BASIC IGNEOUS COMPLEX, MANGERITE, AND HIGH GRADE GNEISSES OF FLAKSTADÖY, LOFOTEN, NORTHERN NORWAY: I. FIELD RELATIONS AND SPECULATIONS ON ORIGINS

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The rocks of Flakstadöy comprise a layered complex of gabbro, norite, troctolite and anorthosite. The main anorthosite body, on the order of 1 km thick, is at the top of the complex. A model for origin of the anorthosite by convection flotation best accounts for the present configuration. Underlying and probably intruded by the basic complex is a thick body of mangerite that may have formed by high grade metamorphism and partial anatexis of adjacent veined and layered gneisses during intrusion of the basic complex. Alternatively, the mangerite may have formed in an earlier episode unrelated to the intrusion of the basic rocks. There is little evidence to suggest that the basic complex and the mangerite are parts of a single co-magmatic group. Highly deformed and metamorphosed veined and layered gneisses on the western and northern parts of the island are interpreted as the oldest rocks present and probably represent an ancient sedimentary (and volcanic?) pile. All events in the formation of the bedrock on Flakstadöy probably occurred in the Precambrian.

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Introduction

Flakstadöy, in the Lofoten group of Northern Norway, is an island with a total area of about 108 square kilometers. (See inset on geologic map, Fig. 1.) The landscape is dominated by cirques, narrow aretes, and horns, with some of the cirques extending to and below sea level. Maximum relief on the island is 937 meters from sea level to the top of Stjernhodet, and numerous peaks exceed 600 meters in elevation. Topographically the island is divided into three major lobes elongated in a north-south direction and almost completely separated from each other by arms of the sea. Helland (1897, p. 321) referred to these, from east to west, as Nappdraget, the Napp lobe; Flakstaddraget, the Flakstad lobe; and Sunddraget, the Sund lobe.

Although this area is well above the arctic circle, warming effects of the Gulf Stream produce a relatively moderate climate with average daily temperatures ranging from a little below 0 °C in the winter months up to about 12 °C in midsummer. There is much precipitation spread evenly throughout the year, the averages in various parts of Lofoten ranging from about 714–

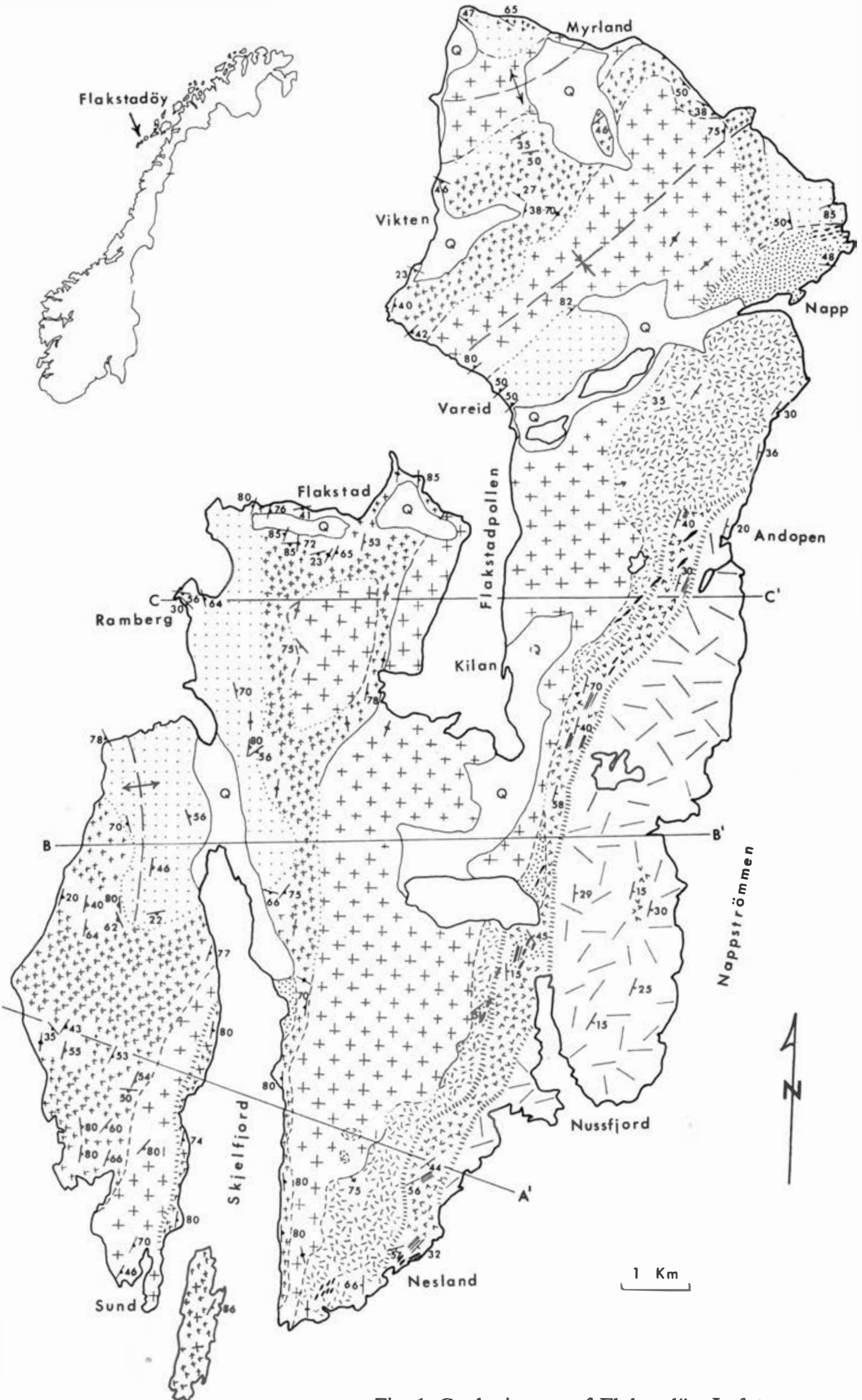

Fig. 1. Geologic map of Flakstadöy, Lofoten.

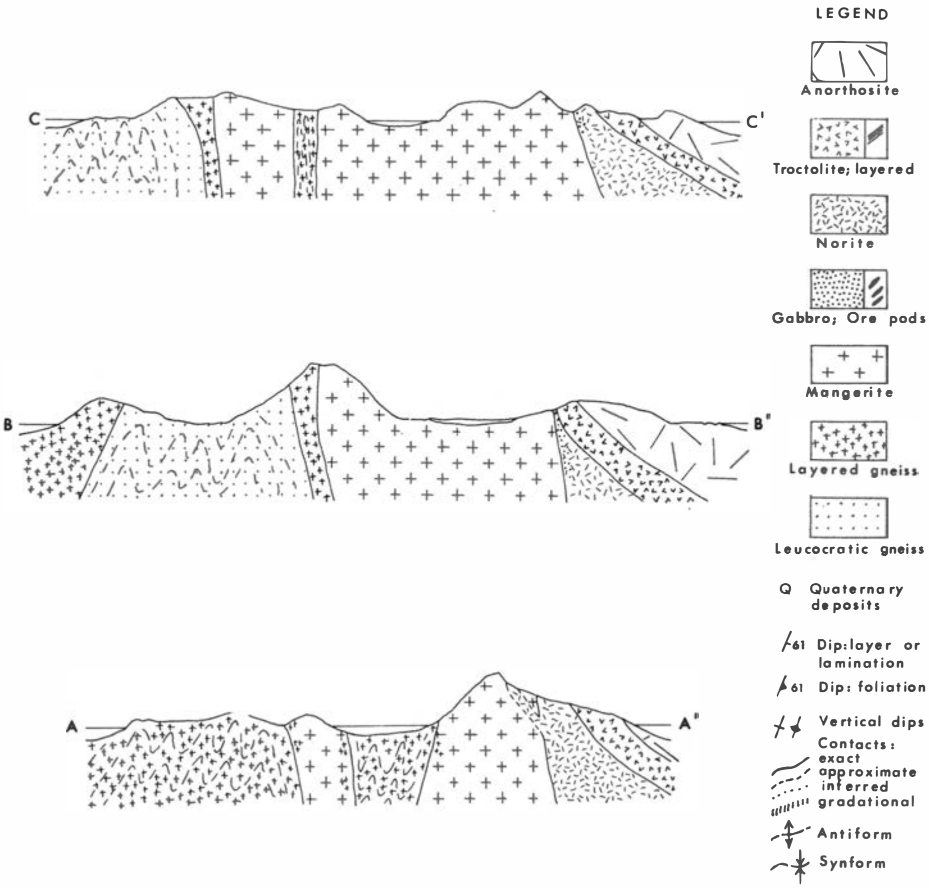

Fig. 2. Geologic cross sections of Flakstadöy. Legend for Figs. 1 and 2.

1523 mm (Helland, 1897, p. 37). On Flakstadöy the midnight sun shines throughout most of June and until about the 20th of July.

Much of the area is covered by an arctic tundra flora. There are few trees, and these, mainly small, deformed birches and alders, are confined to the lowermost slopes. Although it would seem there should be many excellent rock exposures, this is commonly not so because of the widespread and abundant growth of lichens and mosses. The mountains and hills of Flakstadöy and of the other islands in Lofoten are green and mossy making it difficult and in many areas impossible accurately to map subtle textural and mineralogical differences found in the major rock types.

One of the reasons why more studies of the Lofoten group have not been made has been their relative inaccessibility. As recently as 1960 the major islands were still not linked together by any throughgoing system of roads and ferries. With the completion in 1962 of the Lofotvei extending from Svolvær on Austvågöy to Reine on Moskenesöy all of the major islands were connected for the first time. Ferries, which run several times a day, connect the main islands, and gravel roads on each island are now good enough so that in the summer of 1967 it was possible for a Norwegian racing cyclist to travel from Svolvær on Austvågöy to Å at the southern tip of Moskenesöy in less than 4 hours, not including the considerable time spent waiting for and riding on the 3 ferries.

On Flakstadöy itself new roads have recently been built or rebuilt to Nussfjord, Nesland, Sund, and Vikten. These provide excellent, fresh exposures of rocks where previously the only surfaces available for examination were covered with lichens. With the new road system, large parts of each of the major lobes of Flakstadöy have become easily accessible, although some of the steeper coastal areas still cannot easily be visited except by boat. Large parts of Moskenesöy to the west remain accessible only by boat and by the use of alpine climbing techniques. This makes Flakstadöy the westernmost part of the Lofoten group now easily accessible to detailed geological exploration. Vestvågöy and Austvågöy are also well-covered by roads and it is to be hoped that detailed mapping of these islands will soon be undertaken to fill in the gap between the present work on Flakstadöy and Heier's (1960) work on Langöy in Vesterålen.

The present study was begun in the summer of 1967 at the suggestion of Professor T. F. W. Barth while the author was a National Science Foundation Science Faculty Fellow at the Mineralogisk-Geologisk Museum in Oslo. Laboratory work was conducted during the winter of 1967–68 and further field work was continued during the summer of 1968. The primary purpose for which the study was undertaken was to make a detailed petrological study of the gabbro and anorthosite complex on Flakstadöy. However, during the first summer's field work it became apparent that a careful reconnaissance geologic investigation of the rest of the island would be necessary in order to determine properly the overall structural setting of the gabbro-anorthosite complex. The present paper is intended to present a generalized description of the geology of Flakstadöy, to discuss in some detail field relations observed in the gabbro-anorthosite complex, and to speculate on the origin of these rocks.

Summary of major rock groups on Flakstadöy

As shown on the geologic map, Fig. 1, and cross sections, Fig. 2, several major rock groups have been mapped.

1. Forming a band 1 to 3 km wide along the eastern and southeastern shore of Flakstadöy and dipping to the eastward is a basic complex comprising gabbro, norite, troctolite and anorthosite.
2. Immediately to the west of the basic complex, forming a collar underlying the basic complex and ranging in outcrop width from ½ to 3 km is a body of mangerite. In most places the contact between the basic complex and the mangerite is sharp. Present along parts of the contact, however,

are mix rocks. The mangerite is most commonly massive in character, but crude, discontinuous layering occurs in numerous places, and especially near the contact of the mangerite and the veined gneiss to the west.

3. West of the mangerite complex is a group of veined and layered gneisses. In some areas the contact between the mangerite and veined gneisses is somewhat arbitrarily placed, for rocks similar to the mangerite in structure and texture are interlayered with veined gneisses in the contact zone.
4. A second major group of smaller, discontinuous bodies of mangeritic rocks occurs to the west of the main mangerite body. These smaller bodies are elongate, ovoid-shaped, and are associated with highly deformed layered and veined gneisses.

The norite-troctolite-anorthosite complex

The presence of this group of rocks on Flakstadöy was not recognized by either Keilhau (1844) or Pettersen (1881). Vogt first described the complex in 1895, and in 1897 (p. 178) he correctly described the total extent of these rocks on Flakstadöy:

> ". . . fra öens sydspids mellem Skjelfjorden og Nesland, forbi Nussfjord, Langs hele Nappströmmen til lidt nord for Napp og herfra over Vareid-vandene til Vareid og den sydvestre bund af Flakstadpollen, med af-stand 17–18 km fra Nesland til Napp."

In subsequent papers Vogt (1900, 1909) provided information on the general mineralogy of parts of the complex. More recently Carstens (1957) described the titaniferous ore deposits associated with this complex and also included data on the gabbro and anorthosite.

In gross physical aspect, areas in which the norite-troctolite-anorthosite group crops out have several characteristics that distinguish them easily from the mangeritic rocks to the west. Rocks of the norite-troctolite-anorthosite group are relatively more susceptible to weathering than is the mangerite and consequently form a platform at an elevation of about 200–300 meters above sea level, with areas to the west underlain by mangerite rising steeply from the edge of this platform to elevations of 600 to 900 meters (Fig. 3). This is not to say that steep topography does not exist in the norite-troctolite-anorthosite group, for the eastern edge of this plateau of basic rocks does plunge steeply down to Nappströmmen. Numerous cirques with nearly vertical sides also deeply indent the plateau. The hilltops themselves in the basic complex are generally rounded, rough textured, and massive looking, whereas the mangeritic rocks form sharp, angular peaks with smooth-textured, slabby cliffs.

Weathered outcrops of norite, troctolite, and anorthosite are generally gray in color, due both to the gray color of the rock itself and to the abun-

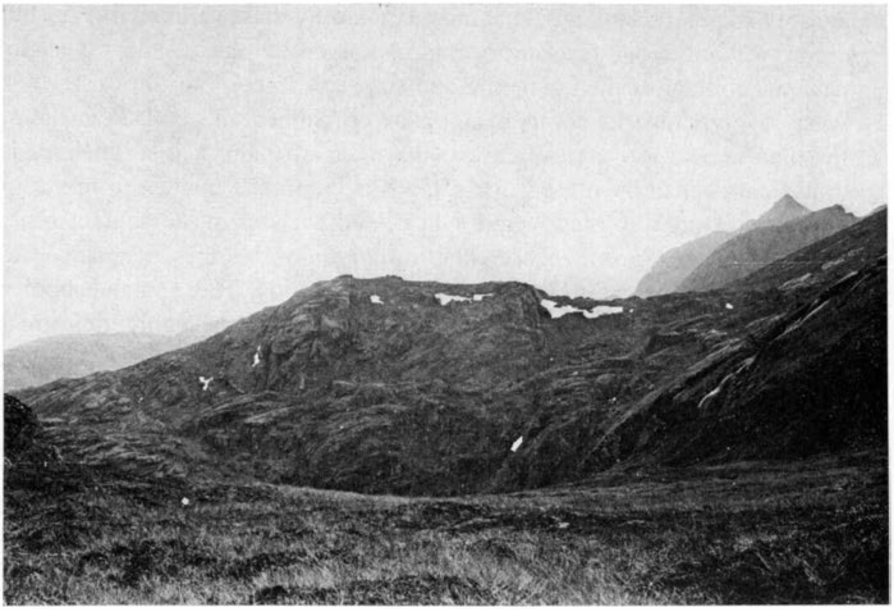

Fig. 3. Platform of rocks belonging to the basic complex (left). High peaks on the right are mangerite.

dant growth on rock surfaces of gray-colored lichens. The mangerite on the other hand is generally brownish on relatively fresh surfaces, or black or green from the mosses and lichens common on its surface.

Many outcrops are present in the norite-troctolite-anorthosite complex. In places there appears to be virtually 100 percent exposure, but unfortunately only a small percentage of these exposures are useful for macroscopic field examination and sampling because of the deeply weathered surface and extensive lichen cover.

Within the basic complex three major, intergradational lithologic units have been distinguished in the field. These are, from bottom to top and overlying the mangeritic rocks located to the west:

A. A lower zone of gabbro and norite.
B. A layered sequence of iron ore-norite-troctolite-anorthosite.
C. An upper zone of coarse-grained anorthositic norite, noritic anorthosite, and anorthosite.

Zone A: The lower gabbro and norite zone

The overall geometry of this zone is not certain, but dips of a foliation formed in places by parallel orientation of (010) of plagioclase crystals (possibly igneous lamination, as defined by Wager & Deer 1939) and in places by a weak compositional layering range from nearly vertical in the complex contact zone at the southern tip of Flakstad lobe to 30–50° eastward in the south central and northern parts of the zone. There are several areas where foliation diverges from these general averages, however. In the central part of the complex near Kilan the lower gabbro and norite are either missing or of very limited extent. This could possibly be a result of faulting, to be discussed later or, more likely, it is a function of level of exposure (see Fig. 2, B–B′). The maximum outcrop width of Zone A is on the order of 2 kilometers.

In composition, Zone A is inhomogeneous, containing at least 5 major textural and compositional types intergradationally mixed together. In some places the gradations are abrupt. Intimate interweaving of fabrics of the various rock types confirms their temporal and genetic association. The various types to be described below all contain plagioclase ($An_{42\text{-}61}$, with an average of about An_{54} as reported in Romey 1969a) as their most abundant constituent with hypersthene, ore (probably titaniferous magnetite) and minor biotite (secondary?) as the chief mafic constituents. Olivine is characteristically absent from the lower parts of Zone A, but appears in the cores of hypersthene grains in the middle and upper parts of the zone. Olivine grains invariably are surrounded by reaction rims probably of orthopyroxene and amphibole and similar to those described by Mason (1967). Weak traces of compositional layering appear within this unit, but these are of limited extent.

Fine-grained norite or gabbro rich in disseminated ore: This rock type, found primarily northeast and southeast of Kilan and west of Nesland is restricted to zones close to the boundary of the norite-troctolite-anorthosite complex and the underlying mangerite. Grain size within the fine-grained, brown-weathering gabbro is about 0.1–0.5 mm. Total mafic content is as high as 35–40 percent, much of this comprising fine-grained Ti-Fe oxide ore. The Napp gabbro, just north of the main norite-troctolite-anorthosite body may be equivalent to this fine-grained gabbro.

Porphyritic norite: Immediately south of the Napp gabbro, on the south side of the valley connecting Napp and Vareid is a remarkable porphyritic norite (Fig. 4). The groundmass of the rock consists of an ophitic intergrowth of plagioclase, hypersthene, and ore grains 0.5–10 mm in length. Within any single area the average size of groundmass grains is likely to be uniform rather than a mixture of all sizes within the range. Set in this groundmass are large numbers of euhedral, rectangular to trapezoidal plagioclase phenocrysts up to about 20 cm long. Phenocrystic and groundmass plagioclase grains are violet-gray in color and exhibit both carlsbad and albite twinning. In some areas the phenocrysts are clearly aligned to form a foliation within the rock.

Coarse norite with ophitic texture: The greatest volume by far of Zone A is composed of gray-weathering, coarse-grained norite with well-developed ophitic texture. A mixture of hypersthene and ore minerals (with minor and possibly secondary biotite) surrounds euhedral, rectangular, gray plagioclase. Most plagioclase crystals show carlsbad and albite twinning. The total

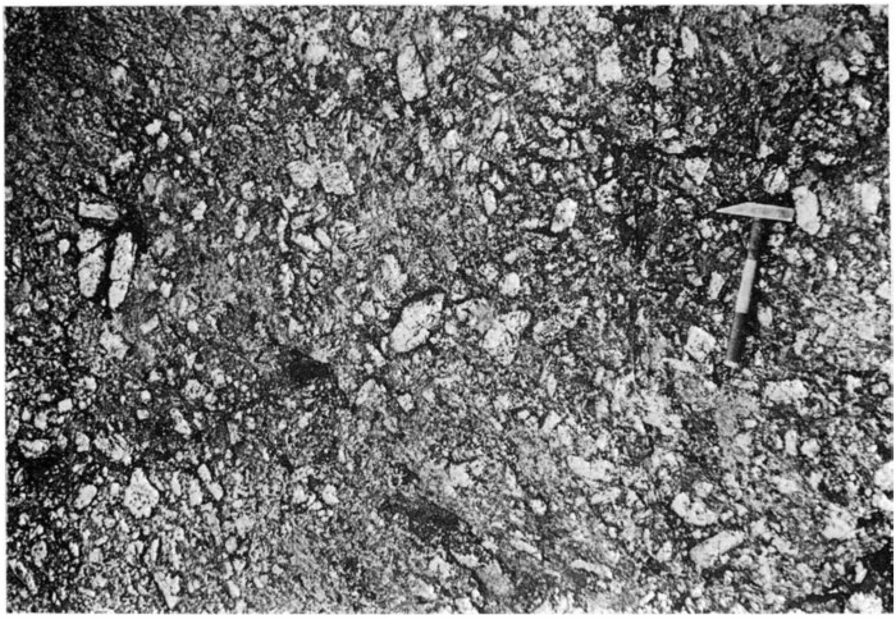

Fig. 4. Porphyritic norite southwest of Napp.

mafic percentage varies greatly from place to place within single outcrops. In some limited areas small patches of virtually pure plagioclase rock with allotriomorphic granular texture occur as leucocratic domains. In other places the percentage of mafic components is as high as 30 percent. The average color index for this rock as a whole is on the order of 20 to 25. Thus the rock may be classified as a leucocratic norite. The average grain size is generally greater than 1 cm, although locally some zones have average grain sizes of 2 to 4 cm. The rock is normally at least sparsely porphyritic, and phenocrysts 8 to 10 cm long are not uncommon. In some places isolated crystals of plagioclase 20 or more cm long have been observed. This rock in places contains 10 or more percent of dispersed ore.

Usually there is no strong preferred orientation of minerals, but in some areas a well-developed igneous lamination formed by the parallel orientation of (010) faces of plagioclase crystals was observed and measured. Some of the plagioclase grains show weak iridescence parallel to the (010) face.

Anorthosite: No major areas of true anorthosite (plagioclase rock containing less than 5 % mafic minerals) occur within Zone A. As mentioned above, the coarse norite contains small intergradational patches of virtually pure plagioclase rock. In addition the norite also contains roughly circular pods of pure, very coarse-grained anorthosite. These pods are discrete, sharply bounded masses of plagioclase and in places the surrounding norite shows a weak concentric foliation. The norite must once have been in a relatively fluid condition while the plagioclase pods were relatively rigid.

Coarse pegmatite pods: Within the main, coarse norite mass, local zones of a pegmatitic nature are common. In these pods, which may be up to a few meters across, are plagioclase crystals 10–20 cm long. The mafic minerals also reach large proportions; one orthopyroxene crystal observed was 40 cm long. The fabric of these pods is gradational into the fabric of the surrounding norite.

Zone B: The layered sequence of iron ore-norite-troctolite-anorthosite

To the east of Zone A, in an area with an outcrop width of about ½ km olivine becomes readily visible and quantitatively important in the modes of rocks. The composition of the plagioclase ranges from $An_{50.5}$ to An_{65} with an average of about An_{53} (Romey 1969a). Well-defined layers are present in this zone, and these provide the best indices within the entire complex to the overall geometry of the norite-troctolite-anorthosite body. Surfaces of layered rocks commonly have a redder weathering color than do the underlying norite of Zone A or the overlying anorthosite of Zone C. This is due both to the reddish brown color of the olivine which is abundant in the layered rocks and also to the secondary reddish oxidation products that form on the surfaces of both ore minerals and olivine.

The composition and structure of layers vary from place to place. In most areas layering results primarily from differences in the relative percentages of orthopyroxene and plagioclase. In these rocks the average grain size may be 2 or more cm, and there is usually well-developed ophitic texture. By far the most clearly visible and continuous layered sequences involve peridotite, troctolite, norite, and ore-rich rock with average grain sizes of 0.5–1

Fig. 5. Graded layers in the troctolite zone, near Nesland.

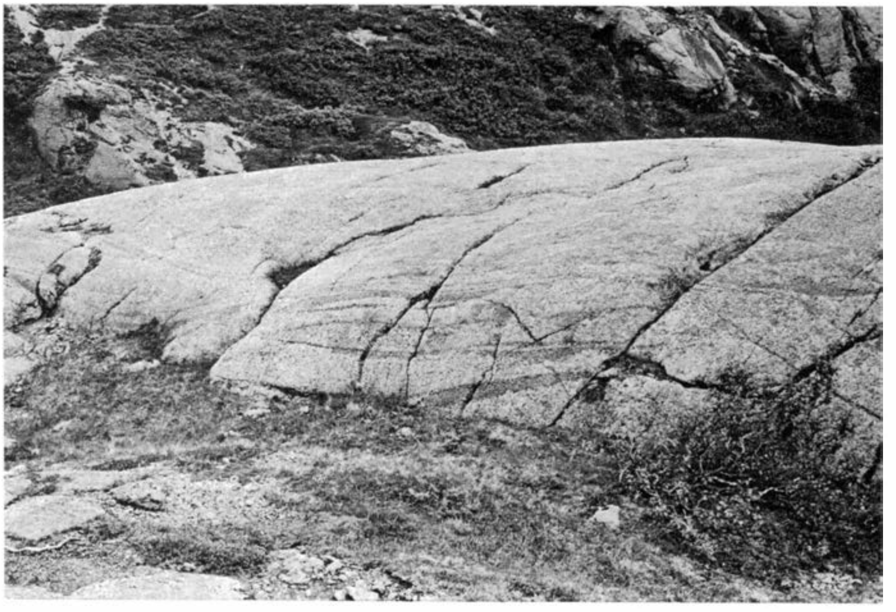

Fig. 6. Lensoid masses of troctolite in layered sequence, east of Kilan. These are interpreted as channel fillings.

cm. In some areas, such as one locality north of Nussfjord, layers of virtually pure Ti-Fe ore up to 2½ meters thick occur at the base of layered sequences.

Within layered zones alignment of (010) of plagioclase grains is common. Both size and density grading can be observed in the better layered zones: fine-grained feldspathic layers are overlain by coarse peridotite layers 1–10 cm thick which grade upward into finer grained, more feldspathic layers. These in turn are again overlain in rhythmic fashion by other layers of similar geometry (Fig. 5). The geometry of arcuate, lensoid peridotite layers seen near Kilan and Nesland strongly resembles that of channel marks commonly seen in stream deposits (Fig. 6) and suggests that the complex is right-side up.

Layering throughout most of this zone dips to the eastward, confirming that the layered sequence overlies the norite zone. However, in the area between Nesland and Nussfjord anomalous layers were observed with west dips. Careful examination of the geometry of these layers shows that they represent small ripples on larger, eastward dipping layers. The presence of such ripples would support the idea of slumping of the layers on the slope, with an inclination up to about 30 degrees, upon which they were apparently deposited.

Good layering cannot be observed everywhere within Zone B partially because of the lack of suitable exposures and partially because of the discontinuous nature of the layers themselves. The layers pinch and swell so that some may be traced only a few meters or at most a few tens of meters along strike until they taper out into the surrounding noritic or olivine-noritic rock. In other places the layers are terminated sharply against shear zones.

In general, well-layered sections were observed to comprise finer-grained rocks than apparently unlayered rocks nearby. It must be mentioned, however, that the coarser-grained rocks have a crumbly weathering surface in which interstitial mafic minerals within the ophitic intergrowth of plagioclase and pyroxene ± olivine weather out, leaving deep pits where lichens quickly take hold. The resulting, crumbly, lichen-covered surface of the coarser-grained rocks may merely have made any existing layers virtually impossible to see in outcrop. On the other hand, the finer-grained layered zones generally have a smoother, more resistant, better polished surface which holds up better against weathering.

Anorthosite in Zone B occurs either as thin layers in layered sequences or as small (maximum diameter a few meters), sharply-bounded pods and masses with the igneous lamination of the surrounding troctolitic material sweeping around them in concentric fashion. Other masses of anorthosite have what would first appear to be a dike-like manner of occurrence forming an anastomosing network of thin feldspathic masses. These are perhaps best interpreted as having formed by a kind of 'autoinjection' when accumulated masses of feldspar and adjacent troctolitic material were of nearly equal mobility and thus appear to intrude one another.

Zone C: Coarse-grained anorthositic norite, noritic anorthosite, and anorthosite (anorthosite zone)

This is the easternmost zone of the basic complex and has a maximum outcrop width of about 2 kilometers. Its eastern edge is against the sea and it is not known how far these rocks might have extended into Nappströmmen. Nappströmmen itself may follow the trace of a fault.

The anorthosite zone is characterized by rocks in which plagioclase is by far the most abundant mineral, ranging in modal percent from about 75 to 98 percent. The anorthite content of the plagioclase, inferred from the refractive index of fused samples, ranges from $An_{50.5}$–$An_{57.5}$, with an average of about An_{55} (Romey 1969a). It is interesting to note that the average anorthite content in Zones A, B, and C of the basic complex is about the same, but the range of anorthite contents in the 3 zones differs. In the norite (lower zone) and troctolite (central zone) the ranges of anorthite contents were, respectively, $An_{42\text{-}61}$ and $An_{50.5\text{-}65}$ while in the anorthosite zone (upper zone) the range was $An_{50.5\text{-}57.5}$. The relatively limited range of anorthite contents in the main anorthosite finds a parallel in the anorthosite of the Adirondacks and many other so-called massif types of anorthosite. The relatively wide range of anorthite contents in the norite and troctolite zones finds a parallel in large layered intrusions such as the Bushveld complex.

Distinctive features of rocks in the anorthosite zone are their massive character, absence of the prominent layering found in parts of the underlying troctolite zone, a low percentage or absence of olivine, and a relatively structureless fabric. The rocks in this zone have a gray, knobbly weathering surface, and clean exposures are rare. The task of distinguishing among anorthositic norite (mafic minerals more than about 15 percent), noritic anorthosite (mafic minerals about 5–15 percent), and anorthosite (mafic minerals less than 5 percent) is complicated by the poor quality of exposures and the intergradational nature of the three rock types.

The particular way in which rock types are mixed deserves careful description because of its probably great genetic significance. In general, there is no such thing in this complex as massive, pure anorthosite. Everywhere there is a mixture of smaller or larger areas within any outcrop which are respectively of anorthositic or of noritic character. The two elements of this mixture are intimately related to each other with clear gradation from one to the other. This texture is common in anorthositic rocks, and has been previously discussed by Romey (1968) and de Waard & Romey (1969a).

In places the rock as a whole contains a greater percentage of noritic domains than it does of anorthositic domains. In such a case the rock would have to be called anorthositic norite. When the anorthosite domains predominate, the rock should be called noritic anorthosite. When only a few noritic domains are present the rock can properly be called anorthosite. In general, it is possible to say that Zone C contains a higher percentage of noritic domains in its western (lower) parts than in its eastern (higher) parts. The anorthositic character of the complex becomes more accentuated towards the top.

In one location the anorthosite is in sharp contact with the underlying layered troctolite sequence, but the interdigitating nature of the contact makes it difficult to say which, if either, of the two rocks has been intrusive with respect to the other. Pods of anorthosite within the layered sequence suggest that the underlying layered sequence was more mobile than the overlying anorthosite at the time when the present configuration became stabilized. In general, the transition from Zone B to C is not thought to be the result of polyphase intrusion within the complex, and the sharp contact exposed in this one area may well represent only local movement of materials above and below the contact, possibly at a time when neither was fully consolidated. (An interesting relationship which deserves further underlining is the position of the anorthosite zone above an olivine-rich zone. Hess (1960) described similar olivine zones at the base of each of the several anorthosite layers in the Stillwater complex.)

In the more pure plagioclase-rock areas, the rocks of Zone C contain many criss-crossing planar zones a few millimeters to centimeters thick in which the anorthosite is bleached and very fine grained. Garnet is locally present in these zones, which are interpreted as mylonites caused by late shearing in the anorthosite when it was nearly or completely consolidated. Within plagioclase masses where there are few noritic domains, large feldspar crystals appear bent and partially granulated. Although such effects of crushing are observed also in plagioclasic pods of Zones A and B and also in placioclase-rich intergradational domains of these same zones, the overall degree of deformation and granulation within the anorthositic rocks of Zone C exceeds that seen in noritic and troctolitic rocks of the complex. This would seem to indicate that anorthositic zones must have been in a more highly consolidated or rigid state earlier than the norite and troctolite. The anorthosite zone is interpreted as a mass composed mainly of plagioclase that accumulated relatively early in the history of the complex.

INCLUSIONS WITHIN THE NORITE-TROCTOLITE-ANORTHOSITE COMPLEX

Several angular inclusions of well-layered gneiss which resembles some of the veined and layered gneiss in the western and northern parts of Flakstadöy have been found within the basic complex. One of these about 30 m long and only about 2 m thick is located well within the complex near the boundary between Zones B and C. Blocks which can more clearly be related to the veined gneiss unit occur in the norite close to the border of the complex. These are taken as evidence in support of an intrusive origin for the basic complex as a whole.

DIKES

Both coarse-grained troctolite and coarse-grained norite dikes have been observed in Zone C, and these are interpreted as being genetically related to the basic complex, indicating the presence of relatively mobile and probably large liquid masses of both norite and troctolite material at a time when the anorthosite-norite mass of Zone C was relatively well consolidated.

Other fine-grained mafic dikes with finer-grained to aphanitic borders are clearly intrusive into the basic complex. Such late dikes have been observed in several localities. Because of their very fine grain size it is inferred that they were intruded at a much higher level in the crust than the level at which the complex as a whole must have formed.

OVERALL EXTENT OF THE BASIC COMPLEX AND POSSIBLE CORRELATION WITH OTHER ROCKS IN NORTH NORWAY

A strong positive gravity anomaly indicated over this area on the Bouguer anomaly map of north Norway (1:1,000,000, Norges Geografiske Oppmåling, published in May, 1963) suggests that relatively dense rocks extend northward up an axis from Moskenesöy through Flakstadöy, and northward to Andöya at the northern tip of Vesterålen. Heier (1960) has described gabbroic and anorthositic rocks on Langöy in Vesterålen. The author has visited the Langöy anorthosite complex briefly and found striking similarities to the Flakstadöy complex. Anorthositic rocks of both Lofoten and Vesterålen represent rock masses of relatively low density, and yet both are clearly associated with a major *positive* gravity anomaly. This strongly suggests that the feldspathic rocks of this area are underlain by a large complex of relatively dense rocks. Indeed in one locality on Flakstadöy, near Andopen, a peridotite mass several meters thick was observed along the northern contact of Zone B. Whether this was a dike or inclusion could not be determined. Of further interest are the reports by Brooks (1966, 1968) that this positive gravity anomaly appears to be connected with the anomaly over Soröya, 270 kilometers northeast of Vesterålen. The basic rocks of Troms and Finnmark, however, seem related to Caledonian events, whereas those of Lofoten and Vesterålen, according to new dates by Heier (personal communication) are Precambrian. There are major age problems involved in any attempt to correlate the basic rocks of Lofoten-Vesterålen with those of Troms-Finnmark.

Mangeritic rocks

The existence on Flakstadöy of K-feldspar-plagioclase-orthopyroxene rocks of low quartz content was recognized by both Keilhau (1844) and Pettersen (1881). Mangerite forms a band along the western side of the norite-troctolite-anorthosite complex. Only at the northernmost tip of the gabbro at Napp where granite, leuco-gneiss, and rocks of the layered gneiss group are in sharp contact with gabbro is the mangerite absent.

The mangeritic rocks forming the main collar around the basic complex are coarse-grained, megacrystic rocks with little or no well-defined foliation. They present a massive, homogeneous aspect (see Fig. 7). The groundmass texture ranges from hypidiomorphic granular to a texture in which grain boundaries have an intricately sutured appearance. In places a tendency toward igneous-like ophitic textures has been observed. The mineral assemblage most commonly observed (verified in preliminary petrographic studies) is mesoperthite, orthopyroxene, clinopyroxene, and ore, with the color index of the rock generally being about 15 or less. In some specimens, mesoperthite, plagioclase, and K-feldspar coexist. K-feldspar fills irregular indentations into the plagioclase in some rocks suggesting that the K-feldspar may have replaced plagioclase. The feldspar occurs as (1) small grains of both plagioclase and K-feldspar from about 0.1–10 mm in largest dimension, some areas containing predominantly the coarser-grained varieties and others containing predominantly the finer-grained varieties; (2) large, rectangular crystals of K-feldspar up to 1 or 2 cm on a side surrounded by a mass of smaller crystals. Whether the large K-feldspar crystals represent phenocrysts or porphyroblasts is debatable. In some specimens, the feldspars show a slight iridescence, rather similar to that observed in larvikite. At a distance from the basic complex the pyroxene in some mangerite samples studied is surrounded by a fine-grained intergrowth of biotite and amphibole

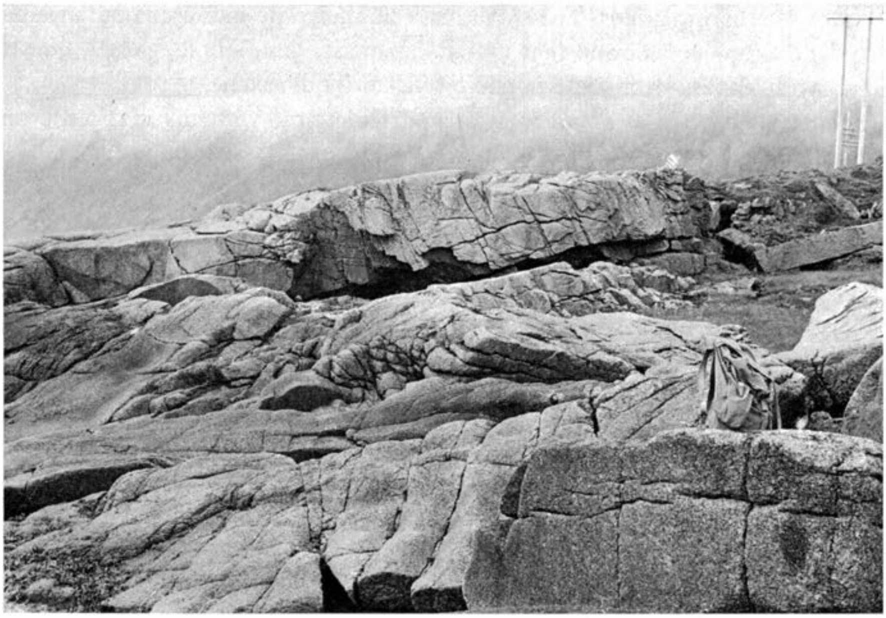

Fig. 7. Outcrops of massive mangerite along the coast south of Vareid.

which are also distributed as tiny crystals throughout the pyroxene. Such intergrowths are inferred to be replacing the pyroxene. In general, the mineral assemblage is that of the granulite facies, with evidence of retrogression to the upper almandite amphibolite facies, especially at a distance from the contact with the basic complex.

Structurally speaking, the mangerite forms the floor of the igneous complex. In most places the contact is sharp, but this is not true everywhere. Near West Nesland the contact shows gradation and conflicting relationships between the basic complex and the mangerite. In that area the author was unable to reach a conclusion about the relationship between the two units. The bulk of the evidence, in other areas and partially in the area around Nesland, suggests that the basic complex was intruded into the surrounding rocks, but that in places assimilation of salic material has complicated the contact relationships. This is a problem which can only be resolved by further large-scale mapping of certain key zones along the contact and by detailed petrographic and chemical studies.

Mangeritic rocks west of the main mangerite and separated from the main body by veined and layered gneisses occur as elongate, ovoid masses parallel to the general regional foliation. The mangerite within these bodies is, in general, similar to that in the main mangerite zone, but foliation and diffuse layering are more common in the isolated bodies. In places, well-layered zones grading into mangerite are found within these smaller mangerite bodies, especially near their contacts with the veined and layered gneiss. Orthopyroxene continues to be present in most places as medium-sized, bronze-weathering grains. The small body of mangerite that crops out around Myrland is an exception in that it contains biotite, amphibole, garnet, quartz and two feldspars with some possible relict orthopyroxene.

The mangerite of Flakstadöy is virtually identical with the mangerite on Vestvågöy just across Nappströmmen, to the east. Massive outcrops of tan-weathering mangerite make up the areas around Balstad and Stamsund. Heier (1960, p. 24) reports other, similar mangerites (which he calls 'porphyroblastic monzonites of Lofoten') in several other localities along the west side of Vestvågöy. Similar rocks also occur on the southeast part of Austvågöy and on the mainland at the Skutvik ferry (Hamaröy). Thus a band of such rocks extends along much of the side of Lofoten facing Vestfjord. Furthermore Heier (1960) described massive monzonite on Langöy, Vesterålen, which appears to be in sharp contrast with anorthosite and gabbro extending from Fleines to Stromfjord and associated with gabbro bodies at Hovden and Bö. Just as on Flakstadöy, the monzonite of Langöy is mainly massive but gradational into layered varieties that Heier labels the 'banded series (charnockite proper)'. Unfortunately, Heier's study involved no detailed work on either the anorthosite or monzonite, and the geometric and petrographic relationships between the two rock groups on Langöy remain to be explored.

Veined and layered gneisses

A complex of gneisses occupies the western and northern parts of Flakstadöy. Near contacts with mangerite the rocks are dominantly bluish gray on fresh surfaces, fine-grained (grain-size ranging from about 0.2–1 mm), and foliated. These gneisses contain fine-grained stringers of granitic material commonly parallel to, but in some places cutting across their foliation. For this reason they are called veined gneisses in this report. Primary minerals in the veined gneiss are biotite, plagioclase, k-feldspar, and pyroxene (some of which is orthorhombic), with some amphibole and quartz. The presence of orthopyroxene would suggest that parts of the veined gneiss complex near to the basic complex can be assigned to the granulite facies (hornblende-orthopyroxene-plagioclase subfacies of de Waard 1965).

Layer-like masses of mangerite up to several meters thick occur within the veined gneisses. At the present map scale the contact between mangerite and veined gneiss in some areas must be rather arbitrarily drawn, for the mangerite interfingers on a fine scale with the gneiss. Angular fragments of what appears to be veined gneiss have been found in both the mangerite and the norite-troctolite-anorthosite complex. These veined gneisses would appear to correspond with the 'banded charnockites' on Langöy, Vesterålen (Heier 1960, p. 34).

The veined gneisses grade westward and northward into a series of well-layered, folded, foliated, lineated schists and gneisses. Excellent coastal exposures of these layered gneisses occur along the north tip of Flakstad

Fig. 8. Multiply deformed, layered gneisses west of Myrland. Besides the folding shown here, this structure plunges both toward the observer and away from the observer, forming a small domal fold culmination.

lobe and the north tip of Napp lobe. Well-exposed layered rocks also can be seen in steep gulleys along the mountainsides south of Ramberg. These rocks are complexly folded, and mesoscopic-scale folds of various sizes are widespread (Fig. 8). Most of the small structures observed have curved axial surfaces.

Dominant rock types within these well-layered gneisses are fine-to-medium-grained biotite-amphibole-plagioclase schist, fine-grained garnet-amphibole-plagioclase granofels, porphyroblastic amphibole-garnet-plagioclase gneiss, plagioclase-K-feldspar-amphibole-epidote-quartz gneiss, and leucocratic plagioclase-K-feldspar-quartz gneiss containing less than 5 percent biotite and epidote. Mineral assemblages at a distance from the basic complex are representative of the amphibolite facies. For purposes of the geologic map only two major groups are shown: (1) a unit including the veined gneisses and the darker-colored blue layered gneisses and (2) a leucocratic gneiss unit. Detailed mapping on a larger scale will undoubtedly yield good information on stratigraphy and both large-and-small-scale fold patterns.

Considerations on the genesis of the norite-troctolite-anorthosite complex

The textures, structures, and mineral assemblages observed within the basic complex, shown diagrammatically in Fig. 9, leave little doubt that this is a major body of igneous rock that has undergone some sort of differentiation process leading to the presently observed layered configuration. The rocks of the complex contain an assemblage of almost entirely anhydrous minerals and the presence of coexisting plagioclase and orthopyroxene suggests conditions of pressure and temperature of the granulite mineral facies at the time of crystallization.

It would appear from the presence in all units of blobby masses of nearly pure plagioclase and from the euhedral to subhedral shapes of plagioclase in Zones A and C that plagioclase was one of the first minerals to crystallize from the melt. Vogt (1909, pp. 90–91) inferred the following order of crystallization for rocks in the Andopen area: plagioclase → plagioclase + magnetite → plagioclase + magnetite + hypersthene (+ biotite + diallage). The sequence proposed by Vogt is based on partial rims of magnetite surrounding large plagioclase crystals from Andopen, the great enrichment in magnetite near plagioclase phenocrysts, and the presence of hypersthene outside of this zone.

The concentration of plagioclase at the top of the complex calls to mind the Nordingrå-Rödö complex of Northern Sweden for which von Eckermann (1938) hypothesized an origin by flotation differentiation of a 'gabbroid' magma. Von Eckermann's analyses (1938, p. 251) of gabbro associated with the Nordingrå anorthosite show a chemical composition related to the hypersthene basalts or tholeites of Yoder & Tilley (1962, pp. 352,

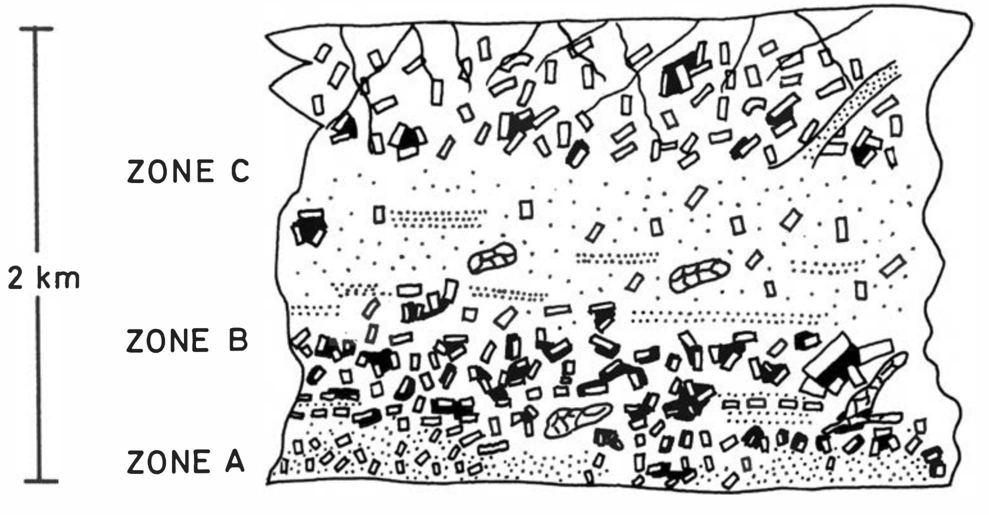

Fig. 9. Diagrammatic cross section through the basic complex. Zone A has gabbro (dense dot-pattern) and porphyritic norite at its base and grades upward into more homogeneous ophitic norite. It contains some large blocks of anorthosite. Rectangles represent plagioclase phenocrysts; black areas represent pyroxene ophitically enclosing plagioclase.
Zone B is finer grained (sparse dot-pattern), contains layers, blocks of anorthosite, and some noritic zones. Zone C is dominantly anorthosite, with noritic domains. This zone is criss-crossed by shear zones and intruded by several troctolite dikes.

361–3). If, as stated by von Eckermann and supported strongly by the author on the basis of a brief field examination of the Nordingrå area in summer, 1968, the anorthosite and gabbro are cogenetic and one adds the anorthosite and gabbro fractions of the Nordingrå area together, the result is an aluminous tholeite. Unfortunately, no exact proportions of anorthosite and gabbro can be established because the bottom of the Nordingrå gabbro is not exposed. However, from von Eckermann's cross section (p. 246) it can be inferred that the minimum gabbro: anorthosite ratio is about 7:5 based on actual surface exposures and 2:1 based on von Eckermann's projection to the bottom of his cross section, which still predicts gabbro. (Martignole (1968) has recently postulated that the anorthosite in a small norite-anorthosite complex near Shawinigan, Quebec originated through flotation of plagioclase in a basaltic magma contaminated by assimilation of pelitic gneisses.)

If plagioclase were the first mineral to crystallize and if it crystallized alone over a long period of time (as Green & Ringwood 1968 have indicated may happen in dry andesite melts at pressures below about 15 kb), then the anorthosite and the plagioclase-rich complex as a whole can be explained by any one of several differentiation models. Plagioclase, however, need not necessarily have been the only mineral to crystallize early as long as the densities of melt, early plagioclase, and other early-forming minerals (ore, orthopyroxene, olivine?) were adjusted in such a way that a separation of plagioclase and mafic minerals could be produced.

Is it reasonable to suspect that the density relationships of phases could have been adjusted in this way? We know little about the density of magmas under high pressure. Barus (1893), Day, Sosman & Hostetter (1914), and Dane (1941) experimentally determined the densities of diabasic magmas to be in the density range 2.55–2.65. These figures have been used in density arguments proposed by Hess (1960), Wager & Brown (1967), and Bridgwater & Harry (1968). Not enough is known about the densities of plagioclases under high pressure either. For bytownite Hess (1960) and Wager & Brown (1967) use a density of 2.68–2.69. However, using more recent information on the thermal expansion of plagioclase, one can find values as low as 2.63 for bytownite. Romey (1969b) has also shown that the density of anorthosite plagioclase of given composition is lower for disordered than for ordered plagioclase, providing an additional reason why one would expect relatively low plagioclase densities for crystals floating in a melt. Bridgwater (in Bridgwater & Harry 1968) uses a value of 2.60 for the specific gravity of labradorite (An_{50}). It seems clear that the density of labradorite at high temperature and pressure must be very close to that of a diabasic or basaltic melt.

If it is assumed that the density of early-formed plagioclase in the Flakstadöy massif was less then the density of the melt, one would expect to find a virtually pure plagioclase rock concentrated at the top of the complex and no plagioclase cumulate masses at the bottom of the complex, However, the situation is not so simple, for there are large glomeroporphyritic masses of pure plagioclase found in both of the lower zones (A & B) in the complex. Furthermore, mafic minerals do occur in the uppermost zones of the complex, although not in great abundance.

If on the other hand the early-formed plagioclase had been much more dense than the melt, one would not expect to find any plagioclase cumulate mass at all at the top of the complex. Instead, there might be a layered complex of Bushveld or Stillwater type.

In a limiting case where the plagioclase had exactly the same density as the magma and there was no convection there would probably be no accumulation of plagioclase anywhere in the complex. Each plagioclase crystal would float at the level where it formed.

Consider a fourth possibility in which the plagioclase has very slightly greater density than the melt. In such a situation the plagioclase would sink slowly, providing the magma was stationary. According to Shaw (1965, p. 126) Stokes' law would accurately describe the rate of settling of individual particles in the viscosity range of interest here.

Stokes' Law: Settling velocity $= \frac{2\,gr^2\,(\varrho_{xl} - \varrho_{magma}}{9\,\eta}$ where g = gravitational constant, r = radius of particles, ϱ = density, η = viscosity of the fluid).

Assuming for the Lofoten body a figure 3×10^3 poises for viscosity, a density of 2.58 for a basaltic magma, and a density of labradorite at high temperature of 2.59, one can use the curves of Hess (1960, p. 142) and Shaw (1965, p. 128) to predict that plagioclase crystals 0.5 cm in diameter would settle at a rate of only 0.02 meters per day, crystals 1 cm in diameter would settle at a rate of 1.6 meters per day, and crystals 10 cm in diameter would settle at a rate of about 165 meters per day.

If the density difference between plagioclase crystals and melt were very small, as is almost certainly true, single crystals or small agglomerate masses of plagioclase slightly denser than the melt would have had their settling velocities cancelled out by the upward velocity of even slow convection currents (see Hess, 1960, p. 136). That there was convective movement in the Flakstadöy magma body can be inferred from the channel and scour marks such as those shown in Fig. 6. Parallel alignment of plagioclase crystals in places also suggests movement of material in a *horizontal* direction at some stages during development of the complex. It is reasonable to expect that there was also movement of hot magma along the upper surface of the complex. Some of the plagioclase that was carried upward would then have been trapped in congealing material in the upper part of the magma chamber.

Plagioclase crystals accreting along the congealing upper surface of the complex would have continued to grow as fresh magma swept past them (adcumulus growth as defined by Wager, Brown & Wadsworth 1960) and new plagioclase crystals would have joined them. The relatively restricted range of anorthite contents in large plagioclases of the anorthosite would support the idea of adcumulus growth. Cementing material between these crystals would likely be noritic material of relatively fine grain size. The texture and structure of Zone C, the anorthosite zone, fit well into this model. However, one might bring up the question of the anorthite content of the phenocrystic plagioclase versus that of the groundmass plagioclase in the noritic 'cement'. One would expect the An content of the phenocrysts to be greater than that of the groundmass crystals. Such a relationship has not yet been established; preliminary studies suggest that the An contents of the phenocrysts and groundmass crystals are nearly equal and certainly not sufficiently different to fit the predicted demands of a liquid line of descent for the groundmass. This apparent equality of An content of phenocrystic and groundmass plagioclases is not something newly discovered; it is characteristic of anorthosite the world over and may have its explanation in some form of equilibration between groundmass plagioclase and phenocrysts. On the other hand it may be that most workers have concentrated too much on studies of phenocrysts and that further studies of groundmass plagioclase in the noritic patches so common in anorthosite may yield a more simple answer to this apparent enigma.

According to Stokes' law, larger, glomeroporphyritic masses of plagioclase would have a much greater settling velocity than the small masses carried upward by convection. Such masses could thus sink in the magma, even in the presence of substantial upward convective flow. The concentration of anorthosite at the top of the Flakstadöy complex and the presence of glomeroporphyritic anorthosite masses in the center and at the bottom of the complex would seem to support this latter model involving upward movement of individual plagioclase crystals by convection in a magma slightly less dense than the crystals and downward sinking of larger plagioclase masses.

This postulated mechanical differentiation by convection flotation of most of the plagioclase and downward sinking of most of the mafic components must have been accompanied by chemical differentiation as well. From the overall bulk composition of the anorthosite zone (C) and the norite zone (A) it can be inferred that the melt must have been gradually depleted in plagioclase components. One effect of selective removal of plagioclase components from the melt would probably be an increase in density of the remaining melt, resulting in an increase in the tendency of plagioclase to float. Another effect would be a decrease in SiO_2 content; the bulk composition of the plagioclase separating from the melt approximated An_{50} ($CaNaAl_3Si_5O_{16}$), giving a Si : O ratio of 5 : 16. Pyroxene separating early from the melt would have a similar ratio, Si : O = 1 : 3. (It must be noted that some olivine also crystallized early from the melt as shown by small remnant cores of olivine surrounded by pyroxene in the lower (norite) zone of the complex.) Thus, in the early stages of crystallization, a relatively high proportion of Si : O was being removed from the melt. Ultimately, the composition of uncongealed magma remaining in the center of the complex should have become deficient enough in silica so that olivine, with a Si : O ratio of 1 : 4, could become a major phase. Preliminary chemical and petrographic data do indeed demonstrate that the rocks in the center zone are olivine-rich and that their bulk chemical composition is more basic than that of rocks in the upper and lower zones. That the center-zone troctolite was the last part of the complex to crystallize is confirmed by dikes of troctolite intruding both anorthosite and norite.

In order to explain the repetition of olivine and plagioclase layers characterizing part of the troctolite zone, one might appeal to a process of limited rhythmic overturn in the melt remaining at the center of the complex. Movements within the liquid of Zone B would also lead to fracturing of the nearly solid overlying plagioclase mass and to the expulsion through these cracks of volatile material from the unconsolidated liquid in the center of the complex. This could account for the bleaching, shearing and local crystallization and garnet formation found in these zones.

Studies of powder diffraction X-ray data have shown that plagioclase in Zones A (norite) and C (anorthosite) is dominantly low transitional in structural state whereas plagioclase in the troctolite is dominantly high transitional (Romey 1969a). Possibly, the more ordered structural states of plagioclases in Zones A and C are a reflection of their having formed early in the crystallization history of the complex, accumulated at the top and bottom of the complex during congealing from the margins of the complex inward, and gradually become annealed (ordered) when held for a long period of time at temperatures just below their crystallization temperature while heat from the hot liquid in the center of the complex was conducted outward through them.

The composition of the magma from which the Flakstadöy complex was derived cannot be estimated without detailed information on the petrography and chemistry of the rocks. Such information will be presented in a later paper. Students of anorthosite-bearing complexes have suggested magmas ranging from gabbroid through granodioritic. Plagioclase cumulates can indeed be derived from any of these, given the right conditions. One of the most important considerations bearing on this problem is the relationship of the basic complex to the surrounding collar of mangerite.

Genesis of the mangerite and its relationship to the basic complex and to the veined and layered gneisses

The association of mangerite or other members of the charnockite suite as described by de Waard (1969) with anorthositic rocks has been widely described and discussed in the literature. Barth (1962) has long suspected that anorthosite and associated acidic rocks (variously called pyroxene syenite, mangerite, monzonite, farsundite, adamellite, and charnockite) are cogenetic. Intimate textural, modal, and chemical gradations among these rocks as shown in some areas by Philpotts (1966) and de Waard & Romey

(1969a, b) would seem to support, or at least not to contradict Barth's idea. On the other hand, Buddington (1939, 1961) maintains, contrary to de Waard & Romey (1969a, b) that the acidic rocks which happen to be spatially associated with anorthosite in the Adirondacks are not part of a single magmatic differentiation series but represent two unrelated magmatic events. Hargraves (1962) concluded that charnockitic rocks associated with the Allard Lake anorthosite were formed by metasomatism and partial fusion of surrounding salic gneisses during emplacement of the anorthosite. Another possibility that needs evaluation is whether or not a body of charnockitic or other acidic rocks associated with anorthosite simply happened to be there already at the time anorthosite was emplaced. A further possibility, that a given anorthosite complex and associated acidic rocks were all formed during some metamorphic event unaccompanied by magmatic intrusions but accompanied by extensive metasomatism, can probably be excluded in discussing the Lofoten complex, for the ubiquitous presence of what are generally interpreted as typically magmatic structural, textural, and mineralogical features in the gabbro-norite-troctolite-anorthosite complex provides good evidence for a magmatic origin. Clearly, in any attempt to form an idea of the composition of the parent magma of a given anorthosite body or of anorthosites in general, one must know in detail the relationship between a basic complex and surrounding charnockitic rocks.

In considering the possibility that anorthositic-gabbroic rocks are cogenetic with the mangeritic rocks in Lofoten, the following factors must be taken into account:

(1) The anorthositic and gabbroic rocks occupy a relatively small area in Lofoten and the mangeritic rocks are extensive. If this were a differentiation suite formed from a relatively basic parent magma one might expect the total mass of the basic rocks to be great and the total mass of the mangeritic rocks to be small. However, it is possible that the volumes of the two rock groups now exposed are not representative of the actual volumes present. The positive gravity anomaly underlying Lofoten and Vesterålen is probably best interpreted as caused by a large mass of underlying basic rocks. If this is indeed so, then the relative volumes of mangerite and basic rocks now exposed must be considered a function of level of exposure and erosion rather than of total volume.

(2) Preliminary chemical and petrographic data indicate an abrupt change in chemical composition and mineral content at the contact between the basic complex and the mangerite rather than the gradation between rock types that might be expected in a comagmatic rock suite. However, the possibility that rocks differentiated from a single magma were emplaced at their present level at slightly different times cannot yet be excluded. Detailed radiometric dating would be necessary to evaluate this matter further.

(3) The anorthositic and gabbroic rocks in most places can be interpreted as intrusive into the mangerite, suggesting that the mangerite was

already present when the basic rocks were emplaced. As described earlier, however, anomalous zones along the contact may contradict this generalization.

(4) The mangerite seems to be more intimately related to the complex of veined and layered gneisses than to the basic complex.

On the whole, considering the evidence now available, the author believes that the mangerite is probably not comagmatic with the basic complex. There is little likelihood that the mangeritic rocks represent a series of intrusions emplaced in a separate magmatic event after emplacement of the anorthosite, in the manner that Buddington (1939, 1961) has suggested that anorthosite and 'syenites' of the Adirondacks are related. Certainly it is possible that the mangerite formed in a separate magmatic event that *preceded* emplacement of the basic complex of Flakstadöy. The possibility that the mangerite developed as a result of metamorphism and partial anatexis of layered and veined gneisses during intrusion of the basic complex deserves serious consideration.

The gradational nature of the contact of the mangerite and veined gneiss and the fine-scale interfingering of these two rock types lead the author to favor the idea that the mangerite has formed as the result of partial melting of the complex of veined and layered gneiss. One problem with this explanation, however, is that the mineralogy of the mangeritic rocks does not necessarily suggest that the overall chemical composition of the rock is equivalent to a minimum melt composition that might be expected from the veined and layered gneisses. Clearly, detailed information on both the relative ages and the chemical compositions of the gneisses and of the mangeritic rocks is needed further to evaluate this problem. An anatectic episode capable of producing the mangerite could have occurred either before the emplacement of the basic complex and be unrelated to this event or it could be directly related to emplacement of a hot basic magma. The relatively small volume of basic rock exposed at the present level might again cast doubt on the latter possibility, but the positive gravity anomaly discussed above may reflect a sufficiently large mass of basic material beneath the present level of exposure to have produced enough heat for metamorphism, metasomatism, and anatexis necessary to account for the large volume of mangerite now exposed.

Structure of the layered gneiss complex

As shown on the geologic map and cross sections (Figs. 1 and 2), a series of north and northeast-trending antiformal and synformal structures can be inferred from the data available. Approximate traces of inferred axial surfaces of the larger structures are shown on the map. It appears from the map pattern and field relationships that these axial surfaces are curved and are moderately steeply dipping. Few horizontal foliation surfaces were observed. Both the larger and smaller structures appear to be elongated, doubly plunging domal and basinal structures. The large northeast-trending antiform shown in the Myrland area may connect with the north-trending structure south of Ramberg. On the Sund lobe many smaller antiforms and synforms were sketch-mapped. The presence of some of these structures is shown on the map by opposing dips. The scale of the map precludes showing these in more detail. Figure 8 shows the nature of some of the small mesoscopic fold structures observed west of Myrland. The geometry of small folds and the overall map pattern lead one to suspect at least 2 distinct episodes of folding. Detailed measurements of foliations and lineations have been made in several small, well-exposed areas (good coastal exposures west of Myrland, south of Vikten, on Flakstad point near the church and near Flakstadlygte, and near Ramberg; other exposures in the central and southern parts of Sund lobe), but these measurements have not yet been studied, and in all probability they will provide too sketchy a basis for an acceptable statistical structural analysis. Further detailed field study and mapping on Moskenesöy and Vestvågöy will be required before the geometry of folding in this area can be satisfactorily established.

Heier (1960) has interpreted structures with similar trends on Vestvågöy and on Langöy as thrust surfaces. Minor zones of mylonite have been found in some places on Flakstadöy, but these are not clearly related to the fold patterns. No evidence in favor of extensive thrusting has yet been found on Flakstadöy and the author suspects that, although minor thrusting is a possibility not to be excluded, the structure is best explained on the basis of complex folding.

Faults

The existence of a prominent north-south-trending set of topographic lineaments in Lofoten and Vesterålen is readily apparent. On opposite sides of three of the major east-west lineaments on Flakstadöy (Nappströmmen, the line Nussfjord-Flakstadpollen, and Skjelfjord) there are possible lateral offsets or peculiar truncations of map units. Unfortunately, formation contacts lie at only a small angle to the lineaments so that offsets of contacts cannot be proved. Heier's (1960) map of Langöy shows similar relationships and one might at least suspect north-south trending faults along the line Möklandsfjord-Jörgensfjord, where anorthosite bodies at Sunnan and Fleines on opposite sides of the line are separated in a right-lateral sense by about 20 kilometers. Along this same line is a zone of strong diaphthoresis. The strong possibility of north-south faults must be seriously considered in any interpretation of the tectonic history of Lofoten and Vesterålen.

Suggested geologic history of Flakstadöy and speculations on regional relationships

On the basis of the evidence available, a tentative sequence of events can be postulated for the rocks on Flakstadöy:

1. A group of sedimentary rocks possibly accompanied by acid volcanic rocks was deposited in this area more than 2,800 million years ago, based on dating of similar rocks in other parts of Lofoten-Vesterålen by Heier & Compston (1969).

2. These rocks were deeply buried and recrystallized in an upper-amphibolite to lower-granulite facies environment between 1,800 and 2,800 million years ago, depending on which of the alternative hypotheses of Heier & Compston (1969) is more nearly correct. From the data of Heier & Compston (1969) it is possible to infer either one or more major metamorphic events, although the preferred hypothesis of these workers involves two separate high-grade events. For the rocks from Flakstadöy which they dated, Heier & Compston (1969, pp. 269–271) find no evidence of an event as old as 2,800 million years, obtaining from 'least altered' samples of monzonite and anorthosite a good isochron for 1,775 ± 30 million years. It must be emphasized, however, that in the determination of this isochron they included no banded granulites, which field evidence obtained in the present study clearly indicates to be the oldest rocks present.

3. On the order of 1,775 ± 30 million years ago (Heier & Compston 1969, pp. 269–271), a large basic complex was intruded along major fracture zones parallel to the regional trend of major folds. This intrusive episode led to partial fusion and local higher grade metamorphism of the layered gneiss. The mangeritic rocks are a result of this episode.

4. As the basic complex cooled, it was differentiated, leading to the zonation in rock types now visible.

5. There is as yet no clear evidence that either the Grenville (= Gothic) or the Caledonian orogenies had any effect on these rocks. However, Rb-Sr data for rocks at Lilleidet, on Vestvågöy, directly across Nappströmmen from Napp and separated by a distance of only about two km from granulite-facies rocks, lie on a possible isochron for 442 million years (Heier & Compston 1969, pp. 271–272). It is possible that age-dating of amphibolite-facies rocks, including biotite schists north of Napp may also give these younger ages, possibly indicating effects of the Caledonian orogeny.

An alternative interpretation would be to subdivide the third step, above, in the following way:

3. Either a magmatic mangerite was intruded into the complex of veined and layered gneisses or these gneisses were subjected to metamorphism of high enough grade so that the mangerite was produced through anatexis.

3a. The basic complex was later intruded in an event unrelated to the formation of the mangerite.

The very ancient ages of the oldest rocks in Lofoten (Heier & Compston 1969; Herr & Merz 1958) probably relates them to rocks of the Kenoran orogenic belts of the Canadian Shield, the pre-Karelian basement of Fenno-Scandia and the Ketilidian rocks of west Greenland (names and ages of orogenic belts based on Holmes 1965, pp. 368–369). The ages of the anorthositic rocks of Lofoten which from the work of Heier & Compston (1969) can probably be inferred to be about 1775 ± 30 million years can be related to the ages of anorthositic rocks of Labrador (Emslie 1965) and possibly of west Greenland and of the Kola peninsula. And yet the Lofoten belt is apparently surrounded on all sides by rocks giving Caledonian ages (Bryhni 1967). This would suggest that Lofoten may form a window of Precambrian showing through the Caledonian thrust plates and unaffected by Caledonian events or by the Grenville (= Gothic) events that affected Precambrian basements along the west coast of Norway (Bergen, Nordfjord, etc.) and South Norway (Egersund, etc.). These windows would be similar in geometry to the Rombak and Raipas windows (Reitan 1960). Furthermore, it could be suggested that the anorthosite at Öksfjord (Krauskopf 1954), south of Stjernöy, may also belong to this group of basement rocks, although it is unknown to the author whether or not these rocks have been overprinted with Caledonian ages. The apparent geologic 'fit' of northwestern Norway and east Greenland is not altered by the presence of this window.

The anorthosite-mangerite rock suite in Lofoten lines up well with the anorthosite-mangerite belts of Labrador, west Greenland, and the Kola peninsula, providing additional circumstantial evidence consonant with the idea that these areas may have formed part of a massive, pre-continental-drift craton (see Hurley & Rand 1969) in the interval from about 2500 million years ago until at least after the formation of this complex was completed and probably until after the Caledonian orogeny had terminated.

Herz (1969) has noted, as have many other workers, that the ages of anorthosites range from about 1100 to 1700 million years. The Flakstadöy body, not mentioned by Herz, would appear to fall into this same age range. However, as indicated by the discussion of anorthosite genesis in this paper, the Flakstadöy body could easily have been derived from a 'normal' basaltic or andesitic parent magma, and there is no need to call upon the production of a special anorthositic magma through some unique cataclysmic or thermal event such as the advent of the earth-moon system to account for its development, or for that matter for the development of many other anorthosite bodies (see Romey 1968).

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