A Structural, Stratigraphic and Petrologic Study of Anorthosites, Eclogites, and Ultramafic Rocks and Their Country Rocks, Tafjord Area, Western South Norway

H. K. BRUECKNER

Brueckner, H. K. 1977: A structural, stratigraphic and petrologic study of anorthosites, eclogites, and ultramafic rocks and their country rocks, Tafjord area, western South Norway. *Norges geol. Unders. 332*, 1–53.

The Tafjord area of the central basal gneiss region contains Caledonized metamorphic rocks divided into two groups. The 1000 m.y. old *Fetvatn Gneiss* is a homogeneous zone of granodioritic gneiss that was intruded as part of a large plutonic complex during the Sveconorwegian Cycle. The heterogeneous *Vikvatn Sequence* contains obvious metasedimentary and metavolcanic rocks and is comprised of the Øyen, Øvste Rødal and Svartegga groups. In addition, anorthosite, eclogite and anhydrous peridotite occur within the Øyen and Øvste Rødal groups. The quartzite-rich Svartegga group may form the basal (Eocambrian) part of the Caledonian supracrustal sequence typical of the eastern basal gneiss region. The Øyen and Øvste Rødal groups would then be Precambrian metasedimentary sequences separated from the Svartegga group by a major unconformity.

Structures include two early generations (F_{1A} and F_{1B}) of east-plunging, reclined, isoclinal, similar folds. Later (F_2) folds generally have a concentric geometry, disharmonic profiles and variable axes. The F_1 and F_2 fabrics are identical to structures in the eastern basal gneiss region that are believed to have formed during the Caledonian orogeny; but the F_{1A} fabric may be a heterogeneous assemblage of Precambrian structures that were homogenized (Caledonized) during intense F_{1B} deformation. The last-formed F_{3A} and F_{3B} subsystems are east-plunging chevron and concentric folds, respectively, that locally re-orient the rocks into the east-west structural trends typical of the western basal gneiss region.

Almandine–amphibolite facies metamorphism, with the first sillimanite isograd passing through Tafjord, accompanied the F_1 deformation. Some Øvste Rødal group rocks contain relics of a possible granulite facies metamorphism.

The anorthosites, ultramafic rocks, and eclogites contain all the structural elements found in the enclosing country rocks and possess retrogressed mineralogies that reflect almandine–amphibolite facies metamorphism at variable P_{H_2O}, and hence were in their present stratigraphic position before or during the development of the F_{1A} fabric. Scattered relict assemblages within these rocks indicate episode(s) of high-grade metamorphism and/or magmatism prior to the almandine–amphibolite facies metamorphism.

H. K. Brueckner, Department of Earth & Environmental Sciences, Queens College of the City University of New York, Flushing, New York 11367, U.S.A.

CONTENTS

Introduction 2
Part I: General geology of the gneisses and metasediments of the Tafjord area 3
Stratigraphy 3
Previous work 3

Fundamental binary subdivision 4
Fetvatn Gneiss 4
Vikvatn Sequence 5
Øyen group 6
Øvste Rødal group 8
Svartegga group 8
Correlations and relative ages 8
Structure 11
Minor structures 11
Folds 11
Brittle structures 14
Boudins 18
Major structures 18
Rødalsegga Reclined Fold 18
Kallskar and Kaldhusseter Reclined Folds 19
Tafjord Synform 19
Discussion 20
Petrology 24
General statement 24
Petrology of the mineral assemblages formed during the development of the F_1 structural fabric 24
Relict assemblages 26
Parent rocks 27
Metasomatism 27
Petrology of the mineral assemblages formed during F_2 faulting and subvertical fracturing 28
Petrology of mineral assemblages formed during mineralization of subvertical fractures 29
Part II: The anorthosites, ultramafic rocks and eclogites of the Tafjord area 30
General statement 30
Form and stratigraphic distribution 30
Structure 32
Minor structures 33
Folds 33
Brittle structures 36
Cataclastic features 38
Boudins 38
Major structures 40
Petrology 40
Anorthosite 42
Ultramafic rock 42
Dunite 42
Garnet peridotite and garnet pyroxenite 43
Eclogite 44
Discussion and interpretation 45
Petrology of mineral assemblages formed during the development of the F_1 structural fabric 45
Relict assemblages 47
Possible origin of the anorthosite, ultramafic rock and eclogite of the basal gneiss region 49
Acknowledgements 51
References 51

Introduction

Tafjord (Fig. 1) is a small fishing and farming community that lies at the head of Storfjord in western Norway (62°22'N, 7°28'E). The area occurs in the middle of the basal gneiss complex of Holtedahl (1944) and contains

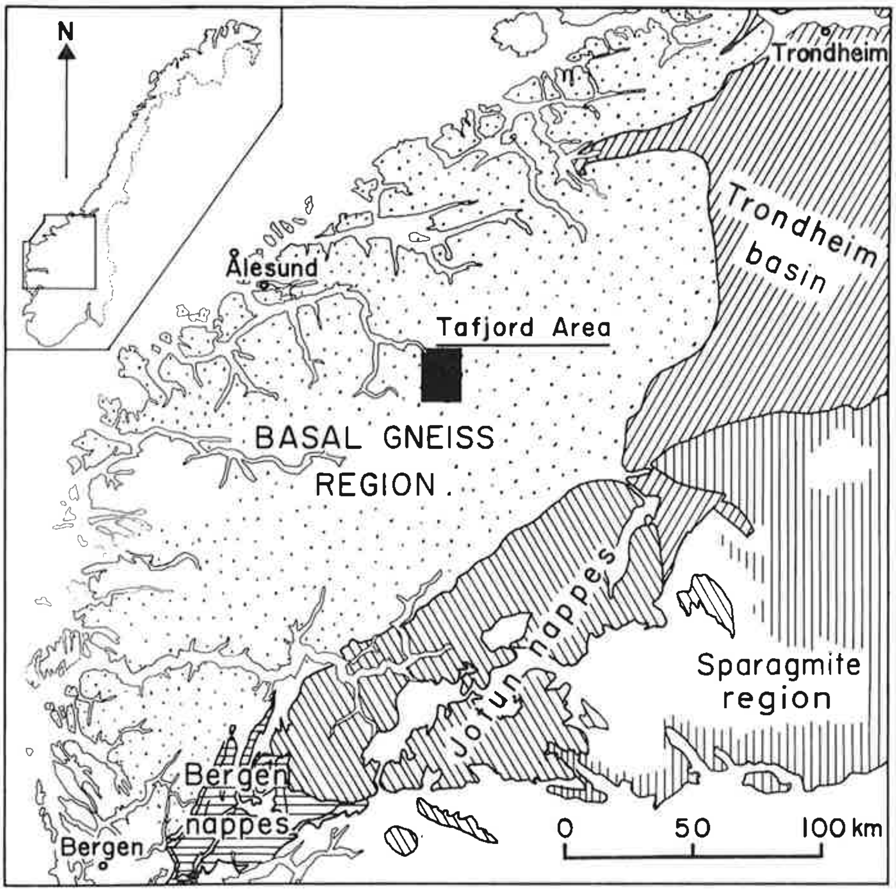


Fig. 1. Simplified geological map of western south Norway showing major tectonic provinces. The insert shows the area of Norway covered by the figure.

Caledonized metamorphic rocks that are stratigraphically, structurally and petrologically transitional between rocks to the east and those to the west. This report is divided into two parts. The first contains descriptions and interpretations of the country rocks around Tafjord. The second part describes the stratigraphic, structural and petrologic relationships of the anorthosites, ultramafic rocks and eclogites to the enclosing country rocks.

Part I. General geology of the gneisses and metasediments of the Tafjord area

Stratigraphy

PREVIOUS WORK

Strand (1949) was the first to describe quartzites in Grotli, just to the south of Tafjord, which he believed represent transformed Caledonian sediments. Gjelsvik (1953) states that these quartzites ('flagstones') can be traced from

Vågå, in Gudbrandsdalen, through Skjåk and Grotli as far as Rødalsegga in the Tafjord area, where, he claims, they fold back on themselves and trend eastward. Muret's map (1960) shows a similar trend for what he terms Eocambrian and Cambro–Silurian metasediments through Tafjord and Grotli. However, he connects these metasediments with ones found near Lesjaskog in Romsdal, rather than with the quartzites in Gudbrandsdalen. O'Hara & Mercy (1963) also describe metasediments in the Tafjord area. Strand (1949) and O'Hara & Mercy (1963) suggest that the east-dipping section trending from Grotli through Vikvatn is right side up. Muret (1960) and Gjelsvik (1953), however, map the section as inverted.

FUNDAMENTAL BINARY SUBDIVISION OF TAFJORD AREA ROCKS

The basic subdivision of the Tafjord area rocks is into two broad units: a series of relatively monotonous banded gneisses and a sequence of more varied rock-types, some of which are metasediments. The monotonous gneisses were formerly called the Grotli Gneiss (Brueckner 1969). However, the subsequent careful mapping by Strand (1969) shows that the Grotli Hotel rests upon a variegated sequence of rock-types that should not be correlated with the monotonous gneisses of the Tafjord area. The Grotli Gneiss is therefore renamed the Fetvatn Gneiss for the area around Fetvatn where the rocks of this unit have been dated (Brueckner et al. 1968). The sequence of more varied rock-types is here called the Vikvatn Sequence for the area around Vikvatn where these rocks are well exposed. The Vikvatn Sequence is divided informally into the Øyen group, the Øvste Rødal group and the Svartegga group.

FETVATN GNEISS

The Fevatn Gneiss is predominantly granitic to granodioritic in composition; the major minerals are quartz, microcline, plagioclase, biotite and some hornblende. On Fetegga, muscovite-biotite schist occurs. Layers of amphibolite occur sporadically, but differ from comparable rocks in the Vikvatn Sequence in that they lack garnets. The gneisses are commonly strongly banded; light layers rich in quartz and feldspar contrast sharply with dark grey layers rich in biotite. In some zones, the biotite is more uniformly distributed, producing a well-foliated light grey gneiss. Rare layers are rich enough in biotite to be called schists. Finally, certain zones contain microcline feldspar augen up to 15 cm long. These augen-rich zones generally grade into the more common augen-free gneisses along strike. Pegmatites occur in scattered localities and are commonly conformable to the foliation. Cross-cutting pegmatites are infrequent.

The distribution of the Fetvatn Gneiss west of Vikvatn is shown on Plate 1. The lower (western) contact has not been observed, and the Fetvatn Gneiss may extend west beyond Geiranger, where the rocks are similar (Gjelsvik 1951, Bryhni 1966). Microcline-rich gneisses occur in the northeastern part of the Tafjord area that are similar in composition and texture to the Fetvatn

Gneiss. Structurally below these gneisses is a series of quartzites and other metasediments similar to those structurally above the Fetvatn Gneiss at Vikvatn. Structural evidence indicates that the stratigraphy in the north-eastern part of the area is inverted relative to the stratigraphy in the western part, and the gneisses are therefore tentatively correlated with the Fetvatn Gneiss.

VIKVATN SEQUENCE

Most of the Vikvatn Sequence also consists of granitic and granodioritic gneisses. The major minerals are quartz, microcline, plagioclase and biotite. These rocks are difficult to distinguish from the Fetvatn Gneiss in the field, but they are texturally more variable, with schists and augen gneisses more common. Banding in the Vikvatn Sequence gneisses is more irregular, and the contrast between the light and dark layers is not as sharp as in the Fetvatn Gneiss. Hornblende is a relatively common accessory mineral in the Vikvatn gneisses, whereas the Fetvatn Gneiss is more hornblende-poor. The presence of garnet in some of the more plagioclase-rich gneisses is especially diagnostic, for it is the only abundant mineral not found in the Fetvatn Gneiss. As these criteria for differentiating the gneisses of the two units are tenuous, much of the mapping and subdividing of the Vikvatn Sequence is based on the presence or absence of the following readily recognizable rock-types.

Quartzites contain up to 80% quartz, with microcline and muscovite as the important accessories. The rock is banded, zones almost free of muscovite alternating with zones rich in muscovite. This layering, plus a parallel schistosity, causes the rock to break up into thin slabs or 'flags' upon weathering. The quartzites may grade into granitic gneisses by addition of feldspar and biotite both perpendicular and parallel to the compositional layering, and are therefore discontinuous along strike.

Calc-silicate rocks occur as thin (up to 0.3 m thick) layers closely associated with quartzites. In addition to quartz and microcline, they may contain diopside, tremolite, epidote, biotite, calcite and scapolite. These minerals give the rocks a greyish-green to dark green color that is easily recognized in the field. Marble occurs at one locality (Øyen).

Two-mica gneisses contain abundant, coarse-grained muscovite in addition to biotite. Banding in the gneiss is weak; the micas are evenly distributed throughout the rock. These rocks are commonly associated with quartzites, and form important stratigraphic markers where quartzites are rare or missing.

Mafic rocks, including amphibolites, garnet amphibolites and hornblende schists, occur as pods, lenses and layers intercalated between plagioclase-rich gneisses. These rocks are commonly strongly lineated, but the development of schistosity is variable. Plagioclase, hornblende and biotite are the most important minerals, but garnets may form up to 40% of the rock. The garnet-bearing amphibolites are easily distinguished from the garnet-free amphibolites in the Grotli Gneiss. They may, however, be confused with retrogressed eclogites.

Syenitic or felsic gneisses are recognized by their light color. They consist of plagioclase and microcline with very little quartz. Mafic minerals make up less than 15% of the rocks, and include biotite, hornblende, clinopyroxene and garnet. The rocks are texturally complex, ranging from well lineated and foliated gneiss to massive rocks with hornfelsic textures containing orbicular clots and schlieren of mafic minerals.

Aluminum-rich rocks are relatively rare in the Vikvatn Sequence. They weather to a strong rusty color and are commonly found near quartzites. They are the only rocks with enough excess aluminum to form kyanite and sillimanite, or corundum, in addition to garnet. Kyanite and sillimanite rarely make up more than 5% of the rock, and cannot be detected without a microscope. Thus, they have been recognized in only a few localities.

Anorthosites, ultramafic rocks and eclogites are closely associated with the metasediments of the Vikvatn Sequence. Their distribution is stratigraphically controlled, and they therefore serve as important stratigraphic markers.

On the basis of the distribution of the rock-types listed above, the Vikvatn Sequence in the western part of the Tafjord area can be divided into three major informal groups.

(1) Øyen group

Fig. 2 illustrates, in structural succession, the rock-types found in the lower part of the Vikvatn Sequence in the Øyen valley. Two relatively thin sections of metasedimentary rocks are separated by a thick zone of microcline-rich gneisses. Three formations are thus recognized.

The Kaldhusseter formation is only 100 m thick in Øyen valley, but it thickens considerably to the south and is on the order of several hundred metres thick east of Kaldhussetervatn. The lowest quartzite of the Vikvatn Sequence is separated from the Fetvatn Gneiss by a zone of rusty-weathering, garnet-bearing and garnet-free plagioclase gneisses. The contact between the Kaldhusseter formation and the Fetvatn Gneiss is placed below the lowest occurrence of garnet-bearing plagioclase gneiss. There are at least two layers of quartzite in this formation. Above the quartzites is a zone of biotite-quartz-plagioclase ± hornblende gneisses, hornblende schists, garnet-bearing plagioclase gneisses and garnet amphibolites (retrogressed eclogites?). Intercalated within this mafic sequence are two anorthosite layers, very thin near Øyen but at least 100 m thick above Kaldhussetervatn, and ultramafics, the largest of which are the Kaldhusseter and Kallskar bodies. The upper part of the Kaldhusseter formation consists, at least locally, of a zone of two-mica gneiss. The Kaldhusseter formation displays great lateral lithologic variation. For example, the quartzites are thin or lacking near Onilsavatn, although they recur south of Kallskaret and are quite thick and common near Vikvatn.

The Nyk formation is about 400 m thick near Øyen, but is considerably thinner to the south. It is predominantly a sequence of biotite-quartz-feldspar gneisses. Near the base, it is rich in biotite-quartz-plagioclase and hornblende-biotite-quartz-plagioclase gneisses, interlayered with less common rusty-weathering schists. The middle and upper parts of the sequence consist

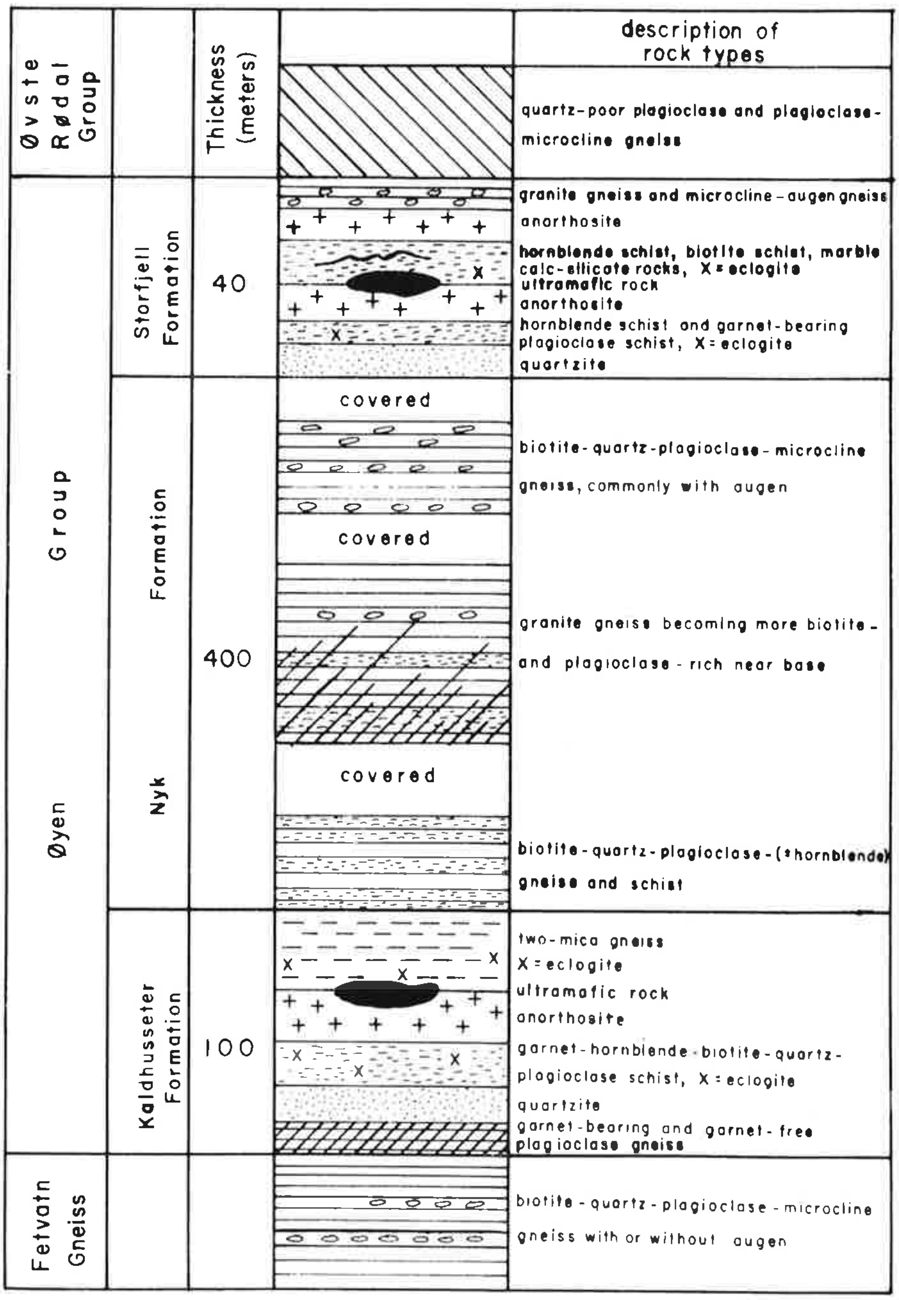


Fig. 2. Structural successions of rock-types, Øyen valley, Tafjord area.

predominantly of biotite-quartz-microcline gneisses, commonly in the form of augen gneisses.

Most of the Storfjell formation is well exposed at Øyen where it is 40 m thick. It also forms most of the rocks found just to the west of the summit of Storfjell. It includes a basal quartzite, two layers of anorthosite, and mafic gneisses. The two anorthosite layers, each about 10 to 15 m thick, are separated by hornblende schists, ampibolites (retrogressed eclogites?), some bands of calc-silicate rocks and a thin layer (up to 10 cm thick) of marble. Ultramafic rocks are associated with the lower anorthosite layer.

(2) Øvste Rødal group

The 700 to 1000 m-thick Øvste Rødal group occurs on the eastern flank of Storfjell, underlies Øvste Rødal, and continues about three-quarters of the way up the western side of Rødalsegga. It consists predominantly of biotite-quartz-plagioclase, hornblende-biotite-quartz-plagioclase and garnet-bearing biotite-quartz-plagioclase ± hornblende gneisses. Microcline-rich gneisses, commonly containing hornblende, occur in scattered localities. The most diagnostic rocks of this group are the felsic or syenitic gneisses. Foliate syenitic gneisses are found throughout the group, but some non-foliate rocks are also found near an abandoned mine south of Zakariasvatn in Øvste Rødal. Anorthosites and ultramafic rocks also occur in this group, but are not as abundant as in the Øyen group.

(3) Svartegga group

The lithologic details and thickness of the Svartegga group are less well known than those of the Øyen and Øvste Rødal groups. The basal part of the Svartegga group is marked by a thick, quartzite-rich section interlayered with aluminum-rich, rusty-weathering schists, hornblende schists, garnet-bearing and garnet-free hornblende-biotite-quartz-plagioclase gneisses and amphibolites. The zone of quartzites is the thickest mapped in the Tafjord area. If the structural interpretations of the Tafjord area are correct, then the quartzites on top of the ridge separating Daurmaalsvatn and Svarteggvatn are the same as those exposed to the west and southwest of Daurmaalsvatn, and are repeated by folding. Above the quartzites is a thick zone of two-mica gneisses and amphibolites. The lower contact of the Svartegga Group is mapped just below the lowest quartzite. Unlike the other two groups of the Vikvatn Sequence, the Svartegga Group does not contain anorthosite, eclogite or unaltered dunite or peridotite, although it does contain hydrous ultramafic rock (serpentinite, talc-tremolite schist) typical of those that occur in the eastern basal gneiss region.

CORRELATIONS AND RELATIVE AGES

Strand (1949, 1969) and Muret (1960) suggested that the rocks of the Tafjord area may be correlated with those in the eastern basal gneiss region which have, in turn, been correlated with unmetamorphosed or weakly metamorphosed rocks in adjacent areas. In Trollheimen, for example, Hansen (1971) differentiates a basement gneiss (Trollhetta Granite Gneiss), a quartzite-rich unit (Gjevilvatn group) and a schist-amphibolite unit (Blåhø group). He correlates the Gjevilvatn group and the Blåhø group with the Eocambrian sandstones and arkoses of the Sparagmite region and the Cambro-Silurian pelitic rocks and volcanics of the Trondheim basin, respectively.

Scott (1967) and Wheeler (1973) describe very similar stratigraphies in areas to the south in eastern Dovrefjell. Scott (1967), however, suggests the additional presence of a sequence of plagioclase-rich gneisses (Tveratind group) that separates the Gjevilvatn and Blåhø groups. Muret (1960) and others suggest that the gneisses of granitic to granodioritic composition underlying the lowermost Eocambrian of early Paleozoic metasedimentary rocks may form part of the Precambrian basement.

It is tempting to correlate the Vikvatn Sequence, with its obviously metasedimentary units, with the Eocambrian and early Paleozoic rocks of the eastern basal gneiss region, and the more monotonous Fetvatn Gneiss with the Precambrian basement. An Rb-Sr whole-rock isochron from samples of the Fetvatn Gneiss collected between Kaldhussetervatn and Fetvatn yields an apparent age of 1000 ± 150 m. y.[1] (Brueckner et al. 1968). Nearly identical ages (Bryhni et al. 1971, Brueckner 1972, Priem et al. 1973) have been determined from very similar rocks (the Jostedal Complex of Bryhni 1966) in the western basal gneiss region. These ages are similar to those obtained on rocks of the Sveconorwegian Province in the Precambrian shield of southern Norway (Kulp & Neuman 1961). Thus, lithology and Rb-Sr whole-rock isochron ages indicate that the Fetvatn Gneiss and the rest of the contiguous Jostedal Complex is Precambrian in age, although this conclusion has been disputed (Hernes 1970).

The correlation of the Vikvatn Sequence rocks with Caledonian supracrustal rocks is less certain. Some broad correlations seem possible. For example, the Øyen group is sufficiently rich in quartzites (particularly near its base and top), calc–silicate rocks, marble and microcline gneisses to be comparable with the Gjevilvatn group in Trollheimen and Dovrefjell, thus supporting an Eocambrian age. However, correlations between the structurally higher units from the three areas are far more tenuous. For instance, the amphibolites and two-mica gneisses in the upper part of the Svartegga group are lithologically similar to Blåhø group rocks in Trollheimen and Dovrefjell, suggesting a possible correlation, but the thick quartzites near its base have no lithologic equivalents above the Eocambrian in the eastern gneiss region.

The lithologies of the western basal gneiss region in general, and the Tafjord area in particular, may differ significantly from those in the eastern basal gneiss region. Carswell (1973) presents extensive arguments that suggest that some of the metasedimentary rocks of the western basal gneiss region had a pre-Caledonian origin. Specifically, Carswell proposes that the metasedimentary sequences that contain associated eclogites, garnetiferous ultramafic rocks, granulites and calcic anorthosites may have originated as long as 2,800 m.y. ago, and may be comparable to the granulite-facies Scourian rocks of the Lewisian basement of northwest Scotland.

Recent radiometric evidence appears partially to support Carswell's (1973) view. Bryhni et al. (1971), Brueckner (1972), Mysen & Heier (1972) and

[1] All quoted Rb–Sr isochron ages are based on or recalculated to the Rb^{87} decay constant of $1.39 \times 10^{-11} yr^{-1}$.

Pidgeon & Råheim (1972) report ages ranging from 1100 to 1860 m.y. from rocks previously interpreted to be Caledonian supracrustal rocks. For example, Pidgeon & Råheim (1972) interpret Rb–Sr whole-rock isochron ages and U–Pb zircon ages of 1600 to 1780 m. y. from the Kristiansund and Frei Groups in Nordmøre as re-equilibration ages during a Precambrian almandine–amphibolite facies metamorphism of the region. This interpretation clearly contradicts the earlier held view (Hernes 1967) that the rocks of the Kristiansund and Frei Groups had a late-Precambrian (Eocambrian) origin. However, an alternative interpretation of these apparent dates is that they are relict ages, dating the terrain from which the sediments were originally derived (Brueckner 1972). Under this model, the metasediments of the Kristiansund and Frei Groups could still have been deposited during the Eocambrian as proposed by Hernes.

Nevertheless, it is clear that the previous assumption that all metasedimentary rocks in the basal gneiss region had a Caledonian origin has to be discarded. It is likely that there are two generations of supracrustal rocks in the basal gneiss region. The metasedimentary and metavolcanic sequences in the eastern basal gneiss region, and possibly those near the large Devonian synclines in the western basal gneiss region (Kolderup 1960a), may have had a Caledonian origin. However, many of the other supracrustal sequences in the western basal gneiss region, particularly those associated with eclogite facies rocks and anorthosites, may have had a Precambrian origin, although there is, as yet, no radiometric evidence for an origin before 1800 m. y. ago. There may, therefore, be two generations of ultramafic rocks in the basal gneiss region and adjacent areas: one associated with the Lower Paleozoic rocks of the eastern basal gneiss region and Trondheim basin (highly hydrated serpentinites and talc–tremolite schists) and another associated with Precambrian supracrustal rocks (unusually fresh dunites and peridotites that may contain garnet peridotites and garnet pyroxenites).

The Øyen and Øvste Rødal groups of the Vikvatn Sequence in the Tafjord area contain abundant eclogites, anorthosites and unaltered ultramafic rocks, and hence may be Precambrian. The Svartegga group differs from the Øyen and Øvste Rødal groups, however, by lacking anorthosites, eclogites, and unaltered dunites and peridotites. Further mapping is required to verify this observation. If true, it would suggest that the Svartegga group may have had a different origin than the Øyen and Øvste Rødal groups. The thick quartzites and two-mica gneisses near its base resemble the Eocambrian metasediments in the eastern basal gneiss region (Hansen 1963, Scott 1967). Muret (unpublished data) maps rocks similar to the Cambro–Silurian rocks of the Trondheim basin stratigraphically above the Svartegga group. It is, therefore, possible that the contact between the Svartegga group and the Øvste Rødal group represents an unconformity between a Precambrian supracrustal sequence to the west, and a Caledonian metasedimentary sequence to the east.

The relation between the (?) 1800 m. y. old Øyen and Øvste Rødal groups and the 1000 m. y. old Fetvatn Gneiss is unclear. The lack of obvious metasedimentary rocks and the generally homogeneous aspect of the Fetvatn Gneiss suggests an igneous origin. Possibly the Fetvatn Gneiss intruded the older Øyen and Øvste Rødal groups of the Vikvatn Sequence as part of a batholith complex (the Jostedal Complex) during the Sveconorwegian Cycle. Recent radiometric evidence by O'Nions & Baadsgaard (1971) and O'Nions & Heier (1972) suggests that relics of 1800 m. y. old (Svecofennian) rock occur in the Kongsberg and the Bamble areas in the eastern part of the Precambrian Shield of southern Norway. They suggest that these older rocks were recrystallized and invaded by younger igneous rocks during the 1000 m. y. old Sveconorwegian Cycle. A similar sequence may have affected the rocks of the Tafjord area prior to the Caledonian orogeny.

Structure

MINOR STRUCTURES

Small-scale folds are the most easily classified structural elements of the Tafjord area, and have been subdivided, following the conventions of Hansen (1971), into systems on the basis of style, associated fabric elements and age relationships. Mineral lineations, schistosities and other structures that are genetically related to folds are discussed in terms of this relationship. Other fabric elements, such as boudins and joints, not related in an obvious way to folds, are discussed separately.

Folds

Table 1 and Figs. 3 through 6 describe the important characteristics of folds of the Tafjord area. The earliest and most pervasively developed structures are the isoclinal folds, mineral lineation and regional schistosity of the F_{1A} subsystem (Table 1). All other structures in the Vikvatn Sequence are superimposed on the F_{1A} fabric. F_{1B} and F_{1A} are geometrically identical, and are classified under the same system. F_{1B} folds are distinguished from F_{1A} folds by the presence of two schistosities; one that wraps around the F_{1B} fold hinge parallel to the rock compositional layering (the relict F_{1A} schistosity), and another weaker one that parallels the F_{1B} axial plane. Although the axes and planes of F_{1A} and F_{1B} folds are close to parallel (Figs. 3 and 4), their senses of asymmetry are opposite throughout most of the Tafjord area. F_{1A} and F_{1B} folds are the same in style, orientation, associated fabric elements and age relationships as two sets of folds found in the eastern basal gneiss region (Banham & Elliott 1965, Scott 1967, Wheeler 1973) where they are associated with the development of two sets of Caledonian major recumbent folds (Scott 1967).

F_{2A} folds deform the elements of the F_1 fabric (Fig. 5A) and are common only in areas where there is a strong difference in the competence of the deformed beds. They are locally associated with east-dipping faults. F_{2B} folds occur locally, and are associated with a weak axial-surface schistosity and mineral lineation parallel to the fold axis. F_{2B} folds are considered to be related

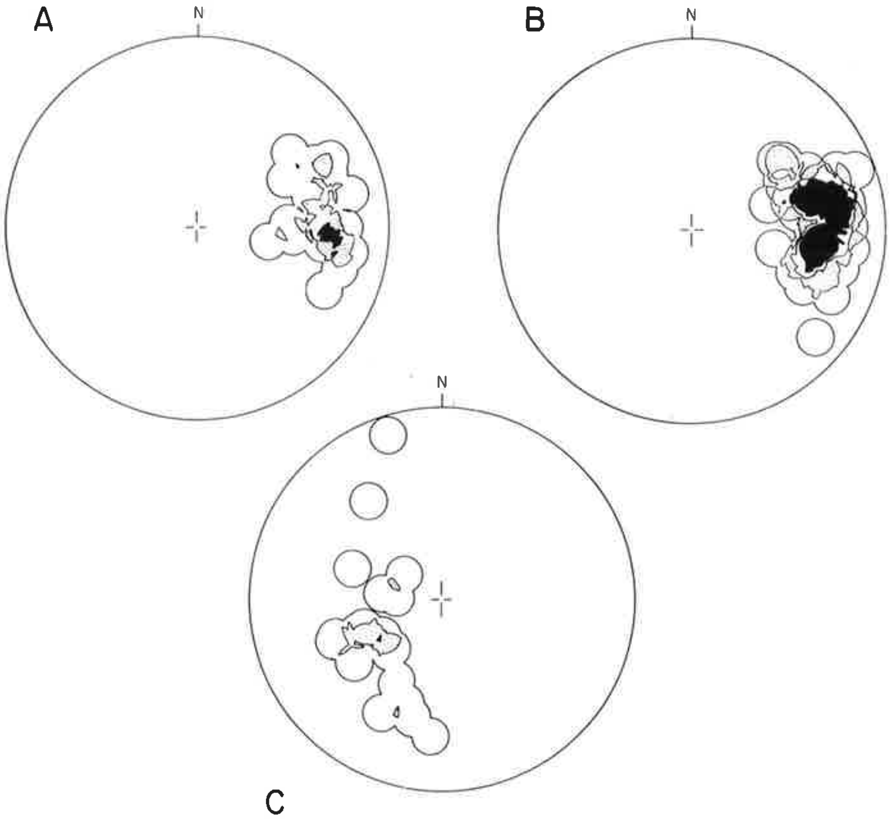


Fig. 3. Orientations of elements of the F_{1A} fabric system. A. 23 F_{1A} fold axes contours: 4.3, 13.0, 21.7% per 1% area. B. 56 L_{1A} mineral lineations; contours: 1.8, 5.4, 8.9, 12.5% per 1% area. C. 20 poles to F_{1A} axial surface schistosities; contours: 5, 10, 15% per 1% area.

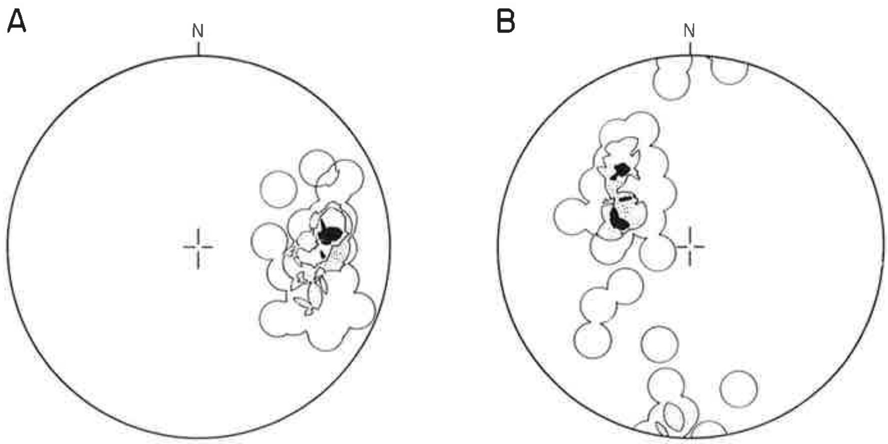


Fig. 4. Orientations of elements of the F_{1B} fabric system. A. 28 F_{1B} fold axes; contours: 3.6, 10.7, 17.8, 25.0% per 1% area. B. 33 poles to F_{1B} axial surface schistosities; contours: 3.0, 9.0, 15.2% per 1% area.

ɔle 1. Minor fold systems and subsystems of the Tafjord area

ɔystem	Sub-system	Description	Orientation* Trend and plunge of fold axes	Strike and dip of axial plane	Related fabric elements
F_1	F_{1A}	Tightly appressed, recumbent, isoclinal folds with similar geometry.	E–W/E	NNW/ENE	Strong axial–surface schistosity and mineral lineation parallel to fold axes
	F_{1B}	Less appressed, recumbent, isoclinal folds. Sense of asymmetry locally opposite to F_1 folds.	E–W/E	NNW/ENE	Weaker axial–surface schistosity and mineral lineation parallel to fold axes.
F_2	F_{2A}	Concentric folds with disharmonic profiles.	Cleft NNW/ENE girdle distribution with north-plunging maximum	Variable	NNW/ENE shear zones with retrograded mineral assemblages
	F_{2B}	Geometry somewhat more similar than F_2 folds.	»	»	Very weak axial–surface schistosity and mineral lineation parallel to fold axes
F_3	F_{3A}	Chevron folds.	Diffuse east-plunging maximum	Variable common east-plunging intersection	Local fracture cleavage
	F_{3B}	Open, concentric folds.	»	»	

ːxcept where reoriented by later-developed structures.

to the far more abundant F_{2A} folds that lack these features. Some F_{2A} fold axes measured near Vikvatn are plotted in Fig. 5B. Although most of the axes plunge north, some are spread into a partial great circle that strikes north-northwest and dips gently to the east, parallel to the local compositional layering. The axes of the F_{2A} folds are variable not only from fold to fold but actually within many individual folds. The undulating hinge of a single fold may vary in orientation by as much as 40°. Fig. 5 plots some of the axes of F_{2A} folds measured at Vikvatn, Øyen and Kaldhusseter. Included in the plots is the shear sense of each fold using the conventions of Hansen (1971). Most, if not all, of the folds along one portion of the girdle defined by the fold axes have the same shear sense, whereas those in the remaining part of the girdle have the opposite shear sense. The gap on the girdle between the last fold of one shear sense and the first fold of the opposite shear sense is defined as the separation arc (Hansen 1971). Also included in the plots of Fig. 5 is the unrotated, previously-formed mineral lineation (L_1) from each locality.

F_{2A} folds have been described in detail and interpreted by Hansen (1971) in Trollheimen (Discfolds) and by Scott (1967) and Wheeler (1973) in

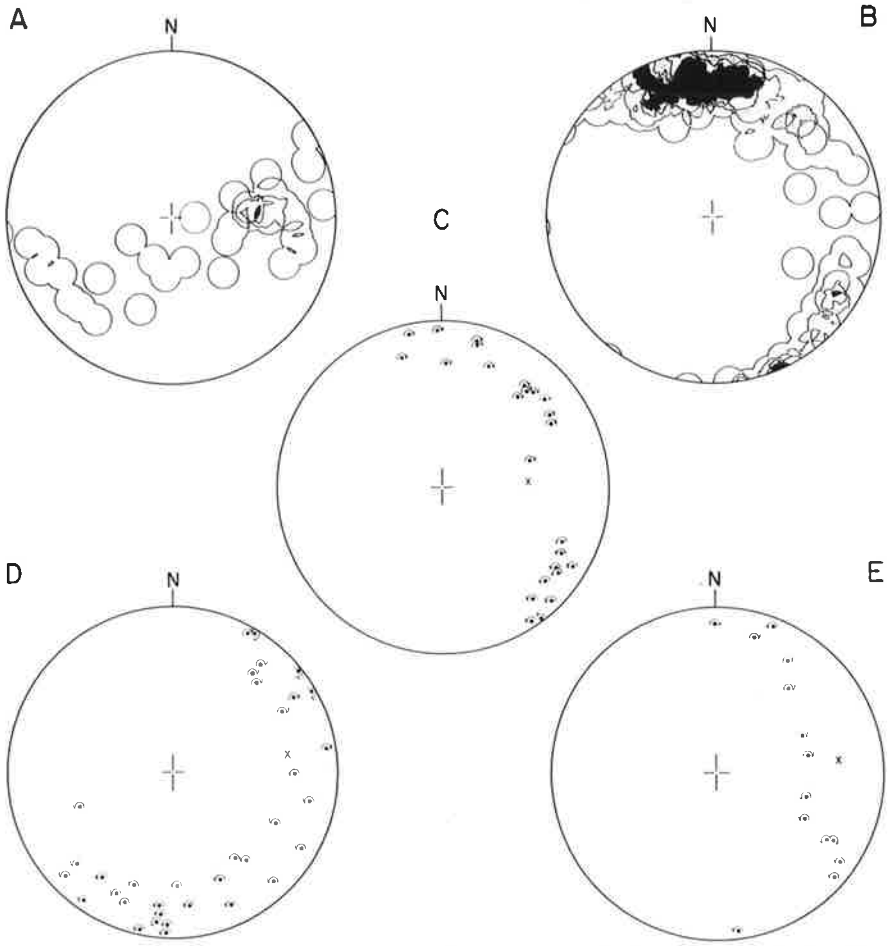


Fig. 5. Orientations of elements of the F_2 fabric system. A. 41 L_{1A} mineral lineations rotated by F_{2A} folds; contours: 2.4, 7.3, 12.2, 17.1% per 1% area. B. 118 F_{2A} fold axes measured near Vikvatn; contours: 0.8, 2.5, 4.2, 5.9% per 1% area. C, D, and E. Orientations and shear sense distributions of F_{2A} fold sets near Vikvatn (C), Øyen (D), and Kaldhussetervatn (E). X = unrotated F_{1A} mineral lineation.

Dovrefjell. Banham & Elliott (1965), Muret (1960), Gjelsvik (1951) and Lappin (1966) refer to them in other portions of the basal gneiss region.

The F_{3A} and F_{3B} subsystems consist of east-plunging chevron folds and open, concentric folds, respectively (Fig. 6). F_{3A} folds are commonly restricted to pelitic rocks, whereas F_{3B} folds occur in both pelitic and psammitic rocks, but tend to be more common in the latter. Otherwise, the two fold types differ only in geometry and are probably genetically related. F_{3A} folds occur in the eastern basal gneiss region (Hansen 1963, Scott 1967).

Brittle Structures

Low-angle fault zones are associated with F_{2A} folds at Vikvatn (Fig. 7). Most of the fault planes parallel the girdle defined by the F_{2A} fold axes and are

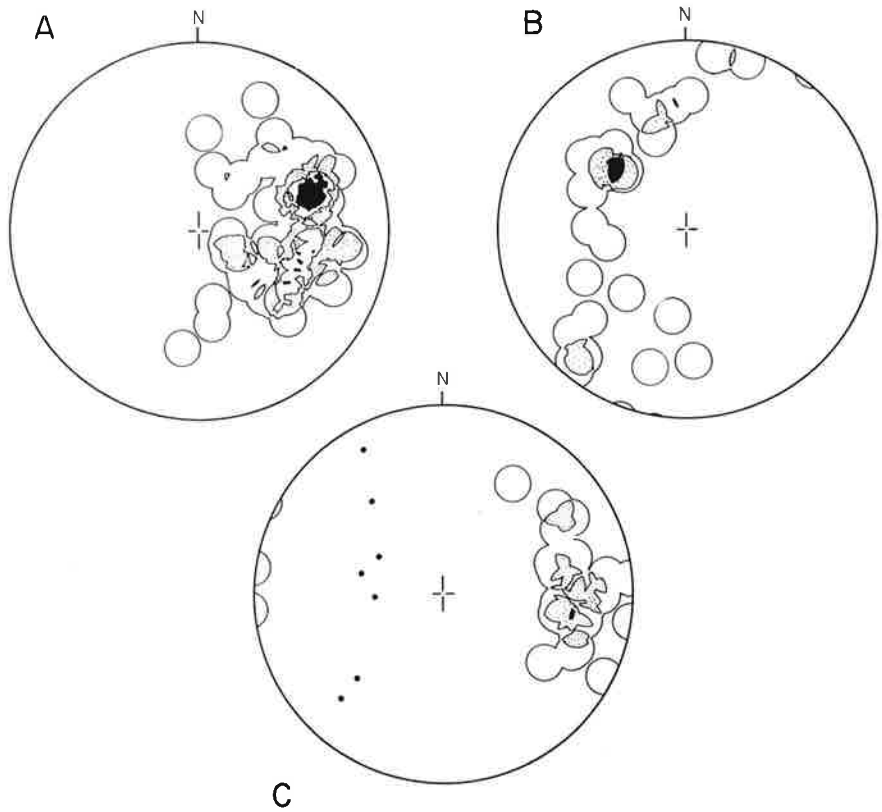


Fig. 6. Orientations of elements of the F_3 fabric system. A. 70 F_{3A} fold axes; contours: 1.4, 4.3, 7.1, 10.0% per 1% area. B. 32 poles to F_{3A} axial planes; contours: 3.1, 9.4, 15.6% per 1% area. C. 23 F_{3B} fold axes; contours: 1.8, 5.4, 8.9% per 1% area. Dots represent poles to F_{3B} axial planes.

concentrated near the limbs or the axial surfaces of the more appressed F_{2A} folds. Some faults are refolded about a north-plunging axis (Fig. 7B), demonstrating the synchronous development of the folds and the fault zones.

The rocks near the low-angle thrusts show cataclastic to mylonitic textures. Thin-sections show granulated, fine-grained streaks of quartz and muscovite that swirl around and between streamlined augen of microcline and saussuritized plagioclase. The long axes of these augen commonly define a lineation plunging to the east. Banham & Elliott (1965) describe a similar association of thrusts, folds and lineations from the Hestbrepiggan area.

Fig. 7C plots 300 quartz c-axes measurements of a sheared quartzite taken from three mutually perpendicular planes and rotated into a plane parallel to the fault plane. The c-axes define two maxima, subtending an angle of about 60° to 70°, perpendicular to the milled augen lineation. The acute bisector of the maxima is contained in the F_{2A} fault plane and parallels the maximum defined by the F_{2A} fold axes. Thus, the quartz fabric bears a symmetrical relationship to the F_{2A} fabric, suggesting that the quartz recrystallized during the F_2 deformation.

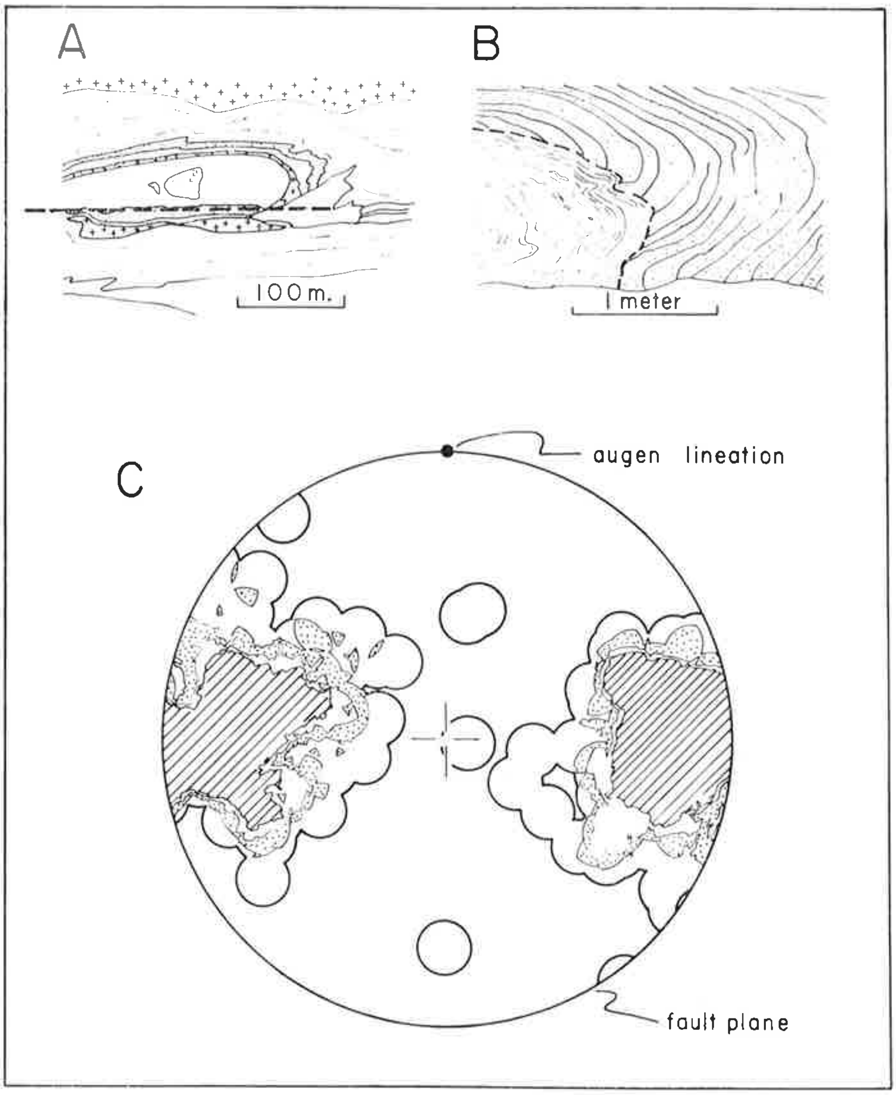


Fig. 7. Shear zone structures associated with 2_{2A} folds. A. The lower limb of a major F_{2A} fold at Vikvatn cut by a low-angle fault. B. Minor F_{2A} fold at Vikvatn associated with folded shear zone. Heavy dashed line = fault. Stippled pattern = quartzite. Crossed pattern = anorthosite. Dashed pattern = schist. Small folded ultramafic lens occur in the core of the major F_{2A} fold. C. 300 quartz c-axes measured in three mutually perpendic ular planes and rotated into plane parallel to F_{2A} fault. Contours: 0.3, 1.0, 1.7, 2.3% per 1% area.

Dilated gash fractures filled with pegmatitic material are exposed just above the water on the eastern side of Tafjord. They strike a little east of north, and dip at 60° to 70° to the west. They form attenuated lenses less than 2 m in length. Generally, they are oriented perpendicular to the regional, east-plunging linear elements of the F_1 system, and presumably formed more or less synchronously. Schmitt (1963) describes analogous structures from the Eiksund area of western Norway.

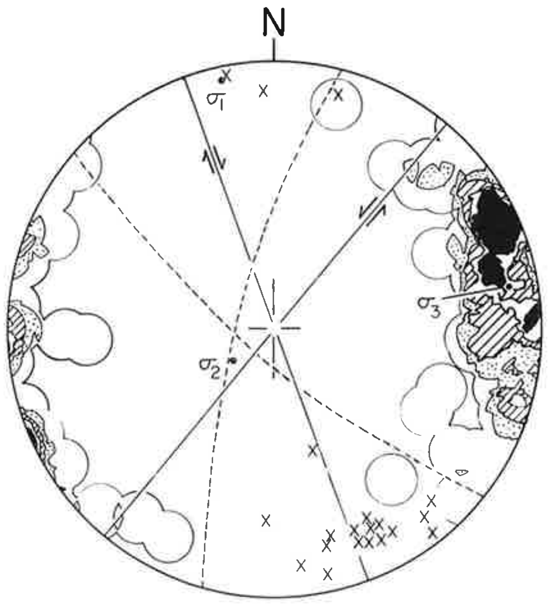


Fig 8. Orientations of 89 poles to late subvertical fractures; contours: 1.1, 3.4, 5.6, 7.9, 10.1% per 1% area. X = slickenside directions within fracture planes. Dashed and solid great circles represent average orientations of conjugate shear fractures from Dovrefjell (Scott 1967) and Eiksund (Schmitt 1963), respectively. σ_1, σ_2, and σ_3 = inferred orientations of principle stress axes.

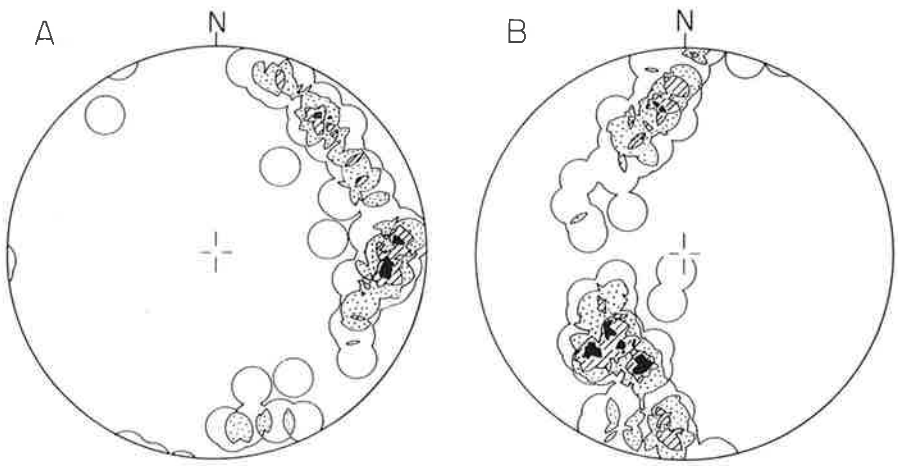


Fig. 9A. Orientations of 60 boudin lines measured throughout Tafjord area; contours: 1.7, 5, 8.3, 11.7% per 1% area.
Fig. 9B. Orientations of 79 poles to foliation measured throughout the Tafjord Synform; contours: 1.3, 3.8, 6.3, 8.9% per 1% area.

Late subvertical fractures are common throughout the Tafjord area (Fig. 8). The fractures are mineralized predominantly with chlorite, epidote and stilbite. Locally, abundant slickensides plunge to the south at gentle angles (Fig. 8), suggesting more or less subhorizontal relative displacements. Subvertical fractures were observed by Scott (1967) in Dovrefjell and Schmitt (1963) in Eiksund. However, both workers found that the fractures tended to form into two sets (Fig. 8), a feature not readily apparent in the Tafjord area.

Boudins

Fig. 9a plots boudin axes measured throughout the Tafjord area. They define a great circle with a weak point maximum plunging to the east. Muret (1960) noticed that boudin lines tend to parallel regional lineations in much of the basal gneiss region. The boudin lines in Trollheimen (Hansen 1963) and Dovrefjell (Scott 1967), however, are preferentially oriented at high angles to the east-plunging regional lineations.

At a road-cut just west of Onilsavatn, a zone of boudined hornblende schist is folded by a F_{1B} fold, suggesting that the boudins formed before or during the F_{1B} deformation. At another road-cut, east of Kaldhussetervatn, F_{1A} folds so highly appressed that the two limbs are in contact with each other have been boudined. On the basis of just these two observations, it appears that the boudins formed synchronously with the fabric of the F_1 system. Locally, pegmatites containing quartz, plagioclase, microcline, biotite and muscovite are concentrated near the hinge lines of boudins, indicating that temperatures must have been high during their formation.

MAJOR STRUCTURES

Plate I presents the distribution of the gneisses and metasediments of the Tafjord area. The attitudes of the various fabric elements of the minor structures are also plotted. Because the stratigraphy in some portions of the area is tentative, the proposed interpretation of some of the major structures must also be regarded as tentative. For this reason, the major structural interpretation is presented separately on Fig. 21.

Rødalsegga Reclined Fold

The Rødalsegga Reclined Fold, named after the ridge where the quartzites of the Svartegga Group fold back on themselves, is the dominant structural feature of the Tafjord area. The Rødalsegga Fold is overturned, as both the eastern and western limbs dip at moderate to steep angles to the northeast. The axial plane of the Rødalsegga Reclined Fold dips east or northeast and strikes northwest from Øvre Huldrekoppvatn along Svartegga, through Rødalsegga to Zakariasvatn. North and west of Zakariasvatn the orientation of the axial plane is variable because of later isoaxial refolding. Throughout most of the Tafjord area, a schistosity parallels the compositional layering on the limbs of the fold. At Huldrekoppen, however, which is near the closure of the fold, the schistosity trends consistently 10° to 15° more north-south than does the compositional layering. The intersection of the schistosity and the compositional layering plunges to the east, indicating that the fold axis of the Rødalsegga Reclined Fold likewise plunges to the east. The structure opens rapidly to the southeast. In the northern part of the area, it appears to open to the north, but this feature is probably due to refolding.

Muret (1960, p. 30) and Gjelsvik (1951, p. 91) mapped a fold with a similar northwesterly closure at Rødalsegga. However, they connect the quartzites found on Rødalsegga with those found just east of Vikvatn. As shown on Plate I, the quartzites near Vikvatn belong to the Øyen group. Thus, there is a thick sequence of metasediments structurally lower than those found on Rødalsegga, explaining, perhaps, why Muret and Gjelsvik thought the lower limb of the Rødalsegga fold was upside down.

The Rødalsegga Fold is refolded by at least two other major structures, and appears to record the earliest deformational event in the Tafjord area. The minor folds of the F_{1A} subsystem are considered to be parasitic to the Rødalsegga Reclined Fold.

Kallskar and Kaldhusseter Reclined Folds

Fold closures have been mapped east of Kaldhussetervatn and south of Kallskar in the western and northwestern part of the Tafjord area, respectively (see Plate I). The upper limb of the fold near Kaldhussetervatn is sheared off by a north-striking fault. The fold at Kallskar is far broader than the Kaldhusseter fold. Both folds have cores occupied by the largest dunite bodies in the Tafjord area. Both structures have complementary recumbent folds to the east, although the one at Kaldhusseter is modified by a fault.

The Kaldhusseter and the Kallskar folds are associated with two schistosities. One schistosity parallels the compositional layering wrapping around the hinges of the folds. Just south of the hinges defined by the contact between the Fetvatn Gneiss and the Øyen group, a schistosity parallels the axial planes of these folds. When traced north, toward the fold closures, this axial-surface schistosity becomes obscured by the schistosity that wraps around the hinges. Apparently, an axial-surface schistosity could develop in the relatively incompetent Fetvatn Gneiss but not in the more competent rocks of the Vikvatn Sequence. The axes of the two folds plunge toward the east at a moderate angle, parallel to the axes of the Rødalsegga Fold. However, the asymmetry of these folds is opposite to that expected if they are parasitic to the north-closing Rødalsegga Fold. They are, therefore, not likely to be genetically related to the Rødalsegga Fold, especially as they fold the schistosity formed during the development of the Rødalsegga Fold. The Kaldhusseter and Kallskar Reclined folds are correlated with the minor folds of the F_{1B} subsystem.

Tafjord Synform

Figs. 3C and 4B show plots of poles to the axial-surface schistosity of F_{1A} and F_{1B} minor folds, respectively. The refolding of the axial-surface schistosities of the F_{1B} folds must be attributed to the development of a later major fold system, the axis of which must plunge to the east. This third system, only weakly developed in the southern part of the Tafjord area, is very strongly developed as a large synform (the Tafjord Synform) in the northern part of the area (Fig. 21).

The Tafjord Synform is most clearly displayed just west of the fjord, where the rocks north of Gjeitvikdalen strike northeast and dip at moderate angles to the southeast, whereas rocks between Gjeitvikdalen and Kallskarhornet strike northwest and dip at moderate to steep angles to the norheast. Poles to regional foliations throughout the synform (Fig. 9B) define a cleft girdle with two elongate maxima subtended by an angle of about 90°. The Tafjord Synform thus appears to be an open fold with slightly curved limbs and a very sharp hinge. While the Tafjord Synform is simple in general, it is complex in detail and contains parasitic synforms and antiforms on its flanks. The fold geometry suggested by Fig. 9B is like the geometry of the F_{3B} folds in that it is an open fold with slightly curved limbs. On the other hand, the cleft girdle and double maxima suggest that the geometry of the Tafjord Synform is like the geometry of the F_{3A} (chevron) folds in having a very sharp hinge.

Muret (1960), O. Holtedahl (1938), H. Holtedahl (1950), Hansen (1963), Scott (1967) and Wheeler (1973) map and describe very large-scale recumbent folds or nappes in the eastern basal gneiss region. Furthermore, large-scale refolding of these early nappes by later recumbent folds has been observed in Dovrefjell (Scott 1967, Wheeler 1973) and Hestbrepiggan (Banham & Elliot 1965). These nappe systems have the same geometries, orientations and age relationships as those in the Tafjord area. Hence, the the formation of the Rødalsegga Reclined Fold and its subsequent refolding by the Kaldhusseter and Kallskar Reclined Folds were processes that may have occurred more or less simultaneously with the folding in the eastern basal gneiss region.

Structural analogues of the Tafjord Synform are rare in the eastern basal gneiss region although the major F_3 folds described by Wheeler (1973) in Dovrefjell may be similar. Banham & Elliot (1965) describe very large, open, east-plunging folds that developed late in the structural evolution of the Hestbrepiggan area. H. Holtedahl (1950) describes a synclinal trough that refolds earlier nappe structures in the Oppdal District. These structures do not, however, appear to have been the most important factor in determining the the structural trends of the eastern basal gneiss region. On the other hand, structural analogues of the Tafjord Synform appear to be common in the western basal gneiss region (Gjelsvik 1951, Hernes 1957, Schmitt 1963). They may have been important in determining the characteristic east-west structural trends in the western basal gneiss region. The Tafjord area, then, may be structurally transitional between the basal gneiss region to the east, characterized by highly variable structural trends, and the basal gneiss region to the west, characterized by east-west structural trends.

DISCUSSION

The following section deals with the deformation environment responsible for the tectonic fabric of the Tafjord area and, where possible, with the stress system responsible for the deformation. The work of Hansen (1963, 1971) in Trollheimen, subsequently applied by Scott (1967) and Wheeler (1973) in Dovrefjell, describes the kinematic behavior of folds, within a variety of geologic materials, that have the same geometry and orientation distribution

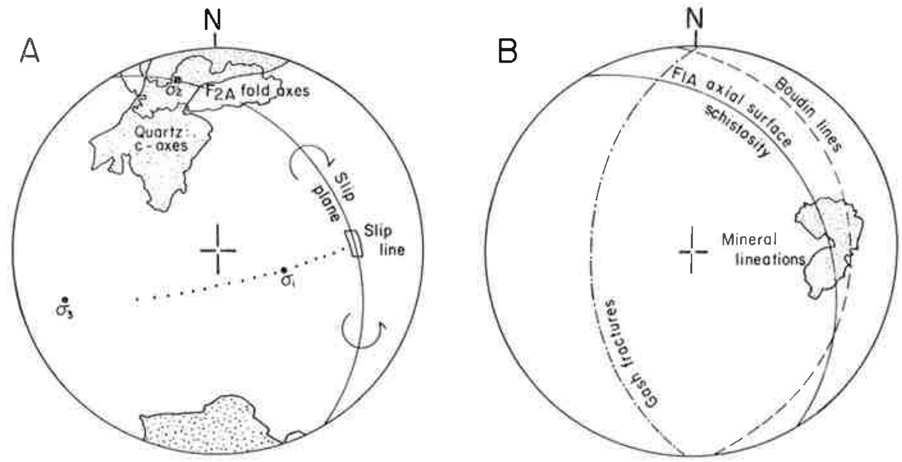


Fig. 10A. Orientations of important fabric elements of F_2 system and inferred orientations of the principle stress axes.
Fig. 10B. Orientations of important fabric elements of F_1 system.

as the F_{2A} folds of the Tafjord area. In all the examined cases, the slip plane is found to parallel the girdle defined by the axes of the F_{2A} folds, and the slip line always falls within the separation arc defined by the reversal of shear sense of the folds within the girdle. Under this model, the slip plane of the F_{2A} fold axes at Vikvatn, Øyen and Kaldhusseter had a north-south strike and dipped east at shallow to moderate angles (see Fig. 5). The slip line plunged at shallow to moderate angles to the east. The shear senses of all the F_{2A} folds in the measured areas indicate that the structurally higher layers moved to the east relative to the structurally lower layers. Slip within the associated F_{2A} faults presumably occurred parallel to the east-plunging direction defined by the cataclastically milled feldspar augen. Thus, two independent methods yield the same slip line and slip plane during F_2 deformation.

Fig. 10A plots some of the important fabric elements of the F_2 system with the principal stress axes derived by Hansen (1971) for identical folds from other environments. The plane defined by σ_1 and σ_3 is perpendicular to the slip plane and contains the slip line. Because the material being deformed possessed a strong anisotropic layering, shear (or flexural folding) was favored parallel to the compositional layering rather than the conjugate shear plane. σ_1 does not necessarily lie at 30° to the slip plane, and its orientation may have varied from close to parallel to close to perpendicular (dotted line in Fig. 10A) to the F_2 slip plane. The intermediate stress axis (σ_2) parallels the statistical maximum defined by the orientation of F_{2A} fold axes (shaded area in Fig. 10A).

The approximate position of the maxima defined by the preferred orientations of the c-axes of quartz grains from a quartzite within one of the F_2 fault zones is plotted in Fig. 10A as the stippled area. The intermediate stress axis (σ_2) is parallel to the acute bisector of the angle between the maxima. The maximum and minimum principal stress axes (σ_1 and σ_3) lie in the plane that bisects the larger angle between the two maxima and are approximately equidistant from the maxima.

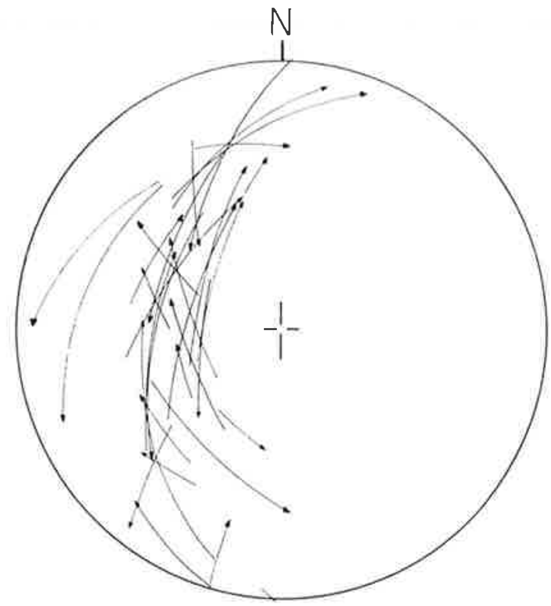


Fig. 11. Orientations of poles to F_3 fold axial planes and poles to nearby unfolded foliation. Arrows originate at poles to the unfolded foliation and point to poles of the F_3 axial planes.

Fig. 10B summarizes the average orientations of the important fabric elements of the F_1 system including gash fractures and boudin lines. The usual folding mechanism invoked to explain folds with F_1 characteristics is by slip parallel to the axial plane of the fold (Turner & Weiss 1963). Analytical techniques exist for determining the slip lines of slip folds that refold earlier slip folds if the hinges of the later folds are not parallel to the hinges of the earlier folds (Turner & Weiss 1963) or if the slip plane of the later folds is not parallel to either the hinge line or the axial surface of the earlier folds (Hansen 1971). Unfortunately, the F_{1A} and F_{1B} folds of the Tafjord area are not only coaxial, but also in most places their axial planes are roughly coplanar as well. Hansen (1971) and others have shown that the slip line during the slip folding can be close to parallel to their hinge lines. That a similar situation may have occurred during F_1 deformation in the Tafjord area is suggested by fragmentary bits of evidence: (1) the fold hinge lines and associated mineral lineations fall within the separation arcs defined by the F_{2A} folds. If the dominant slip direction during F_1 deformation was close to parallel to the F_1 hinge lines (i.e., east-west), then it paralleled the direction inferred for F_2 deformation. (2) Lenticular gash fractures are oriented perpendicular to the east-plunging linear elements of the F_1 system. If they formed by extension perpendicular to the plane they define (in domains where the strain rate was too high to be compensated for by ductile flow), then they indicate rock extension parallel to the east-plunging F_1 hinge lines.

The poles to the axial planes of the F_{3A} folds and the poles to the adjacent unfolded foliations are plotted in Fig. 11. Arrows, pointing to the poles of the F_{3A} axial planes, connect the acute angles between the elements. These arrows indicate that the axial planes are related to the foliations by an east-plunging axis of rotation. However, there is no systematic sense of rotation about this axis, nor is there any preferred angle between the F_{3A} axial planes and the foliation. The attitudes of the F_{3A} planes, plotted on Plate I, vary irregularly and apparently unsystematically from one locality to the next. If the F_{3A} folds record flattening perpendicular to their axial planes, then the direction and degree of shortening do not appear to be related in a simple way to the formation of major structures, as suggested by Hansen (1963) in Trollheimen, or to uniform regional or subregional shortening, as suggested by Scott (1967) in Dovrefjell. Similar problems exist for the F_{3B} folds.

The subvertical, north-south striking fractures are filled with low temperature and low pressure minerals that suggest a time of formation after the main phase of metamorphism. Fig. 8 plots the inferred orientations of the principal stress axes responsible for fracture formation. If the fractures formed by east-west extension, the minimum principal stress axis (σ_3) was perpendicular to the fracture plane, and the intermediate and maximum principal stress axes (σ_2 and σ_1) lay within the plane. The attitudes of the conjugate shear fractures from Dovrefjell (Scott 1967) and Eiksund (Schmitt 1963), also plotted on Fig. 8, are consistent with the subvertical fractures of the Tafjord area and suggest that σ_1 was north–south and horizontal and that σ_2 was roughly vertical.

The F_1 through F_3 fabrics were previously thought (Brueckner 1969) to be the result of Caledonian deformation. If, however, Caledonian supracrustal rocks are lacking in the Tafjord area, there is no *a priori* evidence that any of the structures in the mapped area had a Caledonian origin. It is entirely possible that some or all of the fabric systems formed during one or more Precambrian orogenies. All the folds in the Vikvatn Sequence can be correlated with structures of identical style and orientation in the eastern basal gneiss region where they are believed to deform Caledonian rocks (Hansen 1963, Scott 1967, Wheeler 1973). Furthermore, most K-Ar and Rb-Sr mineral ages from the basal gneiss region range from 372 to 420 m. y. (Broch 1964, McDougall & Green 1964, Brueckner et al. 1968, Strand 1969, Bryhni et al. 1971, Brueckner 1972, Priem et al. 1973), suggesting rather complete recrystallization during the Caledonian orogeny. It would seem likely that recrystallization was accompanied by intense deformation. Evidence from Tafjord indicates that this recrystallization occurred during the development of the F_1 fabric. It is possible, however, that this Caledonian event resulted in the development of the F_{1B} subfabric, and that the F_{1A} subfabric was formed during one or more pre-Caledonian events. Priem et al. (1973) note that the emplacement of the 975 m. y. old Hestbrepiggan Granite in West Jotunheimen post-dated an episode of almandine–amphibolite facies metamorphism associated with the development of NNW–SSE trending folds. The F_{1A} subsystem may be related to these ancient folds, and may include other, possibly older structures. Under this model, these pre-Caledonian structures would have been severely compressed and pervasively rotated to their present configuration during intense F_{1B} deformation. Thus, the F_{1A} subsystem could be a 'homogenized' fabric that was previously a rather heterogeneous assortment of older structures. Further geochronological studies and field work may clarify these problems.

Petrology

GENERAL STATEMENT

Mineral assemblages typical of three and possibly four metamorphic facies are present in the country rocks of the Tafjord area. Because the development of these assemblages can be correlated with the development of some of the structural episodes described in the previous section, the following petrographic descriptions are here discussed in terms of the structural environment in which they formed.

PETROLOGY OF THE MINERAL ASSEMBLAGES FORMED DURING THE DEVELOPMENT OF THE F_1 STRUCTURAL FABRIC

Detailed petrographic data and modes of the Tafjord area rocks are presented elsewhere (Brueckner 1968) and are available from the author upon request. Table 2 lists the mineral assemblages (given in order of decreasing abundance) where one or more of the minerals are clearly aligned parallel to one or more of the F_1 fabric elements. Obvious secondary minerals are ignored. The list thereby defines the phase assemblages in equilibrium with the pressures, temperatures and chemical conditions prevalent during the F_1 deformation.

The minerals and mineral assemblages listed in Table 2 are typical of rocks that formed under conditions appropriate to the kyanite–almandine–muscovite or the sillimanite–almandine–muscovite subfacies of the almandine–amphibolite facies (Turner & Verhoogen 1960). Only six of the one hundred samples of country rock examined under the microscope contain the aluminium silicates kyanite and/or sillimanite. Kyanite typically occurs as small (0.5 to 2.0 mm) euhedral prisms or irregular prisms filled with plagioclase, biotite and opaque inclusions. The prisms are concentrated in biotite–plagioclase-rich laminae and show strong preferred orientation. They lack reaction relationships with other minerals and were evidently in equilibrium with them during the formation of the F_1 fabric. Tourmaline forms large (up to 2.0 cm long) prisms in a few kyanite-bearing schists. Sillimanite forms fibrous, felt-like sheaves intergrown with biotite from which it evidently was derived. The sheaves consist of aggregates of minute, long, slender prisms. Kyanite and sillimanite may coexist within the same plagioclase–biotite laminae, but they are not in direct contact. Muscovite within the two-mica gneisses and the quartzites forms large, euhedral blades that show no signs of alteration to sillimanite. Sillimanite occurs with muscovite in a fine-grained two-mica gneiss near Kallskaret. The sillimanite fibers form irregular clots, however, suggesting that the sillimanite had isomorphically replaced a mineral other than muscovite or biotite.

Kyanite is the stable aluminium silicate in the eastern portions of the basal gneiss region (Scott 1967, Hansen, 1963). The only other reported occurrence of sillimanite in the basal gneiss region is in the Eiksund area on the west coast of Norway (Schmitt 1963). It is found in quartz nodules replacing

Table 2. Equilibrium mineral assemblages formed during the development of the F_1 structural fabric

Granite gneiss	
	quartz–plagioclase–microcline–biotite
	quartz–plagioclase–microcline–biotite–hornblende
Granodiorite gneiss and schist	
	quartz–plagioclase–biotite
	quartz–plagioclase–biotite–hornblende
	quartz–plagioclase–biotite–garnet
	quartz–plagioclase–biotite–hornblende–garnet
	quartz–plagioclase–biotite–diopside–garnet
Syenite gneiss	
	microcline–plagioclase–biotite–clinopyroxene–garnet
	microcline–plagioclase–biotite–hornblende–garnet
Two-mica gneiss	
	quartz–microcline–plagioclase–biotite–muscovite
	quartz–plagioclase–biotite–muscovite
Quartzite	
	quartz–microcline–plagioclase–muscovite–biotite
Calc–silicate rocks	
	quartz–plagioclase–diopside–actinolite
	quartz–plagioclase–diopside–calcite–scapolite
	quartz–plagioclase–diopside–calcite–epidote–scapolite
	quartz–plagioclase–microcline–biotite–diopside
	quartz–plagioclase–microcline–biotite–diopside–epidote
	quartz–plagioclase–microcline–biotite–diopside–epidote–scapolite
	quartz–plagioclase–microcline–biotite–diopside–epidote–scapolite–calcite
Mafic rocks	
	amphibole–plagioclase–quartz
	amphibole–plagioclase–biotite–quartz
	amphibole–plagioclase–garnet–quartz
	amphibole–plagioclase–garnet–quartz–biotite
	amphibole–plagioclase–epidote
Aluminum-rich rocks	
	quartz–plagioclase–biotite–garnet–kyanite ± tourmaline
	quartz–plagioclase–biotite–garnet–sillimanite
	quartz–plagioclase–biotite–garnet–kyanite–sillimanite
	quartz–plagioclase–muscovite–biotite–microcline–sillimanite
	plagioclase–biotite–muscovite–corundum

kyanite and muscovite as well as biotite, indicating that temperatures during sillimanite metamorphism were higher there than in the Tafjord area. There is an increase in metamorphic grade in the rocks of southwestern Norway from the greenschist rocks of the western part of the Trondheim basin to the upper almandine–amphibolite and granulite facies of the western basal gneiss region. Sillimanite-bearing rocks are restricted to the northeastern portion of the Tafjord area. This distribution is consistent with the regional picture, and suggests that the first sillimanite isograd may pass through the Tafjord area.

The mineral assemblages within the granite gneisses, granodiorite gneisses and schists, two-mica gneisses and quartzites are entirely consistent with recrystallization under almandine–amphibolite facies conditions. Textures and mineralogies within the calc–silicate and mafic rocks are heterogeneous, and some of the phases within these rocks could be secondary. Diopside, actinolite, epidote, scapolite and calcite within calc–silicate rocks may be concentrated into laminae, lenticules and irregularly shaped pods, giving the rock a gneissic appearance, or the phases may be homogeneously distributed throughout the rock, giving it a massive appearance. Plagioclase is rather calcic ($An_{42\text{-}56}$). Diopside, scapolite and epidote form either porphyroblasts or subhedral to euhedral apparently primary grains. Actinolite forms long, slender prisms that show no obvious replacement features in a few samples, and hence may also be primary.

The mafic rocks contain mostly hornblende and plagioclase, with or without lesser amounts of biotite, garnet, quartz and epidote. Most of the epidote in the mafic rocks is secondary, but stubby, euhedral prisms in a few samples show strong preferred orientations and no signs of alteration, and may be primary. Garnet within mafic rocks form clean, equant grains homogeneously distributed throughout the rock. Their grain-size is the same as that of other minerals in the rock, and they do not distort the surrounding fabric. They lack reaction rims, and were in apparent equilibrium with the other minerals at their time of formation. These features distinguish mafic rocks stable under almandine–amphibolite facies conditions from the eclogites.

RELICT ASSEMBLAGES

The unusual syenite rocks that occur in the Øvste Rødal group just south of Zakariasvatn may possess assemblages that formed prior to the development of the F_1 fabric. Most syenite rocks are medium-grained, well-foliated gneisses with strong linear fabrics and little or no quartz. Those south of Zakariasvatn have hornfelsic textures. The feldspars of these rocks are segregated into microcline-rich and plagioclase-rich mosaics. The microcline-rich zones contain clots and schlieren of mafic minerals, mostly garnet, biotite, clinopyroxene and opaques. The clinopyroxene forms ill-defined, zoned prisms surrounded by albite–diopside symplectite. Their cores show Schiller structures. Some microcline contains exsolved laminae of albite and hence is a perthite. The clinopyroxene, perthite, and some of the garnet within these rocks may be relict, and their association suggests that the syenitic rocks may originally have been formed in the pyroxene–almandine subfacies of the granulite facies (Turner & Verhoogen 1960). They are, therefore, the only rocks described in this section that may record a higher metamorphic facies than almandine amphibolite.

The syenitic rocks with these possible relict assemblages occur near a titanium-rich mafic intrusion in Øvste Rødal. This body has been investigated by Gjelsvik (1957) who interpreted it as a late intermediate differentiate of a strongly fractionated gabbro. The survival of the relict assemblage within the syenitic rocks may be the result of this intrusion. Almandine–amphibolite facies minerals (hornblende, biotite, some of the feldspar and most of the garnet) in the non-foliate samples show no preferred orientation, suggesting that they formed after the F_1 deformation. Possibly relict minerals like clinopyroxene remained stable longer in the rocks near the intrusion because of a local rise in temperature, the only pyroxene alteration being the formation of a plagiclase–diopside symplectite corona. Not until the temperature around the intrusion dropped to the same level as that of the surrounding gneisses could water enter the system to begin to convert the clinopyroxene to amphibole.

PARENT ROCKS

The granodioritic rocks of the Fetvatn group are broadly uniform in composition and are believed to have had an igneous origin. The lithologies of the Vikvatn Sequence are far more heterogeneous, suggesting more diverse origins for these rocks. Quartzites, calc–silicate rocks and two-mica gneisses are commonly associated and are the most obviously metasedimentary rocks of the Tafjord area. The mafic amphibolites and garnet-amphibolites of the Vikvatn Sequence are believed to have originally been extrusive basalts or intrusive gabbros or dolerites. The syenitic, granodioritic and granitic gneisses of the Øvste Rødal group appear to be the metamorphosed equivalents of an interlayered sequence of trachytes, latites, rhyodacites and rhyolites. A kyanite–sillimanite schist from the same group indicates that a few layers of aluminous sedimentary rocks may have been intercalated between the igneous rocks. Scott (1967) suggests similar origins for the rocks of the Tveratind group in Dovrefjell.

The origin of the syenitic rocks of the Øvste Rødal group may be related to that of the anorthosite. Syenitic or mangeritic rocks are commonly associated with anorthosites through most of the Norwegian Caledonides. The anorthosites of the western basal gneiss region occur with quartz syenites and monzonites (Bryhni 1966, Lappin 1966). The non-foliate syenites in Øvste Rødal occur immediately adjacent to anorthosite, and are the only rocks except for the anorthosites, eclogites and ultramafic rocks that may record a higher metamorphic facies than that of almandine–amphibolite.

The bulk of the Vikvatn Sequence is composed of granite and granodiorite gneiss. The origin of these rocks is obscure. They could initially have been sedimentary (arkoses, for example) or igneous (rhyolitic or rhyodacitic lavas or sialic intrusive rocks). A similar problem exists for schists bearing either kyanite, sillimanite or corundum. These rocks have high aluminum contents suggesting a pelitic metasedimentary origin. However, these same rocks contain very little or no quartz. Pelitic or psammitic sedimentary rocks with less than 25% normative silica are rare (Hopson 1964).

METASOMATISM

There is evidence of limited metasomatism in the Tafjord area, particularly on the scale of a few centimetres. Some potassium migration must have occurred to form the abundant large microcline augen in some of the gneisses. The strong banding in the Fetvatn Gneiss may be the result of ion migration along gradients set up by previously existing chemical differences.

An attractive explanation for the unusually low silica and high alumina content of the kyanite, sillimanite and corundum-bearing schists is that these rocks were desilicified. Most were collected near dunites. For example, a coarse-grained schist containing large porphyroblasts (up to 2.0 cm long) of corundum occurs next to the Kallskar ultramafic body. Just south of Kallskaret, an anorthosite layer contains a lens of very coarse (15 to 20 cm) corundum crystals coexisting with lesser amounts of fine-grained margarite and chlorite. Both corundum localities are adjacent to dunites that contain zoned mineral assemblages at their margins. The reactions that produced a zoned sequence within a fracture in the Kallskar body is described in detail by Carswell et al. (1974). The presence of such minerals as phlogopite, tremolite and enstatite in these zones indicates that ions, including silicon, were introduced into the ultramafic bodies during metamorphism. Thus, the aluminous schists may have been desilicified during metamorphism as a result of their proximity to the dunites.

Several workers in the basal gneiss region (Strand 1949, Muret 1960, Lappin 1966) suggest that potassium and/or sodium migration occurred on a regional scale during the Caledonian orogeny. However, a Rb-Sr isochron has yielded a Precambrian age for the rocks of the Fetvatn group (Brueckner et al. 1968). If these rocks had been subject to metasomatic cation and anion migrations of more than a few centimetres during Caledonian metamorphism, the strontium and rubidium would have re-equilibrated, and the determined isochron would have yielded Caledonian numbers. Hence metasomatism in the Tafjord area is believed to be limited in scale. It is unlikely that the Fetvatn Gneiss was subjected to regional metasomatism during Caledonian metamorphism.

PETROLOGY OF MINERAL ASSEMBLAGES FORMED DURING F_2 FAULTING AND SUBVERTICAL FRACTURING

The rocks in and near the F_{2A} faults and subvertical fractures show signs of recrystallization accompanied by retrograde metamorphism. Rocks in or near fractures display a pink and green color, quite distinct from the normal color away from these zones. Microcline gives the rock its predominant pink tinge; the green colors are caused by chlorite and epidote. Plagioclase is commonly sericitized and saussuritized. Many plagioclase grains possess antiperthitic textures, suggesting that the plagioclase was exsolving to albite and microcline in response to decreasing temperatures. Perthitic textures in the microcline suggest a similar interpretation. Relics of garnet and biotite are surrounded by alteration products, predominantly chlorite. That quartz recrystallized is shown by its strong preferred orientation in the more mylonitized samples from F_{2A} faults (see Fig. 7C).

Chlorite, epidote, muscovite, actinolite, microcline, albite and quartz were apparently stable under the new temperatures, pressures and chemical environments, and are typical of the quartz-albite-muscovite-chlorite subfacies of the greenschist facies (Winkler 1965). Apparently, the F_{2A} fault zones and subvertical fractures originated under these metamorphic conditions. If F_{2A} folds and faults developed synchronously, then F_{2A} folding also occurred under greenschist facies conditions.

PETROLOGY OF MINERAL ASSEMBLAGES FORMED DURING MINERALIZATION OF SUBVERTICAL FRACTURES

Many of the subvertical fracture faces are mineralized. The following lists the mineral assemblages:

epidote
chlorite
stilbite
epidote-chlorite
epidote-chlorite-quartz
epidote-chlorite-Kspar(?)
epidote-chlorite-pyrite
chlorite-quartz-pyrite
epidote-chlorite-quartz-pyrite-Kspar(?)
epidote-clinopyroxene-quartz-calcite-pyrite-garnet-magnetite
tourmaline
epidote-fluorite
epidote-prehnite
epidote-prehnite-calcite
epidote-prehnite-apophyllite
epidote-prehnite-apophyllite-calcite-(chlorite-pyrite)-quartz

Epidote and chlorite are commonly slickensided, indicating that initial mineralization and movement across the fracture faces occurred under greenschist facies conditions. Stilbite occurs as small euhedral tablets or, less commonly, white, radially-arranged, cone-shaped aggregates. Its presence on the faces of some of the fractures indicates that mineralization of the fractures continued down to temperatures and pressures appropriate to the zeolite facies. Banham (1966) reports that stilbite occurs in fractures throughout the basal gneiss region.

The mineral assemblages that include prehnite occur in a single cavity in gneiss east of Zakariasvatn. The minerals within this cavity are arranged in zones. Nearest the walls are long (up to 1.5 cm) prisms or bundles of prisms of epidote that tend to point away from the cavity wall and which are associated with minor amounts of chlorite and pyrite. Toward the cavity center, prehnite either sheathes the epidote or forms euhedral, pale green tablets that range in size from 0.5 to 5.0 mm. Most prehnite is intergrown with very large (up to 4.0 cm) calcite blades, arranged in a box-work structure, and with subordinate amounts of euhedral clear quartz. Finally, small- to medium-sized blocks (less than 1.0 cm) of creamy white apophyllite have grown on top of the prehnite and epidote.

The suggested order of crystallization of the minerals on the fracture faces is:

(1) epidote (± chlorite, quartz, K-feldspar, calcite, pyrite, magnetite, fluorite)
(2) prehnite (± calcite, quartz)
(3) stilbite and apophyllite

The sequence of crystallization is consistent with trends noted by Coombs et al. (1959) for calcium-zeolites and other calcium-aluminum minerals. They also appear in order of increasing hydration. Existing experimental evidence supports the idea that the observed trend reflects cooling.

Part II. The anorthosites, ultramafic rocks, and eclogites of the Tafjord area

General statement

Ultramafic rocks in the Tafjord area were initially described by Vogt (1883). He was also the first to describe the garnet peridotite lens ('garnet-olivine rock') in the large ultramafic mass at Kallskaret. O'Hara & Mercy (1963) and Carswell (1968b) discuss the petrology and petrogenesis of the ultramafic bodies of the Tafjord area, particularly of the garnet peridotite. O'Hara & Mercy (1963) were the first to describe amphibolitized eclogite lenses in the anorthosites and country rocks of the Tafjord area. Carswell (1968b) has shown that some mafic lenses within ultramafic bodies are also retrogressed eclogites. Anorthosites near the Tafjord area were first described by Strand (1949); Gjelsvik (1951) and O'Hara & Mercy (1963) located additional bodies within the area.

Form and stratigraphic distribution

The anorthosite, ultramafic rock and eclogite localities of the Tafjord area are shown on Plate I and Fig. 21. The large anorthosite masses form long, relatively thin, tabular sheets that are generally conformable to the compositional layering of the surrounding country rocks. Smaller anorthosite layers may be as thin as 5 cm. In detail, the schistosity of the anorthosite may be discordant to that of the surrounding gneisses. Furthermore, several rock-types occur at the upper and lower contacts of a single anorthosite body. Granite gneiss (including augen gneiss), plagioclase-biotite gneiss and schist, pelitic schist, hornblende schist and quartzite were observed in either concordant or discordant contact with a single anorthosite body near Vikvatn.

The largest anorthosite bodies occur in the upper and lower parts of the Øyen group. One body can be traced continuously along strike from Grotli, just south of the mapped area (see Strand, 1969), to just northeast of the northern tip of Vikvatn, a distance of 14 km. A thick anorthosite sheet that occurs on top of Storfjell occupies a similar stratigraphic position as the anorthosite near Vikvatn, and may be a continuation of the same body. Thus, it is possible that one anorthosite layer extends continuously along strike for a distance of over 25 km. Yet, its maximum thickness, near Grotli, is about 400 metres (Strand 1969). Other anorthosite bodies have not been traced for such long distances, and several, notably the anorthosite bodies near Kaldhusseter and Kallskaret, evidently pinch out and disappear along strike.

Almost one hundred bodies of ultramafic rocks have been observed in the Tafjord area. Most occur as small elliptical lenses, 5 to 50 m long, in the country rocks. Some ultramafic lenses occur within thick anorthosite layers. The two longest axes of the lenses parallel the schistosity of the surrounding rocks. Two very large ultramafic masses occur at Kallskaret and just east of Kaldhussetervatn. The Kaldhusseter mass is a very flattened lensoid body about 1300 m long and, on the average, 100 m wide. At Kallskaret, lenses as long as 600 m poke out as isolated masses through the surrounding rubble and vegetation. These isolated masses may be connected. If so, the Kallskar body, in plan view, covers an area of approximately 1.5 km^2. All ultramafic rocks possess a strong schistosity.

The retrogressed eclogites of the Tafjord area occur as small (0.5 to 7.0 m long), concordant lenses, generally strung out along specific stratigraphic horizons in the country rock. Mafic lenses and layers of hornblende, plagioclase and partially to strongly altered garnets that occur within the Kallskar and Kaldhusseter ultramafic bodies are petrographically and mineralogically indistinguishable from eclogites that occur as lenses in the country rock (Carswell 1968b). One such lens contains omphacite. Similar mafic rocks occur as boudins and thin, sheared-out layers in the anorthosite at Vikvatn and Kallskaret. Thus, eclogites may occur in the country rocks, anorthosites and ultramafic rocks.

Fig. 21 summarizes the observed distribution of anorthosites, ultramafic rocks and eclogites within the Tafjord area; without exception, they occur within the Vikvatn Sequence. As was noted previously, this association may be restricted to the Øyen and Øvste Rødal groups of the Vikvatn Sequence and may be lacking in the Svartegga group. Further mapping is required to substantiate this suggestion. No ultramafic rocks, eclogites or anorthosites were found in the Fetvatn gneisses.

Serpentinites are characteristically restricted to schists and amphibolites of Cambro-Silurian age in the Trondheim basin (Strand 1961) and eastern basal gneiss region (Muret 1960, Hansen 1963). Scott (1967), however, maps several ultramafic bodies in the Tveratind group of western Dovrefjell. This group is lower, stratigraphically, than the Cambro-Silurian schists and amphibolites, and Scott suggests that its age may be lower Cambro-Silurian or upper Eocambrian. Strand (1966), Gjelsvik (1953) and Muret (1960) map ultramafic bodies in the areas south of Tafjord that are closely associated with quartzites, but the age of these quartzites is not known.

The stratigraphic distribution of ultramafic rocks west of Tafjord is less well known. The view that is tentatively adopted here is Carswell's (1973) suggestion that ultramafic rocks associated with anorthosites, eclogites and

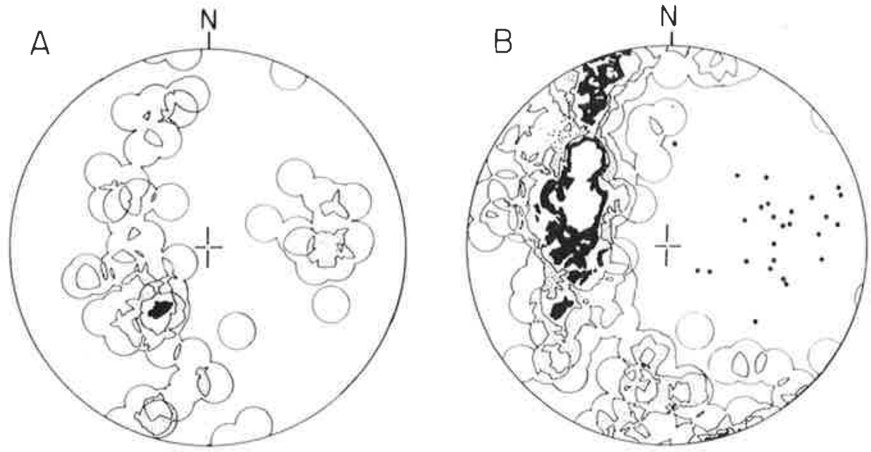


Fig. 12A. Orientations of mineral lineations and poles to foliation in anorthosites. 16 lineations; contours: 6.3, 18.8, 31.3% per 1% area. 59 poles; contours: 1.7, 5.1, 8.5, 11.9% per 1% area.
Fig. 12B. Orientations of 115 poles to foliation in ultramafic rocks; contours: 0.87, 2.6, 4.3, 6.1, 7.8% per 1% area. Dots represent primary and secondary ultramafic mineral lineations.

granulites occur within a Precambrian terrain, whereas hydrated ultramafic rocks not associated with these other rock-types may be intercalated between Caledonian supracrustal rocks.

Structure

Evidence from the Tafjord area indicates that the anorthosites and ultramafic rocks and probably the eclogites were introduced or formed in their present stratigraphic and tectonic position before or during the earliest phase of Caledonian deformation. Fig. 12A and 12B show the attitudes of schistosities and mineral lineations in the anorthosites and ultramafic rocks, respectively, of the Tafjord area. The anorthosite schistosity and mineral lineation are defined by the strong preferred orientation of the hornblende, actinolite, epidote, biotite, muscovite and, to a lesser extent, plagioclase. The schistosity in the ultramafic rocks is marked by flattened olivine grains and the parallel alignment of chlorite flakes. A primary ultramafic lineation is defined by the parallel alignment of actinolite, elongate olivine, enstatite and streaks or trains of olivine and chromite. A secondary lineation, caused by crinkled or streaked chlorite flakes, is more easily seen and may obscure the primary lineation. Both lineations are plotted in Fig. 12B. The poles to the foliations in both figures plot as girdles that closely approximate the girdle defined by the schistosity in the country rocks. The mineral lineations define maxima that coincide with the maximum defined by mineral lineations in the country rocks. The very presence of a strong schistosity and mineral lineation in the anorthosite and ultramafic rock, as well as the parallelism of their orientations to similar features in the country rocks, strongly suggests that the anorthosites and ultramafic rocks have undergone much, if not all, of the deformation suffered by the metasediments of the Vikvatn Sequence.

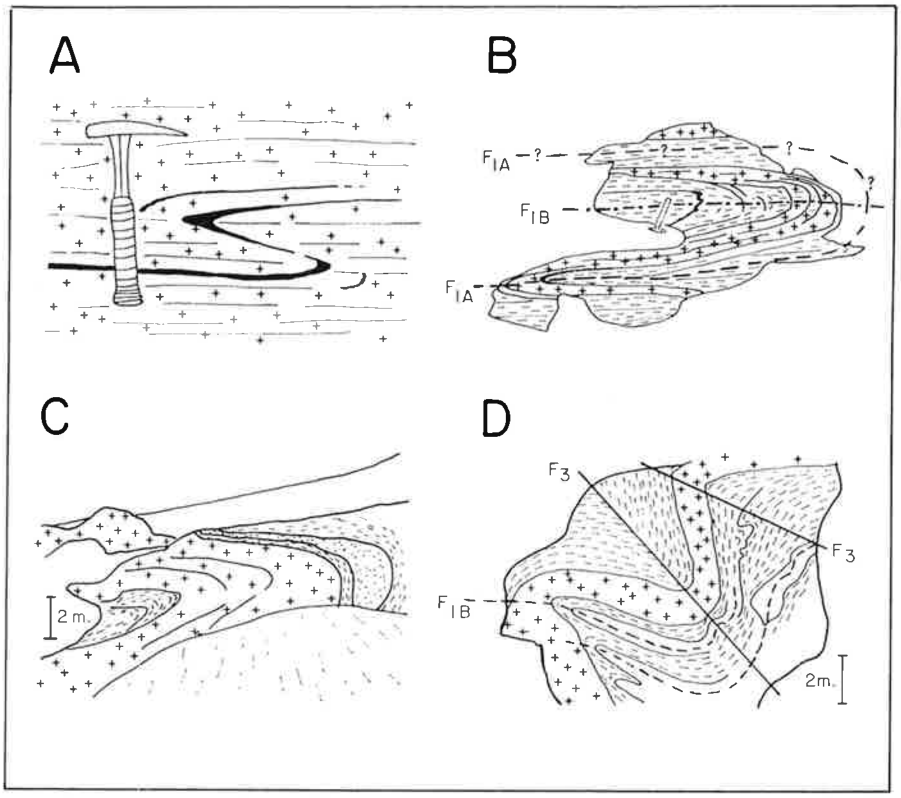


Fig. 13. Examples of minor folds within anorthosites of the Tafjord area. A, F_{1A} fold, Vikvatn; B, F_{1A} fold refolded by recumbent F_{1B} fold, Kallskaret; C, F_{2A} fold, Vikvatn; D, F_{1B} folds refolded by F_3 fold, Kaldhusseter. Crossed pattern = anorthosite. Stippled pattern = quartzite. Dashed pattern = schist. Hammer shaft measures 28 cm.

MINOR STRUCTURES

Folds

Figs. 13 and 14 present examples of minor structures within anorthosites and dunites, respectively, of the Tafjord area. F_{3A} (chevron) folds are common in several ultramafic bodies (Fig. 14B). They deform the pre-existing schistosity and are characterized by relatively straight limbs and sharp hinges. Some contain a fracture cleavage parallel to their axial planes. F_{3B} folds are common only in the Kallskar body where they are closely associated with F_{3A} folds. F_3 folds are rare in most anorthosites. A few of the thinner anorthosite layers, however, contain relatively open folds with east-plunging axes (Fig. 13D).

F_3 folds strongly affect the geometries of the Kaldhusseter and Kallskar ultramafic bodies. The poles to F_{3A} and F_{3B} axial planes (Fig. 15A) define a rough girdle with an axis that plunges toward the east parallel to the direction defined by the fold axes. The F_3 fold axes (Fig. 15A) show a rather wide scatter, and hence the foliations rotated about these axes are deflected and more variable in orientation than foliations unaffected by F_3 folds. The secondary ultramafic lineations formed by streaked and crinkled chlorite flakes

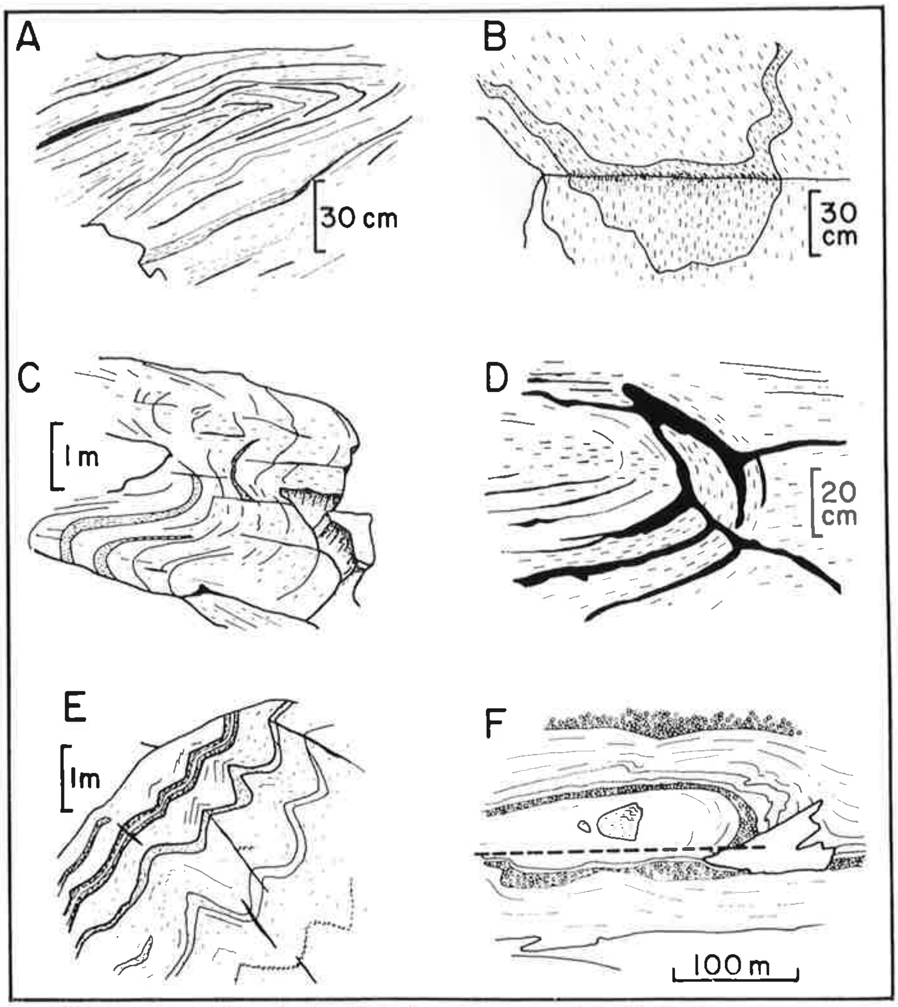


Fig. 14. Examples of minor folds within ultramafic rocks of the Tafjord area: A, recumbent F_{1A} fold, Kallskaret body; B, F_{1A} fold rotated to subvertical orientation by F_3 fold, Kallskaret body; C, F_{1B} folds, Kaldhusseter body; D, F_{1B} fold, Kallskaret body; E, F_{3A} (chevron) folds, Kaldhusseter body; F, small ultramafic lens in the core of major F_{2A} fold, 1500 m north of Vikvatn. Dashed lines represent chlorite schistosity.

and trains of beaded serpentine are plotted in Fig. 15B. They are best developed near the hinges of F_{3A} folds and display the same maximum and amount of scatter as the F_3 fold axes. They are therefore considered related to the F_3 folds. Hence, the small discrepancies in the orientations of the schistosities and mineral lineations between the ultramafic bodies and country rocks, noted by O'Hara & Mercy (1963), are caused by the F_3 folds that occur in the Kaldhusseter and Kallskar bodies, but are not as extensively developed in the predominantly psammitic country rocks.

The anorthosite layers and small ultramafic lenses near Vikvatn contain or are distorted by abundant F_{2A} folds. An example of a folded anorthosite layer,

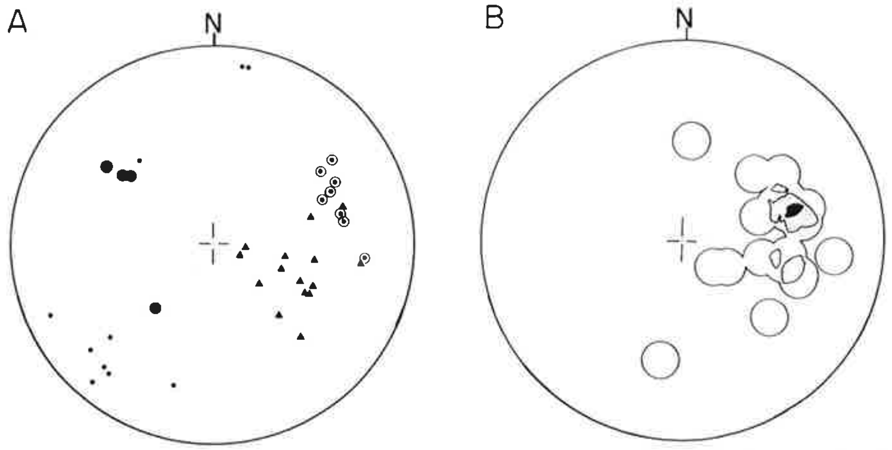


Fig. 15A. Orientations of F_3 fold axes from Kallskaret (triangles) and Kaldhusseter (small dots with circles) bodies. Orientations of poles to axial planes of F_3 folds from Kallskaret (small dots) and Kaldhusseter (large dots) ultramafic bodies.
Fig. 15B. Orientations of 20 mineral lineations related to F_3 folds in Kaldshusseter and Kallskaret ultramafic bodies; contours: 5, 15, 25% per 1% area.

3 m thick, is shown in Fig. 13C. Fig. 14F is a sketch of a large F_{2A} fold (20 – 30 m amplitude) that occurs about 0.5 km north of Vikvatn. Fig. 16A shows the attitudes of some minor F_{2A} fold axes and axial planes in the anorthosite affected by this fold. The tendency of the fold axes to form a north-plunging maximum rather than a north-south striking, east-dipping girdle typical of the country rocks may indicate that the anorthosite could deform by flexural slip only when the axes were in the orientation of maximum shear (i.e., 90° to the slip direction). Fig. 16A also plots the attitudes of the rotated L_1 mineral lineations in the anorthosite. A strongly chloritized ultramafic lens forms the core of the fold. The lens itself is intensely deformed by small-scale F_{2A} folds. The orientations and shear senses of the folds in this lens are plotted in Fig. 16B. The separation arc is 15°, an angle smaller than the separation arcs defined by F_{2A} folds in the nearby metasediments. The folds in the dunite and anorthosite are the same in geometry, orientation and shear sense as the F_{2A} folds in the country rocks, indicating that the dunite and anorthosite were deformed along with the country rocks during F_2 deformation.

F_{1A} folds are visible only in anorthosites and ultramafic rocks that possess a compositional layering. A few anorthosites at Vikvatn, Rønukfjell and Kallskaret contain an apparently primary compositional layering defined by hornblende- and biotite-rich layers and schlieren occurring in the far more abundant leucocratic anorthosite. An example of an isoclinal F_{1A} fold defined by one of these mafic schlieren in the anorthosite at Vikvatn is shown in Fig. 13A. Some ultramafic layers, particularly those at Kallskaret and Kaldhusseter, contain a compositional layering defined by alternating chlorite-rich and chlorite-poor layers. These layers define a few folds, examples of which are shown in Figs. 14A and 14B. The schistosity and primary mineral lineation that pervade every ultramafic mass and anorthosite in the Tafjord area parallel the axial

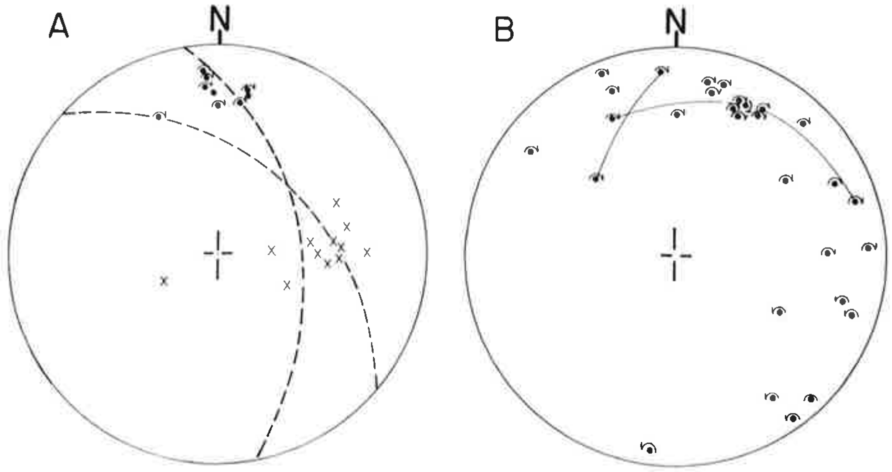


Fig. 16A. Orientations of F_{2A} fold axes (large dots showing shear sense) and axial planes (dashed great circles) in anorthosite. X represents orientations of L_{1A} mineral lineations rotated by a major F_{2A} fold.
Fig. 16B. Orientations and shear sense distribution of F_{2A} folds in small ultramafic lens 1500 m north of Vikvatn.

planes and axes, respectively, of these F_{1A} folds. These folds are the same in geometry and fabric as the F_{1A} folds in the country rocks.

A few isoclinal folds possess two schistosities. Fig. 13B shows a thin anorthosite layer near Kallskaret that was folded by an F_{1A} fold and then refolded by a second isoclinal fold. A few examples of isoclinal folds associated with schistosities from the Kaldhusseter and Kallskar ultramafic bodies are shown in Figs. 14C and 14D. The two schistosities can be seen in ultramafic bodies where there is no compositional layering to define the fold outline. These folds are similar in geometry and fabric to the F_{1B} folds in the encompassing country rocks.

Fig. 17 shows the orientations of the axial surface schistosities and axes of some F_{1A} and F_{1B} folds measured in the Kaldhusseter and Kallskar ultramafic bodies. The axes invariably plunge toward the east at shallow to moderate angles. Although most of the axial surface schistosities strike near north-south and dip toward the east, a few strike east-west and are subvertical. They intersect in a direction that parallels the axes of the F_3 folds in each body (see Fig. 15A) indicating that they were rotated during subsequent F_3 deformation.

Brittle structures
Straight, north-south striking, subvertical fractures with serpentine slickensides occur in many ulramafic bodies (Fig. 18). The fracture planes and slickensides roughly parallel similar structures in the country rocks.

Irregular, anastomising fractures, containing anthophyllite, enstatite, tremolite and chlorite, occur in the ultramafic masses of Kaldhusseter, Kallskaret and Raudnakken. They differ from the straight, serpentine-filled fractures described

Fig. 17. Orientations of axes and axial surface schistosities of F_1 folds in ultramafic bodies of the Tafjord area.

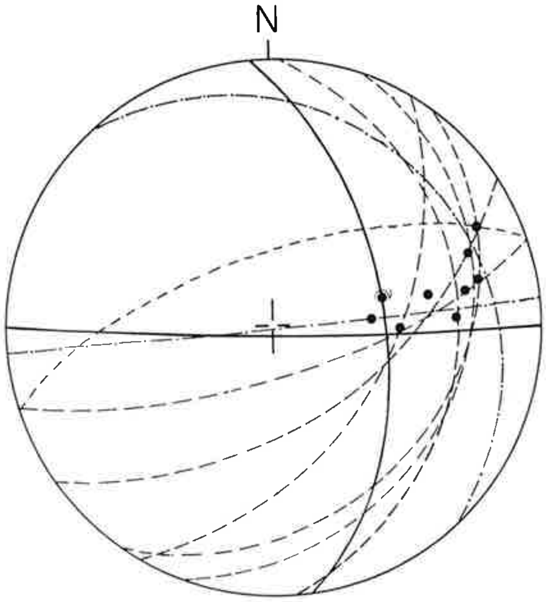
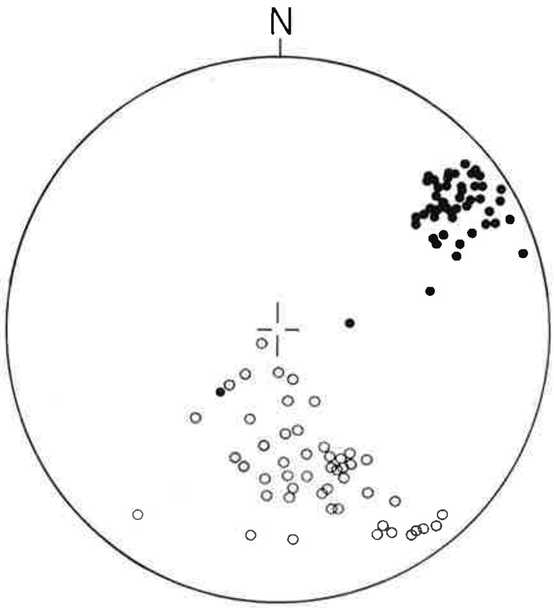


Fig. 18. Orientations of poles to subvertical fractures (large solid dots) and slickenside directions within fractures (circles) in Kaldhusseter ultramafic body.



above in that they are boudined, folded and offset by later fractures. The fractures may parallel or cut across the ultramafic schistosity. The minerals within the fractures rarely display a preferred orientation, and none of the fractures are deformed by F_1 folds. One enstatite-filled fracture is folded by an F_3 fold at Kaldhusseter, indicating that some fracturing occurred prior to F_3 deformation. Evidently, strain rates during deformation were sufficiently high to cause the rock to fracture. Concomitant or later deformation, presumably at lower strain rates, allowed the rock to react in a ductile manner, thereby folding the mineralized veins. That the mineralized fractures display pinch-and-swell structures and are, in many cases, boudined indicates that the dunite was more ductile than the mineral aggregates within the fractures during some phases of deformation.

Cataclastic features

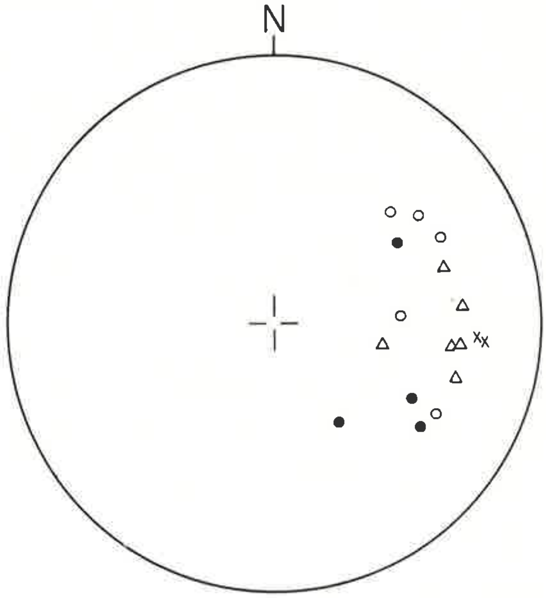
Relict olivine, enstatite, diopside and garnet within the ultramafic bodies show abundant strain features such as undulatory extinction, bent twin planes, micro-fractures, micro-faults and granulated margins. O'Hara & Mercy (1963) believe that these features developed during the intrusion of the ultramafic rocks. However, garnets in the eclogites and labradorite porphyroclasts in the anorthosites also contain cataclastic textures. These features could equally well have developed during *in situ* deformation, and would, in fact, be expected where large differences in the competence between the gneiss and the dunite and eclogite could lead to brittle behavior during deformation.

Boudins

Lappin (1966) and Bryhni (1966) interpret the small eclogites of the western basal gneiss region as boudins. Evidence from the Tafjord area supports this conclusion. The eclogites occur as several lenses along discrete stratigraphic horizons. An apparently primary compositional layering (defined by greater or lesser amounts of garnets relative to plagioclase and hornblende) is commonly truncated by the eclogite-country rock contact, suggesting that the eclogites reacted in a brittle manner to deformation. All gradations can be seen between concordant eclogite lenses that are widely separated from each other by country rocks, to those that had just begun to separate through the formation of boudin 'necks'. Pegmatites and quartz veins fill the fractures within the eclogites and occupy voids left as they pulled apart. The attitudes of boudin lines of eclogites in country rocks and eclogites in anorthosite (Fig. 19) parallel the orientations of boudin lines defined by nearby mafic and leucocratic boudins in the country rocks.

The eclogites are strongly amphibolitized, and the degree of amphibolitization increases toward the margins of each lens. The altered zones completely surround each lens, indicating that amphibolitization occurred after boudinage. In some cases, alteration was extreme enough to form euhedral hornblende and epidote prisms that parallel the attitudes of similar minerals in the country rocks (Fig. 19), suggesting that much of the amphibolitization occurred during the development of the F_1 structural fabric. In a few cases, the F_1 lineation and schistosity are truncated by the eclogite-country rock contact, indicating that the lenses continued to behave as brittle bodies after their amphibolitization. A few strongly retrograded eclogites are folded near their margins, and the gneisses immediately adjacent to them may be strongly folded. Most of these folds formed by flow during boudinage, particularly into the areas vacated as the boudins pulled apart. Their geometries and orientations

Fig. 19. Orientations of some important tectonic fabric elements in the eclogites of the Tafjord area. Mineral lineations in retrograded eclogites (triangles); fold axis in folded eclogite near Kallskaret (x); boudin lines of eclogite lenses in country rock near Kallskaret (large dots) and in anorthosite, southeast shore of Vikvatn (circles).



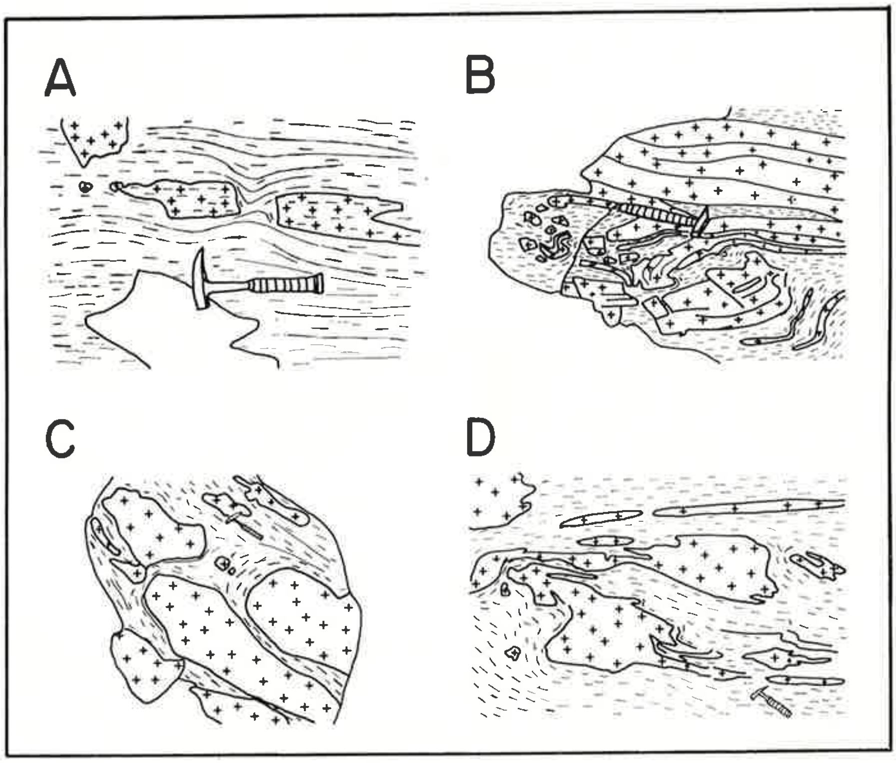


Fig. 20. Examples of thin anorthosite layers near Vikvatn (crossed pattern) showing both brittle and ductile behavior. Dashed lines generally represent biotite schist. Hammer shaft measures 28 cm.

(Fig. 19) are therefore largely controlled by a flow environment that differed from the more homogeneous flow environment that affected rocks remote from the eclogites.

Some of the thin anorthosite layers are also boudined (see Fig. 20). Characteristically, the anorthosite displayed some brittle and some ductile behavior during boudinage (see, e.g. Fig. 20D).

MAJOR STRUCTURES

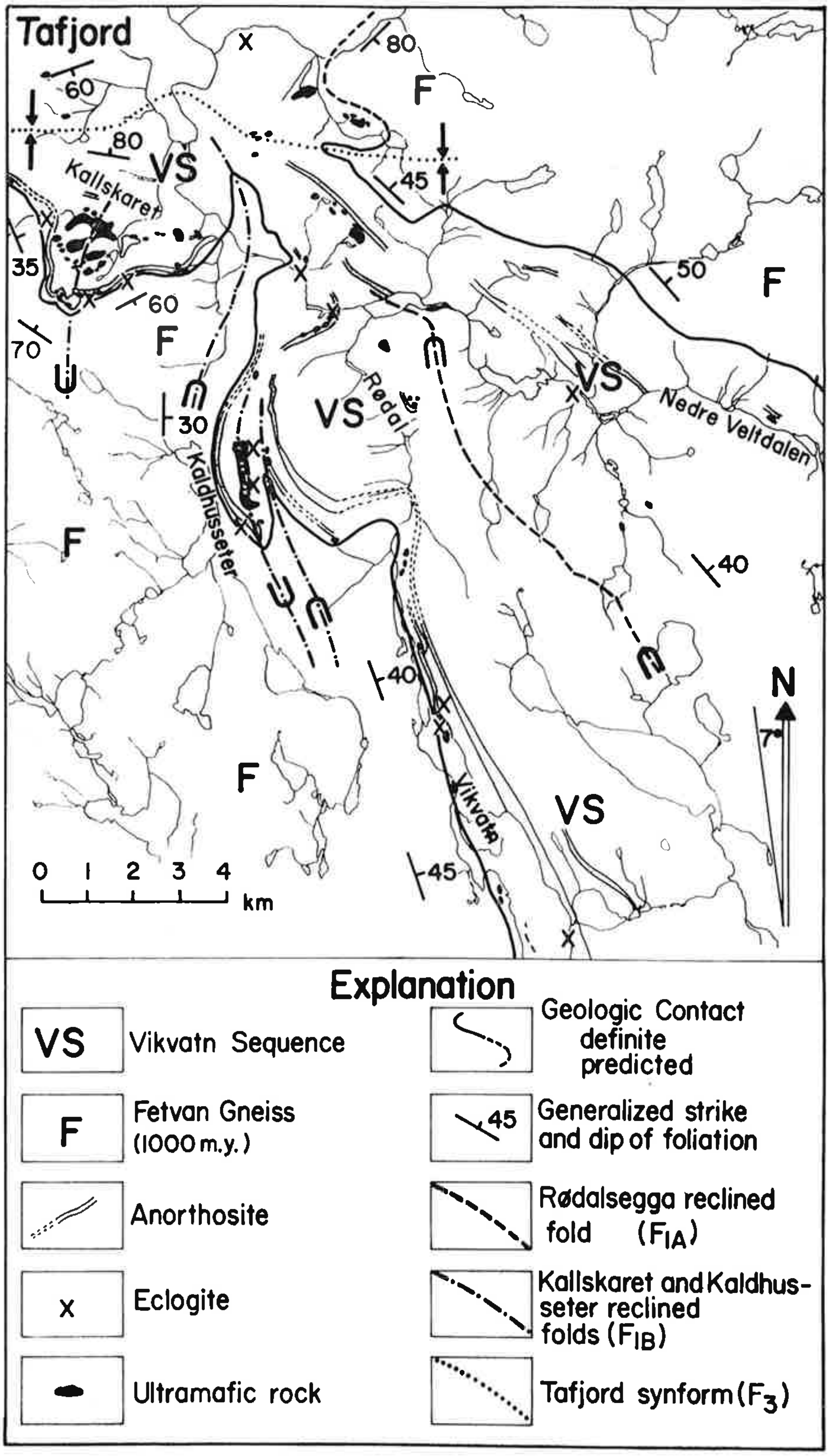
Anorthosite layers and eclogite- and ultramafic rock-bearing horizons are folded about all the major structures of the Tafjord area (Fig. 21). Anorthosites and ultramafic rocks occur in both limbs of the Tafjord Synform (see Plate I) which is tentatively considered to be a large-scale equivalent of F_3 folds. The anorthosite in the basal portion of the Øyen group near Kallskaret can be traced continuously around the large-scale F_{1B} Kallskar reclined fold. Furthermore, an horizon of ultramafic rocks, closely associated with the thick anorthosite layer, and an horizon of eclogites can also be traced around the Kallskar fold (see Plate I). The anorthosite in this fold shows two schistosities. One is a very weakly developed, incipient schistosity, caused by the alignment of a few biotite flakes, that parallels the axial plane of the Kallskar fold. The other schistosity is very strongly developed and parallels the compositional layering in wrapping around the fold nose. It has not been possible to trace an anorthosite layer or a zone of eclogites or ultramafic rocks around the nose of a major F_{1A} fold. Nevertheless, anorthosites, ultramafic rocks and eclogites occur in both the upper and lower limb of the F_{1A} Rødalsegga fold, and hence it is likely that they were present in the country rocks during the formation of this fold.

The basal anorthosite layer in the large F_{1B} fold near Kallskaret is particularly thick near the fold hinge and much thinner near the fold limbs, indicating that it suffered considerable tectonic thinning and thickening during deformation. Similarly, the size of the ultramafic bodies (large at the F_{1B} fold hinges, smaller at the fold limbs) and their distance from each other (close together at the fold hinges, far apart at the limbs) reflect the ductile nature of F_1 folding.

Petrology

The anorthosites, ultramafic rocks and eclogites contain all the visible fold systems and related fabric elements that occur in the enclosing country rocks and therefore have suffered precisely the same deformation suffered by the surrounding country rocks. They must, therefore, have been in their present stratigraphic and tectonic position during or before F_{1A} deformation. The

Fig. 21. Generalized geologic map of the Tafjord area showing distribution of the Fetvatn Gneiss and Vikvatn Sequence, the preferred interpretation of major structures, and the distribution of ultramafic lenses, eclogite boudins and anorthosite layers within the Vikvatn Sequence.


Tafjord
Kallskaret
Kaldhusseter
Rødal
Nedre Veltdalen
Vikvatn
VS
F
N
7°
0 1 2 3 4
km
Explanation
VS Vikvatn Sequence
F Fetvan Gneiss (1000 m.y.)
Anorthosite
x Eclogite
Ultramafic rock
Geologic Contact definite predicted
45 Generalized strike and dip of foliation
Rødalsegga reclined fold (F_{IA})
Kallskaret and Kaldhusseter reclined folds (F_{IB})
Tafjord synform (F_3)

country rocks suffered almandine–amphibolite facies metamorphism during the development of the F_1 structural fabric. Yet the eclogites and ultramafic rocks and possibly the anorthosites retain minerals and mineral assemblages that are generally believed to have formed under different physical and chemical conditions than those typical of the almandine–amphibolite facies. The conditions under which these assemblages either formed or at least survived the almandine –amphibolite facies metamorphism is a major petrogenetic problem. The following section reviews the general petrography of the ultramafic rocks, eclogites and anorthosites of the Tafjord area. Detailed petrographic data and modes of these rocks are presented elsewhere (Brueckner 1968) and are available from the author upon request.

ANORTHOSITE

Ninety-five percent of the anorthosites of the Tafjord area are creamy or chalky white to dark grey bodies containing 75% to 95% plagioclase (An_{45-65}) with hornblende, actinolite, biotite, epidote, chlorite and muscovite as the major accessories. The anorthosite's microscopic appearance is dominated by 0.5 to 1.5 mm-sized, interlocking, anhedral plagioclase grains, at least 50% of which are twinned according to the albite and the albite-plus-pericline laws. Many grains show reverse zoning, and differences of 10%–15% in the anorthite content between the core and the margin of a single grain are not uncommon. The mafic minerals may be concentrated into thin laminae and show very strong preferred orientations, giving some anorthosites a banded appearance. Anorthosite layers at Vikvatn and Kallskaret contain scattered, very large (up to 10 cm), light grey, plagioclase porphyroclasts enclosed in and cut by the normal, finer-grained plagioclase matrix. The porphyroclasts have the same anorthite content (An_{60}) as the grains of the groundmass. They lack, however, any compositional zoning. Cataclastic features, including microfractures, microfaults, banded or irregular extinction, bent twin and cleavage planes, and granulated margins, are far better developed in the porphyroclasts than in the groundmass. The large grains lack foreign inclusions and do not distort the surrounding schistosity, and hence are not porphyroblastic growths.

Mafic rocks occur as random layers, schlieren and boudins in most anorthosite layers. Most are strongly lineated quartz–biotite–hornblende–plagioclase schists. A few weakly schistose boudins and layers contain relict garnet, and may have been eclogites (O'Hara & Mercy 1963). Mafic schlieren consisting almost entirely of actinolite and anthophyllite occur in anorthosites at Vikvatn and Kallskaret. A lens containing very coarse (15 to 20 cm) corundum crystals coexisting with lesser amounts of fine-grained margarite and chlorite occurs in the basal anorthosite south of Kallskaret.

ULTRAMAFIC ROCK

Dunite

Most of the ultramafic rocks of the Tafjord area are fine- to medium-grained dunites containing olivine (80%–90%), enstatite (0–10%), diopside (0–

1%), chromite (0–1%), and variable amounts of chlorite, tremolite, talc, antigorite and magnetite. Chlorite and tremolite form subhedral flakes and prisms, respectively, that define the dunite schistosity and mineral lineation. They generally make up only 5% to 20% of the rock, but lenses made up almost entirely of chlorite and/or tremolite occur in a few localities. The olivine is flattened parallel to the chlorite schistosity and elongate parallel to the tremolite lineation, and shows strong crystallographic preferred orientation. Enstatite and diopside form small, stubby, less well-oriented prisms which are of the same grain-size as the olivine. Most olivine, enstatite and diopside grains extinguish cleanly and show very few strain features. Strung out chromite granules occur in all ultramafic rocks where garnet is absent.

Large (up to 3 cm) olivine and enstatite porphyroclasts, optically similar to olivine and enstatite in the enclosing groundmass, occur in some chlorite-poor masses. They display abundant strain features, including undulatory extinction, lamellar twin bands, granulated margins and bent and fractured cleavage planes, that are missing in the minerals of the groundmass. Original large grains may have been broken into several smaller ones that are separated from each other by the recrystallized groundmass. Inclusions are rare, and the grains do not distort surrounding schistosity.

The margins of the ultramafic bodies may be altered to enstatite-rich rocks, tremolite and/or chlorite schists, talc–tremolite–carbonates and serpentine-bearing rocks. Although entire ultramafic lenses may be converted to these assemblages, most of the alteration zones are thin, commonly less than one metre. The folded, irregular fractures that occur in many ultramafic bodies also contain a variety of minerals and mineral assemblages. The most deformed fractures can contain enstatite, anthophyllite, tremolite and chlorite. Late, less deformed fractures contain carbonate (magnesite and dolomite), talc and tremolite. The very late, undeformed fractures contain fine-grained aggregates of antigorite. The minerals and mineral assemblages that occur in the fractures and at the margins of the ultramafic bodies may be arranged in zones.

Garnet peridotite and garnet pyroxenite

A lens, roughly 25 m long and 8 m wide, of garnet peridotite and garnet pyroxenite occurs conformably within a dunite at Kallskaret. These striking, beautiful rocks are ultrabasic rather than basic in composition (Bryhni et al. 1969, Carswell 1973) and should not be confused with 'true' eclogites that are essentially basic in composition. The garnet peridotites consist of large (0.1 to 1.5 cm) porphyroclasts of red garnet, green diopside, light-green olivine and yellow-brown enstatite set in a yellow-weathering matrix of much finer-grained olivine, tremolite and chlorite. Schmitt (1963) divides the mineral assemblages of the eclogites and garnet peridotites of the Eiksundal area into a 'macro-assemblage', referring to minerals that constitute the coarsest-grained mineral aggregate in a given rock, and a 'corona-assemblage', referring to minerals that occur in the reaction rims of the minerals of the macro-assemblage. A similar division may be applied to the garnet peridotite at Kallskaret.

The macro-assemblage consists of pyrope-rich garnet ($Fe_{27.4}$ $Mn_{.6}$ $Mg_{59.6}$ $Ca_{12.1}$ to $Fe_{17.5}$ $Mn_{.6}$ $Mg_{69.8}$ Ca_{12}; O'Hara & Mercy 1963) diopside, olivine and enstatite. The relative proportions of the macro-assemblage minerals vary, defining a compositional layering that parallels the lens outline. Most of the grains of the macro-assemblage contain strain features such as undulatory extinction, twinning, bent or fractured twin and cleavage planes, mortar textures along fractures and granulated margins. The garnets are surrounded by kelyphite alteration rims containing fine, radiating prisms or fibers of amphibole, chlorite and opaques. Similarly, diopside may be wholly or partially replaced by an amphibole–plagioclase symplectite. Enstatite and olivine, on the other hand, are unaltered but may be granulated and recrystallized.

The corona-assemblage consists of the kelyphites and symplectites around the garnet and diopside, respectively, and the recrystallized groundmass of olivine, enstatite and amphibole. The corona-assemblage may range from very thin reaction rims around the coarse-grained minerals to complete new assemblages including porphyroblastic growths. The rim of the garnet peridotite lens, particularly, is strongly amphibolitized. The minerals in the corona-assemblage display fewer cataclastic features than those in the macro-assemblage. However, they do not show the same strong preferred orientations as similar minerals in the garnet-free dunites. Amphiboles show variable optics (and hence compositions) and occur in different textural domains (kelyphites, symplectites and groundmass). It is doubtful that the minerals within each domain are in equilibrium with minerals of other nearby domains, even within a single thin-section.

ECLOGITE

Eclogites occur as lenses in the country rocks, anorthosites and ultramafic rocks of the Tafjord area. They, like the garnet peridotite, can be divided into a 'macro-assemblage' and a 'corona-assemblage' (Schmitt 1963). In most lenses, the macro-assemblage consists only of garnet with compositions ranging from $Fe_{55}Mg_{16}Ca_{29}$ to $Fe_{43}Mg_{29}Ca_{28}$ (garnet compositions were determined using the optical and x-ray diffraction techniques described by Winchell & Winchell 1961). These garnets are significantly less pyrope-rich than the garnets in the garnet peridotite and provide a means of distinguishing eclogites from garnet peridotites. All grains are fractured, commonly along curved or irregular surfaces. Eclogite garnets are surrounded by reaction rims of secondary products. The degree of alteration increases toward the margins of the eclogite lens, and garnets may be completely altered near the eclogite-country rock contact. These features contrast with those of the evenly distributed, unaltered, euhedral garnets in garnet-amphibolites. Clinopyroxene (?omphacite) occurs as relatively unaltered prisms associated with unaltered garnet in the core of an eclogite lens in the Kaldhusseter ultramafic body. One eclogite lens in the country rock contains clinopyroxene as a subhedral inclusion in garnet.

The corona-assemblage occurs in two textural domains: the kelyphites and symplectites. The kelyphites formed through the alteration of garnet, and consist of fine, radiating prisms and fibers of plagioclase, amphibole, biotite and opaques. Symplectites are fine, worm-like intergrowths of plagioclase and amphibole, or, in a few lenses, plagioclase and diopside, that developed through the alteration of sodium-rich omphacite (Eskola 1921). Other minerals in the eclogite corona-assemblage are zoisite, clinozoisite, epidote and quartz. Similar hydrous minerals in the amphibolites and garnet amphibolites are medium- to coarse-grained, euhedral prisms and flakes that display moderate to strong preferred orientations. Epidote, biotite and amphibole may form oriented porphyroblasts in some of the more recrystallized eclogites.

DISCUSSION AND INTERPRETATION

Petrology of the mineral assemblages formed during the development of the F_1 structural fabric

Most of the anorthosites, ultramafic rocks and strongly retrogressed eclogites contain an F_1 schistosity and mineral lineation that formed during the F_1 metamorphism and deformation. In the anorthosites, for example, the granoblastic fabric of the fine-grained plagioclase and the strong preferred orientation of the hornblende, actinolite, epidote, biotite and muscovite suggest that almost the entire anorthosite mineralogy recrystallized during F_1 metamorphism. Most of these minerals are typical of the almandine-amphibolite facies, and are common in the associated country rocks. Epidote and actinolite are more typical of the greenschist facies, but may be stable in the almandine-amphibolite facies in calcium-rich environments. It is interesting that the calc-silicate rocks are the only other rocks in the Tafjord area that contain primary actinolite and epidote.

The anthophyllite, actinolite, hornblende, chlorite and biotite that occur in mafic schlieren and layers in most anorthosites are strongly oriented parallel to the fabric elements of the F_1 system, and are stable in the almandine-amphibolite facies in sufficiently magnesium-rich rocks (Bryhni 1963). The corundum lens in the basal anorthosite layer south of Kallskaret may have formed through hydrothermal alteration of anorthosite. Water vapor, streaming through a fracture in the anorthosite during a thermal event, could have removed the mobile ions (calcium, sodium, silica), leaving the relatively immobile aluminum to form corundum.

The common dunite mineral association is olivine-enstatite-diopside-chromite-chlorite-tremolite. Chlorite, tremolite and flattened and elongated olivine define the rock schistosity and therefore formed during F_1 metamorphism. The few stubby enstatite prisms show a weak preferred orientation and the same grain-size as the surrounding olivine, lack cataclastic textures, F_1 metamorphism. Diopside, on the other hand, may be partially altered to and show no signs of alteration, suggesting that they too were stable during tremolite. Also, some chromite grains are slightly altered to chlorite and new opaques. Thus diopside and chromite may not have been stable with the adjacent mineral assemblage, but nevertheless survived metamorphism without pervasive alteration.

It was initially proposed (Brueckner 1969) that the assemblage olivine-enstatite-diopside-chromite-chlorite-tremolite could have survived F_1 metamorphism if insufficient water entered the ultramafic system to allow the completion of hydration reactions. Evidence to support this hypothesis comes from the zoned mineral assemblages that occur at some of the dunite margins and in some of the fractures that cross-cut them (Brueckner 1969, Carswell et al. 1974).

The reactions that produce the zoned sequences are described in considerable detail by Carswell et al. (1974) from an enstatite, anthopyhillite, tremolite and chlorite-bearing vein within the Kallskar peridotite. They suggest that the actual zonation reflects variation in the extent of metasomatic penetration of the different ionic species. Water evidently did not penetrate deeply into the ultramafic rock allowing most of the ultramafic bodies to remain essentially anhydrous throughout F_1 metamorphism.

The Tafjord area eclogites and garnet peridotites contain some mineral assemblages that formed during F_1 metamorphism. Both rock-types contain corona-assemblages that are clearly secondary after the primary macro-assemblages. The two important reactions that lead to the development of the corona-assemblage in the Tafjord eclogites are:

Mg-rich garnet → plagioclase + amphibole + opaques

Na- + Ca-rich clinopyroxene (omphacite) → Na-poor clinopyroxene (diopside) + plagioclase → hornblende + plagioclase

The final product of these reactions is an assemblage of plagioclase, amphibole, biotite, quartz, minerals of the epidote group and opaques. Garnet peridotites alter to similar assemblages although those initially containing olivine and enstatite will also contain recrystallized enstatite, tremolite and olivine and will be quartz-free. Most of the minerals of the corona-assemblage are very fine-grained and do not display preferred orientations, and hence are not readily related to the F_1 structural fabric. Furthermore, it is doubtful that the minerals within each textural domain (i. e., within each kelyphite rim or symplectite) are in equilibrium with minerals of other nearby domains. Nevertheless, the corona-minerals and mineral assemblages are typical of those produced during almandine-amphibolite metamorphism, and presumbly formed during the F_1 almandine-amphibolite metamorphism. That a few of the more completely recrystallized amphibole, biotite and epidote porphyroblasts display a preferred orientation that parallels the F_1 fold axes supports this conclusion.

The corona-assemblages are hydrous ones that could have formed only if water entered the system. The presence of secondary biotite indicates that some potassium may also have been introduced. Apparently, however, insufficient water entered the systems during metamorphism to completely alter the macro-assemblages to corona-assemblages. The partially altered garnet peridotite at Kallskaret occurs in the middle of a large dunite mass that may have armored the lens against total hydration. Furthermore, the only eclogite observed to contain primary omphacite occurs in the middle of the Kaldhusseter ultramafic body. Mafic or ultramafic lenses occurring near the margins of dunite bodies consist almost entirely of hydrous assemblages, suggesting that they were too near the dunite-country rock contact to be effectively shielded against the introduction of water.

Relict assemblages

Scattered cataclastically deformed minerals and mineral assemblages (labradorite porphyroclasts in the anorthosite, olivine and enstatite porphyroclasts in the dunite, garnet in the eclogite, and the assemblage pyrope-rich garnet-diopside-enstatite-forsterite in the garnet peridotite) are interpreted as relics that escaped recrystallization during F_1 metamorphism and deformation. These relics suggest that the anorthosites, ultramafic rocks and eclogites of the Tafjord area suffered a possible pre-F_1 metamorphic and igneous history.

The large grain-size and homogeneous anorthite content of the plagioclase relics in the anorthosites are typical of plagioclase in anorthosites of the Adirondack type. Bryhni (1966) suggests that similar coarse plagioclase grains in the anorthosites at Fiska and Sandsøy had a magmatic origin. The macro-assemblages of the eclogites and garnet peridotites indicate crystallization under eclogite-facies conditions. Although there is little doubt that the ultrabasic and basic compositions of these rocks originated in the mantle, there is considerable controversy about where the actual eclogite-facies mineralogies of these rocks formed. O'Hara & Mercy (1963), Lappin (1966, 1967, 1973, 1974), Carswell (1968b), O'Hara et al. (1971) and others suggest that the macro-assemblages of the eclogites and garnet peridotites were formed in the upper mantle. Gjelsvik (1960), Hernes (1954), Kolderup (1960b), Schmitt (1963), Bryhni et al. (1969), Bryhni et al. (1970) and Mysen & Heier (1972), on the other hand, conclude that the eclogite-facies assemblages formed in the crust.

A similar difficulty exists for determining the time of eclogite-facies metamorphism. Geochronological evidence reviewed in an earlier part of this report suggests that the eclogite-bearing Øyen and Øvste Rødal groups of the Vikvatn Sequence may have originated as long as 1800 m. y. ago. McDougall & Green (1964) have determined K-Ar ages of clinopyroxene, amphibole and phlogopite from eclogites and garnet peridotites from the western basal gneiss region. Secondary hornblendes, that formed during the waning of the P-T environment responsible for the macro-assemblage or during a later metamorphic event, yield ages from 1740 to 1850 m. y. K-Ar ages on less well-formed hornblendes and phlogopites from the margins of more strongly retrograded lenses range from 950 to 1170 m. y.; McDougall & Green interpret these ages as indicating a metamorphism in this period. Finally, K-Ar and Rb-Sr ages on micas give ages of about 400 m. y., reflecting the effects of Caledonian metamorphism. Thus, this evidence strongly suggests that the eclogites and garnet peridotites may have suffered two periods of retrograde metamorphism prior to the Caledonian metamorphism. The older K-Ar ages may be anomalous, because of excess argon in the amphibole and phlogopite structures, and have been disputed by O'Hara (1965). Furthermore, there is equally strong geochronological evidence that the actual eclogite-facies metamorphism occurred during the Caledonian orogeny. Krogh et al. (1975) report a U-Pb age of 401 ± 20 m. y. from apparently primary zircons from the Hareidland Eclogite. If this age is valid, it suggests that the older K-Ar ages are the result of excess argon during metamorphism. It remains to be explained, however, why the K-Ar ages from the eclogites give ages that coincide with either the Svecofennian, Sveconorwegian or Caledonian orogenies, but not with the intervals between these events. Field data from the Tafjord area do not resolve these controversies. However, the data provide certain restrictions that must be observed for any satisfactory model for the formation of the eclogite-facies assemblages.

(1) The present spatial distribution, shapes and structures of the anorthosites, ultramafic rocks and eclogites cannot be used to infer mechanisms of intrusion or formation of these rocks since many of these features were acquired during the one, and possibly three, orogenic events that post-dated their formation. For example, the tabular form of most anorthosites led Lappin (1966) to suggest that they were inserted into the basal gneiss region as solid sheets. Yet the present sill-like form of these bodies could equally well be the result of intense tectonic thinning. The same probably is true of the eclogites. Bryhni (1966, pp. 58–59) suggests that they could not have intruded the enclosing gneisses as solid bodies because 'it is difficult to visualize how the thousands of small and large bodies of eclogitic rocks in the gneisses may have been transported any large distance relative to the enclosing gneiss.' Yet their present sizes and numbers are at least partially the result of boudinage during F_1 formation. Lappin (1966) attempted to use the present distribution of ultramafic rocks to derive their intrusion mechanism. He noted that most ultramafic bodies in Sunnmøre occur in east-west trending directions, and suggested that these directions define the shear zone along which the ultramafic rocks intruded. However, ultramafic rocks occur in certain stratigraphic levels that have suffered multiple foldings. That ultramafic rocks occur along east-west zones in Sunnmøre merely suggests that the limbs of large-scale folds there trend east-west.

(2) There is no petrographic evidence from Tafjord that the macro-assemblages formed as a result of F_1 metamorphism. The minerals of the macro-assemblages of the Kallskar garnet peridotite show no morphological or crystallographic relationship to the elements of the F_1 fabric; a feature that would be expected if they formed during F_1 crystallization. Unfortunately, relict garnets in the country rock eclogites are morphologically equant and crystallographically isotropic, and hence little can be said about their relation to the F_1 fabric. The only minerals that do show clear relationships to the F_1 fabric are the corona assemblages. The evidence from Tafjord is limited, however. A major effort should be made with less altered eclogite-facies assem-

blages to see if there is any morphological or crystallographic relationship between eclogite facies minerals and the tectonic fabric in the enclosing country rocks.

(3) If the macro-assemblages of the garnet peridotites had a pre-F_1 origin, and if they formed in their present stratigraphic and tectonic position, the rocks surrounding them might also be expected to contain relict assemblages reflecting the earlier phase of metamorphism. Most of the gneisses and meta-sediments of the Tafjord area do not show such features. However, some of the syenitic rocks of the Øvste Rødal group, just south of Zakariasvatn, contain minerals and textures that suggest a granulite or eclogite facies metamorphism prior to the almandine–amphibolite facies F_1 metamorphism. Similar relics have been described in supracrustal rocks of the western basal gneiss region, and they generally appear to be associated with eclogites, anorthosites and ultramafic rocks (Lappin 1966, Bryhni 1966, Carswell 1968b). Interestingly, Bryhni et al. (1971) interpret Ar^{40}/Ar^{39} data from a pyroxene from 'mangeritic rocks' of the 'anorthosite kindred' as indicating a pre-1550 m. y. origin for these rocks. They suggest that these rocks may occur as tectonic inliers within younger, non-granulitic rocks. It is quite possible that more pre-Caledonian relics occur in the country rocks than have previously been recognized.

(4) The eclogites and garnet peridotite lens of the Tafjord area are found only within the Øyen and Øvste Rødal groups of the Vikvatn Sequence. No eclogites were found in the 1000 m. y. Fetvatn Gneiss, even though rocks of appropriate mafic compositions (i.e., amphibolites) are common within it. The Fetvatn gneisses recrystallized under the same metamorphic conditions as the rocks of the Vikvatn Sequence. It is difficult to envisage a metamorphic process whereby mafic rocks would form eclogites in the Vikvatn Sequence, but not in the Fetvatn Gneiss, during Caledonian metamorphism.

Possible origin of the anorthosites, ultramafic rocks and eclogites of the basal gneiss region

This author still finds merit in Eskola's (1921) suggestion that anorthosites, ultramafic rocks and eclogites may have a common origin. This hypothesis explains the common spatial and stratigraphic association of these rocks, and is consistent with the fact that one rock-type may be found within another. Lappin's (1966) argument that the anorthosites, eclogites and ultramafic rocks intruded the basal gneiss region at different times from different levels of the earth's lower crust and upper mantle negates Eskola's suggestion. However, there is no evidence in the Tafjord area that they intruded at different times. The only restriction on the time of intrusion or formation of these rock-types is that it occurred before the F_1 deformation. The anorthosites, eclogites and ultramafic rocks contain relict minerals and mineral assemblages suggesting that they have undergone a common pre-F_1 metamorphic history.

Additional evidence comes from the striking abundance of apparently genetically related mafic, ultramafic and anorthositic rocks throughout the entire length of the Norwegian Caledonides in such areas as the Jotun nappes, the Bergen arc districts, the islands of Lofoten and Vesterålen, the Lyngen peninsula and, in western Finnmark and northern Troms, the islands of Seiland, Stjernøy and Sørøy and the peninsulas of Øksfjord and Bergsfjord (Holtedahl & Dons 1960). The ages of some of these complexes are still uncertain. However, Heier & Compston (1969) and Brueckner (1971) have determined ages of 1700 to 1800 m. y. for gneisses and granites associated with mafic, ultramafic and anorthositic rocks in Lofoten and Vesterålen. Brueckner (1973) has found some rather inconclusive evidence for a 1600 m. y. event in the salic, mafic and ultramafic rocks of the Seiland Petrographic Province of northern Norway. The few ages that come from these areas are the same as the ages from the basal gneiss region, supporting the hypothesis of a common origin.

It is, therefore, extremely tempting to suggest that the anorthosites, ultramafic rocks and eclogites (or other mafic rocks) of the basal gneiss region had a similar origin, but were caught up in and radically modified by Caledonian or older, pre-F_1 deformation and metamorphism. Some of these bodies display compositional variations that are reminiscent of the kind found in layered intrusives. Schmitt (1963), for example, describes three kinds of compositional layering in the large Eiksundsdal eclogite complex in western Norway: (1) thick, mappable layers of mafic or ultramafic composition, (2) thin layers reflecting variations in the relative proportions of minerals within the thick layers, and (3) rhythmically repeated layers of sharply contrasting mineralogy. The thick layers may be zoned, and vary from ultramafic compositions at one contact to progressively less mafic compositions toward the other. Schmitt suggested that these variations within layers may be the result of fractional crystallization, and therefore interprets the Eiksundsdal complex as a metamorphosed layered gabbro complex.

Much of the evidence for a common origin is admittedly circumstantial, and there are several problems that are not easily explained. To find one rock-type near or within another may not indicate a genetic relationship, but, as Lappin (1966) suggests, may merely be caused by juxtaposition during deformation or intrusion. In the Tafjord area, the total quantity of anorthosite and ultramafic rocks far exceeds that of the eclogites and other mafic rocks. The opposite is true in most mafic–ultramafic layered intrusions. At one stage in their history, some mafic and ultramafic rocks were metamorphosed to garnet–omphacite and garnet–diopside–enstatite–olivine assemblages, respectively. When this event occurred and whether it occurred while the rocks were in their present stratigraphic positions or in the upper mantle is still debatable.

Eskola's (1921) suggestion is most consistent with evidence from the Tafjord area and the rest of the Norwegian Caledonides. It is accepted in this report as a working hypothesis that should not be summarily discarded until strong evidence suggests a contrary view.

Acknowledgements. – This paper is a condensation of my Ph.D. dissertation for Yale University, submitted in 1968. John Rodgers, Edward Hansen and William Scott instigated and encouraged the Tafjord area investigation, and the work profited immeasurably from their guidance during field visits and their discussions and criticisms at home. Daniel Lewis, John Madden, Paul Nunes, Russel Wheeler, William Sacco and Anne Brueckner were valuable field assistants. Richard Armstrong, Anne Brueckner, Neville Carter, Sydney Clark, Edward Hansen, John Rodgers and Trygve Strand critically read this manuscript or portions of it. My personal thanks go to Tony Carswell of the University of Sheffield, with whom I spent many profitable hours in the Tafjord area, and to Professor Trygve Strand of the University of Oslo for his hospitality and encouragement while I visited him in his field area in Grotli.

Mr. and Mrs. Tafjord, the members of the Ålesund–Sunnmøre Turistforening, and the workers of Tafjord Kraftselskap provided hospitable room and board. Tafjord Kraft also provided field maps. Sigma Xi helped with field expenses during the summers of 1964, 1965 and 1966.

REFERENCES

Banham, P. H. 1966: Fault vein mineralization as the result of shearing in Norwegian basement rocks. *Norsk geol. Tidsskr. 46*, 181–191.

Banham, P. H. & Elliott, R. B. 1965: Geology of the Hestbrepiggan area – preliminary account. *Norsk geol. Tidsskr. 45*, 189–198.

Broch, O. A. 1964: Age determination of Norwegian minerals up to March 1964. *Norges geol. Unders. 228*, 84–113.

Brueckner, H. K., Wheeler, R. L. & Armstrong, R. L. 1968: Rb–Sr isochron for older gneisses of the Tafjord area, basal gneiss region, south-western Norway. *Norsk geol. Tidsskr. 48*, 127–131.

Bruecknes, H. K. 1968: *The relationships of anorthosites, ultramafic rocks, and eclogites to the country rocks of the Tafjord area, southwestern Norway.* Unpubl. Ph.D. thesis, Yale Univ. 149 pp. Available on microfilm.

Brueckner, H. K. 1969: Timing of ultramafic intrusion in the core zone of the Caledonides of southern Norway. *Am. Journ. Sci. 267*, 1195–1212.

Brueckner, H. K. 1971: The age of the Torset Granite, Langøy, northern Norway. *Norsk geol Tidsskr. 51*, 85–87.

Brueckner, H. K. 1972: Interpretation of Rb–Sr Ages from the Precambrian and Paleozoic Rocks of southern Norway. *Am. Journ. Sci. 272*, 343–358.

Brueckner, H. K. 1973: Reconnaissance Rb–Sr investigation of salic, mafic and ultramafic rocks in the Øksfjord area, Seiland province, northern Norway. *Norsk geol. Tidsskr. 53*, 11–23.

Bryhni, I. 1963: On anthophyllite and some reaction zones in anorthosite. *Norsk geol. Tidsskr. 43*, 59–71.

Bryhni, I. 1966: Reconnaissance studies of gneisses, ultrabasites, eclogites and anorthosites in outer Nordfjord, western Norway. *Norges geol. Unders. 241*, 68 pp.

Bryhni, I., Bollingberg, H. J. & Graff, P.-R. 1969: Eclogites in quartzo-feldspathic gneisses of Nordfjord, West Norway. *Norsk geol. Tidsskr. 49*, 193–226.

Bryhni, I. & Grimstad, E. 1970: Supracrustal and infracrustal rocks in the Gneiss Region of the Caledonides west of Breimsvatn. *Norges geol. Unders. 266*, 105–140.

Bryhni, I., Green, D. H. & Heier, K. S. 1970: On the occurrence of eclogite in western Norway. *Contr. Mineral. and Petrol. 26*, 12–19.

Bryhni, I., Fitch, F. J. & Miller, J. A. 1971: $^{40}Ar/^{39}Ar$ dates from recycled Precambrian rocks in the Gneiss region of the Norwegian Caledonides. *Norges geol. Unders. 51*, 391–406.

Carswell, D. A. 1968a: Possible primary upper mantle peridotite in Norwegian basal gneiss. *Lithos. 1*, 322–355.

Carswell, D. A. 1968b: Picritic magma-residual dunite relationships in garnet peridotites at Kalskaret near Tafjord, south Norway. *Contr. Mineral. Petrol. 19*, 97–124.

Carswell, D. A. 1973: The age and status of the basal gneiss complex of north-west southern Norway. *Norges geol. Tidsskr. 53*, 65–78.

Carswell, D. A., Curtis, C. D. & Kanaris-Sotiriou, R. 1974: Vein metasomatism in peridotite at Kalskaret near Tafjord, south Norway. *Journ. Petrology 15*, 383–402.

Coombs, D. S., Ellis, A. J., Fyfe, W. S. & Taylor, A. M. 1959: The zeolite facies, with comments on the interpretation of hydrothermal synthesis. *Geochim, et Cosmochim Acta 17*, 53–107.

Eskola, P. 1921: On the eclogites of Norway. *Norsk Vidensk. Selsk. Skr., Mat. Nat. Kl. 8*, 118 pp.

Gjelsvik, T. 1951: Oversikt over bergartene i Sunnmøre og tilgrensende deler av Nordfjord. *Norges geol. Unders. 179*, 45 pp.

Gjelsvik, T. 1953: Det nordvestlige gneis-område i det sydlige Norge, aldersforhold og tektonisk–stratigrafisk stilling. *Norges geol. Unders. 184*, 71–94.

Gjelsvik, T. 1957: Geochemical and mineralogical investigations of titaniferous iron ores, west coast of Norway. *Econ. Geology 52*, 482–498.

Gjelsvik, T. 1960: The differentiation and metamorphism of the west Norwegian dolerites. *Norsk Vidensk. Selsk. Skr. 3*, 1–11.

Hansen, E. L. 1963: *Strain facies of the metamorphic rocks of western Dovrefjell, Norway.* Unpubl. Ph.D. thesis. Yale Univ. 194 pp.

Hansen, E. C. 1971: Strain facies. New York, *Springer Verlag*, 220 pp.

Heier, K. S. & Compston, W. 1969: Interpretation of Rb–Sr patterns in high-grade metamorphic rocks, north Norway. *Norsk geol. Tidsskr. 49*, 257–283.

Hernes, I. 1954: Eclogite–amphibolites on the Molde peninsula, southern Norway. *Norsk geol. Tidsskr. 33*, 163–197.

Hernes, I. 1957: Connections between the Trondheim and Sunnhordland regions, Caledonides of Norway. *Norsk geol. Tidsskr. 37*, 247–255.

Hernes, I. 1967: The late Pre-Cambrian stratigraphic sequence in the Scandinavian mountain chain. *Geol. Mag. 104*, 557–563.

Hernes, I. 1970: Tafjord–Grotli-området og grensen mellom det eldre prekambriske kompleks og den senprekambriske – eokambriske lagrekke. *Norsk geol. Tidsskr. 50*, 261–268.

Holtedahl, H. 1950: Geological and petrological investigations in the northwestern part of the Opdal quadrangle. *Univ. Bergen Årb. 1949, naturv. rekke, 7*, 60 pp.

Holtedahl, O. 1938: Geological observations in the Opdal–Sunndal–Trollheimen district. *Norsk geol. Tidsskr. 18*, 29–53.

Holtedahl, O. 1944: On the Caledonides of Norway with some scattered local observations. *Skr. Norske Vidensk.-Akad. i Oslo, Mat.-naturv. Kl. 4*, 31 pp.

Holtedahl, O. & Dons, J. A., compilers, 1960: Geologisk kart over Norge, Berggrunnskart. *Norges geol. Unders. 208.*

Hopson, C. A. 1964: The crystalline rocks of Howard and Montgomery Counties. *In* The geology of Howard and Montgomery Counties. *Maryland Geological Survey*, 27–215.

Kolderup, N.-H. 1960a: The relationship between Cambro-Silurian schists and the gneiss complex in the deep Caledonides of Sogn and Fjordane, West Norway. *Internat. Geol. Cong., 21st, Oslo 1960, Guidebk. to Excursions A5–C2*, 31 pp.

Kolderup, N.-H. 1960b: Origin of Norwegian eclogites in gneisses. *Norsk geol. Tidsskr. 40*, 73–76.

Krogh, T. E., Mysen, B. O. & Davis, G. L. 1975: A Paleozoic age for the primary minerals of a Norwegian eclogite. *Carnegie Inst. Wash. Yearb. 73*, 575–576.

Kulp, J. L. & Neumann, H. 1961: Some potassium–argon ages on rocks from the Norwegian basement. *N. Y. Acad. Sci. Annals 91*, 469–475.

Lappin, M. A. 1966: The field relationships of basic and ultrabasic masses in the basal gneiss complex of Stadlandet and Almklovdalen, Nordfjord, southwestern Norway. *Norsk geol. Tidsskr. 46*, 439–495.

Lappin, M. A. 1967: Structural and petrofabric studies of the dunites of Almklovdalen, Nordfjord, Norway. *In* Wyllie, P. J. (ed.): Ultramafic and related rocks. New York, John Wiley & Sons, Inc. 183–190.

Lappin, M. A. 1973: An unusual clinopyroxene with complex lamellar intergrowths from an eclogite in the Sunndal–Grubse ultramafic mass, Almklovdalen, Nordfjord, Norway. *Mineralog. Mag. 39*, 313–320.

Lappin, M. A. 1974: Eclogites from the Sunndal–Grubse ultramafic mass, Almklovdalen, Norway, and the T–P history of the Almklovdalen masses. *Journ. Petrol. 15*, 567–601.

McDougall, I. & Green, D. H. 1964: Excess radiogenic argon in pyroxenes and isotopic ages on minerals from Norwegian eclogites. *Norsk geol. Tidsskr. 44*, 183–196.

Muret, G. 1960: Partie S. E. de la culmination du Romsdal, chaîne calédonienne, Norvège. *Internat. Geol. Cong., 21st, Copenhagen 1960, Proc. 19,* 28–32.
Mysen, B. O. & Heier, K. S. 1972: Petrogenesis of eclogites in high grade metamorphic gneisses exemplified by the Hareidland eclogite, western Norway. *Cont. Mineral. and Petrol. 36,* 73–94.
O'Hara, J. J. & Mercy, E. L. P. 1963: Petrology and petrogenesis of some garnetiferous peridotites: *Roy. Soc. Edinb. Trans. 65,* 251–314.
O'Hara, M. J., Richardson, S. W. & Wilson, G. 1971: Garnet–peridotite stability and occurrence in crust and mantle. *Contr. Mineral. and Petrol. 32,* 46–68.
O'Hara, M. J. 1965: On diamantiferous diatremes. *Econ. Geol. 60,* 830–832.
O'Nions, R. K. & Baadsgaard, H. 1971: A radiometric study of polymetamorphism in the Bamble region, Norway. *Contr. Mineral. and Petrol. 34,* 1–24.
O'Nions, R. K. & Heier, K. S. 1972: A reconnaissance Rb–Sr geochronological study of the Kongsberg area, south Norway. *Norsk geol. Tidsskr. 52,* 143–150.
Pidgeon, R. T. & Råheim, A. 1972: Geochronological investigation of the gneisses and minor intrusive rocks from Kristiansund, west Norway. *Norsk geol. Tidsskr. 52,* 241–256.
Priem, H. N. A., Boelrijk, N. A. I. M., Hebeda, E. H., Verdurmen, E. A. Th. & Verschure, R. H. 1973: A note of the geochronology of the Hestbrepiggan Granite in west Jotunheimen. *Norges geol. Unders. 289,* 31–35.
Schmitt, H. H. 1963: *Petrology and structure of the Eiksundsdal eclogite complex, Hareidland, Sunnmøre, Norway.* Unpubl. Ph.D. thesis, Harvard Univ., 323 pp.
Scott, W. H. 1967: *Evolution of folds in the metamorphic rocks of western Dovrefjell, Norway.* Unpubl. Ph.D. thesis, Yale Univ.
Strand, T. 1949: On the gneisses from a part of the northwestern gneiss area of southern Norway. *Norges geol. Unders. 173,* 45 pp.
Strand, T. 1961: The Scandinavian Caledonides – a review. *Am. Jour. Sci. 259,* 161–172.
Strand, T. 1969: Geology of the Grotli area. *Norsk geol. Tidsskr. 49,* 341–360.
Turner, F. J. & Verhoogen, J. 1960: *Igneous and metamorphic petrology.* New York, McGraw-Hill, 694 pp.
Turner, F. J. & Weiss, L. E. 1963: *Structural analysis of metamorphic tectonites.* New York, McGraw-Hill, 545 pp.
Vogt, J. H. L. 1883: Olivinstenen i indre og Søndre Søndmøre. *Nyt Magasin Naturvidensk. 27,* 125–154.
Wheeler, R. L. 1973: *Folding history of the metamorphic rocks of eastern Dovrefjell, Norway.* Unpubl. Ph.D. thesis, Yale Univ., 231 pp.
Winchell, A. N. & Winchell, H. 1961: *Elements of optical mineralogy, part II.* New York, John Wiley & Sons, Inc. 551 pp.
Winkler, H. G. F. 1965: *Petrogenesis of metamorphic rocks.* New York, Springer, 220 pp.

STRATIGRAPHY

Vikvatn Sequence

Svartegga Group

S — Two-mica-quartz-plagioclase gneiss and schist containing lenses of hornblende amphibolite, quartzite, aluminum-rich plagioclase schist, and hornblende-biotite-quartz-plagioclase (± garnet) gneiss at base Sq-quartzite unit, Su-ultramafic body

Øvste Rødal Group

ØR — Predominantly well foliated biotite-quartz-plagioclase (± hornblende and garnet) gneiss and schist and foliated and non-foliated garnet-biotite-hornblende (and/or clinopyroxene)-microcline-plagioclase gneiss. Lesser amount of microcline-rich gneiss, hornblende schist, and amphibolite. ØRa-anorthosite, ØRu-ultramafic body.

Øyen Group

Ø — Predominantly biotite-quartz-plagioclase-microcline and biotite-quartz-plagioclase-(± hornblende) gneiss and schist. Heterogeneous assemblage of quartzite, two mica gneiss, calc-silicate rock, garnet-bearing plagioclase gneiss, and mafic rock at top and bottom of section. Øq-quartzite unit, Øa-anorthosite, Øe-eclogite, Øu-ultramafic body.

Fetvatn Gneiss

F — Strongly banded and weakly foliated, medium-grained, granitic to granodioritic biotite-quartz-plagioclase-microcline gneiss and schist. Zone of microcline-augen gneiss common. Some lenses of hornblende amphibolite.

STRUCTURE

Geologic Contact

- definite
- approximate
- covered
- predicted
- modified by faults

45 / 35 — strike and dip of foliation (perpendicular long and short lines), and strike and plunge of mineral lineation (arrow)

45 / 35 — strike and dip of F_{1A} fold axial surface schistosity and strike and plunge of F_{1A} fold axis

45 / 35 — strike and dip of F_{1B} fold axial surface schistosity, and strike and plunge of F_{1B} fold axis

60 — strike and plunge of major F_{2A} and F_{2A} fold axes

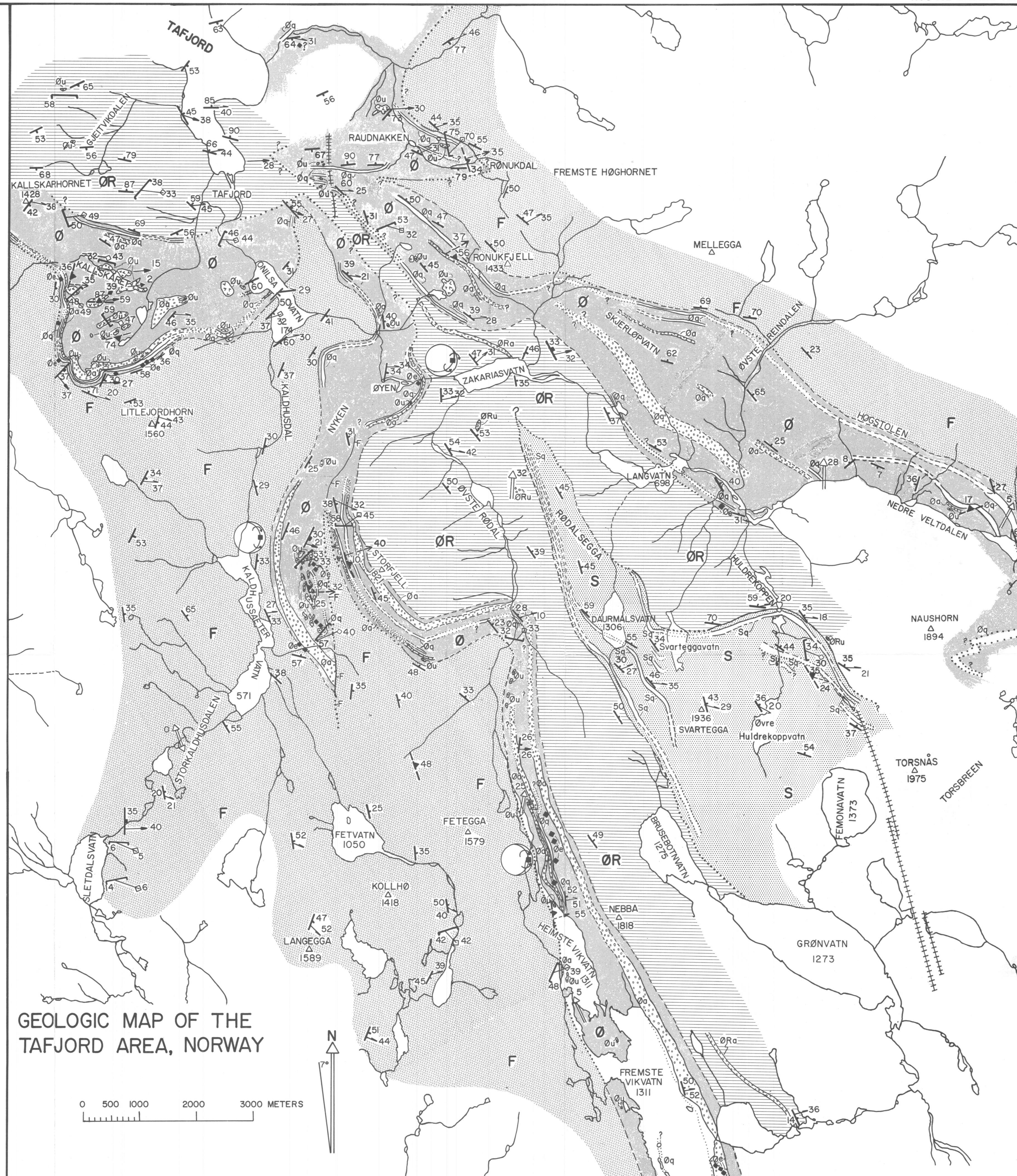
F_{2A} fold set (see text)

1. circle represents an equal area net
2. arcuate line within circle represents the F_{2A} great circle
3. arcuate arrows represent rotation sense of folds within the F_{2A} great circle
4. enlarged portion of the F_{2A} great circle separates rotation sense realms

45 / 35 — strike and dip of F_3 fold axial plane, and strike and plunge of F_3 fold axis

35 — strike and plunge of F_4 fold axis

trace of major subvertical fracture zone on land surface



GEOLOGIC MAP OF THE TAFJORD AREA, NORWAY