Norberg district (Geijer 1936) the writer briefly described a bed of metamorphic "iron-sand" or "blacksand", interstratified in a quartzite. This deposit has later been subjected to a more detailed investigation, and some new field observations relating to the quartzite have also been made. Since the publication of the monograph important new data have, through the work of several investigators, become available concerning the supracrustal sequence of which the quartzite forms a member. The stratigraphic position of this unit, and the conclusions that can be drawn about its nature and that of the iron-sand deposit, make it a suitable starting-point for a discussion of the general development of the supracrustal complex in the light of these new data.

It is, unfortunately, necessary to devote first some space to questions of stratigraphic nomenclature, or misunderstandings may arise. The supracrustal "Archean" sequence in Central Sweden has long been known as the Leptite Formation. In 1920, Sederholm spoke of it as representing "a Svionian type". What he then had in mind was, however, what is now known to characterize only the lower part of the sequence. Svionian was later taken up as a term for the whole geological cycle of which the deposition of these supracrustals formed the first stage (Magnusson 1936 b). Some years ago the term, which was derived in the same way as Cambrian, Ordovician and Silurian, i. e., from the Roman name for the people inhabiting the type area, was discarded, and replaced by Svecofennian, with the same extended meaning. "Svecofennian" was originally introduced by Ramsay (1909) for the denuded mountain chain of Precambrian age whose roots form a belt from Central Sweden to southern Finland. More recent developments have had as a result that Svecofennian now has several different meanings. One speaks of "the Svecofennian orogeny", in accordance with Ramsay's original use of the term. (Absolute datings have shown conclusively that this epoch ended about 1750—1800 m. y. ago.) Through "the Svecofennian cycle" (synonymous with Magnusson's "Svionian cycle") the term is extended backwards in geological time to cover also the supracrustal complex which was deformed during this orogeny. With the time concept thus introduced, it is natural that rocks in other regions which can be shown to have formed during the time interval in question, are also classified as Svecofennian. Finally, Svecofennian (as a noun) has been used in Finland also in a restricted sense, to cover the supracrustal complex of the original "Svecofennian belt" (Simonen 1960). Neither this term nor "the Leptite Formation" will be used here. The Svecofennian is avoided for reasons that ought to be clear from the foregoing, being used in the commonly accepted sense for the cycle. "The Leptite Formation" is not now accepted by Sveriges Geologiska Undersökning (the Geological Survey of Sweden), because this use of "formation" is contrary to general prin-

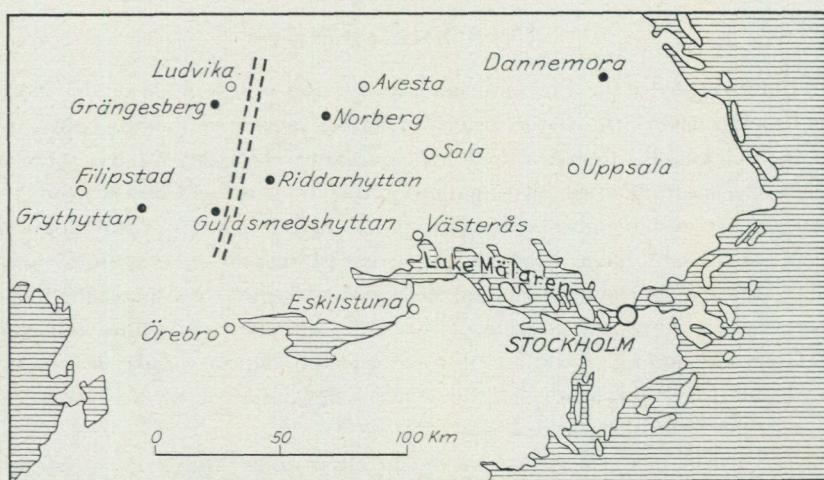


Fig. 1. Sketch map of part of Central Sweden, showing position of localities referred to in the text, and some cities for orientation. The double broken line marks the approximate position of the boundary between "west" and "east" in the later stages of the development of the supracrustal complex.

ciples in stratigraphic nomenclature. Introduction of a new term with only local significance may seem justified, but the writer prefers "the supracrustal complex" as a neutral designation. It is rather unwieldy, but this seems the lesser of several evils.

Names in use for subdivisions of the supracrustal complex of the region are the Grythyttan Series and the Larsbo-Mälars Series. These are sequences of detrital sediments which, in separate parts of the region, form the upper stratigraphic unit of the complex. Below these sediments is a continuous sequence of siliceous volcanics — leptites and hälleflintas — with interstratified limestones and iron ores but virtually without detrital sediments. This is the unit that made Sederholm speak of a Svionian type. To revive this term in its original meaning would, however, be impractical, in view of the greatly extended use that has later been made of it. "The Leptite Series" is a possible term, but it would have a more restricted meaning than the earlier "Leptite Formation", and might thus lead to misunderstandings. Therefore simply the colourless designation, "the Lower Series", will be used provisionally.

The Lower Series extends over the whole region, and in most of it — except the southeastern part — is the one most commonly represented in the present rock surface. The Larsbo-Mälars Series follows on top of it in the central and eastern parts of the region, and the Grythyttan Series occupies the same position in the west. Finally, the youngest member of the supracrustal complex, the Älvestorp conglomerate, has only a very restricted extension at present time, occurring locally in close association with the Grythyttan Series but separated from it by a marked unconformity.

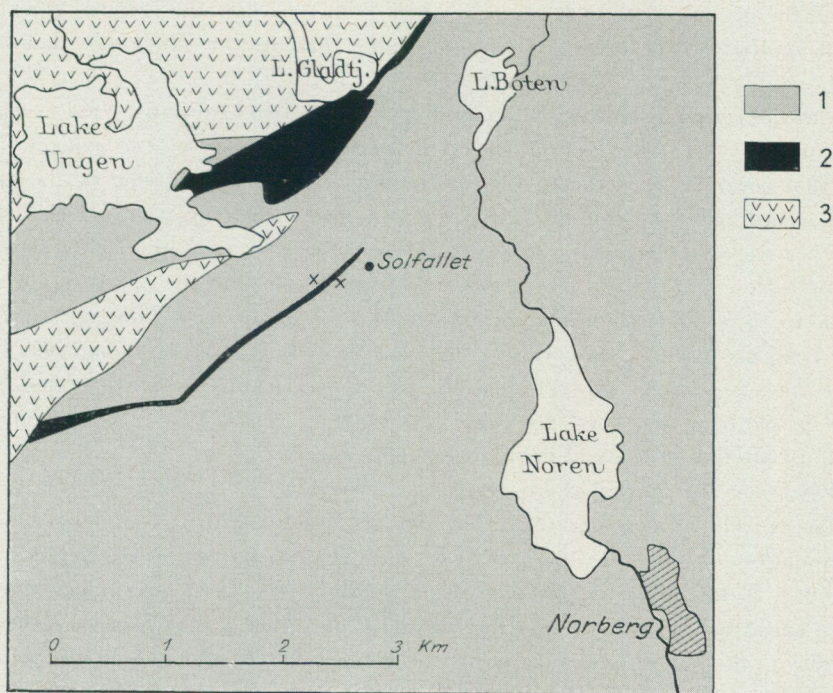


Fig. 2. Geological map of part of the Norberg district (adapted from Hjelmqvist 1946). 1 Leptites (incl. carbonate and iron ore beds); 2 Quartzite; 3 Granite. Between the two crosses is the road section with the iron-sand bed in the quartzite.

THE QUARTZITE

To the west and southwest of the Norberg district the supracrustal complex is chiefly represented by rocks of the Larsbo-Mälar Series, the stratigraphic importance of which was first realized in this tract (original designation: Larsbo Series). Through the work of Högbom (1929, 1930) and Hjelmqvist (1937, 1938, 1946) in this area it has become clear that the Norberg quartzite here forms the basal member of the series. West of Norberg, however, this basal unit is on the whole separated from the main Larsbo area by an intrusive granite (Fig. 2).

There are in Norberg two areas of quartzite (Fig. 2). One is a band, extending SW—NE in the general strike direction of the supracrustal complex. It has a very steep NW dip and a width varying from about 40 m or somewhat more, in the northeastern part, to more than 100 m in the southwest, where it is cut off by the granite. The other area is much larger. While the band appears to represent a single monoclinical sequence, this larger area is probably a complicated syncline. The dip of the bedding has been observed at several points along its southeastern margin and is there always vertical. The surface extension of this area, uninter-

raptured by any infolded units of other rocks, indicates that the original thickness of the quartzite here must have been greater than in the previously described band.

As to the relations between these quartzite areas, two alternatives are possible. They may represent two periods of sedimentation, separated by the volcanic leptites that occur between them (Fig. 2). In this case the band should be the older unit, according to the general stratigraphy in the Norberg district. Or they may have become separated through structural disturbances (now impossible to trace because of the scarcity of outcrops in critical areas), so that the band should correspond stratigraphically to the basal portion of the larger area. These alternatives have been discussed by the writer (1936, and in Geijer & Magnusson 1944), but no decision is possible. An occurrence of quartzite, certainly to be correlated with those in Norberg, has been described by Högbom (1929, 1930) from Kolpebo, about 10 km SW of the end of the band as shown in Fig. 2.

In Norberg the base of the quartzite is exposed at one locality in each of the two areas. The underlying rock is a potassic leptite with quartz phenocrysts. Examination of this rock immediately below the quartzite of the larger area proved it to be somewhat altered, the groundmass between the quartz phenocrysts consisting mainly of quartz and mica, thus indicating some weathering previous to the deposition of the quartzite. However, it is evident that this weathering extends only a little below the surface of the leptite, and the latter shows no disturbance or desintegration, the quartz phenocrysts clearly having kept their relative positions. Both localities give the impression that the quartz sand was deposited on a solid and even rock surface. Conditions are similar at Kolpebo, as described by Högbom (1929, 1930), with the difference, however, that the lower portion of the quartzite there contains some pebbles of vein quartz, up to a walnut's size, and a few small fragments of leptite, while nothing of the kind has ever been noted in the Norberg quartzite.

The quartzite in Norberg often shows bedding, with units down to some centimeters in thickness; distinct cross-bedding has also been observed. Though quartz is always the predominating constituent, other minerals, notably feldspar and mica, are usually present. The purest varieties contain about 94 per cent SiO_2 (Geijer 1936). The feldspars are sodic oligoclase and microcline, and are always in small amounts only. The mica is generally colourless in thin section, or slightly bluish, and has a small axial angle; apparently it is a phengite. Biotite is rather rare. Epidote, apatite, opaque ore minerals, zircon, titanite and calcite have been noted occasionally.

The texture of the quartz mass is crystalloblastic. The grains are somewhat elongated, thereby giving the rock a schistosity; most of the grains have a long dimension of about 0.3 mm. No clear indications about the original, detrital grain size have been observed in samples of the ordinary quartzite, but in the quartz-rich seams interstratified in the iron-sand bed the original size was gener-

ally about 0.2 mm (compare below, and Fig. 4). The texture of the ordinary quartzite makes it probable that the same applies to most of this material. The feldspars are, as a rule, in somewhat smaller grains. The mica, in small plates, mainly occurs between the quartz grains.

The iron-sand bed occurs in the quartzite band, near a place called Solfallet (Fig. 2). Here the quartzite was exposed when a road was straightened out and levelled, the rock forming a threshold, a little higher than the surrounding leptites. The road runs roughly E—W, making an angle of about 50° with the strike of the quartzite, which is exposed in the cut for a length of 34 m, with a very steep dip NW. The quartzite is here of a very light yellowish gray colour and in the fresh cut, from a little distance, looked like a sandstone. Cross-bedding on a small scale is sometimes marked by dark seams, one or a few millimeters in thickness, which consist of feldspar, micas, epidote and opaque grains, and also contain notably much apatite but no quartz.

THE IRON-SAND BED

Occurrence

The bed of iron-sand is well exposed in the southern wall of the cut, and stratigraphically at least 5 m above the base of the quartzite. Here its maximum thickness is 15 cm, including interstratified seams of quartzite. It has also been identified in the northern side of the cut but is only weakly developed there. A

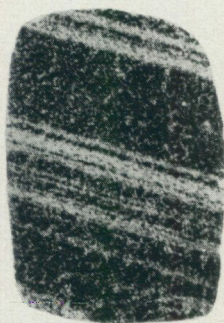


Fig. 3. Thin section from iron-sand bed, Solfallet, Norberg. Nat. size. Shows the internal stratification.

magnetic survey of the covered surroundings of the cut, with a Schmidt balance, was kindly undertaken by Dr. Sture Werner. It showed highest intensity around the exposure in the southern wall of the cut, for an extent of a few meters. A continuation can be traced southwestwards for about 20 m as a zone of very much weaker attraction. Northeastwards, there is instead a more abrupt fall in intensity, the lean continuation of the bed which has been exposed on the north-

Table 1
Iron-sand bed, Solfallet, Norberg. Analyst R. Blix

SiO ₂	24.36	0.4056
TiO ₂	6.96	0.0871
ZrO ₂	0.87	0.0071
Al ₂ O ₃	4.99	0.0490
Fe ₂ O ₃	38.18	0.2391
FeO	17.84	0.2483
V ₂ O ₃	0.07	0.0005
MnO	0.10	0.0014
MgO	0.50	0.0124
CaO	3.07	0.0547
Na ₂ O	0.91	0.0147
K ₂ O	0.44	0.0047
P ₂ O ₅	0.32	0.0023
H ₂ O < 105°	0.14	
H ₂ O > 105°	0.94	0.0522
	99.69	

ern side of the cut not being magnetically discernible. The geological interpretation of these data will be discussed under "conditions of deposition".

Chemical composition

A hand specimen representing the total thickness of the bed has been analysed by R. Blix, at the Mineralogical Department, Swedish Museum of Natural History. The results, together with the molecular proportions, are presented in Table 1.

Petrography

Stratification. The iron-sand bed exhibits a regular alternation of seams in which heavy minerals greatly predominate, and thin quartzitic ones (Fig. 3). The seams rich in heavy minerals which form the bulk of the bed, generally vary in thickness from one or a few mm to several cm; the quartzitic ones range from some tenths of one mm to about 2 mm. In the latter, heavy minerals are sometimes scattered in certain stratification planes.

Quartz in the quartzitic seams forms parallel elongated grains, mostly of a size about 0.20 to 0.30 by 0.10 to 0.20 mm. Between them is often a "packing" of scaly mica and feldspar (compare below). When a thin section is seen in ordinary light, one sometimes gets the impression that the quartz units represent somewhat deformed detrital grains (Fig. 4). In other cases, deformation has proceeded further, resulting in a relation of length to thickness about 3:1.

In the seams of heavy minerals, similar quartz grains may appear, but there are also much smaller grains which, at least in part, are clearly of secondary origin.

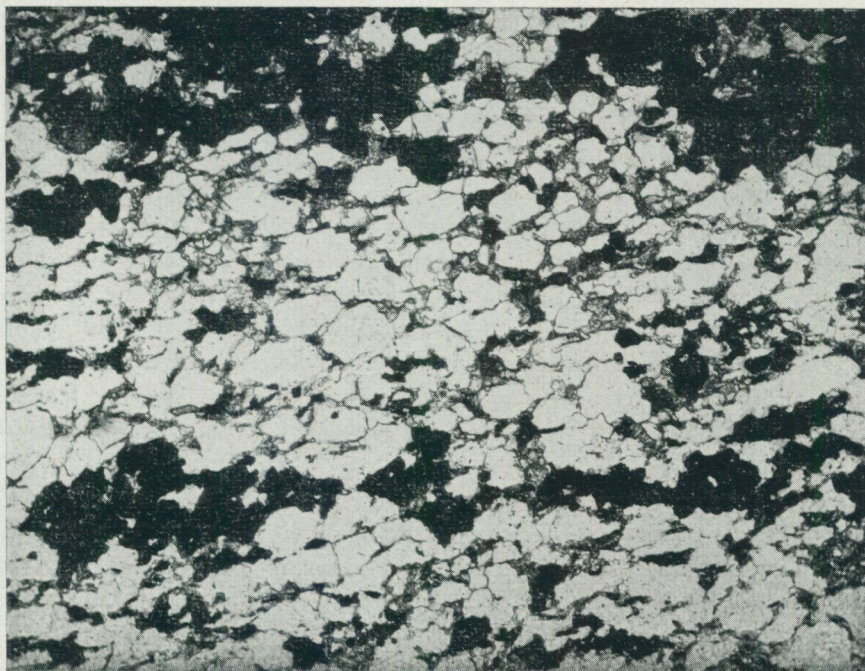


Fig. 4. Thin section from iron-sand bed, Solfallet, Norberg. $\times 33$. Shows texture of quartz and magnetite seams.

Feldspars and micas. In the quartzitic seams of the iron-sand bed, both the feldspars normally found in the quartzite — microcline and sodic oligoclase — have been identified in crushed samples. In all the thin sections examined, however, no feldspar has been determined. The interstices between the quartz grains are filled by colourless mica, mostly in small scales. Between them one can often distinguish an optically homogeneous substance with low refraction and low birefringence. This is probably a feldspar, but further identification is made impossible by the abundant mica scales. Mica of similar, sericite-like habit, and in larger plates is also noted here and there in the seams of heavy minerals.

Small scales of a greenish-brown biotite, often chloritised, are occasionally noted, both in the quartzose and in the iron-rich layers.

Epidote. Small grains of an epidote mineral with low birefringence (clinzoisite) occur with fine-grained titanite in one section. Reddish brown grains, weakly pleochroic, are often noted but their total quantity is very small. These epidote minerals evidently are new-formed in the rock.

Allanite (orthite). Metamict allanite is rare, but has been noted in several thin sections. It has a yellowish brown or greenish brown colour and forms elongated, well rounded grains, up to 0.21 mm in size. The shape makes it clear that these grains are detrital.

Tourmaline. Grains of blue tourmaline have been observed in a few cases. Since seams of new-formed tourmaline have been noted in another part of the same quartzite band (Hjelmqvist 1946), these probably are secondary.

Apatite is a common mineral in the layers of "heavies". From the analysis, it is estimated to make up about 0.75 per cent (by weight) of the bed. The shape generally is somewhat elongated, always in the plane of stratification. Grain size varies considerably, from about 0.02 to more than 0.30 mm. Most grains, however, are about 0.20 mm or more in length. The largest one observed measures 0.32 by 0.25 mm. In detail, the shape is rounded. Presumably these grains are detrital. However, a few of them have small outgrowths at the ends, in the direction of the *c* axis; these are probably secondary accretions. Only the magnetite grains sometimes encroach upon the apatite and exhibit crystal faces against it, but in other cases the textural relations are determined by the form of the apatite grain.

Zircon is the most interesting mineral of the iron-sand bed, although its quantity is not great — about 1.30 per cent according to the analysis. Every grain has kept its detrital shape practically intact, and secondary growths have not been noted.

When first looking over the thin sections, one gets the impression that there is represented a number of distinctly different types of zircon. Closer study shows, however, that types which look well defined when occurring as separate grains, are connected with each other by transitional forms and by zonal intergrowths. The number of types is thus reduced to two or three. One of them, rather variable in colour, birefringence, and zoning, is by far the most common.

This predominant type is mostly colourless in thin sections, or has a very light yellowish brown tinge, but a number of grains are brownish rose, wholly or in part. Most grains are full of irregular cracks and also of fine fissures. It is therefore difficult to ascertain when inclusions occur, and impossible to state their nature. One can only say that minute inclusions sometimes occur in great numbers. Birefringence is generally of a strength normal for unaltered zircon, but zoned crystals frequently have isotropic cores or shells, and in some cases the whole grain is isotropic or nearly so.

Zoning occurs in several different forms. Thus one may see a shell of isotropic substance in a crystal that otherwise exhibits normal birefringence. Sometimes a crystal is built up of a number of shells with sharply marked boundaries but without any appreciable differences in optical characters. In such cases, the shells sometimes are thicker on the pyramidal faces than in the prism zone.

Quite frequently, especially in the outer part of a crystal, zoning takes the form known as lamination, a succession of very thin shells. When viewed with one nicol only, such laminated portions often show a very light pinkish colour or a weak but quite distinct iridescence, apparently due to refraction ("Becke's bright line"). Generally, birefringent shells alternate with isotropic ones. In the

former, interference colours are often lower than normal. This, however, does not necessarily indicate a lower birefringence; it may instead be due to the thinness of the shells. Sometimes no isotropic shells are discernible. In such cases one might suspect that the break between the shells is of a physical nature only, as indicated by Claus (1936) for similar occurrences. But it may simply be that the isotropic shells are extremely thin. Often almost the whole crystal is laminated, a simple core making up only a very small portion of the grain.

An interesting case is that of a cross section that shows a square isotropic core and a very weakly birefringent mantle through which cracks radiate from the unfissured core.

That the brownish rose-coloured zircon and the more common, colourless to light yellowish brown forms are closely related is illustrated by several intergrowths, as a grain with a brownish rose core and a colourless mantle.

While this example illustrates an intergrowth of forms of different colour, there are also found grains with an intermediate tinge. Isotropism is at least as common in the brownish rose zircons as in the others, but many grains have a normal birefringence. A certain difference between the two colour variations, however, appears to be that the brownish rose zircons are less often fissured and carry fewer inclusions.

We must conclude, apparently, that these different colour variations of zircon may have formed together.

Another type is represented only by about half a dozen grains in the 16 thin sections, mostly irregularly rounded and slightly above average size. In thin sections the substance appears gray from abundant microfissures and small inclusions. Birefringence is low. In one case, a core of this nature is surrounded by a laminated mantle of isotropic and birefringent shells. This intergrowth links the type in question with those previously described, although it otherwise appears quite distinct from them.

Finally there occurs, rather sparingly, a type that is quite distinct from the rest of the zircon grains. It is a curious fact that this type is not rare in some sections, but entirely lacking in others. It is colourless, clear without microfissures and inclusions, but occasionally shows a crack or two. It differs, therefore, from the other zircons in a way that calls to mind the designation "water clear" or "limpid" that have been used by a number of investigators to describe a certain type of zircon. Birefringence is always high. No zonal growths or lamination have been observed.

Zircons separated from the crushed rock, when seen under the binocular microscope, generally appear white, grayish, or light yellowish brown. Darker grains, when broken, are translucent with a brownish purple colour.

The zircon grains normally separate from the rock without damage or are simply broken in two. Observations on this material, combined with the study of the thin sections, make it possible to reconstruct fairly well the original crystal

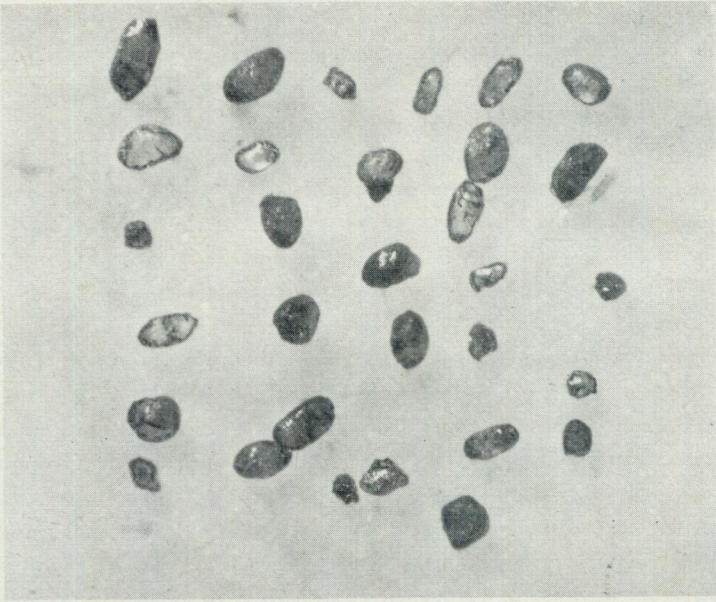


Fig. 5. Zircons isolated from iron-sand bed, Solfallet, Norberg, $\times 55$. Shows variations in original size, colour, and degree of abrasion. (A little magnetite or quartz from the matrix still adheres to some grains).

shape, although practically all grains are more or less abraded. So far as can be judged in this way, the original shape has, with very few exceptions, been that of the simple tetragonal prism combined with the unit pyramid. Other crystal faces may, of course, sometimes have been developed as quite subordinate forms but have been obliterated by even moderate abrasion. A steeper pyramid form has been identified with certainty in one case forming the end of a crystal. The crystals are always comparatively short. Measurements on a great number of grains indicate an average relation of length to width about 1.6:1. This, however, is a minimum figure, as abrasion always tends to reduce length more rapidly than width. To judge from comparatively well preserved grains, the ratio has generally been somewhat above 2:1, and in some cases has been about 3:1, or even slightly greater.

In their present shape the zircons, as already stated, always show signs of abrasion, but the degree varies from a moderate blunting of crystal edges (with the prism faces still brilliantly reflecting) to the production of almost spherical pellets. Most grains are strongly rounded but still retain an elongated shape. A figure for the degree of rounding has been obtained when measuring the dimensions of zircons as seen in the thin sections. Of 306 grains, not less than 151 were found to be approximately circular in outline. Many of them are cross sections and only indicate that the prism zone has been abraded to this form, but many

have a high birefringence, showing that the whole elongated crystal has been reduced to a spherical grain. The "water clear" type, except for one elongated splinter, always occurs in this form. Fig. 5 shows examples of the shapes of zircons separated from the rock, and that they are in all respects similar to those from recent beach sands.

The original size of the zircons is a factor of importance for the discussion of the source rock of the sediment. A fairly good idea has been obtained by measurements in thin sections, this method having been chosen for the comparison with zircon sizes in leptites (compare below). Obviously, in this way only minimum figures are obtained. In six sections, a number of 306 grains were measured, as already mentioned. The 151 with circular outlines range in diameter from 0.06 to 0.225 mm, with an average of 0.12 mm. In the other group, of 155 grains, elongated, or approximately square cross sections, length ranges from 0.06 to 0.42 mm with 103 of them lying between 0.15 and 0.22 mm; the average is 0.18 mm. Width varies from 0.03 to 0.20 mm, with an average of 0.11.

The largest of these grains, a lenticular one, measures 0.42 by 0.20 mm. It is clear, however, from certain grains in thin sections not included in this count, that zircons considerably surpassing the dimensions indicated have formed part of the source material of the rock. Thus a pyramidal fragment indicates a crystal probably a good deal over 0.5 mm in original length. Another fragment, in the shape of an equilateral triangle with a side length of 0.31 mm, is cut almost perpendicularly to the *c* axis and thus indicates a *shortest* dimension for the original crystal surpassing 0.31 mm.

Magnetite, ilmenite, titanite, and rutile. These minerals in the iron-sand bed are interrelated in such a way that it is practical to describe them together.

Of the four, magnetite is by far the most common. Even allowing for some uncertainty in calculating mineral composition from the analysis quoted above, it is clear that the percentage of magnetite is about 50 at least, and a comparable figure seems to be valid also for other sections of the bed. The proportions of ilmenite and titanite are difficult to estimate. Measurements under the microscope would give no reliable figures as the ilmenite largely occurs in very thin lamellae and the titanite is most unevenly distributed. Nor is an exact calculation from the analysis possible. A roughly approximate figure for the titanite can, however, be obtained from the amount of CaO, 3.07 per cent. Of this figure, 0.40 is accounted for by apatite, and an equal figure may reasonably be assumed for the sum of plagioclase and epidote minerals. One arrives then at the approximative figures, titanite 7 and ilmenite 7.4 per cent. The latter figure, however, leaves much too little FeO for the magnetite. This cannot be fully explained by assuming that rutile (which is being neglected) occurred in unusual quantities in this particular sample, and that all MgO (+ MnO) — but no Al₂O₃ — belongs to the magnetite. Possibly some Fe²⁺ in the magnetite became oxidized when the sample was ground for analysis.

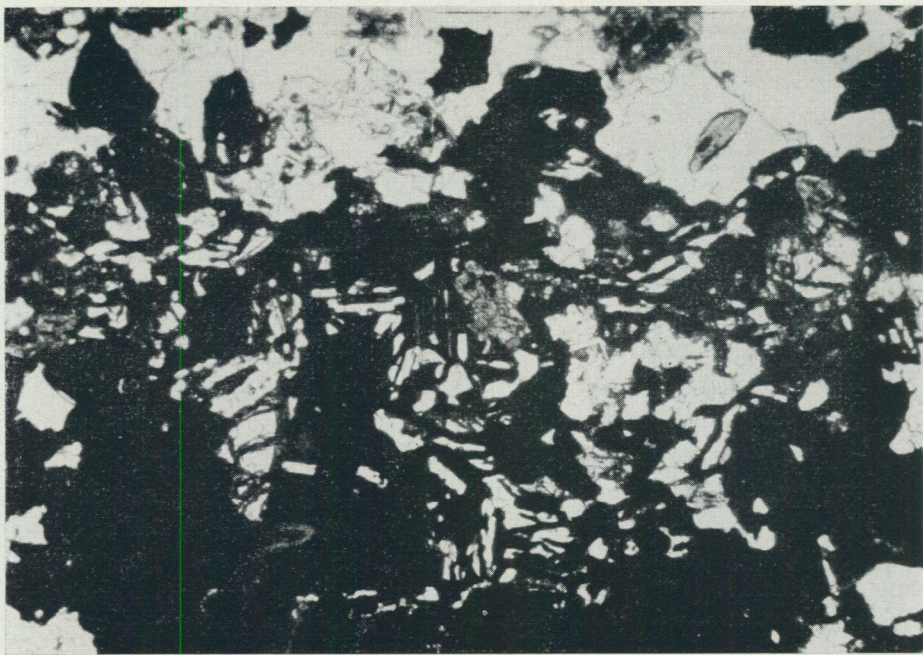


Fig. 6. Thin section from the iron sand bed, Solfallet, Norberg, $\times 75$. Black is magnetite and ilmenite, white quartz. The section shows the recrystallization forms of magnetite and the skeleton remains of detrital ilmenite grains filled by quartz.

The magnetite very frequently shows well-developed crystal outlines. There are no indications of detrital shapes. The crystals mostly range from 0.02 to 0.10 mm, occasionally a good deal more; they are often united into grain groups. Zircon is the only mineral that invariably exhibits unchanged detrital shapes towards the magnetite. Sometimes cracks in the latter radiate from an enclosed grain of zircon. A more or less clearly detrital form, although not so marked as in the case of zircon, is exhibited also by apatite and metamict orthite, and occasionally by quartz.

Ilmenite occurs in two different ways. One, representing at least half the amount of this mineral in the rock, is texturally characteristic and conspicuous. It is in the form of thin lamellae, arranged in systems that indicate a crystal unit, and with their interstices filled by other minerals. It is clear that these intergrowths represent partial pseudomorphs of ilmenite grains. The interstices between the ilmenite lamellae may be filled by titanite, and by quartz, and some mica. Rutile also occurs, although much less frequently than titanite and quartz; it has been observed to form a coating on the ilmenite lamellae, or sometimes to fill the whole space between them.

Sometimes the ilmenite forms only one set of parallel lamellae, and there are even cases where replacement in this pattern has affected only a part of the

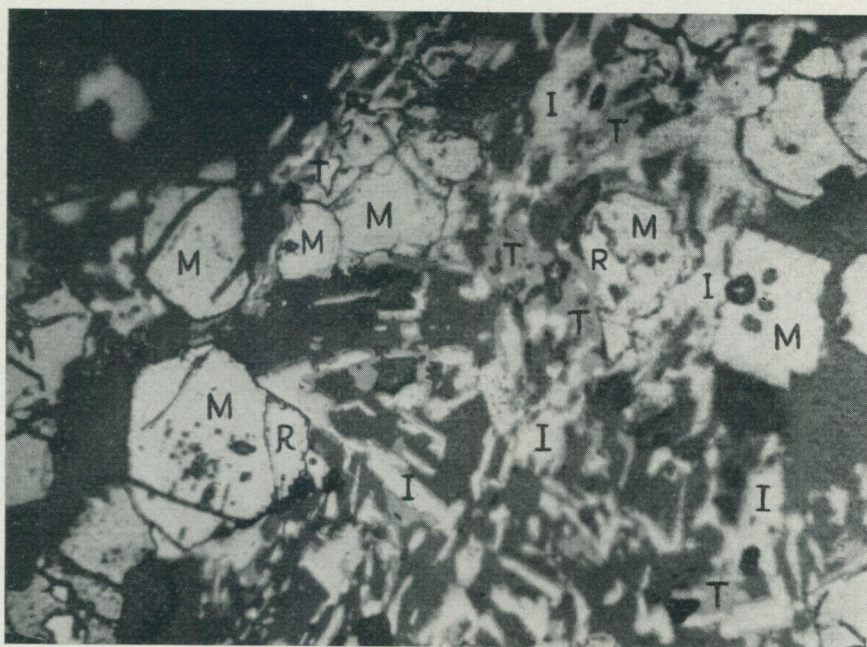


Fig. 7. Polished section, iron-sand bed, Solfallet, Norberg. \times ca 150. White is magnetite (M), ilmenite (I) and rutile (R); gray is titanite (T), black is quartz.

grain. Otherwise, where the new-formed product is only titanite, the relations are typically those of the replacement of ilmenite by leucoxene,¹ the alteration having proceeded along the rhombohedral or basal planes of the ilmenite, gradually thinning the partitions. The observed facts illustrate, for instance, Rosenbusch's (1905) description of the process: "Zeigt der Ilmenit den schaligen Bau nach (1011) oder (0001), so folgt die Pseudomorphose diesen Schalen und ersetzt von den Schalengrenzen aus mehr und mehr den Ilmenit, der auf schmale Blätter beschränkt wird und zuletzt gänzlich verschwindet."

What distinguishes the pseudomorphs in the iron-sand bed from the ordinary process is only that there has been much more movement of the dissolved substances. Fine-grained quartz with some plates of colourless mica has often largely or wholly taken the place of the titanite in filling the remaining skeleton structure of ilmenite. It is also characteristic that titanite is very unevenly distributed, practically lacking in some sections but plentiful in others.

Fig. 6 illustrates a typical aspect of the pseudomorphs when the interstices are filled with quartz, and Fig. 7 shows a more advanced stage in the alteration of an ilmenite grain, and several different alteration products.

¹ There is no need to discuss here the various minerals which are known to occur as the product called leucoxene. It is sufficient that titanite frequently occurs in this form.

The other form of ilmenite is directly intergrown with magnetite, but always only in a simple pattern. Occasionally it forms a band across a crystal of magnetite, or occurs as rounded or irregular inclusions in it. In other cases it may form a rim around a magnetite grain, the boundary between the two being either a crystal face, or irregular and frayed. Ilmenite in this peripheral position is often replaced by rutile, in part or wholly.

Rutile is very rarely seen except in close association with ilmenite of either mode of occurrence. No grain has been observed that may be interpreted as detrital.

It is evident that the magnetite (and the associated ilmenite) has crystallized in the rock. As regards the ilmenite pseudomorphs, they can only be interpreted as remnants of detrital grains of ilmenite. Their shape and size may perhaps have changed somewhat through partial collapse when the dissolved substance has been carried away without having its place immediately taken by new-formed minerals. But essentially the pseudomorphs must indicate the size of the original, detrital ilmenite grains, giving about 0.2 mm as a normal dimension.

Pyrite. Megascopically, pyrite has been noted as small crystals which in part line a fissure. A few grains have also been observed microscopically in the rock; in one case, the pyrite crystal occurs in the centre of a grain of zircon.

Absent minerals. It may be noted that such common heavy detrital minerals as rutile and garnet have not been observed. The fact that flakes of gold are occasionally observed in beach concentrations of heavy minerals caused an assay to be made, but the result was negative, the amount of gold, if any, being less than 0.1 p. p. m. (S. G. U. laboratories).

Metamorphism

The metamorphic state of the supracrustal complex in Norberg corresponds to the amphibolite facies. Before reaching this stage, however, the rocks may have undergone diagenetic and perhaps also other changes.

To diagenetic processes is probably to be ascribed the more or less extensive alteration that has befallen the original ilmenite grains. Their replacement, to the most varying extent, by titanite, fine-grained quartz, rutile and mica indicates the work of solutions moving in the still easily permeable rock. Some units show a collapsed structure, a sign that deposition did not always keep pace with solution. Also the irregular distribution of the titanite supplies an argument for referring the alteration to this stage. Furthermore, the ilmenite assumed to be of later crystallization (compare below) does not appear to have been affected by this alteration. The supposed accretions on apatite grains may perhaps also date from the diagenetic stage.

With regard to the later metamorphism, some mineral species are wholly unaffected, some are modified, and at least one is wholly changed. Conspicuously

unaffected are zircon, allanite, and apatite. Typically modified are the quartz grains, displaying a secondary, somewhat elongated shape while still retaining, for the most part, the individuality of the original, detrital grains. Wholly new-formed at this stage is the magnetite. It shows a euhedral habit towards most other minerals, including the ilmenite pseudomorphs, but never towards such minerals as zircon. There is nothing conclusive to show that the detrital material was in the form of magnetite, but this is decidedly the most probable alternative. The low but not negligible content of vanadium in the analysed sample also supports this interpretation.

The immediate association of the new-formed magnetite with some ilmenite suggests that the detrital magnetite was titaniferous and that on recrystallization, which presumably took place at a lower temperature than the original formation, the contained ilmenite — whether already exsolved or not — took up the peripheral and other positions in relation to the new-formed magnetite grain that are now observed. Such separation of these two minerals, under varying conditions of metamorphism, has been described by a number of authors (e. g. Ramdohr 1945, Vaasjoki 1947, Gjelsvik 1957).

Conditions of deposition

Evidently, we may envisage the environment at the formation of the iron-sand bed at Solfallet as a beach of rather fine, quartz-rich sand, which with a thickness of at least 5 m rested on an even rock floor. Concentrations of heavy minerals on a beach of this nature generally take the form of strips approximately parallel to the shore line and with a width that is only a small fraction of the length and a thickness with a similar relation to the width. From recent beach concentrations of "heavies" the following dimensions may be quoted. Wassmund (1938) reports from the coasts of Germany the figures: thickness 5—50 cm, width 5—20 m, length along the shore 200—2000 m (in one exceptional case even more). From Florida, Martens (1928) reports thicknesses of 2—2½ feet and widths of 25—35 feet. Other authors give comparable data. Thin laminae of quartz sand are often interstratified, as at Solfallet.

In view of the figures just quoted it seems probable that the strike length of the iron-sand bed at Solfallet does not much differ from the width of the deposit and that, in consequence, the longest dimension of the bed — and thus also the direction of the former shore line which it parallels — makes a rather large angle with the present surface. The magnetic data lead to a further paleogeographic inference which, though admittedly uncertain, deserves mentioning. With the present distribution of the Larsbo sediments (and disregarding the possible effects of folding and subsequent erosion) it might be natural to think of the west as the seaward side. However, the character of the iron-sand bed,

with its more gradual decrease in thickness westwards, appears to point in the opposite direction. Instructive in this respect is an excellent photograph showing a strip of iron-sand forming on a beach in the Taranaki district, New Zealand, and published in the *National Geographic Magazine* 1936 (Vol, 69, p. 194). This picture shows that the thick part of the strip has a distinct margin on the seaward side but inland is continued by a wide sheet formed by tongues of much thinner black sand. A photograph from a Florida beach, published by Martens (1928, Fig. 36), though less distinct in the details shows quite clearly the same pattern. To conclude from the magnetic map of the occurrence at Solfallet that the sea lay to the east is perhaps going too far, but the comparison with the recent examples quoted — both from regions known for remarkable beach concentrations of heavy minerals — is suggestive, and may at least serve as a warning against accepting the alternative.

Both Martens (1928) and Wassmund (1938) argue that the recent deposits studied by them are essentially the work of storm waves acting upon dune sands, in which, however, some other form of concentration may have already taken place. In Norberg, bedding in the quartzite, where observable, never seems to indicate that the rock is a dune deposit. This negative evidence is, however, not conclusive, and one must a priori allow for the possibility that such thick beach sands may have given rise to dunes above ordinary high-water mark, particularly as no binding vegetation can be assumed. Obviously, erosion of dunes will supply an abundance of sand for the waves to work over, and from the scarcity of heavy minerals in the Norberg quartzite it is clear that the iron-sand bed represents a highly effective sorting of a great quantity of sand.

It is worth noticing that the character of the iron-sand bed is quite different from that of the thin dark seams in the quartzite which were referred to above. This indicates a difference in origin, probably that between the action of the breakers of a violent storm, and the swash of the ordinary surf on the beach.

THE SOURCE OF THE DETRITAL MATERIAL

In the writer's earlier treatment of the Norberg quartzite (Geijer 1936) the opinion was expressed that a sediment of this nature could not be derived from the underlying volcanic suite (Lower Series), and that the source must be sought in a terrane of granitic nature. However, there appears to be, among those interested in the geology of the region, a tendency to assume that the sediments of the Larsbo-Mälar Series were essentially, if not wholly, derived from the volcanics of the Lower Series. In view of the important differences in the interpretation of the evolution of the supracrustal complex that a choice between these alternatives implies, it is necessary to consider the evidence more in detail.

If an accumulation of quartz grains is believed to derive from siliceous volcanics one must, of course, first assume that the quartz phenocrysts in the latter

have been set free from the surrounding matrix, which may happen in either of two ways. The grains may have been washed out from a fine-grained, unconsolidated pyroclastic deposit. Or chemical weathering of any kind of rock with such phenocrysts may have altered the rest to a clayey mass, the quartz phenocrysts being unaffected, as always under such conditions.

Products of the first-mentioned process are, in fact, known to occur in the Lower Series. Törnebohm's (1878) "quartzitic hälleflinta" at Dannemora is the best example (see also Geijer & Magnusson 1944). But these accumulations of quartz phenocrysts — largely still recognizable as such by their shape and by corrosion embayments — are local products without any quantitative significance. A rock of possibly similar origin, but occurring on a greater scale, has been described by Lundegårdh (1956) from the Uppsala map area. Though Lundegårdh's view that it may represent a pyroclastic deposit is a possible interpretation, the rock, with quartz grains in a felt of mica, is so peculiar, and so unlike the "quartzitic hälleflinta" of Dannemora, that it can hardly be reckoned with in discussing the origin of the Norberg quartzite. A very weighty argument against the derivation of the latter from unconsolidated pyroclastics in the amount of quartz involved. Since it seems clear that the quartzite occurrence at Kolpebo was originally continuous with that in Norberg, one has to reckon with a strike length of at least 15 km and an average original thickness certainly above 50 m. Sequences of siliceous volcanics have been studied in many parts of the world, and if ever a quartz sediment of such magnitude had been proved to have arisen through the washing out of quartz phenocrysts from unconsolidated pyroclastics, the fact would certainly have been widely quoted.

The quantitative aspect of the problem weighs heavily also against the other possible alternative, namely, derivation from a weathered terrane of quartz-porphyritic volcanics of the Lower Series. Very deep-going chemical weathering would be required to account for the volumes of quartz sand, and of such weathering no indications have been found in the region, apart from the superficial part of the substratum in Norberg, as described above.

A negative feature in the quartz grains of the quartzite also deserves mentioning, although it cannot be accepted as a proof against a phenocrystal origin — the complete absence of corrosion embayments. Such features often occur in the phenocrystal quartz grains of the Lower Series, anywhere in the region, and may be on the whole preserved even when metamorphism had led to considerable deformation of the grain (Geijer 1936), but no trace has been observed in the quartzite. This may, of course, be referred to abrasion, but certainly does not indicate any near source.

Assuming a land surface of a granitic nature as the source of the detritals, however, it follows that the material for the basal sediments of the Larsbo-Mälars Series, at least so far as the original Larsbo area is concerned, are of extraneous origin. This, together with the great amount of quartz in the later Larsbo-Mälars

sediments, makes it reasonable to assume such a foreign derivation for a very large part of the latter also. Products of the erosion of the Lower Series are, however, also identified. Thus local conglomerates, otherwise derived wholly from the immediately underlying rock, also contain some pebbles identifiable as volcanics of the Lower Group (Hjelmqvist 1937, 1938, 1946). And the conglomerate occurring south of Lake Väringen, about 15 km NNE of Örebro, may be in part made up of material of this nature, so far as one can judge from the brief description that is available (Gumælius 1873). Also part of the more fine-grained sediments may be derived from rocks of the Lower Series. On the whole, however, the Larsbo-Mälars sediments, at least north of Lake Mälaren, appear to consist predominantly of other material.

Of the minerals in the iron-sand bed, allanite and zircon are of most interest in this connection. Metamict allanite is a common, though always quantitatively insignificant, constituent of granitic rocks, but has not, to the writer's knowledge, ever been identified in the volcanics of the Lower Series. Nor is the unaltered mineral (orthite), which is locally encountered in this series, an original constituent, but stems from a time later than the deposition of the quartzite. Also the size of the detrital allanite grains points to a plutonic rock as the most probable source.

As to the zircon, interest centers on the grain size. Zircon is well known to be a widespread though always quite subordinate constituent of igneous rocks, particularly siliceous ones, and the sometimes expressed opinion that it should be rarer in extrusive rocks than in those of deep-seated consolidation is clearly irrelevant in weighing the alternatives discussed here. On the other hand, it is a priori probable that zircons will grow to greater dimensions in a plutonic rock, and observations appear to bear this out. Therefore it is of interest to consider the zircons of the iron-sand bed from this viewpoint, even if it is clear from the outset that no definite conclusion as to their provenance will be reached.

Measurements on zircons in the iron-sand bed have been given above (p. 15). While this material, because of the abraded state of the grains, does not permit any statistical calculation of the original size distribution, it seems clear that the majority have surpassed 0.2 mm in length, and some grains even 0.5 mm. For comparison, 300 thin sections of leptites and hälleflintas of the Lower Series from practically all parts of the region, have been studied. Most sections did not contain any zircon at all, or only a very small grain. Lengths (in the section) above 0.1 mm were noted in 21 grains, the two largest of which were slightly above 0.2 mm. Most noteworthy, however, is a grain, in leptite from an island in Lake Långban, seen in an oblique cross section, 0.15 by 0.15 mm; this grain has a typical lamellar structure.

In the writer's opinion, this comparison shows that derivation from the volcanics of the Lower Series is not excluded but that such a concentration of comparatively large zircon grains in the iron-sand bed would have required that

an enormous local supply of "raw material" was worked over by the waves. Certainly a derivation from a rock of (approximately) granitic nature is much more *probable*.

The occurrence of different types of zircon in the bed does not appear to have any bearing on the source problem. In fact, there are even many reports of two types, one of them often "water clear", occurring together in a granite body (Groves 1930, Brammall 1928, Boos 1935, Bruce & Jewitt 1936, Pabst 1938), and sometimes also a third type (Mackie 1928, Chatterjee 1929).

While thus the original minerals so far considered all point to a granitic rock surface as the most probable source area, this does not apply to the most common ones, magnetite and ilmenite. Certainly, both these minerals may be found in granites and related rocks, but already the proportion to zircon appears to make this alternative improbable. A more likely source are basic plutonics, which may, of course, have occurred together with granitic ones. The dacitic volcanics of the Lower Series also represent a possible source, but a less probable one in view of the evidence of the other constituent minerals.

THE LATER STAGES IN THE EVOLUTION OF THE SUPRACRUSTAL COMPLEX¹

To one who, as was the case of the writer when first encountering the Norberg quartzite, is familiar with the rocks of the Lower Series and the absence among them of any identifiable detrital sediments, the contrast between the underlying volcanics and this accumulation of quartz sand is most impressive, particularly as the quartzite appears to conform to the structure of the underlying unit. It is evident that the base of the quartzite marks an abrupt and important change in the development of the supracrustal complex, and the relations between the quartzite and the Lower Series become a most important problem in the study of the complex. To throw light on this problem it is convenient to consider first what can be concluded about the conditions under which the Lower Series was built up.

The very widespread occurrence of chemical sediments — limestone and iron ore — in the Lower Series makes it clear that practically every part of the region was, at least for some time, covered by water. Whether these surface waters were inland lakes or part of the sea cannot be determined. For the sake of simplicity they will here be spoken of as the sea.

The most remarkable indication of a continuous and extensive basin of deposition that existed during any stage of the evolution of the Lower Series is furnished by the thick deposit of limestone and dolomite, often containing manganiferous

¹ Many of the facts and their stratigraphic interpretations reported in this section have already been compiled in an earlier work by Magnusson and the writer (Geijer & Magnusson 1944). They are here repeated in order to present a complete picture.

iron ores (before metamorphism carbonatic), which in various parts of the region occurs near the top of the series (compare, for part of the region, Geijer & Magnusson 1944, Pl 1). It is found in the Viker syncline in the southwest and in the practically contiguous syncline at Guldsmédshyttan (Fig. 1) NE of the latter, and in the Norberg district, the total distance being about 100 km. It reappears in Garpenberg, about 30 km NNE of Norberg. The gaps in the chain are long, even if thinner beds of the same nature are taken into account, but may be explained by axial culminations and by later granitic intrusions. Furthermore stratigraphic identification with a bed E of Filipstad is practically certain (Geijer & Magnusson, 1. c.). And at Dannemora, far to the east (Fig. 1), a virtually identical development is met with at what is, in all probability, approximately the same stratigraphic level.

The distribution of quartz-banded iron ores also indicates deposition in extensive basins, even if not of any comparable size. The belt with such deposits in Norberg has a continuous strike length of about 10 km, and at Riddarhyttan, 30 km southwestwards in the strike direction, analogous ores are encountered. A particularly noteworthy feature is that in both districts the lowermost beds, in contrast to those in higher stratigraphic positions, contain also a certain amount of manganese. Another example is a belt rich in such sedimentary iron ores which extends NE of Ludvika for about 20 km (compare Geijer & Magnusson 1944, Pl. 56).

It is thus reasonable to assume that most of the area, perhaps all of it, was at least for considerable periods covered by the sea. From the carbonate beds it seems probable that the waters thus occupying the region formed a rather shallow shelf sea. The origin of these carbonate beds — limestone, and dolomite of possible primary nature — has long puzzled geologists working in the region, because these sediments are interstratified with siliceous volcanics low in calcium, and without any complementary deposit derived from a land surface subjected to weathering. Several authors, including the present writer, have discussed a possible contribution of carbonates to the sea by thermal waters. It has also been realized, however, that reference to land deposits from such springs (travertine) does not give a satisfactory answer, since all known occurrences of this nature appear to derive from limestone beds in the underground which are being dissolved by the thermal waters. Comparison with travertine also ignores the fact that conditions of deposition on a land surface, where the water spreads out as a thin sheet and easily loses CO_2 , are quite different from those resulting when the springs issue at the bottom of the sea.

However, Sundius (1923) long ago pointed out biological processes as a possible explanation of the deposition. And whatever the source of the carbonate may be, it now appears most probable that its precipitation was due to biogenic factors. This was first indicated some years ago, when absolute datings showed that there was probably no essential difference in age between the Svecofennian

and the Karelian, in eastern Fennoscandia. For the dolomite beds of the Karelian, in both Finland and Sweden, frequently contain algal reefs, and the same possibilities for plant life must be assumed to have existed during the building-up of the supracrustal complex of Central Sweden. Algal reefs are very rare there — the writer has looked in vain for such features in a great number of carbonate beds — but clear examples have recently been discovered in the large dolomite body at Sala (T. Eriksson and B. Collini, personal comm.).

This scarcity of direct evidence can hardly be regarded as an argument against accepting biogenic precipitation as the most probable interpretation for the carbonate beds in the supracrustals of Central Sweden. Even in most of the Karelian occurrences, reef structures can be discerned only in part of the beds, and it is reasonable to assume that the other parts were due to a geochemical process of the same kind. In Central Sweden one has also to reckon with recrystallization which has usually been fairly strong, and which might easily have effaced such structures.

If one thus assumes that the precipitation of carbonates has been due to the life processes of algae, it follows that the region at the time in question formed a shallow shelf sea, since such processes require a certain amount of light and therefore cannot proceed at a greater depth than about 200 m at most. This, in turn, would lead to the conclusion that the bottom of the sea was gradually sinking during the building-up of the Lower Series. In view of the assumed shallowness of the sea it is, however, noteworthy that there has not been observed any evidence, in the form of shore conglomerates or the like, for the existence of volcanic islands during this period.

The volcanics rarely give evidence of either a submarine or a supramarine origin. Only the spilites form an exception, and these are rare in the Lower Series, and occur there only at what is believed to be a rather high stratigraphic level. The calcite-impregnated scoriaceous spilites certainly have formed under water, being in their original development wholly analogous to lavas intercalated in the Devonian sediments of Devon and western Germany. They tell nothing about the depth, however, except that it must have been less than 2000 m (compare Rittman 1960); in fact, such rocks in the Lahn-Dill region of Germany are assumed to have been extruded in a very shallow sea (quoted from Schneiderhöhn 1941). Additional evidence of the subaqueous origin of our spilites is reported by Koark (1962), who near Falun found traces of pillow structure. For the siliceous rocks, on the other hand, no reliable criteria are known. Harker (1889) in his description of the Bala volcanics in Wales discussed this lack, and little progress has been made since his time. The inferred extent and evenness of units in the Lower Series has always been difficult to explain because of the high viscosity of such products, but the recent concept of submarine ignimbrites ("pyroturbidites", Fitch 1966) may present a plausible interpretation.

The arguments presented above for a submarine origin of the rocks of the Lower Series do not, as a rule, apply to its uppermost section in the central and eastern parts of the region. For this section, with dacitic and andesitic types, has no intercalations of carbonate beds or iron ores. However, in Garpenberg beds of scoriaceous, originally calcite-drenched spilites, testifying to submarine conditions, are associated with the above types (Geijer & Magnusson 1944, Pl. 35).

A general idea of the frequency of volcanic activity in the region can be obtained from the relations between the chemical sediments and the volcanic products. Both limestone beds and quartz-banded iron ores often alternate with beds belonging to the latter. Thus, for instance, at Riddarhyttan there is often an intimate alternation between such iron ore and an even-grained potassic leptite which is, chemically, almost identical with a massive, quartz-porphyritic type in the neighbourhood (Geijer 1923) and which may therefore be assumed to represent ash falls, or possibly an intermittent influx of material swept in from some area of unconsolidated ash. But there are also thicker ore beds without any such interruptions, and in the analogous area in Norberg (compare p. 24) pure ore beds may reach at least 10 m in original thickness; similar figures for iron ores are also found elsewhere in the region. Carbonate beds may show even greater thicknesses; at least in the Viker-Norberg zone mentioned above, original uninterrupted thicknesses of 50 m or more are present. For both ores and carbonates the maximum original figures are impossible to estimate because of deformation, and the figures cited may well fall far short of the actual maxima. The rate of deposition of the chemical sediments is unknown, but clearly the above figures are inconsistent with the idea of an almost continuous volcanic activity in the region that one easily forms subconsciously when working in this predominantly volcanic terrane.

Turning then to conditions at the base of the quartzite in Norberg, it is clear that the even surface of potassic leptite on which the quartzite rests cannot be the original, normally rough surface of a lava flow of this composition, nor can it be the top of an unconsolidated pyroclastic deposit. Evidently, between the formation of this substratum and the beginning sedimentation of sand upon it, there was a period during which some weathering took place, indicating land conditions, and also a certain amount of abrasion. The fact that products of this erosion are so very inconspicuous is a sign that the amount removed has been small. As to the time occupied by this interval, no evidence is available. It may have been short, not longer perhaps than the pauses in volcanic activity, evidence of which is the unbroken sequence of chemical precipitates, as shown above, or it may have been of quite another magnitude.

As to the contact relations at the base of the Larsbo-Mälars Series in the southern part of its extension — broadly speaking, south of Lake Mälaren — nothing definite is known at present, in spite of the important work done in

recent years by several investigators. In some parts, the basal sediments seem to be similar to the quartzite at Norberg and Kolpebo. This appears to be indicated by the fact that Lundegårdh (1959), in the Eskilstuna map area, found a quartzite bed to occur approximately at the base of the series. Generally in these parts, however, metamorphism and intrusions make it difficult to discern features so easily masked as those observed at the contact in Norberg and at Kolpebo.

In the western part of the region, where the upper unit of the supracrustal complex (apart from the local Älvestorp conglomerate) is represented by the Grythyttan Series, a slight break can also be discerned in the corresponding position (compare Geijer & Magnusson 1944, and literature quoted there). However, at some occurrences of Grythyttan argillite, as in the syncline at Guldsmedshyttan (Geijer & Magnusson 1944, p. 275 and Pl. 17), there is no such break at the base, but instead a zone of interstratification. A notable difference between, for example, Norberg and the Grythyttan and Guldsmedshyttan districts is that in the latter the thickness of the supracrustals between the heavy carbonate bed repeatedly referred to, and the base of the detritals, is much less than in Norberg, where also the metadacitic volcanics occur in this position, an element that is lacking in the west.

Just as there are no proofs of any important erosional or structural break at the base of the Larsbo-Mälars Series, this level does not appear to mark any abrupt change in the volcanic activity, at least not in its frequency. Certainly, volcanics greatly predominate in the Lower Series but form only a subordinate element in the Larsbo-Mälars Series of the original Larsbo area. It is necessary to remember, however, when comparing the amounts of volcanic and sedimentary material, that the rate of accumulation of such detrital sediments that make up the Larsbo-Mälars Series in these parts must have been many times greater than the deposition of the chemically precipitated sediments of the Lower Series.

In the chemical character of the volcanic products, however, there is a difference between the original Larsbo area and the neighbouring occurrences of the Lower Series, though a tendency to a similar development is evident already in the uppermost part of the latter. Thus in the Norberg district, as is general in the central and eastern parts of the region, this section is not made up exclusively of metarhyolitic types, as is the case further west, for instance in the Grythyttan and Filipstad districts. Instead metadacites occur, together with generally potassic metarhyolites, in comparable amounts. In the Larsbo area, west of Norberg, the sediments contain subordinate pyroclastic beds of dacitic, or, less commonly, basaltic nature (Hjelmqvist 1937, 1938, 1946). As Hjelmqvist has pointed out, the metadacitic beds are closely similar in composition to a metadacitic lava of the Lower Series at Riddarhyttan. But the potassic metarhyolites of the latter group have not been identified with certainty as intercalations among the Larsbo sediments. The map of the Norberg district (Geijer

1936, compare also Fig. 2 here) shows that metarhyolites occur between the two bodies of quartzite, but, as stated above, this may be due to structural disturbances.¹

As to the eastern part of Central Sweden north of Lake Mälaren, towards the coast of the Baltic and the Åland Sea, where metadacitic and meta-andesitic types are conspicuous among the volcanics, it was until recently uncertain whether any such occurrences belonged to the Larsbo-Mälär Series. Within the last few years, however, a detailed study by Lundqvist (1962) in the northern part of the archipelago, and Stålhös' (1964) mapping of a larger area around Stockholm, have furnished important information. It appears from these investigations that volcanic activity continued in these parts during the building up of the Larsbo-Mälär Series, with an intensity hardly less than in the time of the Lower Series, and that its products ranged from basic types through andesitic and dacitic ones to rhyolites. The metadacites apparently correspond chemically to those of the Lower Series, and to the pyroclastic beds of the Larsbo area. The rhyolitic members are generally characterized by the combination of microcline and oligoclase, in greatly varying proportions. They include types in which microcline predominates so much over oligoclase as to make the rock fully comparable to types among the potassic leptites of the Lower Series; a good example is Stålhös' sample no. 170 (Stålhös 1964, p. 14). Counterparts to the remainder are to be found in the Lower Series, presumably always in its upper portion. This applies even to those types in which the amount of microcline is insignificant, and which ought thus to be designated oligoclase leptites.¹ Examples of typical oligoclase leptites may, from the writer's own experience, be quoted from the Öster-Silvberg district (Geijer 1965) and the Stripa mine (Geijer 1938). In both cases they are associated with potassic types, while rocks with intermediate but varying proportions between microcline and oligoclase are common near Falun (Geijer 1917).

It is of interest to compare conditions in the west, as in the Grythyttan and Filipstad districts, where the volcanics of the Lower Series are overlain by the sediments of the Grythyttan Series, and sediments of the character of the Larsbo-Mälär Series are not present, with what has here been reviewed about the central and eastern parts of the region. In the west, the uppermost portion of the

¹ Alternation of a red leptite with quartz phenocrysts and a mica schist of the Larsbo Series has been reported from the shores of Lake Norn, about 20 km NNW of the quartzite locality at Solfallet (Hjelmqvist 1946). However, at one of the localities mentioned, the landing on the eastern shore, examination by Prof. F. E. Wickman and the writer showed only leptite, with narrow, strongly schistose mica-rich zones. According to Prof. S. Hjelmqvist (personal comm.) the same may be the case with the other locality cited, the "mica schist" of the field assistant's observations referring to such zones and not to mica schist of sedimentary origin.

¹ Some authors (occasionally also the present writer!) have described such rocks from the Lower Series as "soda leptites". This use of the term ought to be avoided, since it was originally introduced for rocks consisting almost wholly of albite and quartz (thus, in another terminology, quartz-keratophyres). To speak of the oligoclase leptites as "soda leptites" gives a wrong picture of the chemical evolution of the volcanism, and of the stratigraphy of the Lower Series.

Lower Series, below the Grythyttan sediments, is made up almost exclusively of potassic metarhyolites (which in the Grythyttan district are still in the hällflinta stage, almost unaffected by metamorphism). The dacitic and andesitic types known from the more easterly parts of the region to occur at this general stratigraphic level are absent.¹ From these hällflintas next below the Grythyttan sediments, Sundius (1923) has reported the occurrence of interstratified seams with a composition indicating a partial derivation from argillaceous products. Especially noteworthy is the occurrence of graphite in such seams.

Above these hällflintas follows the graywacke zone, built up chiefly of unsorted detrital material from the underlying volcanics but containing also fragments of graphitic argillite, similar to the rock forming the lower section of the Grythyttan argillite (compare below), and thus showing that sediments of this nature were already present in the neighbourhood. In this graywacke zone, however, interstratified tuffs with a glass shard texture and a composition corresponding to a rhyolite with an intermediate proportion of potash and soda feldspar also occur (Sundius 1923). At other places, beds of spilite occur in the graywacke zone (Magnusson 1925). On top of this zone follows the "Grythyttan slate", a thick sequence of mainly argillitic sediments, the lower part of which is regularly graphitic and contains some beds of scoriaceous spilite. The slate is otherwise wholly devoid of any volcanic elements.

Sundius (1923) and Magnusson (1925) both state that they have not been able to find any direct proofs of an unconformity in the stratigraphic sequence, but also point out that the nature and distribution of the detrital Grythyttan sediments indicate that changes of level must have taken place in the region during the time covered by the rocks in question. As seen from the foregoing, this is also what can be said of relations between the Larsbo-Mälär Series and the Lower Series. There is, however, the very marked difference that in the western districts the earliest detrital deposit — the graywacke zone — is derived locally from the underlying supracrustals, while in the case of the Larsbo-Mälär Series the basal member clearly stems from outside the region.

As to the volcanism, it ceases rather abruptly in the Grythyttan Series, siliceous products being represented only by the ash tuff in the basal graywacke, while spilites occur also higher up in the sequence, but not above the lower part of the main, argillitic member of the series.

The relative age of the Grythyttan and Larsbo-Mälär Series — a most important point in the development of the supracrustal complex — cannot be decided directly, as they occupy different parts of the region (compare survey map in Geijer & Magnusson 1944, Pl. 2, and also Fig. 1 here). The synclines in which the Grythyttan Series is preserved are spread over the western part,

¹ Such rocks are found locally in the west, viz., at Grängesberg, but there they belong, according to Magnusson (in Geijer & Magnusson 1944), to the lowest known part of the Lower Series, which determination places them in a unique stratigraphic position.

within a west-east width of about 50 km. The shortest distance between any such occurrence and rocks of the Larsbo-Mälär Series appears to be east of the Stollberg syncline, about 2.5 km E of Ludvika (Fig. 1). This distance is about 11 km, much of which is occupied by an intrusive granite (Hjelmqvist 1937). The distance was, of course, originally greater, before folding and resultant shortening, but the influence of the granite cannot be estimated.¹

One has, in fact, strong reasons for believing that the existence of two mutually exclusive sedimentary series is an original feature, or, in other words, that their areas never overlapped. For it seems clear that the Larsbo-Mälär Series never extended west of the present boundary, while the complete absence, in this series, of any material derived from the Grythyttan Series appears significant. There also seems to be a tendency for the sedimentation of the last-mentioned series to decrease eastwards, i. e. towards the boundary.

The opinion has been expressed (Magnusson 1963) that the Larsbo-Mälär Series should be the younger one, and that it was deposited during a more advanced stage of the Svecofennian folding than the Grythyttan Series, the sediments of which were regarded as deposited in local synclines. This opinion is based on a much wider experience of both series than any other geologist possesses. Yet it is difficult to accept the arguments cited in support of it, concerning chiefly the nature of the sediments, as definite proof of the supposed relationship. One must agree, however, that the reverse age relations appear to be excluded. Therefore the alternative would be that the two groups are of broadly contemporaneous deposition, representing different facies of one sedimentary formation. This age relation cannot be proved, any more than the one assumed by Magnusson can, nor would the writer claim for it any greater probability than for the latter. But it is desirable to keep the question open.

A review of a number of features in the supracrustal complex, some of them hitherto little discussed, will define the problem clearer and also appears to lead to a somewhat increased understanding of the evolution of the complex and of the initial stages of the Svecofennian folding.

One such feature concerns the stage at which a contrast between "west" and "east" in the evolution of the complex first became discernible (Fig. 1). Of fundamental importance for the study of the Lower Series has been the stratigraphy of the volcanic pile in the west, as demonstrated by Sundius (1923) and Magnusson (1925), with a lower section made up of sodic metarhyolites and an

¹ Possibly, rocks of the two series come even closer in an area about 10–15 km NNE of Örebro (Fig. 1). In addition to typical Larsbo-Mälär sediments, including conglomerate (compare above), Törnebohm (1883) reported a stripe of dark schist or slate, which he compared to the Grythyttan argillite. The writer has not had opportunity to visit the locality. It is not clear from Törnebohm's paper, however, whether this rock really is to correlate with the Grythyttan argillite, as he compares it also to an occurrence E of Rosersberg, halfway between Stockholm and Uppsala, and this one, which the writer knows from a personal visit, is a typical representative of the Larsbo-Mälär Series. It appears that Törnebohm only wished to emphasize the common detrital-sedimentary origin of the various occurrences.

upper one dominated by potassic metarhyolites. This stratigraphy is probably valid, at least broadly, for the whole region. In any case it can be traced eastwards, quite independently of the boundary which is so marked in the later development of the complex. It gives, for example, a reasonable interpretation of relations in the Norberg district (Geijer 1936). That there are wide areas where it cannot be proved is irrelevant, since these are generally ones where the sodic types are not observed, presumably because they lie below the present erosion surface. An illustrative example may be cited from Riddarhyttan, where potassic types are widespread and sodic leptite is represented only by a small patch — which, however, is found in the interior of a large anticline the rocks of which, with this exception, have been wholly altered to a cordierite-rich mica schist (Geijer 1923). And further eastwards, at Dannemora, the distribution of potassic hällflintas, associated with the large carbonate occurrence and its manganiferous iron ores, and of typical sodic hällflintas in the neighbourhood, is fully consistent with the assumption of a similar stratigraphy, even though scarcity of outcrops makes proof of it impossible to obtain.

It is stratigraphically a little above the characteristic carbonate bed and its ores that the contrast between west and east sets in. In the east, best demonstrable in Norberg and Garpenberg, then follows the volcanism with largely dacitic products, no counterpart to which can be traced in the west. The stratigraphic level at which this turn in the nature of the volcanism occurs, approximately coincides with the base of the Grythyttan Series in the west, or with the period of erosion which there preceded the deposition of the basal graywacke. No traces of dacitic ash material is reported from the Grythyttan sediments that can be correlated with this volcanism, so comparatively near to the east. On the other hand, as already stated, dacitic eruptions continued in the east, up into the Larsbo-Mälars Series, apparently unaffected by the changed conditions that led to the building up of the sedimentary sequence.

It is of interest in this connection to consider also the Guldsmedshyttan syncline (for position, see Fig. 1). Here, in marked contrast to the Norberg part of the same belt (compare p. 24), the Grythyttan argillite follows at approximately the stratigraphic level discussed. Thus no dacitic or related volcanics occur between this sediment and the thick carbonate bed of the underlying Lower Series, only metarhyolitic material. But there is, on the other hand, no break, corresponding to conditions at the base of the Grythyttan Series elsewhere, for the graywacke zone is lacking and the argillite is conformably associated with the underlying leptites by a zone of interstratification. This conformity appears to support the view, expressed by both Sundius and Magnusson, that the break found at the base of the Grythyttan Series in those parts where they have studied it, is of comparatively little importance.

The conditions under which the sediments of the two different series were deposited, and the first aspects of the crustal movements which were to cul-

minate in the Svecofennian orogeny, present interdependent problems. For the west, Sundius and Magnusson have assumed that deposition of the graywacke took place in depressions that were developing locally, a view that explains both the clearly local derivation of these sediments and their sometimes greatly varying thickness. But when it comes to the chief member of the series, the argillite, the situation is different. This deposit cannot be assumed to derive, at least not essentially, from the neighbourhood of the areas where it is now preserved. Its nature indicates instead a great influx of detrital products, largely clayey, from outside the region. This again requires the existence of a much more extensive basin of deposition, the deeper portions of which may well have been those parts where downfolding had begun and made possible the accumulation of graywacke. It is not, of course, necessary to conclude that the whole western part of the region was then covered by a continuous sheet of water, but the picture of a number of local, restricted basins cannot be applied to this stage. That the argillite must, in any case, have covered considerable areas around the now remaining synclines is evident also from the thickness of the deposit in the latter. With figures that may, according to the data reported by Sundius, have reached 1000 m, and certainly cannot have been far less, an abrupt decline down to zero cannot be assumed. The character of the locally overlying Älvestorp conglomerate also furnishes proof of the once much wider extension of the argillite. At the same time, however, it indicates that the thickness of the latter was probably much less even at a comparatively short distance from the present areas, thus supporting the view that these synclines represent the chief centres of deposition. To illustrate this, it is necessary to consider briefly the characters of the Älvestorp conglomerate, as described by Sundius (1923).

The conglomerate is found only in the two synclines near Grythyttan (which are the type locality of the Grythyttan Series). It rests on the older supracrustals, almost everywhere on the Grythyttan sediments, with a very distinct unconformity. Its material is wholly derived from the local supracrustals, about 70 to 80 per cent being made up of Grythyttan argillite and the rest including, although only with a small percentage, even pebbles derived from comparatively low levels in the volcanics of the Lower Series. The thickness has not been ascertained but is certainly much less than that of the Grythyttan Series. The synclines of the latter were already developed, and the conglomerate was deposited on their eroded surface.

The character of the conglomerate points at rapid erosion, the products of which were transported to the basin of deposition by streams with a steep gradient. Obviously, the argillite pebbles cannot be derived from the basin where they have come to rest, but from adjacent high land areas which were covered by the same rock. On the other hand, if this mantle of Grythyttan sediments outside the preserved synclinal areas is assumed to have had the same thickness as in the latter, the amount of erosion becomes so great as to make it

improbable. Even without this assumption, however, the erosion must have removed, at least locally, thicknesses of hundreds of meters. The topography of which the conglomerate bears witness, is consistent with the idea that the source of the pebbles was in areas elevated by folding and complementary to the Grythyttan synclines. However, after deposition the conglomerate, too, was folded, and, as is evident from Sundius' studies, its deformation fits well into the pattern of the earlier folding.

The sequence of events, accordingly, can be assumed to have been as follows. First, depressions began to form, marking the first stage of the crustal movements which were to reach their climax in the Svecofennian orogeny. There had been some denudation of the Lower Series before, but apparently not on any important scale. In the basins formed, local material was deposited, forming the graywacke zone. Further depression of the area led to its being covered to a great extent by a continuous body of water, into which finer, largely clayey erosion products were brought in from more distant sources. Probably, however, the original depressions still continued to form the deepest portions of this more extensive area of sedimentation. Next came a compression by folding, which probably followed essentially the same pattern as the development of the original basins. The folding not only further depressed the floor of the basins but also elevated ridges that rose high above the water surface. This may be designated as the Grythyttan phase of the folding. Then came the period of intense erosion of the raised areas, and the deposition of the Älvestorp conglomerate. It is probable that folding was then still in progress, but apparently erosion prevailed for a time over uplift. And finally further folding took place, which affected also the Älvestorp conglomerate — the Älvestorp folding phase.

The tectonic pattern of the preserved areas of Grythyttan sediments suggests that folding increased somewhat in intensity from west to east. But east of the Älvestorp occurrences near Grythyttan it is impossible to trace the relations between sedimentation, folding and erosion that have been discerned at Grythyttan.

Turning to the province of the Larsbo-Mälars Series one encounters a different development, but also, unfortunately, conditions much less favourable for tracing the sequence of geological events. Our knowledge of relations at the base of the sediments is admittedly limited, but it does not indicate any important erosional or structural unconformity against the Lower Series, not more, anyway, than can be discerned at the base of the Grythyttan Series. There is no sign that the development of the basin of sedimentation began with the forming of local depressions, as was the case in the west. Instead, such a sediment as the Norberg quartzite, representing a great amount of detrital material from foreign sources, suggests deposition in an extensive body of water. The same is indicated by the iron-sand bed, since the storm wave action which is inferred to have produced it requires a long "fetch" and, in consequence, a wide sheet of open water.

Doubts have been expressed above that the bulk of the Larsbo-Mälär sediments was derived from areas of the supracrustals, upfolded during this sedimentation. In support of this attitude may be cited not only the great amounts of quartz in the sediments but also the small part played in them by coarse products such as conglomerates. A comparison with the Älvestorps conglomerate is suggestive. It appears more probable, then, that the intense folding of the Larsbo-Mälär Series did not begin until later, when sedimentation in the geosyncline was approaching its close.

The above sketch of the evolution of the supracrustal complex is based on facts, compiled for the most part from the important works of other investigators. But where the author's view differs on any important point from the opinion expressed by other writers on the subject, it is only in one case that it can claim to be proved — the largely extraneous derivation of the Larsbo-Mälär sediments. For the rest it is a chain of tentative conclusions and alternative interpretations which, in the opinion of the writer, are sufficiently well-grounded to deserve serious consideration in future study of the complex.

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