NORGES GEOLOGISKE UNDERSØKELSE NR. 266

ÅRBOK
1969

REDIGERT AV

MAGNE GUSTAVSON
STATSGEOLOG

OSLO 1970
UNIVERSITETSFORLAGET

INNHOLD – CONTENTS

ÉVOLUTION STRUCTURALE DE LA RÉGION NISSER–VRÅVATN
(Norvége méridionale)

THE PRECAMBRIAN ROCKS OF THE TELEMARK AREA IN SOUTH CENTRAL NORWAY VIII. STRUCTURAL DEVELOPMENT OF THE NISSER—VRÅVATN REGION

par
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Abstract.

Tectonic and petrographic studies of the Precambrian rocks in the Nisser—Vråvatn region of Telemark have led to the recognition of the following principal features:

— The basement of the «Telemark suite» does not appear in the studied region; it seems that all the migmatites are products of the granitization of the supracrustal rocks (acid lavas, basic lavas and quartzitic sediments).
— Three phases of folding have been demonstrated: B_1, B_2, B_3. The phase B_1 is synschistose and isoclinal.
— A later granitic remobilisation occurs in the shear zones of B_2. It is accompanied by the «mise en place» of molybdenum-rich pegmatites.
— The last folding B_3 disorientates former structures, in such a way that the lineations L_2 associated with B_2 are deformed following a plane «locus of L_2». The angle $L_2 \wedge L_3$ is variable.
— The tectonic axis «a» of the last folding (B_3) is vertical.
— The angle between the tectonic axis «a» of B_3 and the surfaces which were refolded was variable, which explains the different developments og B_2 and B_3.
— A consequence of the three phases of folding was the formation of domes and basins.
— In one of these basins, «the Vrådal basin», there was a «mise en place» of a microdioritic complex and of exogenous material filling the fractures around the basin.

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Résumé.

L'étude pétrographique et tectonique des roches précambriennes de la région Nisser—Vråvatn du Telemark nous a permis de reconnaître les faits suivants:

— Le soubassement de la suite du Telemark n'affleure pas dans la région étudiée; il semble que toutes les migmatites sont le résultat de la granitisation des roches supracrustales (laves acides, basiques et sédiments quartzitiques).
— Trois phases de plissement ont été mises en évidence: B_1, B_2 et B_3. La phase B_1 est synschisteuse et isoclinale.
— Une remobilisation granitique tardive apparait dans les zones de cisaillement du plissement B_2; elle est accompagnée par la mise en place de pegmatites riches en molybdène.
— Le dernier plissement B_3 désoriente toutes les structures antérieures, de telle façon que les linéations L_2 associées à B_2 sont déformées suivant un plan «lieu de L_2». L'angle $L_2 \wedge L_3$ est variable.
— L'axe tectonique «a» pendant le plissement B_3 est vertical.
— L'angle entre l'axe tectonique «a» pendant le plissement B_3 et les surfaces qui vont être replissées est variable, ce qui explique les différents développements des plis B_2 et B_3.
— La conséquence de ces trois plissements est la formation de structures en dômes et en bassins.
— Dans un de ces bassins «le Bassin de Vrådal» il y a mise en place d'un massif microdioritique et d'un matériel volcanique remplissant les fractures qui entourent le bassin.

Préambule.

Avant de commencer l'exposé des résultats de nos investigations, nous voulons donner, ci-dessous, la définition des termes géologiques utilisés dans le texte et signaler sur une carte la localisation des affleurements étudiés (carte des affleurements, Fig. 14).

Définition des termes géologiques utilisés dans le texte:

Abréviations:

S_f — stratification des roches supracrustales.
S_1 — schistosité de flux, plan axial des plis synschisteux (B_1).
S_2 — schistosité associée au plissement postschisteux (B_2).
B_1 — plissement synschisteux; linéations associées (L_1).
B_2 — plissement tardif (première génération); linéations associées (L_2).
B_3 — plissement tardif (deuxième génération); linéations associées (L_3).
G_1, G_2, G_3, G_4 et G_5 — générations des pegmatites.

Antiforme (Wilson, G. 1961):
Structure plissée ayant la forme géométrique d'un anticlinal, mais pour laquelle l'âge relatif des roches est inconnu. Cette structure présente une fermeture vers le haut.

Axe de pli (Wegmann, C. E. 1928):
L'axe d'un pli ou d'une surface «S» plissée est une ligne qui, déplacée parallèlement à elle-même dans l'espace, engendre un pli.

Axe β d'un pli:
C'est l'axe défini statistiquement; il est donné par le pôle de la zone qui contient les projections stéréographiques des pôles de «S».

Axe tectonique «a» (McIntyre, D. B. 1956):
Axe parallèle à la direction du mouvement pendant le plissement.

Axe tectonique «b» (McIntyre, D. B. 1956):
Axe de rotation.

Axe tectonique «c» (McIntyre, D. B. 1956):
Axe perpendiculaire au plan de mouvement «ab».

Déformations syngénétiques:
Déformations non tectoniques produites au cours de la sédimentation.

Diagrammes π S:
Diagrammes construits avec les projections des pôles de «S».

Enclaves de minéraux (Wegmann, C. E. 1935):
Catégorie d'enclaves mise en évidence pour certaines migmatites; ce sont des minéraux: zircon, tourmaline, etc., qui proviennent des roches anté-migmatitiques ayant résisté à la granitisation.

Enclaves phasmatoclasiques (Hupé, J. 1951):
Ces enclaves représentent des vestiges de diaclases, de fractures ou de surfaces de laminage développées au sein du granite et reprises par une granitisation postérieure.

Fissures:
Terme général, où sont englobées toutes les fractures sans considérer l'action de déplacement.

Fissures soudées (Wegmann, C. E. 1959):
Les fisures soudées sont des diaclases, des failles, ou d'anciennes zones mylonitiques qui ont cessé de jouer un rôle au point de vue de l'anisotropie.

Foliation:
La foliation se rapporte soit à des structures primaires (type syngénétique) soit à des structures secondaires (type tectonique). La foliation S est, dans la région Nisser—Vråvatn, le résultat de trois surfaces structurales: stratification, schistosité et litage.

Granitisation (Read, H. 1956):
Processus ou ensemble de processus pour lesquels des roches solides se transforment en roches à caractère granitique sans passer par une phase magmatique.

Linéation (McIntyre, D. B. 1956):
Terme non génétique qui exprime différents types de structures linéaires d'une roche:
— l'axe d'un pli mésoscopique ou macroscopique,
— l'intersection de deux schistosités,
— l'axe des lentilles de quartz d'exsudation,
— l'intersection d'une schistosité et de la stratification,
— l'alignement des minéraux,
— l'étirement corpusculaire.

Métamorphisme (Dietrich, R. V. & Mehnert, K. 1961):
Réajustement minéralogique et / ou structural de roches solides, sans passage par

une phase liquide, à changements de température et / ou de pression et / ou de chimisme. Les réajustements par altération supragénétique et par cimentation sont exclus.

Métasomatose (Dietrich, R. V. & Mehnert, K. 1961):
Remplacement d'un ou de plusieurs éléments par d'autres d'origine étrangère qui ont migré en solution ou sous la forme d'ions libres.

Migmatite (Wegmann, C. E. 1935):
Groupe génétique de roches, dans lequel sont englobés tous les gneiss même homogènes, ayant acquis un faciès feldspathique soit par injection d'un magma soit par métasomatose soit par ultramétamorphisme.

Migmatites hétérogènes:
Migmatites pour lesquelles la roche mère et la roche nouvelle sont macroscopiquement bien visibles.

Migmatites homogènes:
Migmatites pour lesquelles la roche mère et la roche nouvelle sont macroscopiquement difficiles à séparer.

Ordres de grandeur (Weiss, L. E. 1957):
Microscopique—champ qui peut être examiné sous le microscope; champ d'une coupe mince.
Mésoscopique—embrasse tout ce qui peut être mesuré avec certitude et continuité, depuis l'échantillon jusqu'à l'affleurement où une continuité peut être suivie.
Macroscopique—embrasse des champs de quelque taille où une continuité des structures ne peut pas être suivie.

Plis synschisteux:
Les plis synschisteux correspondent, dans ce travail, aux plis de clivage de De Sitter.

Plis tardifs:
Les plis tardifs sont considérés comme synonymes de plis postschisteux ou tardi-cinématiques.

Roches granitiques (Dietrich, R. V. & Mehnert, K. 1961):
Terme général comprenant toutes les roches à texture phanéritique et constituées par du quartz et des feldspaths.

Roches supracrustales:
Roches formées à la surface de l'écorce terrestre. Elles peuvent être soit sédimentaires, soit volcaniques.

Synforme (Wilson, G. 1961):
Structure plissée ayant le forme géométrique d'un synclinal mais pour laquelle l'âge rélatif des roches est inconnu. Cette structure présente une fermeture vers le bas.

Introduction.

Il existe deux grandes formations de terrains précambriens en Norvège. L'une, au nord (dans la province du Finnmark), l'autre, au sud (entre la côte et la limite septentrionale de la province du Telemark). Ces terrains ont subi des déformations à différents niveaux structuraux et présentent une tectonique superposée. Dans la formation précambrienne méridionale (Fig. 1), il est possible de considérer trois blocs (Barth, T. F. W. 1960) séparés par des zones de fractures: la région du Telemark, la région de Bamble et la région de Kongsberg. Ces blocs ont joué comme des unités plus ou moins indépendantes et de ce fait, ils présentent des histoires géologiques partiellement différentes.

Ce travail expose les résultats de l'analyse d'une partie de la région du Telemark, en particulier le secteur Nisser–Vråvatn, situé dans la zone centre-Est de cette région (Planche III). Ce secteur d'une superficie

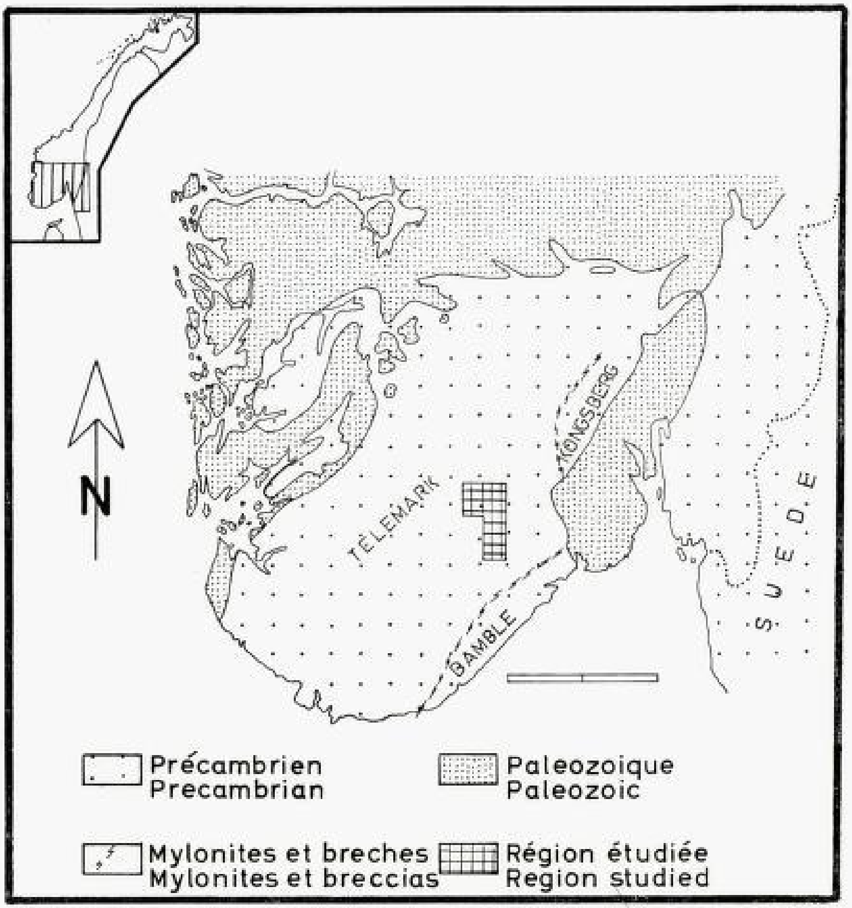

Fig. 1. Précambrien, Sud de la Norvège.
The Precambrian in South Norway.

voisine de 600 Km² est principalement constitué par des migmatites et par des laves (acides et basiques) auxquelles des sédiments quartzitiques (quartzites et conglomérats) sont associés. Dans la partie NE, près du village de Vrådal (Planche I) existe une structure circulaire, où affleurent des roches granitiques, des roches supracrustales et des roches intrusives basiques. L'ensemble des roches supracrustales (laves et sédiments quartzitiques) constitue, semble-t-il, les reliques plus ou moins granitisés de la suite du Telemark, dont d'épaisseur aurait été, d'après Dons (1960), d'environ, 4.000 mètres.

Dans ce résumé, nous cherchons à décrire la suite des événements géologiques dont nous avons trouvé les traces dans les affleurements et les roches étudiées; les rélations chronologiques seront présentées par rapport aux déformations tectoniques, et aux transformations pétrographiques.

Histoire geologique du secteur Nisser—Vråvatn.

Soubassement de la suite du Telemark.

Nos investigations nous ont conduit à conclure que le complexe anté-supracrustal, soubassement de la suite du Telemark, n'affleure pas dans le secteur Nisser—Vråvatn. Ce soubassement formées dans les zones inférieures de l'écorce terrestre (domaines de hautes températures et hautes pressions) serait remonté vers des niveaux supérieurs en subissant une intense fracturation. Profitant de cette fissuration, on assiste alors

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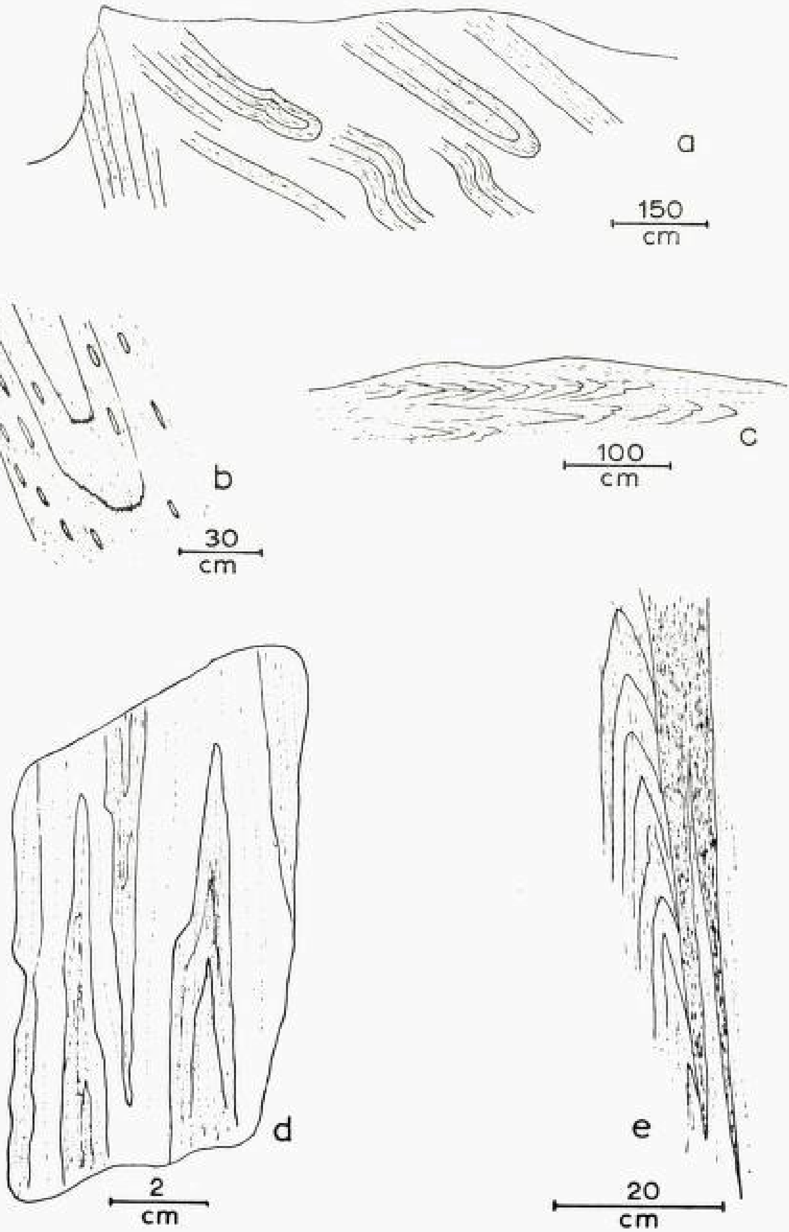
Fig. 2. Pegmatites de la génération G_1.

a) Discordante de la stratification et déformée ptygmatiquement. Nissedal, affleurement 11, Fig. 14.
b) Peu déformée quand parallèle à la stratification, mais avec déformation accentuée quand oblique. Nissedal, affleurement 12, Fig. 14.
c) Plis des veines de quartz d'exsudation dans une lave basique, admettant comme plan axial la schistosité régionale. Nordaviki, affleurement 2, Fig. 14.

Pegmatites of generation G_1

a) Cross-cutting the stratification and ptygmatically deformed. Nissedal, exposure 11, Fig. 14.
b) Slightly deformed when parallel to the stratification, more deformed when oblique. Nissedal, exposure 12, Fig. 14.
c) Folds of exudation quartz veins in basic supracrustal rocks, having as axial plane the regional schistosity. Nordaviki, exposure 2, Fig. 14.


a
b
c


a
150
cm
b
30
cm
c
100
cm
d
2
cm
e
20
cm

à la mise en place des roches volcaniques de la suite du Telemark. Entre certains épisodes volcaniques, quelques niveaux sédimentaires se sont déposés; l'un d'eux est particulièrement riche en fer, et a donné naissance à l'exploitation de Söftestad.

Dépôt de la suite du Telemark.

Le dépôt des roches supracrustales est marqué par des plans de stratification (S) bien visibles. Dans certaines de ces roches, la stratification est soulignée par les alternances des laves de composition chimique différente et par des alternances de roche volcanique et de roche sédimentaire. Les structures syngénétiques (ripple-marks, slumping, figures de compaction, stratifications entrecroisées), très abondantes dans la suite du Telemark, surtout dans les quartzites, caractérisent un milieu de sédimentation agité. La disposition des stratifications entrecroisées dans les quartzites de Nordaviki, nous a conduit à envisager l'existence d'un plissement isoclinal.

Mise en place des pegmatites G_1

Après le dépôt de la suite du Telemark des quartzo-pegmatites et des veines de quartz d'exsudation se sont mises en place. Ces quartzo-

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Fig. 3. Plis synschisteux isoclinaux dans les roches supracrustales.

a) Affleurement de Vrå (1, Fig. 14). Surface verticale, plis synschisteux isoclinaux et plis tardifs dans une lave acide.
b) Affleurement de Nordaviki (2.1, Rig. 14). Plis synschisteux dans un niveau conglomératique.
c) Affleurement de Buskardnt. (3, Fig. 14). Plis synschisteux dans une lave basique.
d) Affleurement de Ånevn. (4, Fig. 14). Plis isoclinaux dans une lave acide (échantillon).
e) Affleurement de Ånevn. (4, Fig. 14). Alternance de lave acide et basique plissées et aplaties.

Synschistose isoclinal folds (Cleavage folds of de Sitter) in supracrustal rocks.

a) Exposure at Vrå (1, Fig. 14). Vertical wall. Synschistose isoclinal folds B_1 and later folds in an acid lava.
b) Exposure at Nordaviki (2.1, Fig. 14). Synschistose folds in a conglomerate bed.
c) Exposure at Buskardnt. (3, Fig. 14). Synschistose folds in basic lava.
d) Exposure at Ånevn. (4, Fig. 14). Isoclinal fold B_1 in a specimen of basic lava.
e) Exposure at Ånevn. (4, Fig. 14). Alternating acid and basic lavas, folded and flattened.

pegmatites sont soit concordantes, soit discordantes à la stratification. Dans le premier cas, elles se présentent actuellement encore comme peu déformées; par contre, là où elles étaient obliques à la stratification, elles montrent actuellement des plis dont la géométrie peut aller jusqu'aux veines ptygmatiques (Fig. 2). Ces quartzo-pegmatites, les plus anciennes de toutes, forment une génération antérieure au plissement et au métamorphisme; nous l'appelerons génération G_1. Les veines de quartz d'exsudation de même âge, qui annoncent le métamorphisme (Fig. 2c) sont très abondantes dans les laves de composition basique et y sont presque toujours plissées; le plan axial de ces plis est parallèle à la schistosité régionale des roches supracrustales.

Plissement synschisteux B_1

Dans la chronologie des déformations l'événement le plus important qui a pu être mis en évidence est le plissement isoclinal de la suite du Telemark.

Il s'observe à tous les ordres de grandeur. Ainsi, macroscopiquement, à l'échelle de la carte géologique (Planche I) il correspond aux alternances des quartzites et de laves que l'on voit dans la partie N de la carte géologique. A l'échelle de l'affleurement (ordre de grandeur mésoscopique), plusieurs localités peuvent être considérées comme caractéristiques:

- L'affleurement de Vrå (Fig. 3a) où les laves acides montrent une schistosité axiale S_1 associée à des plis isoclinaux.
- L'affleurement de Buskardnt. (Fig. 3c) où des plis isoclinaux sont visibles dans les laves basiques.
- L'affleurement de Nordaviki (Fig. 3b) où les niveaux conglomératiques plissés isoclinalement montrent une schistosité axiale S_1 et un allongement très marqué des galets de quartzite.
- L'affleurement de Ånevn (Fig. 3e) où l'on peut voir des plis complètement étirés et aplatis, donnant une série d'alternances de matériel volcanique acide et basique.

L'existence de ce plissement se confirme également à l'échelle de l'échantillon (Fig. 3d); ainsi dans des laves acides il a été possible de constater le plissement isoclinal d'un niveau de vermiculite (échantillon C—6/64).

Ce plissement de style isoclinal (plissement B_1) est accompagné d'une schistosité axiale S_1, qui donne une linéation d'intersection L_1 parallèle à l'axe des plis synschisteux lorsque S_1 est oblique à la stratification (charnière des plis); le régime tangentiel intense a cependant fait disparaître la plupart des charnières de plis.

Sous l'action de ce plissement les roches supracrustales ont été réorientées, elles se sont déformées sans rupture, et les phyllites se sont disposés d'une façon plus ou moins planaire. Les quartzo-pegmatites (G_1) et les veines de quartz d'exsudation, mises en place avant le plissement, se sont trouvées déformées lorsqu'elles étaient discordantes à la stratification ou à la schistosité généralement confondues.

Granitisation générale de la suite du Telemark.

Simultanément au plissement B_1, les roches de la suite du Telemark commencent à prendre un faciès granitique: elles acquièrent une anisotropie marquée par l'apparition d'une foliation. Les plans structuraux (stratification et schistosité) qui comme nous l'avons vu sont presque toujours parallèles, facilitent la mise en place du matériel granitique. A la suite de ces phénomènes, les roches supracrustales granitisées montrent deux types principaux d'anisotropie que l'on peut reconnaître sur les échantillons.

Type A — caractérisé par trois faces non équivalentes:

1 — la face S_1, ou plan de schistosité, matérialisé par la disposition des minéraux phylliteux où il est possible de voir une linéation L_1.

2 — la face perpendiculaire à S_1 et parallèle à L_1, qui montre un fort allongement des yeux de feldspaths.

3 — la face perpendiculaire à S_1 et à L_1 où l'orientation des feldspaths est peu marquée.

Type B — caractérisé par deux faces équivalentes:

1 — la face S_1, ou plan de schistosité, marqué comme précédement par la disposition des phyllites.

2 — les faces perpendiculaires à S_1 qui sont caractérisées par une alternance assez régulière de lits quartzo-feldspathiques et de lits de minéraux phylliteux ferro-magnésiens.

L'anisotropie des roches qui s'exprime par la foliation des migmatites ne peut être bien expliquée qu'en admettant une origine mixte: une recristallisation mécanique et une recristallisation mimétique de la stratification.

Granitisation et plissement synschisteux sont localement des phénomènes synchrones. En effet il est possible de trouver dans les migmatites des plis isoclinaux qui admettent la foliation comme plan axial.

Bilan chimique lors de la granitisation.

L'étude microscopique de plusieurs échantillons nous a permis de saisir les échanges chimiques qui ont accompagné l'introduction du matériel granitique. Celui-ci permet la transformation des roches supracrustales, tout particulièrement celles de composition basique, en roches granitiques. Les résultats peuvent être présentés de la façon suivante:

1 — Une métasomatose potassique est responsable de la décalcification du plagioclase, le potassium remplace le calcium de l'anorthite et permet la cristallisation d'un plagioclase plus albitique (Härme, M. 1958):

$$3(CaO \cdot Al_2O_3 \cdot 2SiO_2) + K_2O \longrightarrow K_2O \cdot Al_2O_3 \cdot 6SiO_2 + 2Al_2O_3 + 3CaO$$

2 — L'augite que nous trouvons encore sous la forme de reliques dans les migmatites se transforme en amphibole qui conserve parfois la morphologie des pyroxènes:

$$8[(Fe,Mg)O \cdot SiO_2] + H_2O \longrightarrow 7(Fe,Mg)O \cdot 8SiO_2 \cdot H_2O + (Fe,Mg)O$$

En utilisant l'alumine disponible, la métasomatose potassique provoque également la transformation de l'amphibole en biotite qui se solde par la mise en liberté de (Fe,Mg)O:

$$7(Fe,Mg)O \cdot 8SiO_2 \cdot H_2O + K_2O + Al_20_3 + H_2O \longrightarrow (Fe,Mg)O + 2SiO_2 + K_2O \cdot 6(Fe,Mg)O \cdot Al_2O_3 \cdot 6SIO_2 \cdot 2H_2O$$

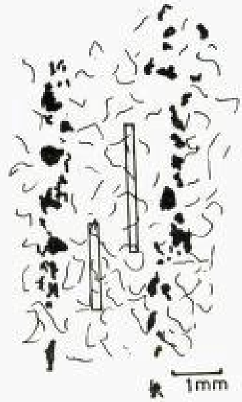
3 — La formation de l'épidote apparaît également liée à la métasomatose potassique. Elle se développe soit à partir de l'amphibole, soit à partir du plagioclase. D'autre part, l'épidote apparaît comme un sous-produit de la chloritisation de la biotite, phénomène qui s'observe dans toutes les coupes minces.

4 — La métasomatose potassique est accompagnée d'une métasomatose siliceuse. Celle-ci est mise en évidence par les myrmékites qui apparaîssent en grand nombre dans les roches granitisées. La structure des myrmékites peut nous renseigner sur la nature de la métasomatose siliceuse. A cet égard nous devons signaler que les gouttes de quartz vermiculaire ne traversent jamais les plans de macle des feldspaths. La transformation de l'amphibole en biotite, la réaction entre le microcline et le plagioclase, ou le remplacement des feldspaths alcalins par du calcium peuvent représenter, en partie, la source de la silice.

Les équilibres chimiqes montrent que les minéraux mafites sont presque totalement remobilisés au cours de la granitisation avec mise en liberté de CaO, de Fe et de Mg. Ces éléments, ainsi mis en liberté ont, dans certains cas, contaminé les solutions granitisantes. Ils ont contribué à la cristallisation de l'hématite, de la magnétite et du sphène dans les pegmatites qui accompagnent et qui, en grande partie, produisent la granitisation des roches supracrustales (Fig. 13). Ces pegmatites se mettent en place parallèlement à la stratification et à la schistosité; leur mise en évidence est difficile: elles passent d'une façon imperceptible aux roches encaissantes. Cependant, sur le terrain, la présence d'enclaves du paléosoma permet souvent de les déceler. Ces pegmatites constituent la génération G_2.

Enclaves de minéraux.

Bien que la plupart des minéraux ferro-magnésiens soit digérés par les ichors, certains résistent à tous les phénomènes de la granitisation et donnent naissance à un type d'enclaves caractéristiques. Dans la coupe mince C-48/64 (Fig. 4), par exemple, la roche formée par des alternances de lits quartzo-feldspathiques et de lits ferro-magnésiens, montre que les minéraux accessoires ne sont pas exclusivement liés aux lits mélanocrates, mais qu'ils se trouvent aussi dans les lits quartzo-feldspathiques, où ils sont disposés, d'une façon plus ou moins régulière. Il s'agit principalement de zircon, de sphène et d'apatite, provenant directement des roches antémigmatitiques; ces minéraux ont résisté aux phénomènes de la granitisation. Ils forment des «enclaves de minéraux» que Wegmann signalait déjà en 1935.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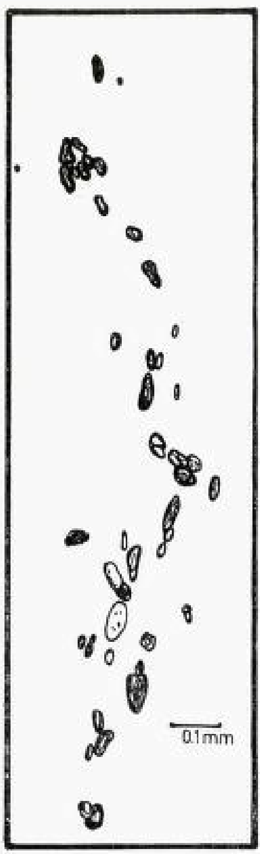
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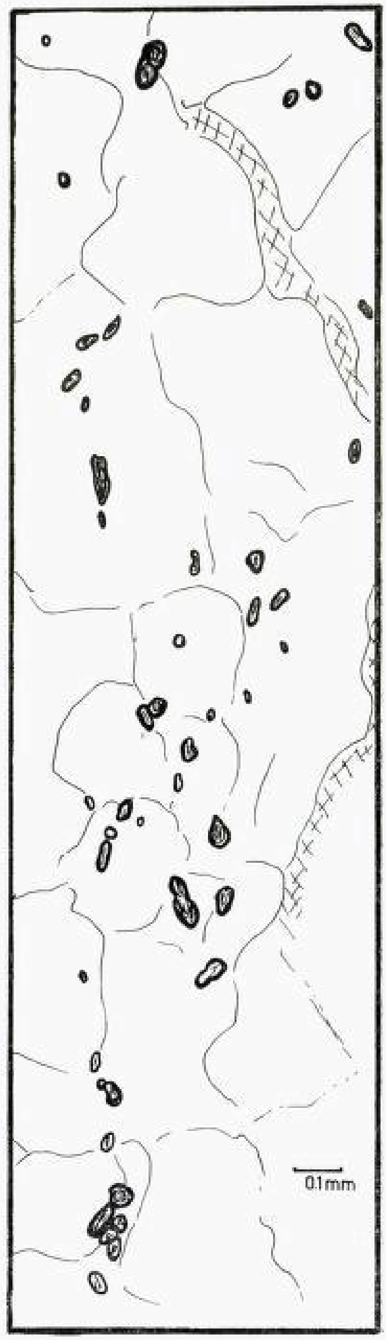
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Conséquences de la granitisation.

Dans le secteur étudié le soubassement ne semble pas affleurer. Il est possible qu'une des principales conséquences de la granitisation des roches supracrustales a conduit à l'éffacement de la limite entre soubassement et la base de la suite du Telemark. A titre d'hypothèse, on est conduit à suggérer que toutes les migmatites résultent de la transformation des roches supracrustales de la suite du Telemark, car:

a) On ne peut mettre en évidence aucune discordance entre les migmatites et les roches supracrustales.
b) Dans l'ensemble de la région il est toujours possible de reconnaître, dans les migmatites, des reliques de roche supracrustales, particulièrement des reliques de composition basique.
c) La morphologie des cristaux de zircon des migmatites, sauf dans quelques cas particuliers, est identique à celle des cristaux de zircon des laves supracrustales. L'une des exceptions est constituée par d'anciens niveaux quartzitiques granitisés dans lesquels la morphologie des zircons est caractéristique des roches sédimentaires. Cette étude des zircons a été entreprise sur la base de 14 échantillans de migmatites. L'analyse des cristaux de zircon s'est faite statistiquement sur la base de 300 déterminations pour chaque échantillon.

En récapitulant les événements géologiques considérés jusqu'à présent, nous avons:

A) Formation et évolution d'un complexe anté-supracrustal, qui n'affleure pas dans la région.
B) Dépot des roches supracrustales et mise en place de la première génération de pegmatites, accompagnée de quartz d'exsudation.
C) Plissement synschisteux et métamorphisme.
D) Granitisation avec mise en place de la deuxième génération de quartzo-pegmatites.

Fig. 4. Enclaves de minéraux.

Des minéraux accessoires (Zircon, Apatite, Sphène) provenant directement des roches anté-migmatitiques et ayant résistés à la granitisation.

«Enclaves de mineraux» is a term introduced by C. E. Wegmann 1935, during his migmatite studies. It refers to accessory minerals (zircon, apatite, sphene) which have survived the granitization and are directly derived from the original pre-migmatite rock.

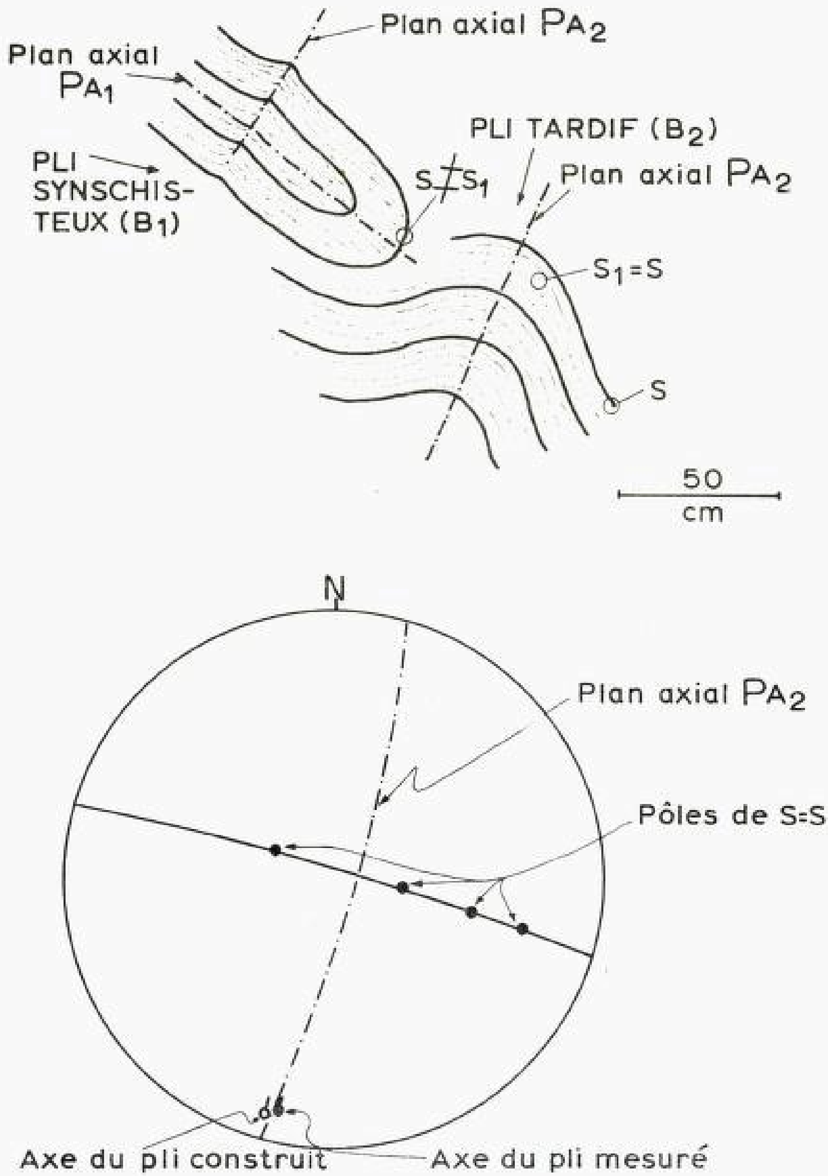


Fig. 5. Détail de l'affleurement de Vrå (voir Fig. 3a).
Influence du plissement tardif B_2 sur le plissement synschisteux B_1 démontrée par la désorientation du plan axial P.A. 1.

Detail of the exposure at Vrå (see Fig. 3a).
Influence of the later folding B_2 (with axial plane P.A. 2) on the synschistose folding B_1 (with axial plane P.A. 1) shown by disorientation of P.A. 1.

Plissement B_2.

Dans la chronologie des événements géologiques, après la granitisation générale de la suite du Telemark, il est possible de mettre en évidence un nouveau plissement, B_2. Celui-ci a une direction moyenne N–S. Il désoriente et en même temps efface les structures synschisteuses. Ce phénomène s'observe à différents ordres de grandeurs:

a) A l'échelle de la carte géologique (schéma tectonique, Planche I) la désorientation des structures synschisteuses est évidente. Le synforme de la mine de Søftestad (plis B_1) est brusquement désorienté par un antiforme du plissement B_2 de telle façon que sa direction axiale change de NE–SW à NW–SE.
b) A l'échelle de l'affleurement, la figure 3a (affleurement de Vrå) nous offre une désorientation du plan axial des plis synschisteux par des plis B_2. Un détail de cet affleurement (Fig. 5) montre mieux cette désorientation: le plan axial des plis B_1 est replissé; d'autre part, la projection stéréographique des éléments structuraux de l'affleurement montre que la géométrie des plis B_2 est différente de celle des plis B_1. Le plan axial des plis B_2 est proche de la verticale; l'inclinaison des flancs est faible et la schistosité de fracture S_2 est mal marquée.

Schistosité S_2.

La schistosité S_2 qui accompagne le plissement B_2 n'apparaît avec netteté que dans les lits de roche supracrustale basique. Dans les lits volcaniques acides, elle est très mal marquée. L'affleurement de Sinnes (Fig. 6), qui est constitué par des alternances de lave acide et basique à plans de stratification assez redressés, montre dans les lits basiques un plissotement de S_1 qui est accompagné d'une schistosité S_2; une linéation d'intersection L_2 se forme sur les plans de schistosité ($S_1 = S$). La projection stéréographique des éléments structuraux de l'affleurement montre que les linéations L_2 (axes des plissotements mésurés) sont parallèles à l'intersection des plans de schistosité S_1 et S_2 (voir diagramme, Fig. 5).

Mise en place des pegmatites G_3.

Le plissement B_2 est accompagné d'une remobilisation granitique et de la mise en place de pegmatites riches en molybdène. Celles-ci forment chronologiquement la génération G_3. Elles occupent des positions bien

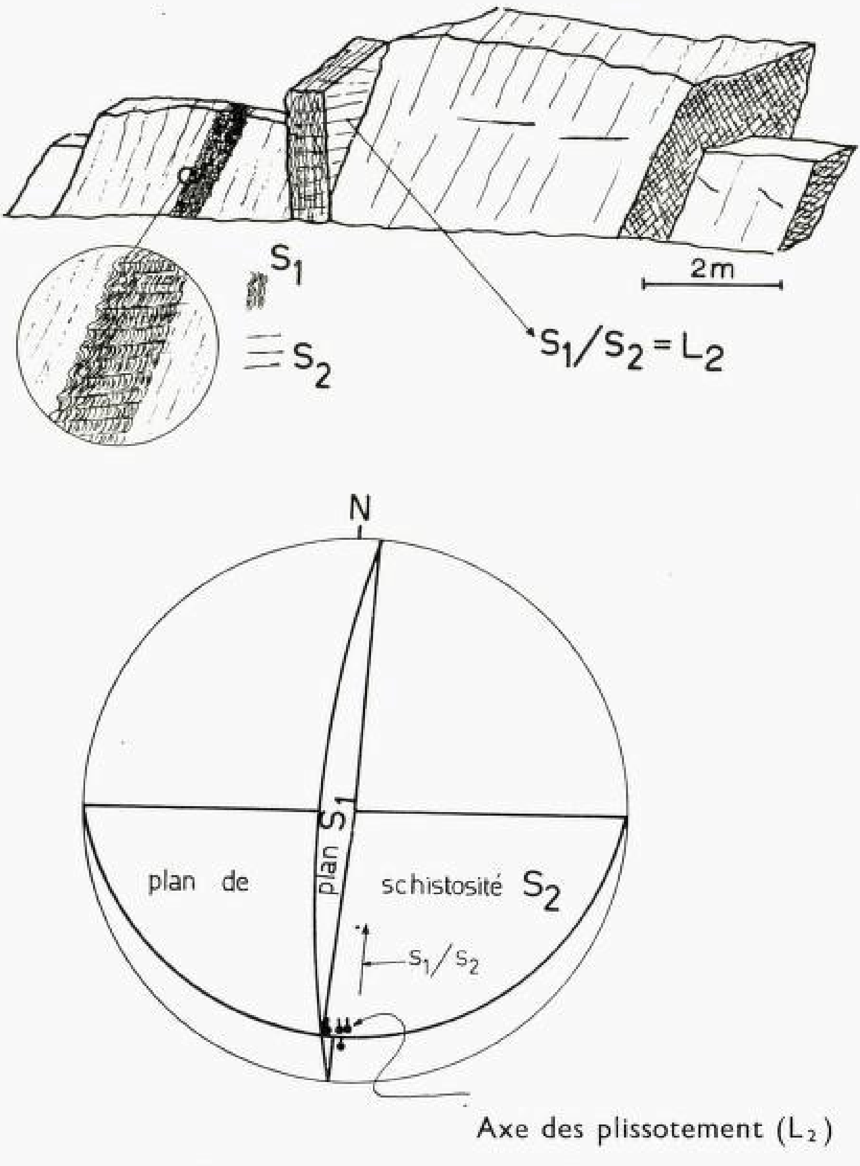


Fig. 6. Affleurement de Sinnes (5, Fig. 14).
Schistosité de fracture S_2. Les axes des plissotements de la schistosité S_1 sont parallèles à l'intersection $S_1/S_2 = L_2$.
Exposure at Sinnes (5, Fig. 14).
Fracture schistosity cleavage S_2. The axes of small folding shown by the schistosity S_1 are parallel to the intersection S_1/S_2 e.g. L_2.

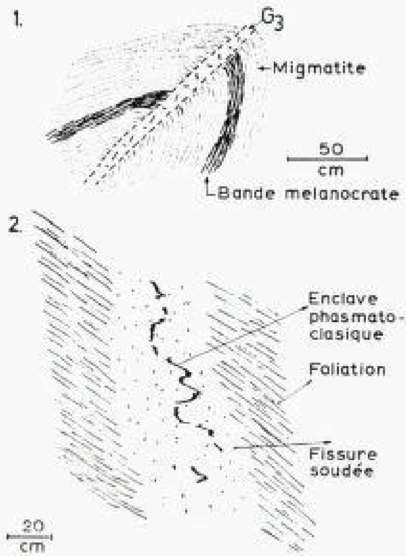


Fig. 7. Remobilisation granitique associée à des zones de cisaillement (Fig. 7₁) et à d'anciennes fissures (Fig. 7₂). Affleurements 6 et 9 de la Fig. 14.

Granitic remobilisation connected with shear zones (Fig. 7₁) and former fissures (Fig. 7₂). Exposures 6 and 9 of Fig. 14.

Migmatite.
Basic band.
Enclave phasmatoclasique (defined by Hupé, 1951) = a former plane now composed of micaceous minerals.
Fissure soudée (defined by Wegmann, 1959) = healed fissure.

déterminées; elles se sont mises en place de préférence dans les zones de cisaillement des plis B2 (Fig. 7, 1), ou dans des fissures formées après la granitisation générale (Fig. 7, 2). Ces dernières qui sont d'anciennes diaclases furent soudées par ces pegmatites; on y voit des enclaves phasmatoclasiques (Hupé, J. 1951) constituées par des alignements de minéraux micacés (chloritisés) plus ou moins plissés.

A la fin du plissement B2, le secteur Nisser—Vråvatn devrait être constitué par des plis à flancs peu redressés dans la partie E, tandis que des plis à flancs proches de la verticale constituaient la partie W (particulièrement près de Nordaviki affleurement 2, Fig. 14). Nous verrons plus tard les raisons de cette affirmation.

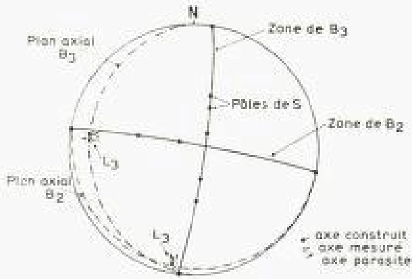


Fig. 8. Affleurement de Holman (10, Fig. 14).
Plis tardifs de la génération B_2 et B_3.

Road section at Holman (10, Fig. 14).
Later folds of the generations B_2 and B_3.

Plissement B_3.

Par la suite le secteur subit un nouveau plissement (plissement B_3). Les plis antérieurs (plis B_1 et B_2) sont désorientés et les linéations rectilignes sont tordues d'une façon caractéristique. La désorientation des plis B_2 par la nouvelle génération des plis B_3 se constate à différents endroits, mais l'affleurement de Holmann (Fig. 8 et 9) est certainement le plus caractéristique. Il n'est malheureusement plus dans sa position originale; placé près d'une importante zone de fracture tardive, il a subit en bloc une désorientation qui n'a fort heureusement pas perturbé les rélations spatiales qui y existent entre les différentes parties. Cet affleurement est constitué par des laves acides à plans de stratification bien marqués renforcés par des alternances d'associations minérales différentes. Parallèlement à la stratification existe une schistosité S_1.

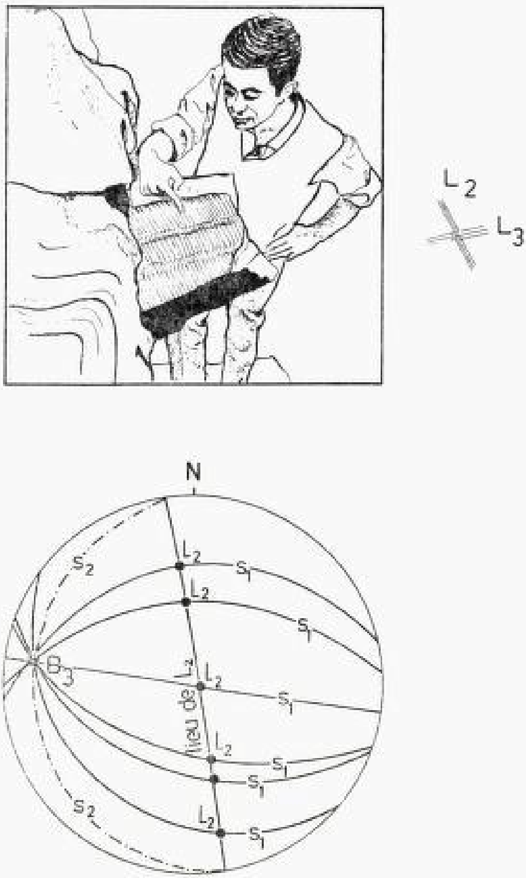


Fig. 9. Détail de l'affleurement de Holman.
Les linéations L_2 sont désorientées suivant le «lieu de L_2». L'angle $L_2 \wedge L_3$ est variable.

Detail of exposure at Holman.
The lineations L_2 are disorientated following «locus L_2». The angle $L_2 \wedge L_3$ is variable.

Des plis postschisteux (plis B_2 et B_3) à géométrie différente sont également bien visibles: les plis très couchés B_3 présentent des axes E—W, parallèles aux linéations L_3 (axes des plis mésoscopiques, voir diagramme stéréographique Fig. 8); les plis B_2 ont une direction axiale N—S, de même que les linéations associées L_2 (axes des plis mésoscopiques et alignement de minéraux) sont désorientés par les plis B_3. Les axes des plis B_3 sont presque contenus dans les plans axiaux des plis B_2.

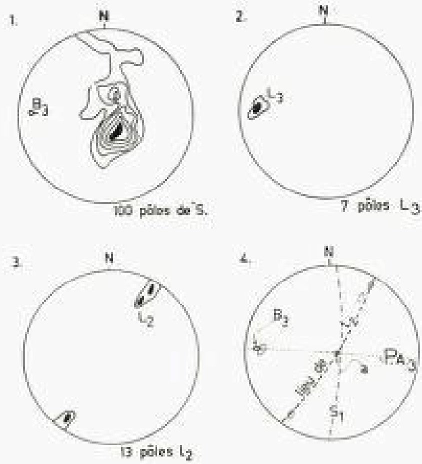


Fig. 10. Géométrie de l'antiforme de l'Elibu (Fig. 14).
The geometry of the Elibu antiform (Fig. 14).

Distorsion des linéations L_2 par le plissement B_3*)

L'observation attentive du flanc inférieur du pli B_3 de la figure 9 permet de reconnaître l'interférence du plissement B_3 sur le plissement B_2. Les linéations L_2 données dans la figure par la direction du doigt sont désorientées par les linéations L_3 (axes des plis mésoscopiques) associées au dernier plissement; cette désorientation n'est pas faite au hasard; sur la projection stéréographique (Fig. 9) nous pouvons constater que l'angle entre L_2 et L_3 est variable et que les pôles de L_2 tombent sur une zone dite «lieu de L_2». Les linéations L_2 sont désorientées par les linéations L_3, mais de telle façon qu'elles se trouvent toujours sur un plan.

La désorientation rencontrée se trouve non seulement dans les structures mésoscopiques, mais aussi dans les structures macroscopiques. L'antiforme de Elibu (Fig. 10) est un pli B_3; avec 100 mesures de $S = S_1$, nous obtenons un axe β de pli de direction E—W ayant une inclinaison axiale assez faible. Les projections stéréographiques des pôles des axes des plis mésoscopiques (L_3) (Fig. 10, 2) se concentrent autour de la projection de l'axe construit β, ce qui nous permet de dire que ces linéations L_3 sont associées au plissement qui a produit l'antiforme de Elibu. Par contre, les projections stéréographiques des pôles des axes des

*) Nous faisons abstraction du plissement isoclinal synschisteux B_1 dans l'étude des structures d'interférence, car son influence sur ces structures est très faible.

plis mésoscopiques B_2 tombent sur un grand cercle de la projection (voir Fig. 10, 3). Sur le diagramme synoptique de l'antiforme de Elibu (Fig. 10, 4) on constate que les linéations L_2 sont désorientées par le plissement B_3, mais qu'elles sont toujours contenus dans un plan; on voit également que le plan axial des plis B_3 est vertical et que l'angle entre L_2 et L_3 est variable. Un examen plus attentif du diagramme synoptique permet même de déterminer la position spatiale de l'axe de mouvement pendant le dernier plissement (axe tectonique «a»); il doit être contenu dans le plan axial des plis B_3 et dans le «lieu de L_2». L'interférence de ces deux plans (Ramsay, J. G. 1960) donne la position de l'axe tectonique «a» pendant le plissement B_3; il est ici proche de la verticale.

Interférence des plissement B_2 et B_3.

Ces deux plissements présentent les caractères suivants:

1 — directions axiales perpendiculaires,
2 — l'axe tectonique «a» pendant le plissement B_3 est vertical,
3 — l'axe tectonique «a» du plissement B_3 est à peu près contenu dans les plans axiaux des plis B_2.

D'après Ramsay (1960) cette situation conduit à des interférences donnant des structures en dômes et en bassins. Celles-ci ont été retrouvées sur le terrain et sont représentées sur le schéma tectonique de la Planche I. Nous constatons cependant que dans la partie E du lac Nisser les structures d'interférence sont très allongées suivant la direction E—W, ce qui signifie que les axes dominants sont les axes B_3. Par contre, dans la partie W du lac les mêmes structures s'allongent de préférence suivant une direction N—S, ce qui nous conduit à reconnaître que les axes B_2 y sont dominants. Ce fait résulte (Ramsay, J. G. 1960) de l'angle qui a existé entre l'axe tectonique «a» et la surface qui est replissée pendant le dernier plissement. Nous devons admettre que les plis B_2 dans la partie W de la région du Nisser avaient des flancs plus redressés que ceux de la partie E, car là, l'axe tectonique «a» du plissement B_3 s'est toujours maintenu vertical. Ceci est particulièrement vrai pour la région de Nordaviki (extrêmité NW de la carte géologique): les plis B_2 possèdent des flancs très inclinés qui ne subissent par conséquence l'influence du plissement B_3 — l'angle entre l'axe tectonique «a» et les surfaces de stratification y étant presque voisin de zéro —. Dans ce cas, il y a uniquement une déformation des linéations sans avoir formation des plis, à part les zones de charnières conservées.

Dans le secteur Nisser—Vråvatn, le plissement B_3 déforme donc les structures antérieures et y produit une série irrégulière de structures en dômes et en bassins qui s'accompagne, dans les dômes par la mise en place de granites diapiriques. La structure diapirique la plus importante est celle de Hægefjell qui constitue topographiquement un des points les plus hauts de la région. Le matériel qui la constitue est pour la plus grande partie de nature granitique à texture pegmatitique. Il est tout à fait distinct du matériel des migmatites avoisinantes: l'absence d'une foliation, permet sur le terrain une rapide distinction.

Evolution du bassin de Vrådal.

La structure en bassin la plus caractéristique est celle de Vrådal; elle mérite une mention spéciale, car toute une série d'événements ont pris place postérieurement à la formation du bassin. Nous écartant du travail de Sylvester (1964) nous suggérons l'évolution suivante:

- Formation du bassin (Fig. 11, 1);
- Intrusion d'un massif microdioritique qui produit un métamorphisme de contact, en même temps que des enclaves énallogène (Fig. 11, 2);
- Un intense fracturation autour du bassin permet la mise en place d'un matériel eudogène basique semblable aux roches supracrustales. Ces roches s'opposent à celles de la suite du Telemark auxquelles elles sont discordantes (Fig. 11, 3);
- Le massif microdioritique se transforme en quartzo-monzonite par intense feldspathisation potassique, avec formation d'enclaves homoeogènes (Fig. 11,4);
- Cette feldspathisation efface le métamorphisme de contact provoqué par le massif microdioritique et donne naissance à une roche porphyroïde à partir de n'importe qu'elle roche qui entoure le bassin de Vrådal. La roche porphyroïde (quartzo-monzonite porphyroïde) est caractérisée par la présence d'énormes cristaux de microcline dans une pâte assez variable au point de vue de la texture et de la composition minéralogique qui résulte de la feldspathisation de roches d'origines variées: la roche basique intrusive, les migmatites et les roches supracrustales.

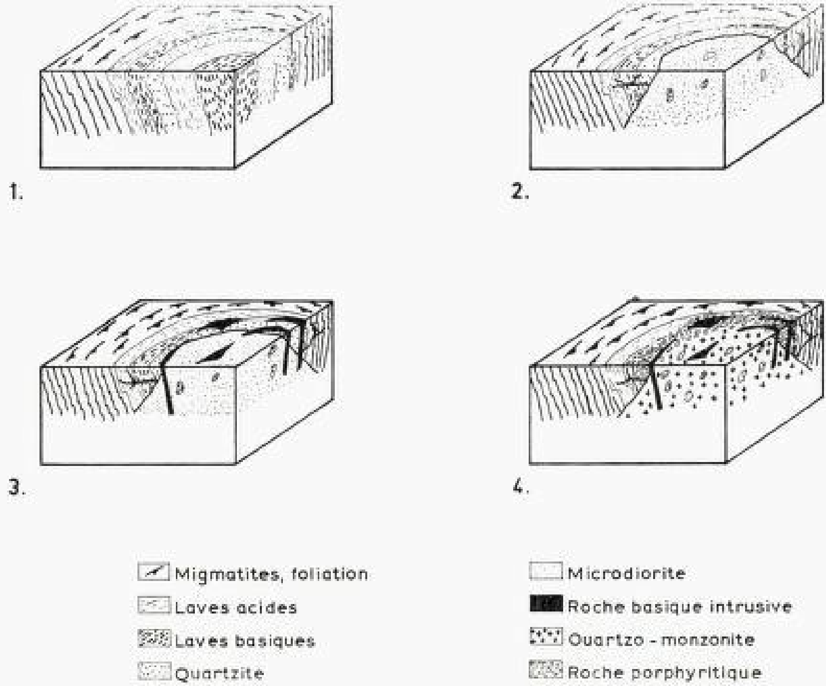


Fig. 11. Evolution du Bassin de Vrådal.

1) Formation de Bassin.
2) Intrusion microdioritique.
3) Mise en place de la roche basique dans les fractures.
4) Feldspathisation avec formation de la quartzo-monzonite et de la roche porphyritique.

Evolution of Vrådal Basin.

1) Formation of a basin.
2) Microdioritic intrusion.
3) Emplacement of a basic rock in fractures.
4) Feldspathisation with formation of quartz-
Migmatites, foliation.
Acid lavas.
Basic lavas.
Quartzite.
Microdiorite.
Intrusive basic rock.
Quartz-monzonite.
Porphyritic rock.

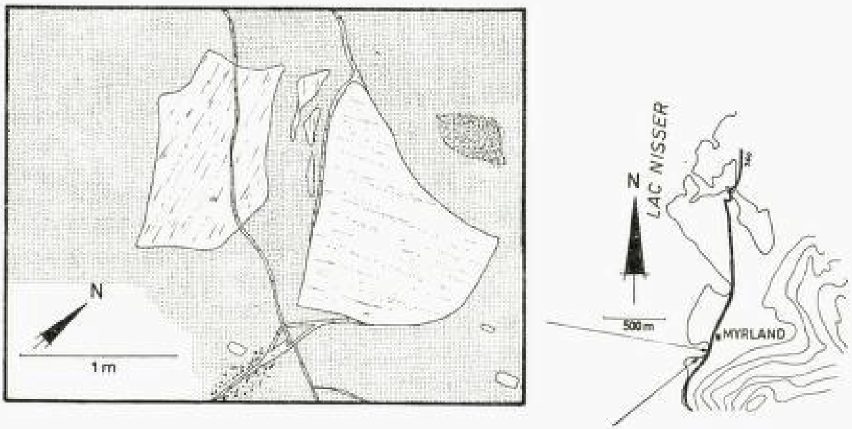


Fig. 12. Enclaves énallogènes déplacées de migmatites dans la quartzo-monzonite. Affleurement de Myrland (7, Fig. 14).

Displaced xenoliths of migmatite in quartz-monzonite. Exposure at Myrland (7, Fig. 14).

Mise en place des dernières générations de pegmatites (G_4, G_5).

Après la suite des événements qui ont affecté le bassin de Vrådal, nous devons signaler dans l'ordre chronologique, la mise en place des pegmatites de la génération G_4; elles sont concordantes avec la foliation des migmatites, mais recoupent la quartzo-monzonite; cela permet de les différencier de celles des générations antérieures qui sont coupées par cette roche (Fig. 13). Par la suite on note la formation des fissures les plus récentes et la mise en place des pegmatites tardives; elles sont discordantes par rapport à la foliation des migmatites et à la schistosité.

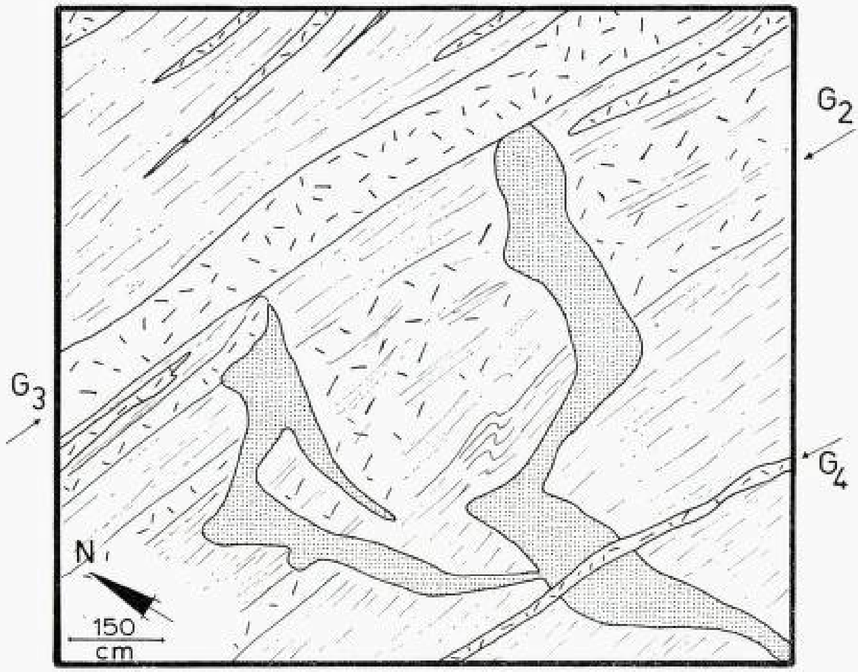
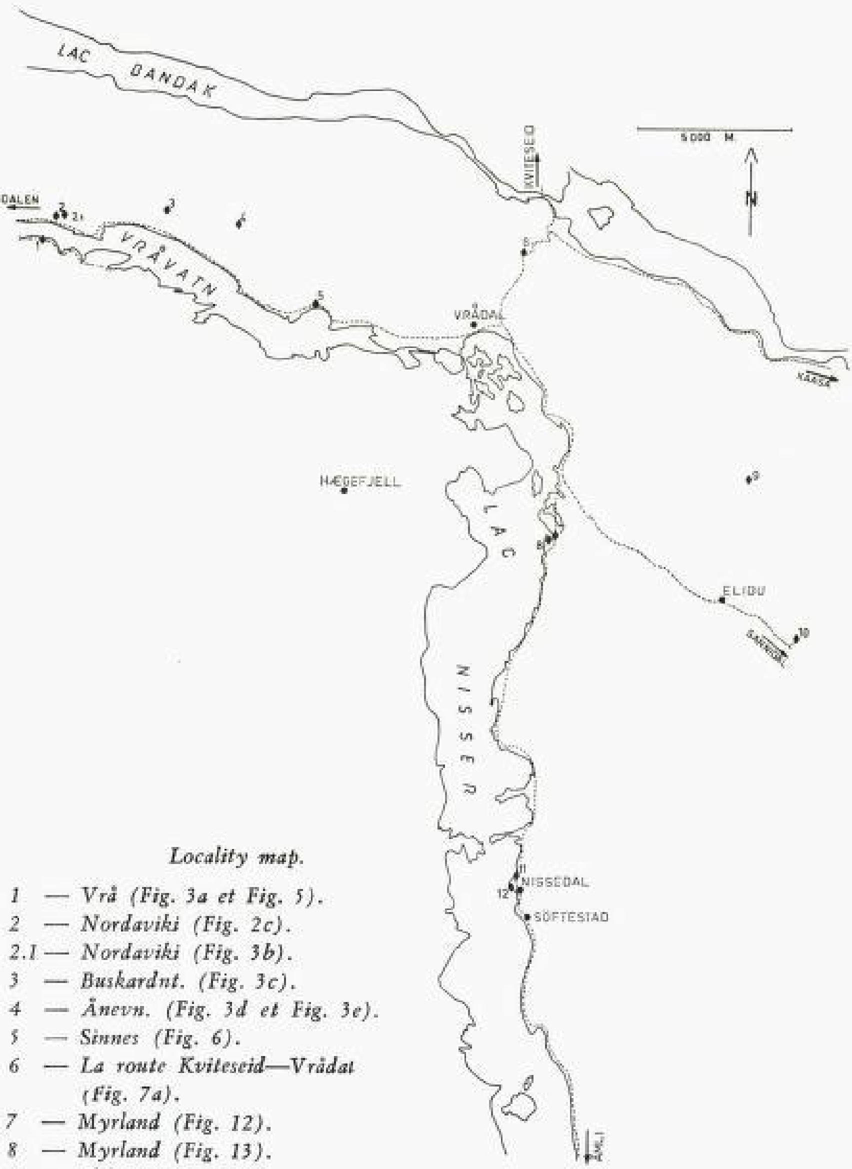


Fig. 13. Quartzo-monzonite discordante des migmatites. Pegmatites G_2 associées à la granitisation générale. Pegmatites G_3 antérieures à la mise en place de la quartzo-monzonite et pegmatites postérieures G_4. Myrland (8, Fig. 14).

Quartz-monsonite cross-cutting migmatite. The pegmatites G_2 associated with the general granitization. Pegmatites G_3 older and pegmatites G_4 younger than the quartz-monzonite.

Evolution morphologique.

Après les derniers phénomènes profonds toute la région remonte vers la surface. Elle sera alors soumise à l'action des agents d'érosion. C'est à la fin du Précambrien que le secteur évolue en pénéplaine. Nous trouvons un rajeunissement de la pénéplaine ainsi que des dislocations plus ou moins importantes comme conséquence des mouvements tectoniques postérieurs. Au quaternaire, l'inlandsis s'installe au Telemark comme dans tout le nord de l'Europe. Il se déplace suivant une direction N 110—120° E (Planche III). Son mouvement n'est pas trop influencé



Locality map.

1 — Vrå (Fig. 3a et Fig. 5).
2 — Nordaviki (Fig. 2c).
2.1 — Nordaviki (Fig. 3b).
3 — Buskardnt. (Fig. 3c).
4 — Ånevn. (Fig. 3d et Fig. 3e).
5 — Sinnes (Fig. 6).
6 — La route Kviteseid—Vrådal (Fig. 7a).
7 — Myrland (Fig. 12).
8 — Myrland (Fig. 13).
9 — Elibu-nord (Fig. 7b).
10 — Holman (Fig. 8 et Fig. 9).
11 — Nissedal (Fig. 2a).
12 — Nissedal (Fig. 2b).

Fig. 14. Carte des affleurements.

par la structure géologique. Localement la présence de fractures favorise l'action de l'érosion glaciaire, surtout dans la formation de lacs de surcreusement. A la fin de la période glaciaire les glaciers n'ont plus la force de détruire les obstacles: ils s'adaptent alors à la topographie précambrienne rajeunie. Il arrive qu'à certains endroits une topographie anormale prenne naissance: l'orientation du lac Nisser en est le plus bel exemple.

Au terme de ce travail, nous tenons à exprimer nos remerciements:

— à la Fondation Calouste Gulbenkian qui par l'octroi d'une bourse nous a permis de venir étudier à l'Institut de Géologie de Neuchâtel, ainsi qu'en Norvège;
— aux professeurs C. E. Wegmann, J. A. Dons et J. P. Schaer sous la direction desquels ce travail fut élaboré;
Nous remercions également le professeur C. Teixeira, ainsi que Monsieur J. Meia chef de travaux à l'Institut de Géologie de Neuchâtel, pour l'aide qu'ils nous ont apporté.

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Original manuscript received in October 1968, revised manuscript received and accepted for publication in August 1969.

Pl. I

Geological Map

Legend:
Quartzites and conglomerates
Acid lavas and associated porphyries
Basic lavas and agglomerates
Quartz-monzonite
Basic intrusive rocks
The Fjone breccia
Visible contact
Interpreted contact
Granite
Porphyric quartz-monzonite
Homogeneous migmatites (foliation)
Heterogeneous migmatites (foliation)
Pegmatitic aplite-complex
Pegmatite
Transformation aureole
Contact not visible

Tectonic sketch:
Synform of 1st folding
Antiform of 1st folding
Synform of 2nd folding
Antiform of 2nd folding
Synform of 3rd folding
Antiform of 3rd folding
Zones of antiforms
Zones of synforms

Pl. II

Map showing the «S» geometry.

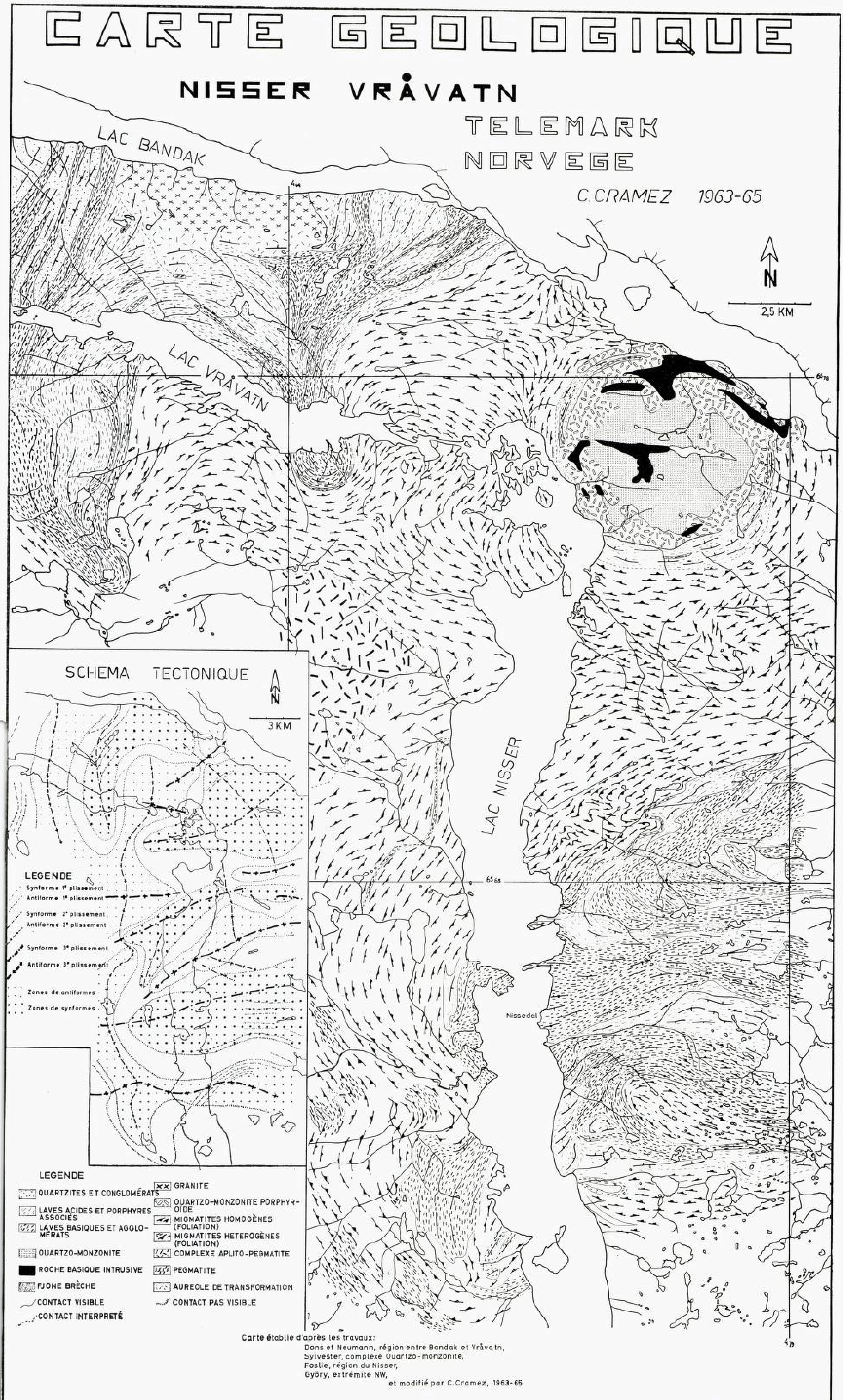
Each of the four regions (I, II, III, IV) is marked in the map by its own pattern. In region I the B_2 axis is dominant. Region II is the Vrådal basin (The Vrådal pluton of Sylvester, 1960). In region III the folds B_2 and B_3 are equally well developed. Region IV is dominated by B_3 axis. Three diagrams lack curves due to insufficient number of observable elements.

Pl. III

Discontinuity lines, Morphology.
Legend:
Discontinuity lines
Glacial striae
Lakes
Contours with intervals 430, 790, 880, 970 m. a.s.l.

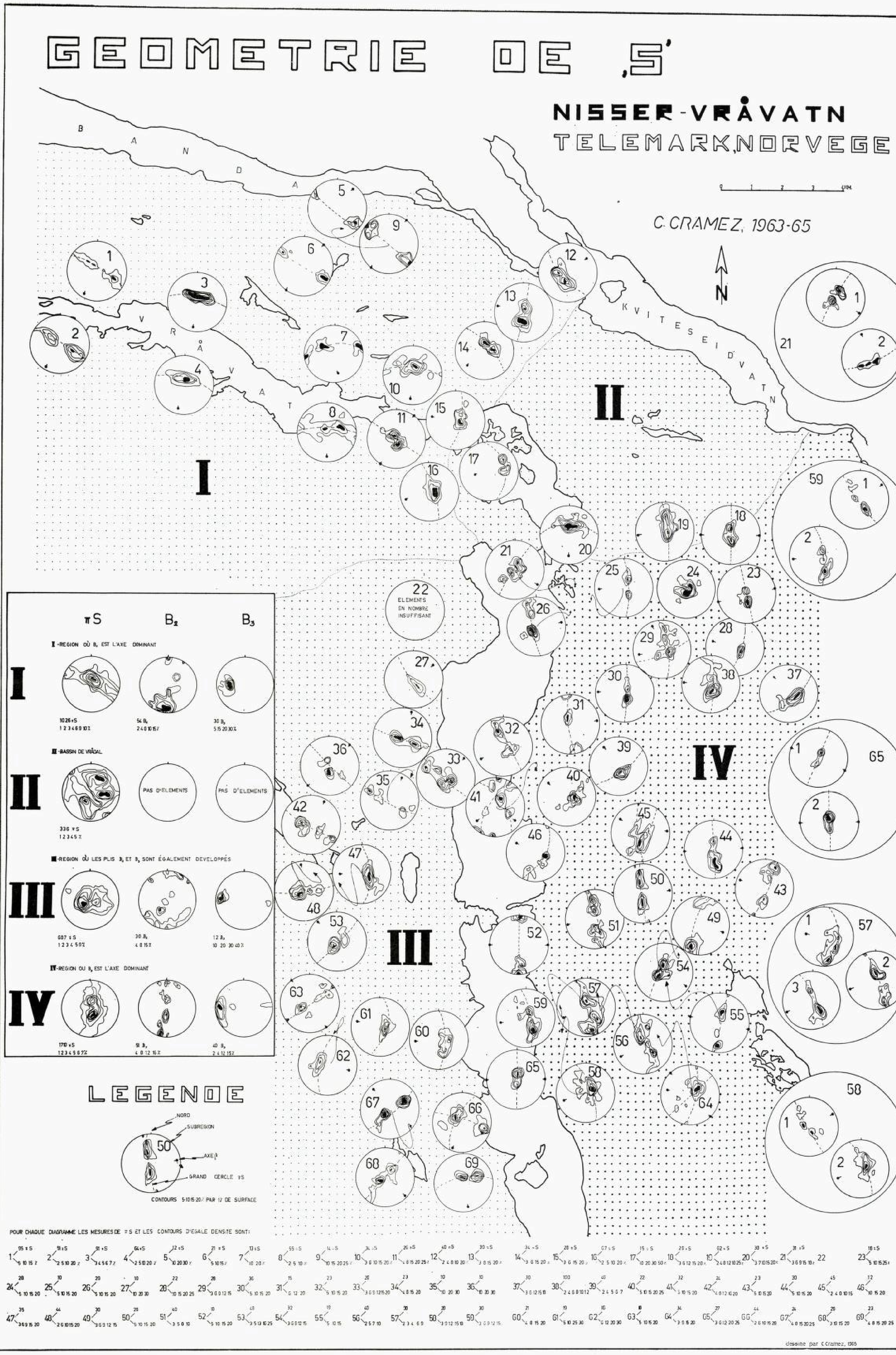
Inset map based on map by O. Holtedahl & B. G. Andersen 1960.

Glacial striae
Water divide
Present glacier
Studied region

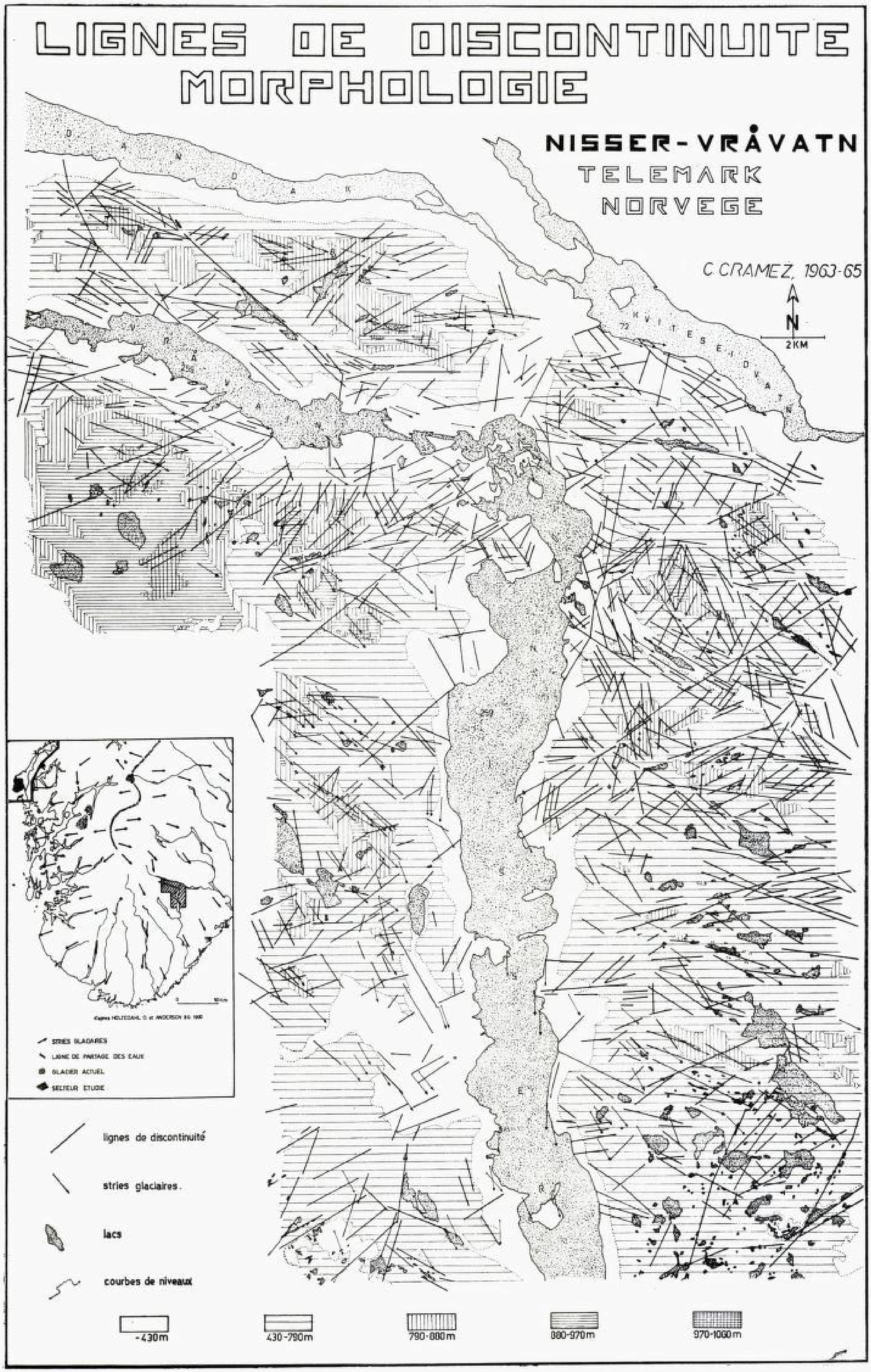


Carte établie d'après les travaux:
Dons et Neumann, région entre Bandak et Vråvatn,
Sylvester, complexe Quartzo-monzonite,
Foslie, région du Nisser,
Györy, extrémité NW,
et modifié par C. Cramez, 1963-65

PL. I


GEOMETRIE DE S'
NISSER-VRÅVATN
TELEMARK, NORVEGE
C. CRAMEZ, 1963-65
N
BANDAK
VRÅVATN
KVITESEIDVATN
I
II
III
IV
22 ELEMENTS EN NOMBRE INSUFFISANT
πS
B2
B3
I - REGION OÙ B2 EST L'AXE DOMINANT
II - BASSIN DE VRÅDAL
PAS D'ELEMENTS
III - REGION OÙ LES PLIS B2 ET B3 SONT ÉGALEMENT DEVELOPPÉS
IV - REGION OÙ B3 EST L'AXE DOMINANT
LEGENDE
NORD
SUBREGION
AXE β
GRAND CERCLE πS
CONTOURS 5-10-15-20% PAR 1% DE SURFACE
POUR CHAQUE DIAGRAMME LES MESURES DE πS ET LES CONTOURS D'EGALE DENSITE SONT:
dessiné par C. Cramez, 1965

PL. II



LIGNES DE DISCONTINUITE
MORPHOLOGIE
NISSER-VRÅVATN
TELEMARK
NORVEGE
C. CRAMEZ, 1963-65
N
2 KM
STRIES GLACIAIRES
LIGNE DE PARTAGE DES EAUX
GLACIER ACTUEL
SECTEUR ETUDIE
lignes de discontinuité
stries glaciaires.
lacs
courbes de niveaux
-430m
430-790m
790-880m
880-970m
970-1060m

PL. III

SOME GEOPHYSICAL PROFILES IN ØSTFOLD

By

G. Lind[1]) and S. Saxov[1])

Abstract.

Based on detailed gravimetric observations three profiles are examined, the profiles running some 20 km west—east from Jeløya, Østfold fylke. The magnetic values are taken from maps published by NGU. Geological samples have been collected and the physical properties of the samples have been determined in the laboratory. By application of GIER—ALGOL programmes models have been studied.

The known geological displacement which is estimated to about 1000 m between Jeløya and the main-land is not seen on the gravimetric picture in the northern profile, while the two southerly profiles give displacement values of 800—900 m. In the northern Jeløya a heavy body of a thickness of about 800 m is calculated. A smaller basic body of 100 m thickness is located in the main-land on profile II. Granitic bodies of thickness from 350 to 2000 m are found in profiles II and III. The magnetic values support the interpretation in the Jeløya area, while we have no distinct magnetic anomalies in the main-land.

Introduction.

In a previous paper (Ramachandra et.al. 1967) a preliminary report was given concerning some geophysical measurements in Jeløya. Since that time the gravimetric surveying has continued and by the time of writing the coverage is some 1500 gravity stations in about 450 km^2. Magnetic measurements on the ground have been carried out only in connection with detailed investigations of special geological problems. For this study the aeromagnetic maps published by NGU have been applied.

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Fig. 1 is a location map as far as the three profiles are concerned. In the figure we have marked the sample localities. The geological samples were mainly collected in the 1968 season in such a way that if an outcrop occurred at the gravimetric reading place a sample was taken.

Geology.

The area investigated can geologically be divided into two main groups, the precambrian and the paleozoic-permian. The precambrian rocks cover the area except the island Jeløya, which is situated in the western part of the area. They consist mainly of strongly metamorphic rocks, gneisses of various kinds, with minor occurrence of plutonic rocks like gabbro and peridotite. The rocks have been transformed during several deformations and migmatites are common. At the end of the last deformation granites were emplaced, one of them being the Våler granite, in the eastern part of the area. (See Fig. 3).

The precambrian geology is not known in details, but a new mapping of the area has been commenced (Berthelsen 1967), and it is then to be hoped that more detailed and correct geological information will be available within a reasonable time in order to obtain a more correct geophysical interpretation. For the present study use has been made of the geological maps by Rekstad (1921), Brøgger and Shetelig (1926), and Gleditsch (1960).

The other group of rocks in the area is the above mentioned paleozoic-permian rocks of Jeløya. They belong to the Oslo field and are down-faulted in the precambrian. The main faultline runs north—south between the island and the main-land. For further details on this group see Brøgger and Shetelig (1926) and Ramachandra et.al. (1967).

In table 1 the rock-samples, which were taken at the gravimetric stations where outcrops occurred, are listed together with their physical properties as they are determined in the laboratory. Density has been determined in the usual way, and an Oersted meter has been applied for the determination of susceptibility and remanent magnetism. The accuracy for the density values is better than 0.01 gr/cm^3, while the accuracy for susceptibility and remanent magnetism is not better than 0.00002 c.g.s. units. The location of the samples are plotted on Fig. 1. As far as rhomb-porphyry, basalt, and sandstone are concerned the samples are all collected in Jeløya. Previously (Ramachandra et.al. 1967) physical properties have been given for the same rock-types originating from Jeløya. The Våler granite, migmatite, biotite-hornblende gneiss, quartz-feldspar gneiss, and amfibolite (gabbro and metagabbro) are collected in the main-land. Even if the samples cluster in groups we believe that the physical properties obtained suffice this preliminary study as representative for the geological formations present.

Table 1

Sample No.	Rock-Type	Locality	Density gr/cm^3	R.M. c.g.s.	Suscept. c.g.s.	Q
302	Våler Granite	Langøen	2.60	<0.00002	<0.00002	
1026	—	Haugen	2.61	—	—	
1065	—	Ven	2.61	—	—	
1068	—	Turen	2.60	—	—	
1070	—	Røstad	2.59	—	—	
1071	—	Turen	2.60	—	—	
312	Migmatite	Brasenbogen	2.79	—	—	
315	—	Dillingøen	2.62	—	—	
317	—		2.68	—	—	
318 A	—	St. Kvernø	2.62	0.00005	0.00107	0.10
325	—		2.67	<0.00002	<0.00002	
326	—	Fæøen	2.62	—	—	
1101	—	Oppegård	2.67	—	—	
1474	—	Høiaas	2.84	—	—	
1673	—		2.68	0.00060	0.00006	21.52
9314	—	Laursbakken	2.67	<0.00002	<0.00002	
9107	—	—	2.74	—	—	
9301	—	Patterød	2.67	—	—	
9101	—	Norødegård	2.63	—	—	
314	Biotite-hornblende gneiss	Dillingøen	2.68	—	—	
319	—	Bjørnø	2.69	—	—	
529	—		2.64	—	—	
667	—		2.71	—	—	
721	—	Henes	2.65	—	—	
766	—	Rød	2.74	<0.00002	<0.00002	
1077	—		2.69	—	—	
1147	—	Sjulerød	2.69	—	—	
9404	—	Noretj.	2.73	—	—	
9103	—	—	2.86	—	—	
9405	—	Norødegård	2.80	—	—	
9314	—	—	2.67	—	—	

Sample No.	Rock-Type	Locality	Density gr/cm³	R.M. c.g.s.	Suscept. c.g.s.	Q
316	Quarz-feld-spar gneiss	Dillingøen	2.64	0.00009	0.00148	0.13
318	—	Kvernø	2.60	$<$0.00002	$<$0.00002	
310	—	Osienrødøen	2.63	—	—	
327	—	Nesengen	2.65	—	—	
1498	—		2.60	0.00009	0.0015	0.11
1499	—	Kjaita	2.62	$<$0.00002	$<$0.00002	
1500	—		2.62	$<$0.00002	0.00043	0.07
9313	—	Laursbakken	2.63	$<$0.00002	$<$0.00002	
9102	—	—	2.64	—	—	
9302	—	—	2.62	—	—	
9303	—	Patterød	2.61	—	—	
9114	—	—	2.62	—	—	
9117	—	—	2.63	—	—	
9120	—	—	2.58	—	—	
9309	—	—	2.65	—	—	
9121	—	—	2.64	—	—	
9312	—	—	2.59	—	—	
164	Amfibolite		3.08	—	—	
222	—	Isdam	3.02	—	—	
1080	— (gabbro)	Risheim	3.07	—	—	
1148	—	Veidal	3.08	—	—	
1468	— (gabbro)	Kaabel	2.88	—	—	
1478	—		3.10	—	—	
9104	—	Europaveg 6	3.07	—	—	
9105	—	—	2.99	—	—	
9316	—	—	2.98	—	—	
9109	—	—	2.96	—	—	
9304	—	—	2.92	—	0.00007	
9110	—	—	2.99	$<$0.00002	$<$0.00002	
9306	—	—	3.04	—	—	
9113	—	—	2.92	—	—	
9305	—	—	3.07	—	—	
9110	—	—	2.98	—	—	
9307	—	—	3.11	—	0.001	
9118	—	—	2.96	—	$<$0.00002	
9119	—	—	3.00	—	—	
9403	—	—	2.95	—	—	
9108	— (meta-gabbro)	Noretj.	3.01	—	—	
9106	— (gabbro)	—	3.22	—	—	

Sample No.	Rock-Type	Locality	Density gr/cm³	R.M. c.g.s.	Suscept. c.g.s.	Q
9112	— (meta-gabbro)	—	3.11	—	—	
9115	— (meta-gabbro)	—	2.96	—	—	
J. 7	Rhomb-Porphyry	Jeløya, Reier	2.68	0.0005	0.0029	0.32
J. 9	—	—	2.62	0.00006	0.0003	0.47
J. 1	Basalt	— Kongshavn	2.88	0.0023	0.0045	1.0
J. 2	—	— —	2.79	0.0028	0.0025	2.3
J. 3	—	— Kullebunden	2.77	0.0016	0.0003	12.5
J. 5	—	— Renneflot	2.77	0.0025	0.0024	2.2
J. 8	—	— Singelsbukta	2.84	0.0097	0.0033	6.1
J. 13	—	— Kippenes	2.77	0.0048	0.0004	24.9
J. 6	— (agglom-erate)	— Kongshavn	2.68	0.00071	0.0030	0.49
J. 11	— (agglom-erate)	— Englevik	2.71	0.0027	0.0019	2.9

Gravity.

The gravimetric readings have been carried out by means of Worden gravimeters W 142, WP 148, WM 653, WM 681, and WM 779. In order to establish a base station net system one or two of the Worden gravimeters have been employed together with LaCoste & Romberg gravimeter 54. A report on the base station net system is under preparation and will be published elsewhere. However, stations previously included in the European Gravimeter Calibration Line (Kejlsø 1958, Saxov 1958 and 1966, Sømod 1957) are included in the base station net system.

All the gravimetric observations are referred to Oslo Fundamental Gravity Station with G=981.92815 Gals. A density value of 2.67 gr/cm³ has been employed in the computation of the Bouguer correction. The coordinates are taken from the topographical/economic maps in the scale of 1 : 5 000 or 1 : 10 000. The elevation values are taken partly from bench-marks and polygon-points, the heights being known in millimetres or centimetres, and partly from dot-points in the topographical maps, the heights then being in metres or half-metres. No topographic correction is applied. Theoretical gravity values are taken from the tables by Andersen (1956).

In Figs. 2, 3, and 4 the profiles are presented. It is not possible to obtain a proper regional gravity trend from the present gravity map (NGO 1960) due to too few gravity stations and a non-representative coverage of the area in question. The regional gravimetric trend has therefore been drawn graphically in the usual manner, see e.g. Dobrin (1960). This means that the regional trend concerns the area close to the profiles. The residual effect is thus the difference between the computed Bouguer anomalies and the regional trend. The residual anomalies are caused by shallow bodies in the outer crust. In the interpretational studies due respect is taken to the density value which according to table 1 and to previous values (Ramachandra et.al. 1967) can be summarized as follows:

Table 2.

Rock-Type	Sample No.	Density Range gr/cm^3	Density Mean gr/cm^3	Stand. Dev. gr/cm^3
Jeløya				
Rhomb-porphyry	18	2.56—2.68	2.60	0.04
Basalt	23	2.62—2.88	2.76	0.06
Sandstone	5	2.63—2.69	2.66	0.02
Main-land				
Våler granite	6	2.59—2.61	2.60	0.01
Amfibolite	24	2.88—3.22	3.02	0.08
Migmatite	13	2.62—2.84	2.68	0.07
Biotite-hornblende gneiss	12	2.64—2.86	2.70	0.06
Quarz-feldspar gneiss	17	2.58—2.65	2.62	0.02
Migm.-biot.-Quarz	42	2.58—2.86	2.66	0.06

In comparison with the values listed by Ramachandra et.al. (1967) it is seen that the addition of two samples to the rhomb-porphyry group has caused no change in the mean density value. Likewise for the basalt group, where 8 additional samples have increased the total number from 15 to 23, the mean density value being unaltered as 2.76 gr/cm^3. As far as sandstone is concerned no additional samples have been collected. The spread in the density values is also the smallest in that group.

Looking into the samples from the main-land we have only 6 Våler granite pieces. However, collected within a larger area the granite seems to be very uniform. The amfibolite group has a large spread, the mean value, however, seems to fit well with the conventional value. Concerning the migmatites and the gneisses we have given values for each group as well as for the total group. Evidently the quarz-feldspar gneisses have a smaller spread and are lighter than the two other groups. We have, however, felt that for the present study it would be reasonable to consider migmatites and gneisses as one group. That means that we count on three groups in the main-land, granite with a density value of 2.60 gr/cm^3, amfibolite with 3.02 gr/cm^3, and migmatite/gneiss with 2.66 gr/cm^3.

By inspection of the gravimetric profiles we find on profile I four anomalies numbered from A to D. Anomaly A has a counterpart in the magnetic anomaly I, and it is also seen in the previous paper (Ramachandra et.al. 1967). Anomalies C and D have the character of a fault structure, however, they may also be effects from geological structures which seems to be supported by magnetic anomaly J. Anomaly B has no analogous magnetic anomaly and is not correlated with any known geological evidence.

In profile II we have three interesting anomalies, E, F, and G. As was the case in profile I gravity anomaly E in Jeløya corresponds to magnetic anomaly K. No magnetic anomalies are seen in the eastern part of the profile. There exists a possibility that one or two of the small gravimetric peaks between anomalies E and F correspond to anomaly B in profile I.

Even if the regional gravimetric trend is seen in profiles I and II it is more clearly demonstrated in profile III. This profile has consequently been deciding in the determination of the regional gravimetric effect. The residuals are all small except anomaly H. Once again a small peak to the east of anomaly H is seen analogous with the two other profiles.

Magnetics.

The magnetic data applied in this study originate from aeromagnetic maps published by NGU (1966/67). The measurements are carried out partly by means of a flux-gate magnetometer type AN/ASQ-3A partly by a proton magnetometer type ELSEC 592. The flight-lines are orientated east-west, the distance between flight-lines being about 1000 m. The heights are between 100 and 150 m. The values shown on Figs. 2, 3, and 4 are the total magnetic force in gammas.

By inspection of the magnetic properties in Table 1 we find that granite, biotite-hornblende gneiss, and sandstone give no response at all, that of migmatite, quarz-feldspar gneiss, and amfibolite only 2, 4, and 2 samples out of 13, 17, and 24 respectively give magnetic indication — and that they all show very weak magnetic properties. As far as rhomb-porphyry and basalt are concerned we have the same picture as found previously (Ramachandra et.al. 1967) that basalt is stronger magnetized than rhomb-porphyry which is rather weak magnetically. The following table summarizes the values:

Table 3.

Rock-Type	Sample No.	R.M. — cgs units Range	R.M. — cgs units Mean	Suscep. — cgs units Range	Suscep. — cgs units Mean
Jeløya					
Rhomb-porphyry	12	0.00006—0.00129	0.00067	0.00004—0.0029	0.00055
Basalt	21	0.00032—0.01484	0.00247	0.00008—0.0065	0.00299
Main-land					
Migmatite	2	0.00005—0.00060	0.00032	0.00006—0.00107	0.00062
Quarz-feldspar gneiss	4	0.00002—0.00009	0.00007	0.00002—0.0015	0.00088
Amfibolite	2			0.00007—0.001	0.00054

It is a general trend in the aeromagnetic profiles that the magnetic curves are very uniform in the main-land with exception of the small anomaly J in profile I and a weaker, more broad, anomaly M in profile III. As mentioned earlier anomaly J seems to be correlated with a geological border, while apparently no gravimetric anomaly is corresponding to anomaly M. That only small magnetic disturbances are to be found in the main-land is in confirmation with the magnetic properties of the rock samples. The magnetic anomalies I, K, and L are presumably all related to the down-faulted rocks at Jeløya.

Some of the permian lava samples were taken orientated and the magnetic results show a reversed magnetization direction for these samples. This fact may contribute to the explanation of some of the peculiar shapes of the magnetic curves.

Interpretation.

To avoid mistakes or misinterpretations due to terrain effects only gravimetric anomalies where terrain effects are supposed not to disturb the anomaly picture have been treated. It has to be pointed out that due to the scarcity of geological evidence some of the models shown are interpreted by application of the geophysical anomaly curves only and they should therefore be looked upon with this point in mind. The calculations have been carried out by application of a GIER-ALGOL programme developed by Henkel (1969) for two-dimensional bodies making use of the formula according to Talwani et.al. (1959). In cases where the anomalies are not of true two-dimensional shape end-corrections according to Nettleton (1940) have been applied.

Turning now to a discussion of the residual anomalies seen in the three profiles it must be stated that none of the four anomalies A to D in profile I can be directly correlated to known geology. The narrow positive anomaly A of about 2 mgals has been considered as due to a local thickening of the basalt, however, the relief could also be due to a disturbance from a heavier body of the essexite type. The shape of anomaly A suggests a shallow structure. The total magnetic field curve with anomaly I shows strong disturbance at the same locality and the sample J. 8 (basalt) shows typical magnetic properties. The model applied is based on a density contrast of $+ 0.10$ gr/cm^3. The fit between the measured and the calculated points is not too satisfying for the present model, which gives a thickness of about 800 m. It may therefore be more reasonable to assume that the density contrast is a little higher. East of anomaly A runs the geological fault-line between the younger rocks in Jeløya and the precambrian in the main-land. The gravimetric field does not show much relief in this respect, probably mostly owing to too few measurements in connection with the strait.

The gravimetrically fault-like anomalies (B, C, and D) in the gravity curves in the eastern part of profile I have no known geological explanation. The anomaly features C and D have been interpreted as a vertical contact between a heavy basic body and a granitic body. Both types of these rocks are known from the area concerned, however, their exact location is unknown so far. By application of a density contrast of $+ 0.30$ gr/cm^3 the basic body (anomaly C) is estimated to have a thickness of about 100 m, while the granitic body (anomaly D) with a density contrast of $- 0.07$ gr/cm^3 shows an asymmetric character with a thickness of about 800 m in the western end and of about 300 m in the eastern end. Anomaly B could be interpreted in a similar way as has been done for anomaly C and the result would be a smaller asymmetric basic body.

When we now turn to profile II it is of interest to note that gravimetric anomaly E consists of a broad negative anomaly of 4—5 mgals with a distinct positive peak of about 3 mgals almost in the middle. The negative anomaly is probably due to the down-faulted paleozoic sedimentary rocks and has been interpreted this way. The density contrast is not known except for the upper devonian sandstone, but the displacement of about 800 m by application of a density contrast of -0.10 gr/cm^3 is in accordance with the geological estimation of about 1000 m. It is seen that the calculated points fit well with the measured curve. The influence from the lavas has not been taken into consideration as the lavas are believed to be thin thus their gravimetric effect is of smaller amplitude. The positive gravimetric peak inside anomaly E could be due to the basalt, but terrain effects may change the picture why no interpretation has been carried out. The terrain effect problem is the reason why the fit of the calculated curves and the measured curves have not been driven too far.

The anomaly K of the total magnetic field force shows a broader positive value with a negative value in the middle. The western positive peak could be correlated with the basaltic formation in Jeløya. This anomaly coincides with the gravimetric anomaly. According to the method described by Bean (1966) the depth to the disturbing body (basalt?) is about 50 m below the surface. The sloping values on the western flank of the magnetic anomaly is probably due to the sandstone. The eastern feature of anomaly K is related to the displacement.

Gravimetric anomalies F and G have been correlated with known outcrops of granite, the Våler granite. The borders against the gneisses are unknown in details and have been fixed from the gravimetric profiles. Using a density contrast of -0.07 gr/cm^3 the granitic body from anomaly F seems to be very regular with a thickness of about 350 m, while the anomaly G granitic body is more irregular of shape. The maximum thickness is about 2000 m.

As stated earlier the regional gravimetric trend is demonstrated most clearly in profile III and the trend dominates the picture. Except anomaly H the residual curve is slightly undulating. Anomaly H has a similar shape as anomaly E except that H has a large negative value in the eastern end. That may be due to lesser terrain effect combined with thicker quaternary cover. The geological conditions in anomaly E and anomaly H are very similar and by application of a density contrast of -0.10 gr/cm^3 we obtain a thickness of about 900 m. When we analyse the aeromagnetic anomaly L according to the method by Bean (1966) we get a depth of about 200 m to the top of the disturbing body.

Acknowledgement.

The authors wish to thank Professor, Dr. St. Skjeseth, Geological Institute, NLH, for facilitating the field work, chief geodesist Th. Sømod, NGO, for information and mapping material, director I. Aalstad, NGU, for aeromagnetic mapping material, docent, Dr. E. Tengström, Uppsala University, who put gravimeters WM 681 and LCR 54 at our disposal, director, Professor, Dr. E. Andersen, Danish Geodetic Institute, who put gravimeter W 142 at our disposal, Professor, Dr. T. Barth, Oslo University, who put gravimeter WM 653 at our disposal, and director I. Aalstad, NGU, who put gravimeter WP 148 at our disposal.

Part of the gravimetric measurements in 1967 and 1968 was carried out by cand. real. I. Ramberg, and H. Henkel, M.Sc., is responsible for part of the gravimetric readings from 1968.

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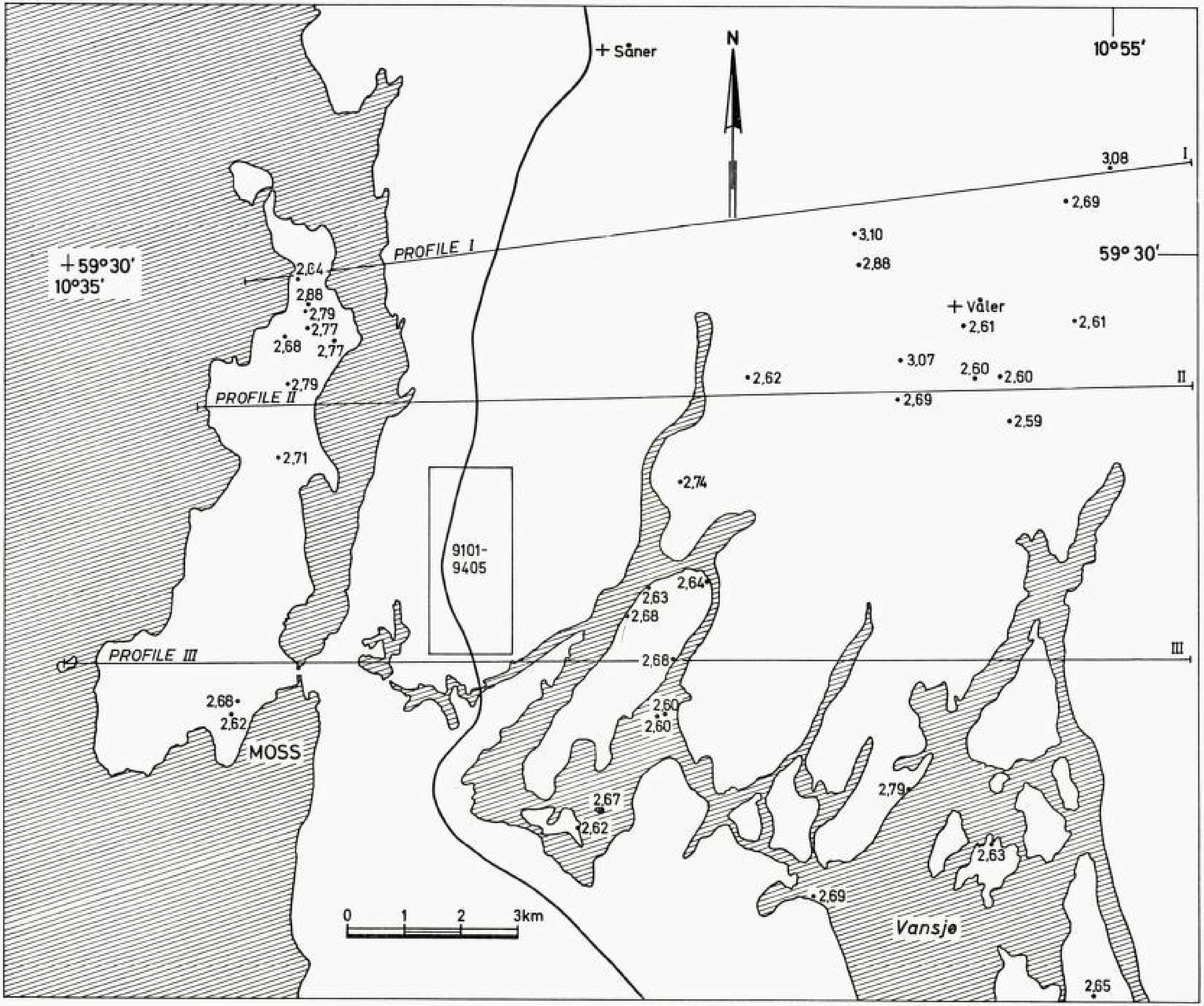

Fig. 1. Location map for profiles I, II, and III is shown. Density values are given.

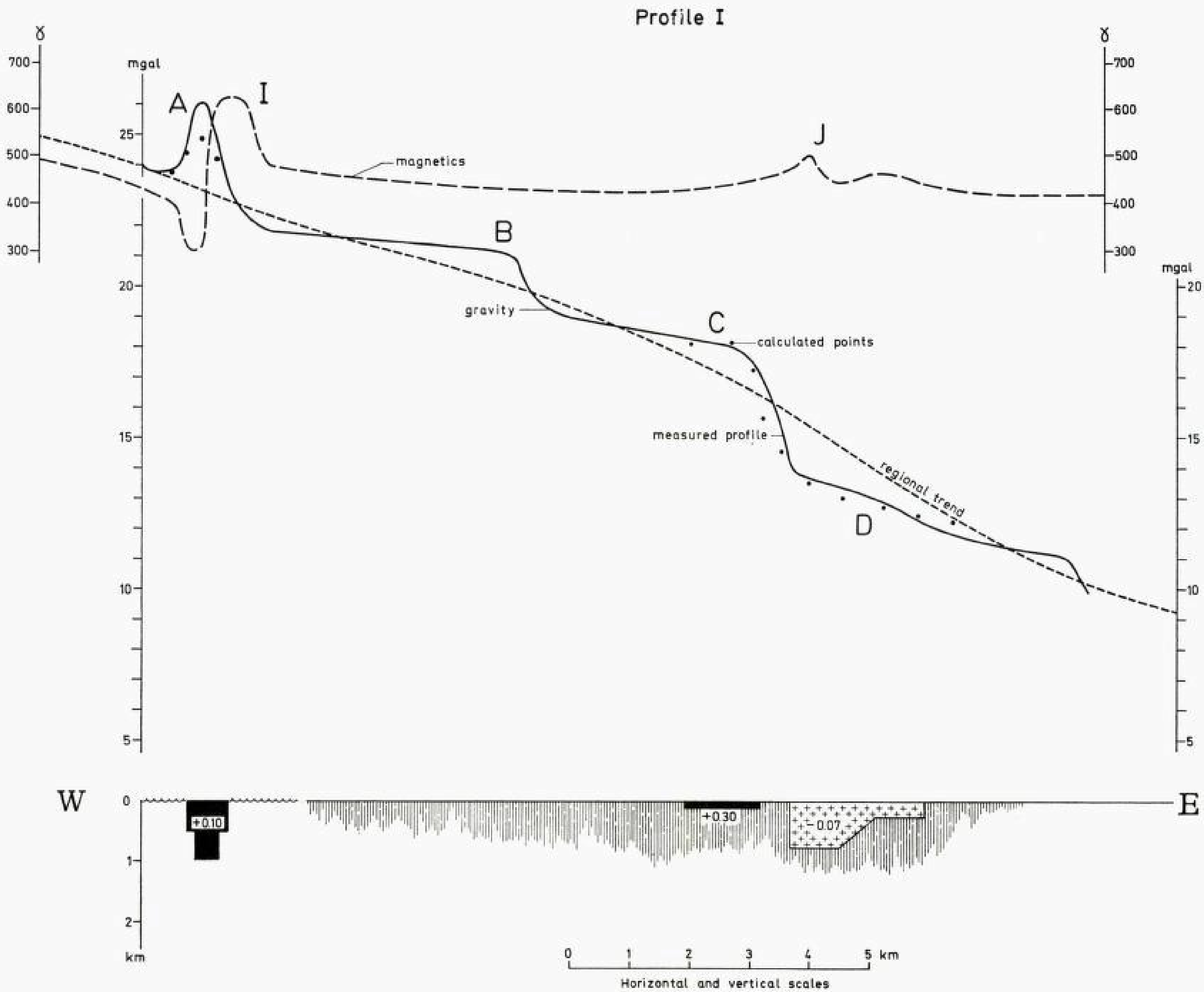

Fig. 2. Profile I showing the gravimetric and magnetic values. In the lower part the geological bodies are shown.

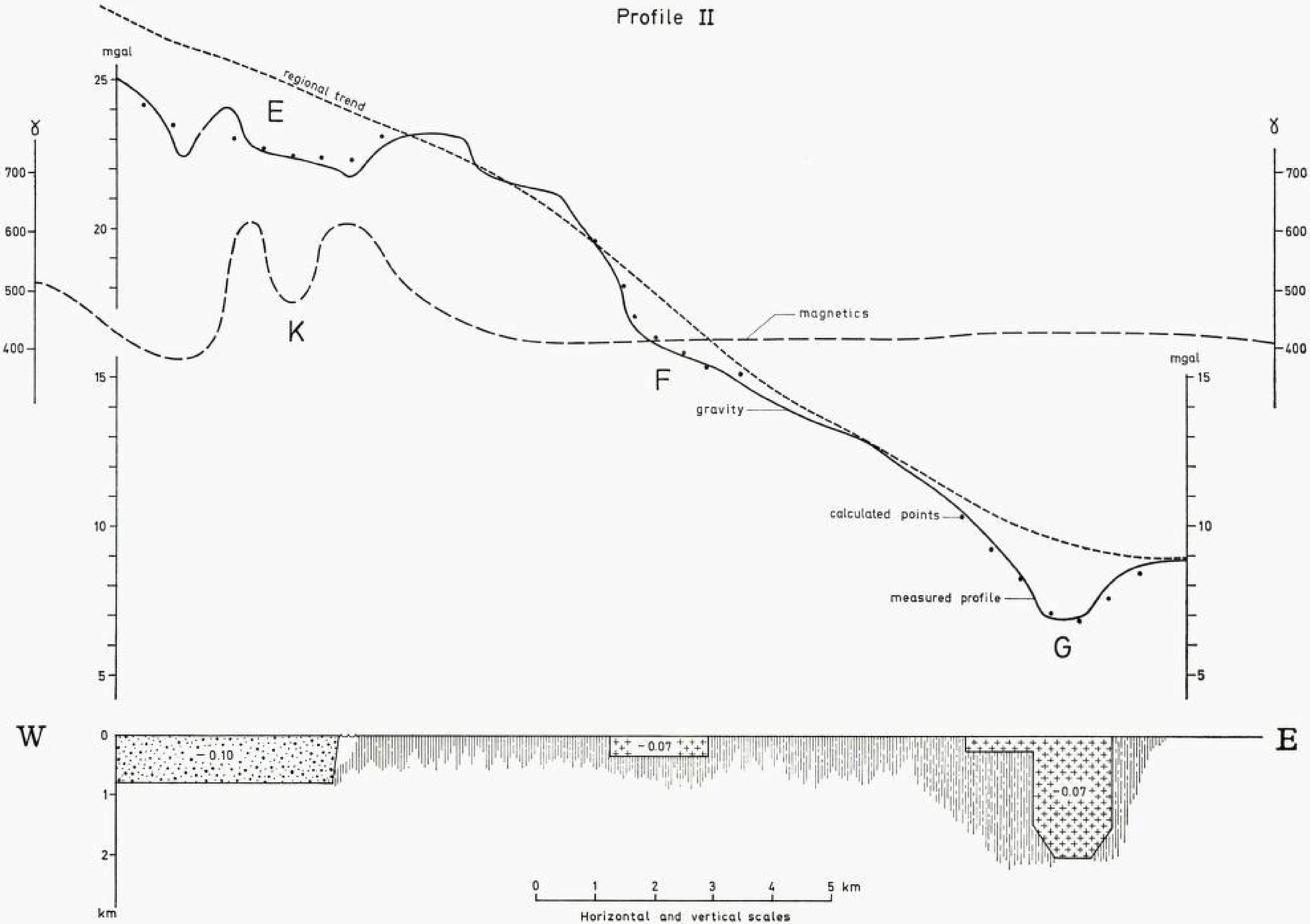

Fig. 3. Profile II showing the gravimetric and magnetic values. In the lower part the geological bodies are shown.

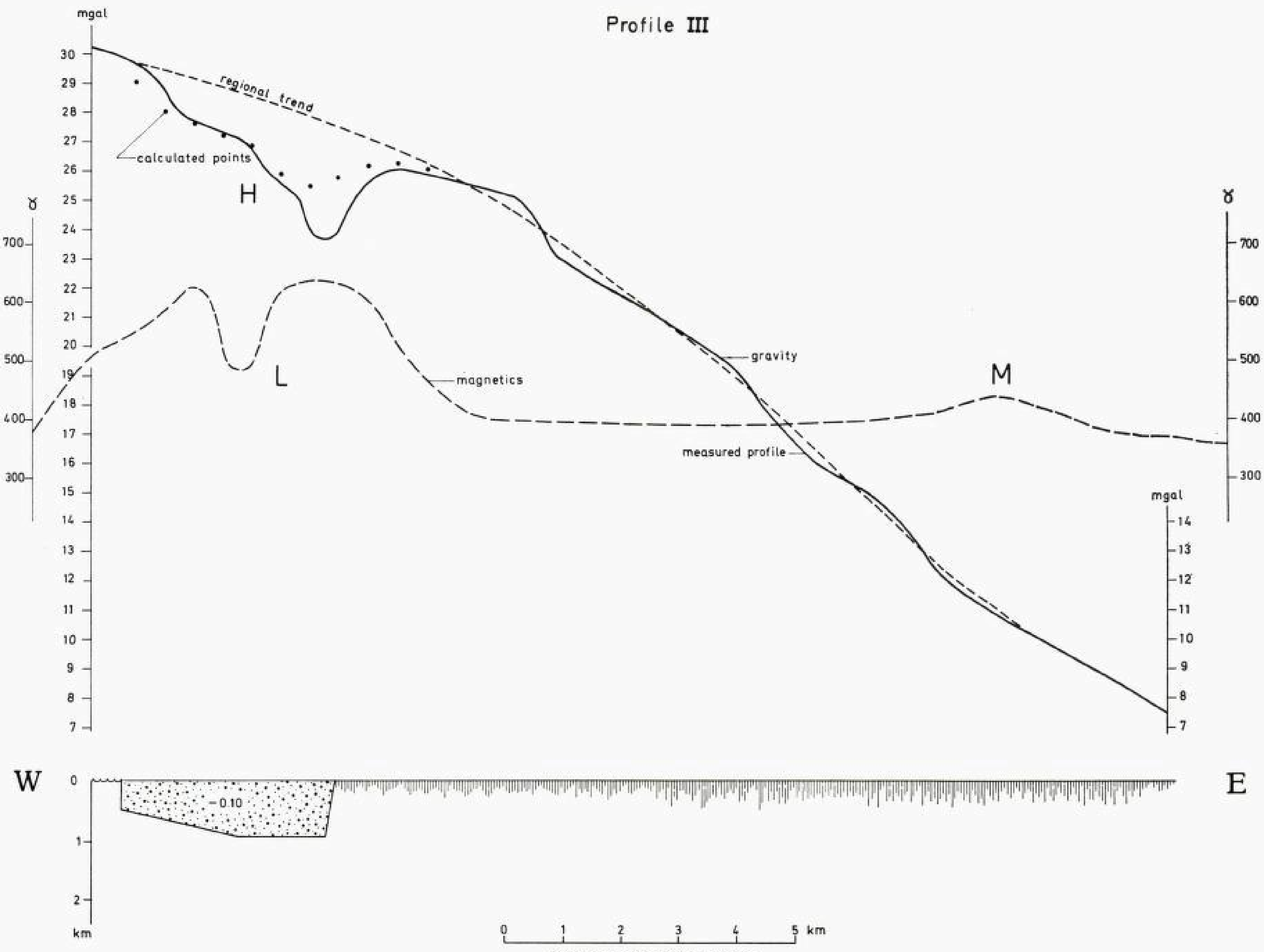

Fig. 4. Profile III showing the gravimetric and magnetic values. In the lower part the geological bodies are shown.

AEROMAGNETIC INVESTIGATIONS ON THE CONTINENTAL SHELF OF NORWAY, STAD-LOFOTEN (62—69°N)[1]

By
Knut Åm[2]

Abstract.

An aeromagnetic isogam map based on measurements by Norges geologiske undersøkelse 1965—67 is presented and interpreted.

The map reveals several interesting geological features, the most important of which is the existence of a large sedimentary basin situated parallel to the coast with its axis 120—150 km from the coastline, having widths above 200 km and maximum depths to magnetic basement exceeding 10 km.

Introduction.

After having gained a considerable amount of experience during years of aeromagnetic surveying in Norway, Norges geologiske undersøkelse, Geofysisk avdeling started to look seawards in 1962. Having successfully done some offshore mapping in Skagerak 1962—63 on behalf of Universitetet i Bergen (Aalstad and Sellevoll, in preparation), the Institution felt prepared to face the problems of the vast and practically unexplored Norwegian shelf areas. The geological knowledge concerning this part of Norway was rather sparse at that time, amounting to a few erratic boulders and a small downfaulted area of Mesozoic age (Ørvig, 1960). This indicated the possibility of younger sediments existing in the continental shelf outside the marginal chan-

1) Publication No. 1 in NTNF's continental shelf project.
Publication No. 16 in the Norwegian geotraverse project.

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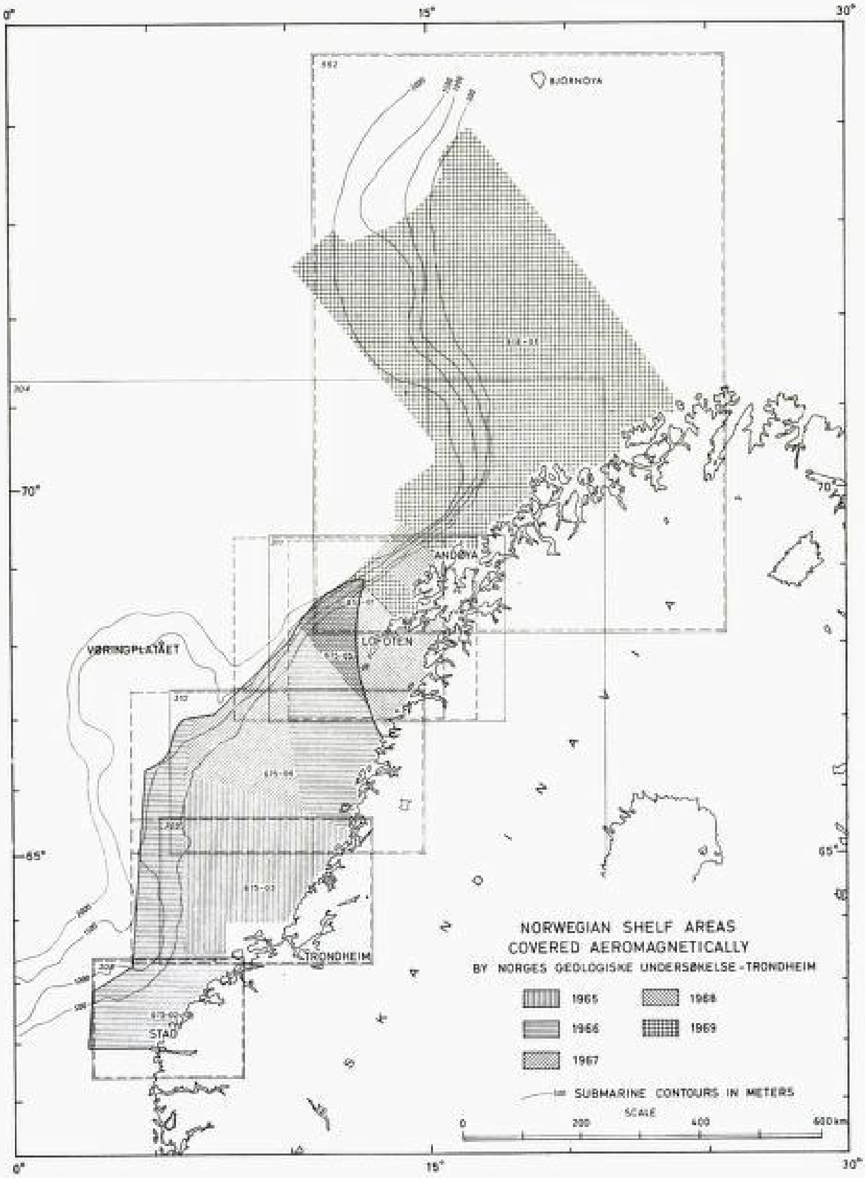

Fig. 1. Location map.

nels along which the Scandinavian landmass was supposed to have been uplifted during Tertiary times (Holtedahl, 1960b). The results of a seismic refraction profile (Ewing and Ewing, 1959, Profile F-8) definitely settled the speculations by showing a sedimentary sequence of at least 4.5 km overlying a doubtful basement. Later (Manum, 1966), Upper Cretaceous sediments were dredged from the continental rise west of Andøya (69° 30′N, 15° 40′E).

In 1963 10 profiles were flown normal to the coast between Stad and Lofoten (62—68°N). The profiles showed a definite change in magnetic pattern when passing from land to sea, and a rough interpretation in terms of depth to magnetic basement gave an elongated basin lying parallel to the coast and with maximum depths exceeding 7 km. With these encouraging results the survey was extended further to the north to include 20 profiles between Lofoten and Senja (68—69°30′N) in 1964.

After these reconnaissance flights, detailed surveying was started in 1965 and is still going on in 1970. The measurements have been conducted by the Institutions section for airborne measurements headed by Henrik Håbrekke. In 1965—66 the measurements were supported financially by NAVF (The Norwegian Research Council for Science and the Humanities). From 1967 surveying has been financed by NTNF (The Royal Norwegian Council for Scientific and Industrial Research). The areas flown until now, covered by approximately 85 000 line kilometers, are shown in Fig. 1. The area with which we shall be dealing in this paper is situated between Stad and Lofoten and is indicated by solid outline .

Acquisition and presentation of data.

The intensity in magnetic total field was measured with an Elsec 592 proton magnetometer with potentiometric recorder installed in an aeroplane. The Elsec proton magnetometer reads the total magnetic field, recycling every 1.7 sec, with an accuracy of ±1 gamma.

For navigation purposes two Loran A receivers (Japanese LR 700) were installed in the aircraft and tuned to two different Loran A transmitter combinations on the Norwegian coast and at Jan Mayen. A fiducial marker on the potentiometric recorder was actuated every time a perpendicular Loran A-lane was crossed. Crosspoints were obtained from the second Loran A receiver tuned to the Loran A transmitter combination producing the perpendicular lanes. The accuracy of navigation is set to «better that 500 meters». By correlating Loran A navigation with drafts in the coastal areas, discrepancies were found to be less that 200 meters.

The profiles measured are identical to Loran A-lanes running approximately north-south in the actual area. The distance between profiles was chosen equal to the distance between lanes displaced by 20 microsec, i.e. 4—5 km. Tie-lines, with an average spacing of approx. 50 km, were flown perpendicular to the ordinary lines in order to compensate for diurnal variations. In addition the total field was registered continuously throughout the survey period by means of a stationary magnetometer in Trondheim. The corrected field values are estimated to be accurate within limits ±5 gammas.

The corrected field values were reduced to 1965-values and four isogam maps on scale 1 : 350 000 with contour interval 20 gammas were constructed. In order to remove a normal and present the anomalous field the International Geomagnetic Reference Field (Fabiano and Peddie, 1969) was subtracted. The residual magnetic field was contoured and redrawn on a 1 : 1 500 000 scale, the result of which is presented here.

Interpretation.

A magnetic isogam map mainly reflects the distribution of magnetite in the ground. As this distribution is principally governed by igneous and metamorphic processes, the magnetic map reveals gross geological features. For instance, magnetic trends and displacement of anomaly axes may reflect tectonic lines and directions.

As only shelf areas are covered by this survey, a regional picture (U.S. Naval Oceanographic Office, 1967) is presented in Fig. 2. From this map it is possible to distinguish between five magnetically different regions.

1. Strong and linear oceanic anomalies.
2. Strong, irregular, and narrow anomalies over Shetland (61°N, 1°W) and to some extent Vøringplatået (67—68°N).
3. Strong, irregular, and broad anomalies along the coast south of Trondheim.

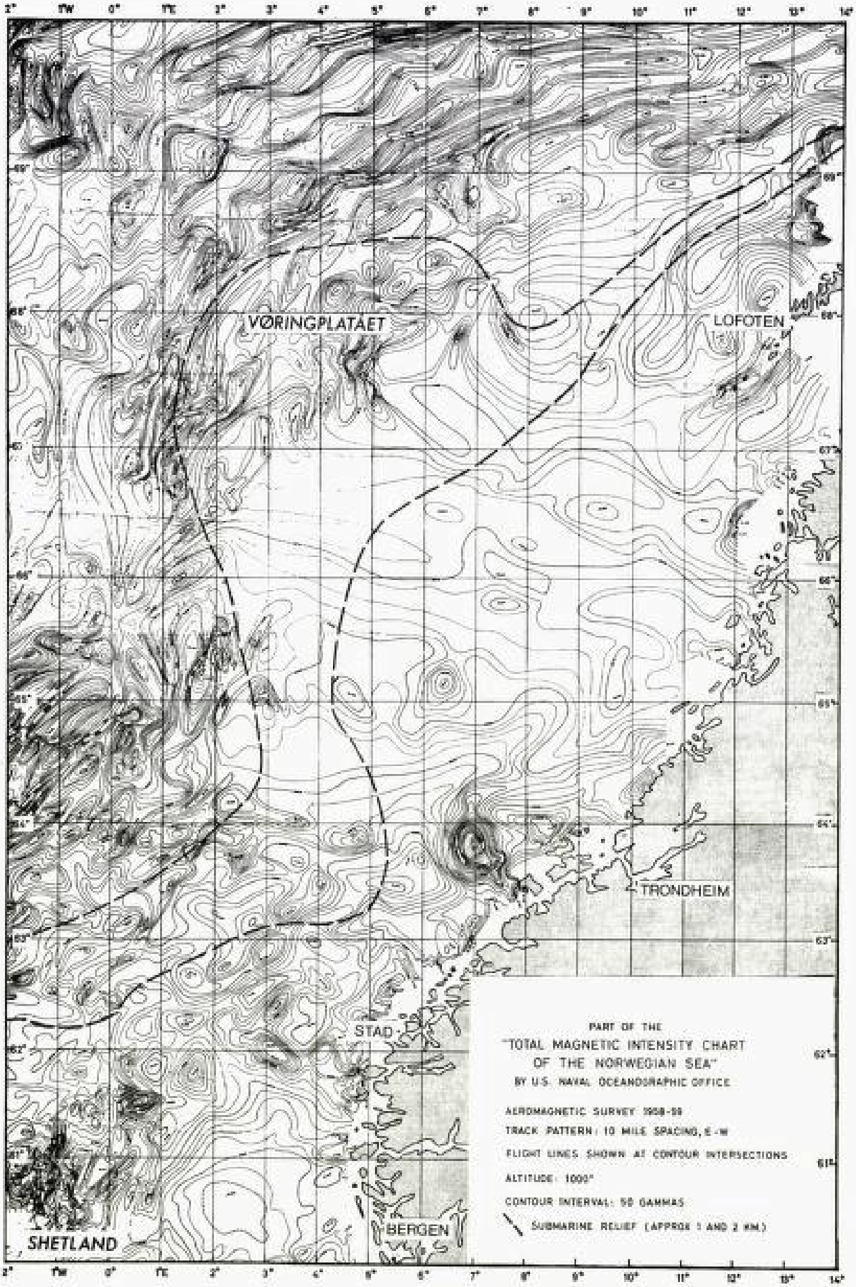

Fig. 2. Aeromagnetic map showing regional features. Reproduction permitted by the U.S. Naval Oceanographic Office.

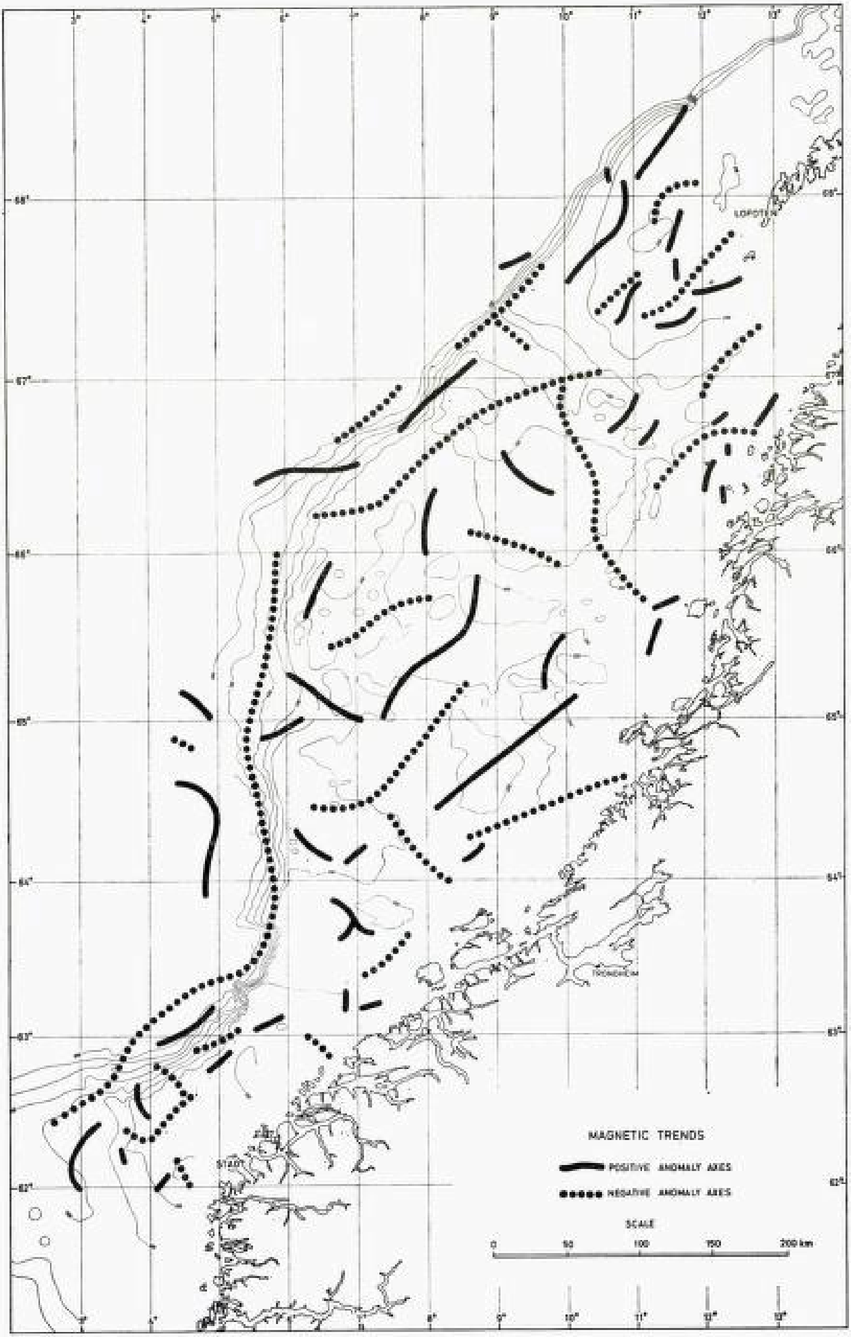

Fig. 3. Map showing magnetic trends.

4. A weak and featureless picture covering the shelf between Trondheim and Lofoten.
5. Very large anomalies in the Lofoten area.

The detailed magnetic isogam map (under separate cover at the end of this book) shows several interesting features.

Magnetics trends.

A close inspection of the map reveals several magnetic directions. For those not familiar with aeromagnetic maps a separate figure containing all the visible directions has been drawn (Fig. 3). The negative axis along the continental edge is peculiar and may lead to comparison with similar anomalies at or near continental slopes in other parts of the world (Taylor et. al., 1968). A direct comparison is, however, not valid as the anomaly in our case is negative, weak, and irregular as opposed to a regular, positive anomaly of several hundred gammas along the eastern continental margin of the United States.

Lofoten.

The Lofoten island group consists of granulite facies charnockitic rocks (Strand, 1960, p. 261). A gravimetric Bouger anomaly of 130 mgals is associated with this province (Norges Geografiske Oppmåling, 1963), and from the isogam map it can be seen that a large aeromagnetic anomaly also occurs over the Lofoten massif. This anomaly clearly shows that the Lofoten granulites continue towards southwest. The similar magnetic anomaly to the west indicates another mass of charnockitic rocks lying parallel to Lofoten and having approximately the same size. A major NW-trending break is indicated in both topography and magnetometry along Trændjupet (67°N, 10°), cutting the Lofoten anomalies to the south. The westernmost mass seems however, to continue on the southern side of the break. A gravimetric Bouger anomaly of 115 mgals (Raw data from Service Central Hydrographique, 1968) associated with the magnetic anomaly strongly supports this interpretation.

Marginal channels.

Another feature is the change in magnetic pattern from narrow to broader anomalies at or near the marginal channels parallelling the coastline. Olaf Holtedahl has drawn attention to this feature and its possible significance in a number of publications, a summary of which will be found in Holtedahl (1960b). Igneous activity is indicated by the small, but distinct, magnetic anomalies along these channels, especially north of 64°N. The aeromagnetic profiles have been thoroughly searched for signs of eventual displacements, but no conclusive evidence has been found. More detailed work is required to prove or disprove the existence of faulting along these depressions. In fact, a limited area has already been measured in detail and is now being interpreted.

Frøyabanken.

The very strong and large anomaly at Frøyabanken (63°50'N, 7°E) deserves some discussion. A study of the analog records shows that it is built up by the growing together of several narrow anomalies having irregular orientations. Also associated with the anomly is a gravity high of more than 60 mgals (Ole Bedsted Andersen, personal communication, 1968).

The most likely explanation for this large magnetic anomaly is that it is due to a magnetic equivalent to the Seiland gabbroic province (Strand, 1960, p. 275) situated at 4—6 km depth. Over the Seiland province there are strong and irregular magnetic anomalies mainly due to ultrabasic layers and lenses in the gabbro, which is itself practically nonmagnetic (Norges geologiske undersøkelse, 1966).

A detailed aeromagnetic survey of the anomaly at Frøyabanken has been conducted and a thorough interpretation is now being done.

Magnetic basement.

The most important result of aeromagnetic surveys over sedimentary basins is the determination of depth to magnetic basement. Significant magnetic anomalies are almost exclusively due to magnetization contrasts in the basement, and less frequently to igneous rocks within the sedimentary column. The form of these anomalies is related to the depth to the top of the magnetic masses. Hence, by studying the form of the anomalies it is possible to determine depths to the magnetic basement which is the same as the thickness of the sedimentary cover. The results are critically examined before doing a generalized contouring of the basement surface, giving less weight to uncertain determinations and values differing to much from the others.

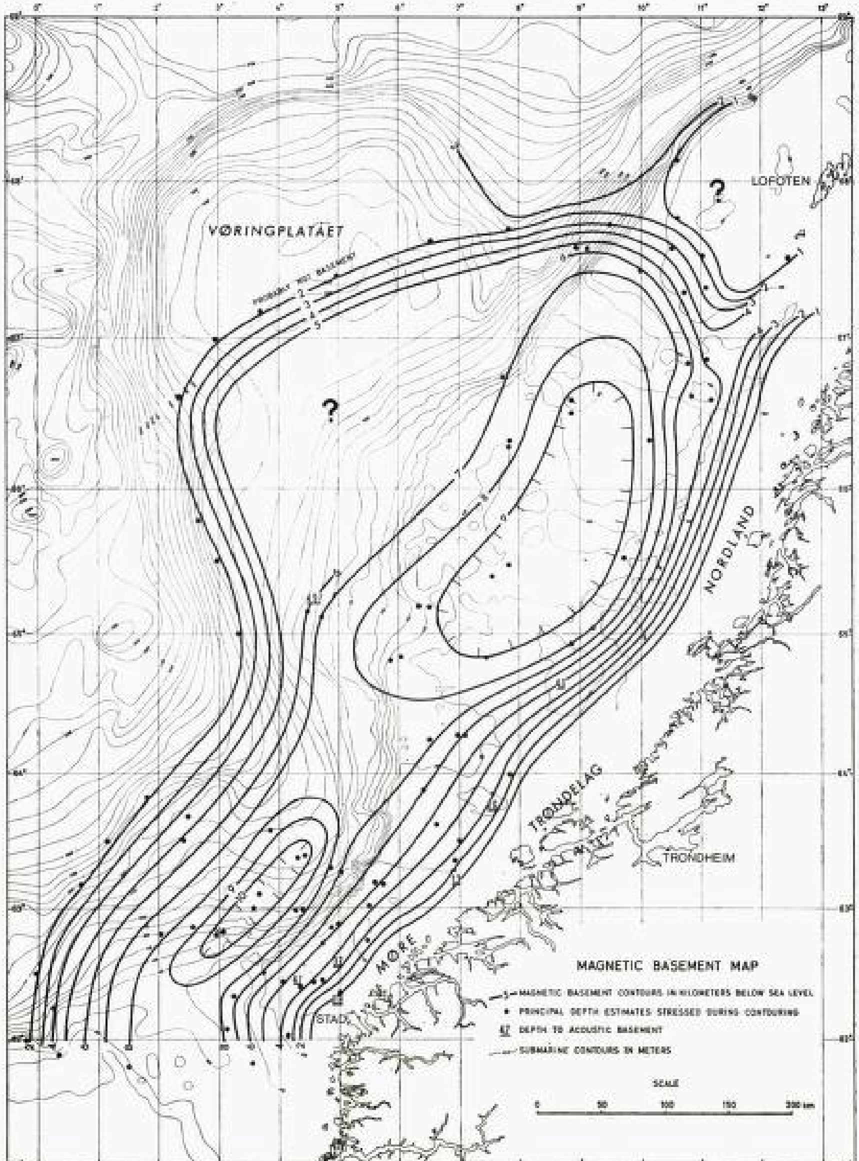

Fig. 4. Final interpretation map. Seismic data from Ewing and Ewing, 1959, Kvale et. al., 1966, and Sellevoll et. al., 1967. Submarine relief after Eggvin et. al., 1963.

In the present case the interpretation has been done directly on the analog records by graphical methods which are to be published in near future (Åm, in preparation). Of already published methods the one of Bean (1966) approaches most closely to those used here.

To get a complete picture of the basin the «Total Magnetic Intensity Chart of the Norwegian Sea» (Fig. 2) and the accompanying microfilm records have been interpreted in the areas not covered by the present survey (An interpretation of this map has already been published (Avery et. al., 1968), but that publication is not concerned with the shelf areas or with any kind of depth determinations). The final interpretation is given in Fig. 4, where the numerous depth determinations inside the 1 km contour have not been plotted.

This interpretation shows the existence of a large sedimentary basin situated parallel to the coast with its axis 120—150 km from the coastline. Along Nordland (65—67°N) the axis is situated not far from the center of the shelf with maximum depths exceeding 9 km. A culmination (7—8 km depths) is indicated outside Trøndelag (64°N). Outside Møre (63°N) the basin deepens again reaching depths of more than 10 km with the axis on the continental rise about 60 km off the shelf.

The central part of Vøringplatået (67—68°N, 4—7°E) is characterized by narrow negative anomalies assumed to be of shallow volcanic origin, masking completely the basement anomalies. In spite of this the depths to these shallow sources have been used when contouring as if they referred to basement. The inner half of Vøringplatået is similar in magnetic pattern to the shelf area inside. The magnetic basement reaches depths of about 6 km along the continental edge and seems to be at least 5 km deeep on the inner part of Vøringplatået.

The basement contours at Haltenbanken (64°40'N) represent a compromise. The large magnetic anomaly has a causative body at 5—8 km depth, while the small anomalies inside are caused by shallow bodies. These anomalies are evidently not due to the same magnetic marker. The shallow anomalies have been assumed to represent basement, and only depth determinations from the outer flank of the large anomaly have been considered when contouring. It is perhaps more likely to assume the shallow effects to be due to igneous dykes in the sediments. In this case the basement contours should be drawn 30 km closer to the coast in the actual area.

Discussion.

The fitting together of the Northern Continents in their predrift positions (e.g. Bullard et.al., 1965), does not allow for the existence of Vøringplatået which has, nevertheless, been considered a subsided part of the shelf by several authors (Demenitskaya and Dibner, 1966). As can be seen from Fig. 4 both the continental rise outside Møre (63°N) and at least the inner part of Vøringplatået should be considered of continental origin and allowed for when trying to fit the continents together. Concerning the central part of Vøringplatået, Johnson et. al. (1968) have given seismic evidence for a buried ridge probably related to volcanism. This interpretation is supported by the magnetic picture.

An earlier publication (Ivanov, 1967) indicating depths to magnetic basement exceeding 20 km deserves a closer examination. The depth estimates are based on the assumption that the magnetic bodies can be treated as lines of poles (Ivanov, 1967, p. 367). This is an unrealistic simplification leading to depths being far too great. Another factor of similar influence is the lack of knowledge about magnetic strike directions in the case of widely spaced profiles.

The seismic refraction profile of Ewing and Ewing (1959, Profile F-8), will be found in Fig. 4 at 65°15′N, 04°35′E. It is a one way shot with an apparent basement velocity of 4.1 km/sec. This low velocity may be due either to basement not having been reached or to downdip shooting. In both cases the depth tends to increase, thus giving a better agreement with the magnetic depth.

The publications of Kvale et. al. (1966) and Sellevoll et. al. (1967) show depths to acoustic basement (5.3 km/sec) which are in striking agreement with the depths to magnetic basement (Fig. 4). The only discrepancy is found at Haltenbanken (64°40′N, 8°50′E). The reason for this disagreement may be sought in the seismic depth determination (one way shot), in the magnetic determination, or in the fact that the two methods are not referring to the same physical parameters. Of these reasons the last is considered the most probable. It can be separated into two related possibilities:

1. The large magnetic anomaly at Haltenbanken may originate from far below the top of the basement.
2. The acoustic basement may be high velocity sediments, e.g. Devonian sandstones and conglomerates occurring in the costal areas inside

(Holtedahl, 1960a), the shallow magnetic effects inside the marginal channels being due to basic dykes in the sediments.

To be able to choose between these two alternatives additional work is required. The second alternative is, however, favoured due to the change in magnetic pattern at 64°20'N, 10°E.

Acknowledgements.

I thank Inge Aalstad, Director of Norges geologiske undersøkelse, Geofysisk avdeling for allowing me to publish the results of a large amount of work to which my contribution has been of minor importance.

The U.S. Naval Oceanographic Office is thanked for the permission to reproduce part of the «Total Magnetic Intensity Chart of the Norwegian Sea» and to publish interpretations made on that map and the accompanying microfilm records.

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NB! Magnetic isogam map under separate cover at the end of this yearbook.

MORAINES IN THE HEMNEFJORD AREA, WESTERN NORWAY

By
N. P. Lasca[1])

Abstract.

Four radiocarbon dates are used to date moraines in the inner Hemnefjord area. The radiometric ages of the moraines range from approximately 10,720 to 11,310 years B.P. and probably correlate with the Ra moraines in Southern Norway and the Tromsø—Lyngen and Skardmunken moraines in Northern Norway. Evidence suggests that by 11,300 years B.P. the Hemnefjord war partially open to the sea. The Hemne substage's ice either extended to the coast on both sides of an open fjord system, or local glaciers persisted in the outer fjord when the Hemne substage's ice was in the inner Hemnefjord.

Introduction.

The Hemnefjord area (63° 17′ N. lat., 9° 10′ E. long.) is situated at the southwestern boundary of Sør-Trøndelag, Norway (see Figure 1). The coastal mountains and off-shore islands, including Hitra and Frøya, of the area were extensively eroded by Pleistocene glaciers which exposed large areas of bedrock. The bedrock in the area was mapped by Ramberg (1966, 1968) and consists primarily of various types of gneiss and schist, amphibolite, occasional meta-diabase and dolerite, and rarely mylonite and marble. The Quaternary and Holocene deposits are generally restricted to the main valleys and low-lying areas. The deposits consist primarily of late-glacial and post-glacial marine clays, outwash sediments and morainal debris.

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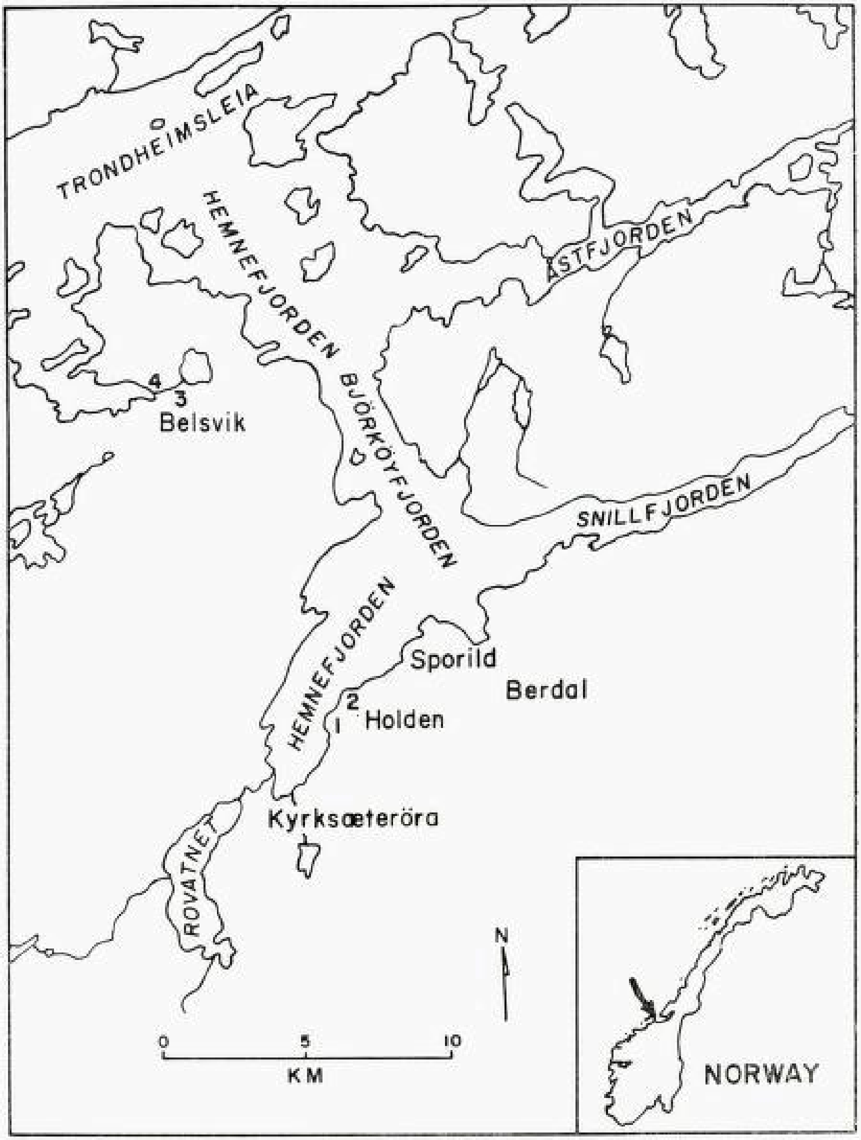

Figure 1. Map of the Hemnefjord area, West Norway. Numbers indicate dated shell localities. See Table 1.

Previous work.

There is no detailed work on the Quaternary deposits of the Hemnefjord area, although extensive work on the marine shorelines was done by Undås (1942) and the positions of various moraines in the Hemnefjord area were described (Undås, 1935). Although no absolute dates were available Undås suggested several alternative positions for the moraines in the Hemnefjord area correlative with the Ra moraines of Southern Norway:

1. based on strandline studies and the size of the moraines, there are Ra-Salpausselkä moraines in the inner Hemnefjord (Undås, 1935).
2. based on «. . . a reliable marine limit and surf limit . . . the Ra-moraine lies outside Krokstadaune [inner Snillfjord]. The moraine may there be situated either in Hemnefjord or at the mouths of the tributary valleys, where there are great local moraines (Undås, 1942).»
3. based on the marine limit, the distal edge of the Ra-moraine lies outside (seaward) of the islands in Møre and Trøndelag (Undås, 1963).

Undås (1942) investigated the shorelines and terraces in the Hemnefjord area, while Øyen (1910, 1914) studied the marine molluskan faunas from the clays in the Trondheimsfjord region which included Hemnefjord. Rekstad (1922) in a summary of the isostatic adjustment of the Scandinavian peninsula, indicated an adjustment of about 100 m on South-Central Hitra and greater than 125 m in the inner Hemnefjord area (Figure 1).

Present work.

The present work was begun in 1965 for the expressed purpose of dating the moraines in the Hemnefjord area. A reconnaissance study of the surficial geology was completed in 1968. In general, the moraines shown on Holtedahl and Andersen's glacial map of Norway (Andersen, 1965, p. 94—95) are correctly located. It is suggested that the Hemne moraines [Kyrksæterøra (Hemne), the outer moraines at Holden, Sporild and Berdal] of the inner Hemnefjord were deposited during the Hemne substage, and that (1) either local glaciers present in the outer Hemnefjord area formed local moraines, or (2) that Hemne substage glaciers extended to the outer coast, but the Hemnefjord was open to the sea. The Hemne moraines usually consist of two parallel ridges, the outer of which is large

Table 1:

Radiocarbon dated shells from the Hemnefjord area, Sør-Trøndelag, Norway. Shells were dated by the Radiological Dating Laboratory, Norwegian Institute of Technology (NTH), Trondheim, Norway.

Loc. No. and NTH Lab. No.	*Loc.*	*Field Alt.*	*Species*	*C-14 age yrs. B.P.*
Loc. 1 T-551	Ice-contact delta c. 1 km south of Holden, Sør-Trøndelag.	21 m	*Pecten islandicus* (Müller) *Astarte elliptica* (Brown) *Cyprina* sp. cf. *c. islandica* (Linné) *Saxicava* sp. cf. *s. arctica* (Linné)	11,310 ± 140
Loc. 2 T-552	Ice-contact delta c. 1 km north of Holden, Sør-Trøndelag, near smelting plant.	20 m	*Pecten islandicus* (Müller) *Astarte elliptica* (Brown) *Macoma calcarea* (Chemnitz) *Acmaea* sp. cf. *A. testudinalis* (Müller)	11,300 ± 150
Loc. 3 T-553	Ice-contact deposit Belsvik, ½ km west of southwest corner of Heimvatn.	46 m	*Mya truncata* Linné *Pecten islandicus* (Müller) *Astarte elliptica* (Brown) *Macoma calcarea* (Chemnitz) *Saxicava* sp. cf. *S. arctica* (Linné) *Liocyma* sp. cf. *L. fluctuosa* (Gould)	10,720 ± 110
Loc. 4 T-554	Glacial-marine clay near Belsvik. 1 km west of the southwest corner of Heimvatn.	15 m	*Pecten islandicus* (Müller) *Mya truncata* Linné *Mytilus edulis* Linné *Macoma* sp. cf. *M. calcarea* (Chemnitz) *Astarte* sp. cf. *A. elliptica* (Brown)	11,250 ± 180

and distinct. There were no absolute dates on the age of any moraines in the Hemnefjord area until 1965-66 when several shell localities were found that related to moraine sequences. In particular, *in situ* shells were found in ice-front delta deposits which are in contact with and directly in front of a moraine 1 km south of Holden, at Holla, and near the smelting plant at Holden (Table 1). At Holla, the delta beds abut the distal side of the moraine and are disected by a post-glacial river where the shells are exposed (see Figure 2). At Holden, the shell-bearing beds were exposed during construction of the smelting plant. Additional shell material was found at Belsvik in near-ice deposits and in glacial-marine clays (Table 1).

The marine shell material was dated by the Radiological Dating Laboratory of the Norwegian Institute of Technology (NTH), Trondheim, Norway, using a radiocarbon half-life of 5570 years. The equipment and counting techniques used by the laboratory are described by Nydal and Sigmond (1957), Nydal (1959, 1962, 1968), and Nydal and others (1964). The localities, species, altitudes and dates are summarized in Table 1. Plotted altitudes are field altitudes taken with Paulin aneroid and are not corrected for eustatic rise in sea level. Accuracy for altitudes between 5 and 25 m is estimated to be ± 1 m, and above 25 m, ± 2 m.

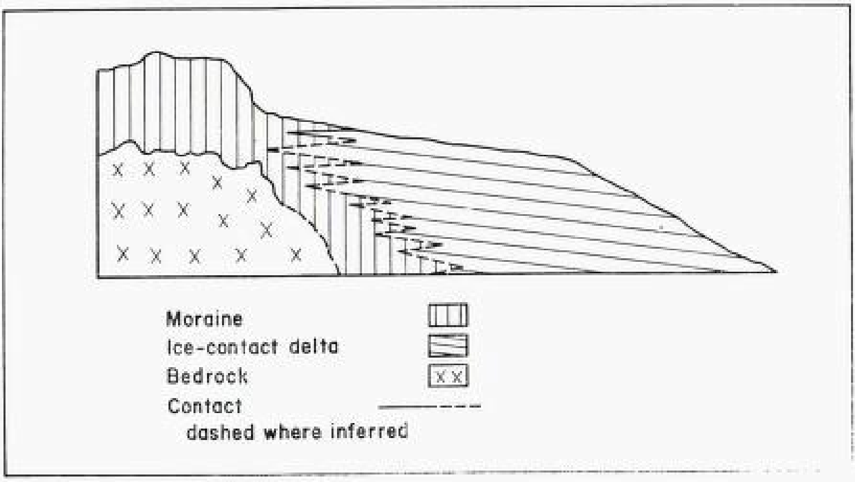

Figure 2. Diagramatic cross-section illustrating the relationship between the moraine and ice-contact delta deposits in which in situ *shell material was found and dated. See Table 1.*

Correlation and Conclusions.

Holtedahl (1928) investigated the ice-front and moraine deposits in the Trondheimsfjord district and noted two glacial substages, the Ørlandet substage and the Tautra substage. In the Trondheimsfjord district these are an outer and inner moraine system respectively. Andersen (1965)suggested that the Tautra substage probably corresponded to the Ra-Salpausselkä substage. Similarly, in the Hemnefjord area there are two distinct moraine systems. It seems likely that the inner moraine system, the Hemne moraines, correlates with the Tautra moraine system and that Andersen's suggestion is correct.

The following conclusions are presented: (1) the marine shell material in the Hemnefjord area is suggestive of a cold (arctic and boreal waters) sea and was probably deposited in front of retreating ice. (2) The marine character of the fossils at Holden, indicates that Hemnefjord–Bjørkøyfjord–Hemnefjord was at least partially open to the sea by about 11,300 B.P. (3) At least 21 m of emergence occurred since 11,300 B.P. It is tentatively suggested that emergence is related almost entirely to isostatic adjustment due to glacial unloading. Emergence probably began prior to 11,300 B.P. for three reasons: higher sea levels are indicated by delta and beach deposits to altitudes of 136 m; undated shell material is found at altitudes up to 88 m; and, some adjustment probably occurred as the ice thinned prior to encroachment of the sea upvalley. (4) Based on the positions and radiometric dates of shell material near Belsvik, it is suggested that either local glaciers persisted in the outer fjord area during the Hemne substage, or that the fjord was open to the sea with ice extending nearly to the sea on both sides of the fjord during the Hemne substage. (5) The Ra moraines in Southern Norway and the Tromsø–Lyngen and Skardmunken moraines of Northern Norway (Andersen, 1965, 1968; Holmes and Andersen, 1964) are large, distinctive moraines. The radiometric ages of the Ra moraines range from approximately 10,000 to 11,200 years B.P., of the two phases (Younger *Dryas* and Older *Dryas*) of the Tromsø–Lyngen moraines from approximately 10,200 to 11,700 years B.P., and of the Skardmunken moraines from approximately 10,390 to 11,500 years B.P. (Andersen, 1965, 1968; Holmes and Andersen, 1964). Considering these dates two interpretations of the Hemne moraines are possible. (A) If the Hemne substage's ice extended nearly to the sea (see 4 above), the radiometric dates of the Hemne moraines range from approximately

10,720 to 11,300 years B.P. and the moraines are correlative with the Ra, Tromsø–Lyngen, and Skardmunken moraines. (B) If local glaciers existed in the outer fjord during the Hemne substage, then the Hemne moraines are late Allerød age (approximately 11,300 years B.P.) and were deposited between the two phases of the Tromsø–Lyngen substage in the north, and before the Ra substage in the south. Therefore, any moraines in the Hemnefjord area correlative with the Ra substage must lie to the east of the Hemne moraines. Further investigations are in progress to clarify the issue.

Acknowledgements.

I wish to thank B. G. Andersen and the Institute of Geology, the University of Oslo, for their help and cooperation in the project. A. Reite, of the Geological Survey of Norway (NGU), kindly read and critiqued the manuscript. The project was funded under a NATO research fellowship and the Graduate School of the University of Wisconsin–Milwaukee through a faculty research grant.

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First manuscript received in April 1969, revised manuscript received and accepted for publication in November 1969.

GLOBULITH
A NEW TYPE OF INTRUSIVE STRUCTURE, EXEMPLIFIED BY METABASIC BODIES IN THE MOSS AREA, SE NORWAY[1])

By

Asger Berthelsen[2])

Abstract.

Intrusion of basic magma into a regionally preheated gneiss complex has given rise to oddly shaped, globular to botryoidal, intrusions with associated contact anatexis and contact deformation whereby pseudoconcordant contacts developed on a local scale. The term globulith is introduced for this unusual type of intrusive structure. It is suggested that osmotic pressure contributed to the special intrusive mechanism of the globuliths. The structures of certain hyperites are compared with those of the Moss globuliths, and the analogy between globuliths and near-surface associations of basic and acid magmas are discussed.

Introduction.

Detailed mapping and structural studies of the Precambrian rocks of the Moss area were started by the author in 1965 in connection with field courses for geology students and as part of NGU's mapping programme in SE Norway (Berthelsen, 1967 a and b).

The predominant rock type in the Moss area (see fig. 1) is a pink, medium- to fine-grained biotite gneiss with conformable coarser migmatitic (? venitic) veins and an overall granitic composition. This rock type builds up a thick and rather monotonous series most probably representing original acid volcanics. These pink gneisses contrast clearly with a series of banded, grey gneisses carrying hornblende in addition

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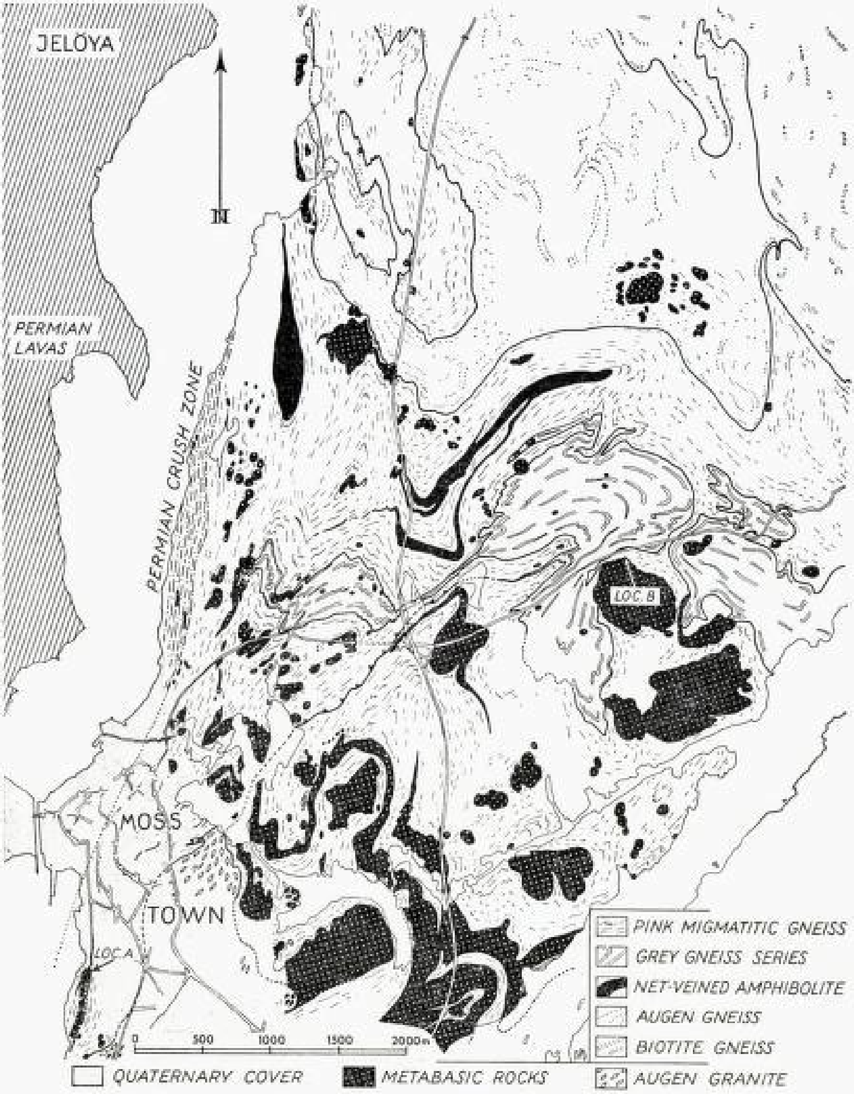

Fig. 1. Geological sketch map of the Moss area. Mapping by students participating in university field courses 1965—1969, L. Skov Andersen, P. Appel, B. Hageskov, K. Secher, M. Ghisler, E. Schou Jensen, I. Madirazza and A. Berthelsen. Compilation by the author. Amphibolites of the grey gneiss series are not differentiated. Metabasic rocks mostly stand for globuliths (with or without superposed deformation) but may include minor amounts of gneiss and other country rocks. Due to scale, Iddefjord pegmatites are not shown. Topography based on photogrammetric maps (1 : 5.000) kindly placed at disposal by Moss kommune.

to biotite and with common thin bands and thicker stratiform sheets of amphibolite. In some places garnetiferous gneiss bands also occur. Both the grey gneiss and the associated amphibolite are equigranular rocks with grain size about 1 mm, and migmatitic veins and schlieren are generally absent. The chemical composition of the grey gneiss proper corresponds of that of an average graywacke and this gneiss may be of sedimentary origin. An intrusive origin (as sills) of at least some of the amphibolitic bands and sheets is suggested by occasional slightly transgressive contacts.

Locally, for example around Lauersbakåsen, net-veined foliated amphibolite of medium to coarse grain and a thin layer of quartzite are complexly interfolded with the pink gneiss (see fig. 1; Berthelsen, 1967 a). Further to the north and the east the common pink gneiss gives way to augen gneisses with rod-like augen, and medium-grained grey, biotite gneisses.

In addition to these rocks, which are all believed to have been derived from original geosynclinal formations of sedimentary, extrusive or shallow-intrusive origin, there occurs a varied suite of basic rocks intruded under plutonic or abyssal to hypabyssal conditions. Their emplacement took place at different times, prekinematically, synkinematically, and postkinematically, in relation to at least two phases of folding.

Some of these basic bodies show such unusual shapes and contact relations that a separate description of their field features is justified.

Globulith, definition of the term.

«So ein Ding muss ich auch haben».

In order to characterise the unusual features of some of the basic bodies of the Moss area, a new term, *Globulith*, is introduced as a supplement to the existing vocabulary for types of intrusive structures. Although Daly (1933, p. 105–110) reserved chonolith as a «sack name» for those discordant igneous bodies which would not be covered by the then existing more specific terms (such as lopolith and laccolith) he also mentioned that once an additional type had been defined, it would automatically be removed from the «chonolith sack». The author considers the field features of some of the basic bodies of the Moss area to be so characteristic and unusual that the adoption of a special term is justified.

In accordance with the German saying quoted above, the author hopes that the new term may serve to direct attention to the various tectonic and petrological problems arising from the recognition of this new type of intrusive structure.

Definition: Globulith is defined as an intrusive body or a group of closely associated bodies of globular or botryoidal shape and with almost concordant contacts resulting from the effects of the intrusion/s on its/their immediate surroundings.

The Moss Globuliths.

The globuliths of the Moss area are made up of metagabbro, metadolerite and fine-grained metabasic types—all of which may grade into amphibolite—intruded into gneisses. The size of the intrusions varies from less than 10 metres to about one kilometre across. Several generations occur. The youngest globuliths were intruded after the last phase of folding affecting the region, but before the formation of pegmatite dyke swarms related to the emplacement of the late Dalslandian Iddefjord/Bohus granite. The older globuliths have preserved their primary structures less well due to superimposed tectonics. The full significance and the peculiarity of the Moss globuliths, therefore, were not realized until, during the field season of 1968, the postkinematic nature of some, i.e. the youngest, bodies was established.

The peculiarity of the Moss globuliths is not only their odd shape, but depend also on their special contact relations. Although the overall patterns of the larger intrusions are clearly discordant to the general country rock structures, their contacts are almost concordant or completely so on outcrop scale.

The pseudo-concordant nature of the contacts of the small as well as the big intrusions leads to the impression that they have become emplaced by forceful intrusion. The composition and corresponding densities of the intrusives and of the invaded rocks rule out the possibility of piercement diapirism, and another easy explanation, boudinage must be dismissed—once the postkinematic nature of some of the globuliths has been established.

The strong effect exerted by the globuliths on their immediate surrounding country rocks comprises not only contact metamorphism and metasomatism but also *contact anatexis* and *contact deformation.*

The term *contact anatexis* is used here in the sense of Rittmann (1967) as a designation for the local, contactbound anatectic phenomena related to the heat given off by a cooling and solidifying magma body. In this way the term is distinguished from regional anatexis which is related to, or is an extension of, regional metamorphism (Winkler, 1968).

Contact deformation refers to contortion, and formation of new textures and structures in the wall rocks and the outermost parts of the intrusion caused by stress differences directly due to the specific intrusive mechanism of the globuliths.

The plastic to fluid style of the structures produced by contact deformation around the Moss globuliths suggests a genetic relation between contact anatexis and contact deformation, i.e. softening of the wall rock through contact anatexis could be regarded as necessary for the development of contact deformation. However, intrusion of basic magma into unconsolidated, water-soaked sediments may also give rise to globulithic structures with contact deformation. An example of this high-level type of globulith is probably to be found in the gabbro intrusions recorded by Bondesen (in press) from the Ketilidian geosynclinal formations of SW Greenland.

Although all the basic rocks of the Moss globuliths have been exposed to regional metamorphism and have become wholy or partly adjusted to amphibolite facies conditions, they usually show well preserved igneous textures except in their most marginal parts. According to primary grain size and texture, metabasalt, metadolerite and metagabbro can be distinguished. In spite of some metamorphic alterations along the margins, chilled contacts are often noticeable. In the metagabbros patches of gabbro pegmatite are common. The mafic minerals of the gabbro pegmatite (now generally uralitic hornblende) often show a characteristic branching growth towards the interior part of the pegmatite.

Several metagabbros also show well preserved igneous banding and lamination. Banding of presumably primary origin has also been noticed in some metadolerites, but is as a rule less evident or less well developed in the medium-grained types.

It is probable that a simple relation exists between the grain size of the basic bodies and the degree of mobilization of the wall rocks. Coarse-grained metagabbro bodies without prominent chilled borders show less contact effects than bodies of finer grain or with prominent chilled margins.

So far, however, no clear relation between the size of an intrusion and its grain size has been established. Even small bodies may consist of extremely coarse (and even banded) metagabbro. A thin zone of amphibolitic migmatite usually separates such bodies from the surrounding gneiss, rendering the contacts more or less conformable.

The intrusion of the Moss globuliths appears to have taken place in a regionally preheated rock complex, i.e. under plutonic conditions with pressures and temperatures (and P_{H_2O}) not much below those required to start regional anatexis.

This assumption is made in order to explain the fact that even globuliths with diameters of less than 10 metres caused partial or complete anatexis of the wall rocks adjacent to the contacts. Some postkinematic globuliths intruded into pink, migmatitic gneiss (of granitic composition) are thus surrounded by a thin shell of porphyric granite of contact-anatectic origin. Porphyric granite also transects the marginal and interior parts of the basic bodies in the form of irregular veins and dykes. Where intensive, the back-veining may have caused a development of veritable intrusion breccias (where the intrusive rock is intruded by the mobilized wall rock). Because of contact deformation, some intrusion breccias have turned into oriented agmatites, as for example SW of Øreåsen. In other cases, the anatectic melt has formed hybrids with the remaining basic magma (e.g. at Dyrevegen at the Moss—Rygge municipal border).

However, not all intrusion breccias, where basic rocks are cut by acid veins, need be due to local contact anatexis. The acid phase may have been generated at an earlier stage and at a deeper level during the ascent of the basic magma, and the anatectic phase may thus have become allochthonous. Such an origin is invoked for those breccias where the quantity of the acid material is great and where signs of local contact anatexis are scarce.

As shown in fig. 2 the basic component may also take a pillowlike shape, where each pillow is distinctly scalloped. This structure appears identical to the basic pillows in granophyres of the Austurhorn intrusion, Iceland, as described by Blake, Elwell, Gibson, Skelhorn and Walker (1965), and it may indicate that basic magma chilled against allochthonous acid magma. In passing, it should be mentioned that intrusion breccias, where strongly tectonised, have lost most of their characteristic features and look like amphibolite-banded grey gneiss. Only when viewed along the axis/lineation can the fragmental shape of the amphibolite «bands» be discerned. These sheared breccias are convergent to the

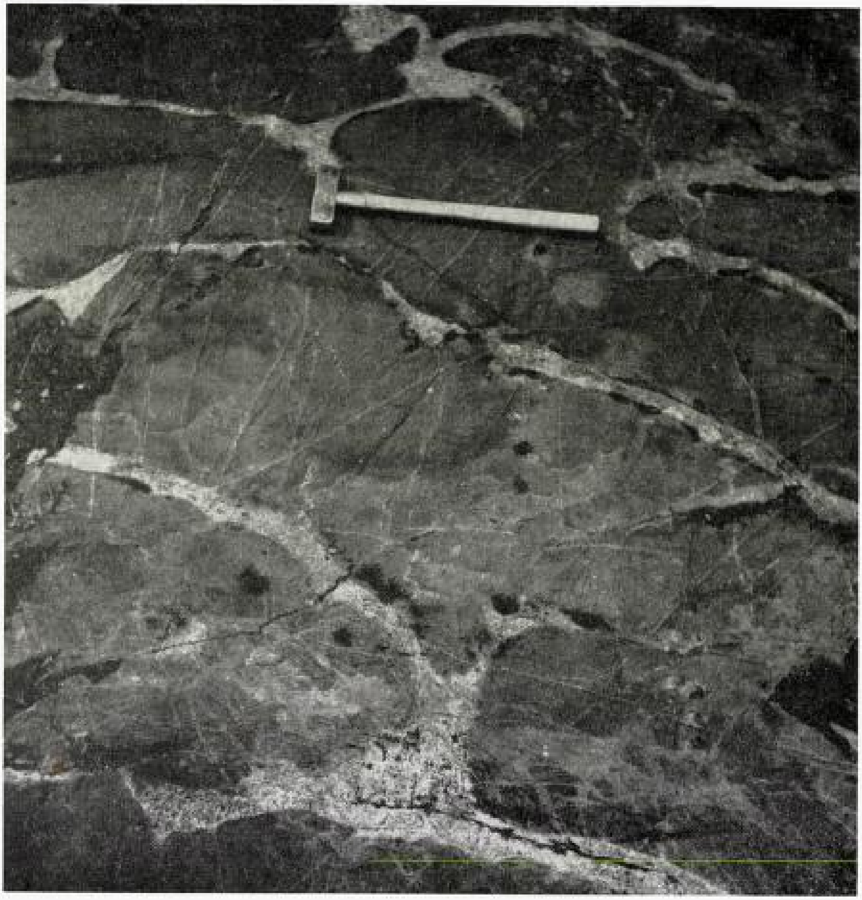

Fig. 2. Pseudopillows in a globulith at Gamlevegen (between Moss and Kambo). The basic «pillows» probably chilled against an anatectic, acid magma; compare Plate 7 a, b, and c and Plate 8 b in Blake et al. (1965).

rocks of the amphibolite-banded, grey gneiss series, which latter, however, is held to be of supracrustal origin.

In order to illustrate some of the typical features of globuliths more clearly, two localities are selected for description:

A) The profile at Klevevegen, southern part of Moss town.

B) The quarry south of Vålervegen, about 4 km east of Moss town.

The Klevevegen profile.

The profile at Klevevegen (see fig. 3) runs almost parallel to a N–S to NNE–SSW trending zone rich in globulithic intrusions of basic rocks. The zone is about 60 metres broad and is overlain and underlain by fairly uniform migmatitic pink gneisses with constant strikes and dips and with axes of small fold and lineations (e.g. rodding in the migmatitic veins) plunging uniformly to the west.

Due to the proximity in the nearby fjord of the Permian fault which borders the Oslo graben to the east, the basement gneisses near the coast show a steeper dip (50–60°) than further inland. Obviously a flexuring of the basement complex prceded or accompanied the Permian faulting which otherwise resulted in the production of a prominent crush zone which enters the coast north of Moss town and is extensively exposed in the coastal hills further north. Between Moss on the mainland and Jelöya a throw of 1500–2000 m has been estimated.

Apart from the monoclinal tilt accompanying this faulting the structures of the migmatitic pink gneisses surrounding the globulithic zone may be ascribed to the last phase of folding affecting the Moss region (the V_2-folding of Berthelsen, 1967 a).

The globulithic zone comprises a large number of individual globuliths of metadolerite grading into finer grained metabasic rocks towards their margins. Along the actual contact foliated amphibolite may also be developed. The structures of the country rock, the usual migmatitic pink gneiss, are much disturbed.

The shape of the separate globuliths may be ball-shaped, ovoid or complex bulbous. Their size varies from 50 to only a few metres. Their contacts are quite sharp, and immediately bordering the contact a thin shell of more or less sheared porphyric granite is generally found. The granite passes gradually into the surrounding gneiss which may carry augen up to a metre from the contact.

In the largest (and northernmost) body in the profile a dyke of porphyric granite cuts with knife-sharp contacts deep into the metabasic rocks. In the southern part of the profile it can be seen how the granite shell at the contact gives off a protrusion cutting obliquely into the basic body.

Another striking feature to be noticed in the profile is the manner in which the structures of the country gneiss have been forced to accommodate the external shape of the individual basic bodies. The otherwise constantly west-plunging axes and lineation have become accentuated through stretching and sweep plastically around the basic globuliths hereby locally attaining vertical to just overturned attitudes. Their deflection is clearly dependent on the bulbous shape of each basic body, and it disappears some distance off the contact.

It could also be argued that the entire structural setting of the globulithic zone should be explained by assuming a tectonic break-up of a once coherent sheet of basic rocks in connection with the last phase of folding whereby the axes and lineation in the «brecciated» zone became oriented according to the local shape of individual lumps of basic rock. Two facts, however oppose this explanation: 1) each globule shows clear signs of chilled contact quite regardless of the contact-methamorphic alterations, and 2) scalloped contacts devoid of axial control.

The porphyric granite of the shells surrounding the globuliths and the cross-cutting veins and dykes may show a sort of flow structure or preferred orientation of the feldspars. The augen occurring in the gneiss just outside the granite shell, however, also show a preferred orientation and this feature points to a close time relation between contact anatexis and contact deformation. In an exposure above and east of the profile, a dyke of porphyric granite cutting metagabbro has been exposed to late shearing together with the surrounding metabasic rock, but as a whole the post-intrusive tectonic influence appears negligible.

It may therefore be concluded that the deflection of the structures of the gneiss may be ascribed to the local pressure exerted by the swelling of the globuliths during their emplacement and we thus have an example of contact deformation of the wall rock around postkinematic intrusions.

An upper age limit for the emplacement of the globuliths at Klevevegen is given by the occurrence of cross-cutting, eastdipping pegmatite dykes belonging to a regional swarm related to the late Dalslandian Iddefjord/Bohus granite which forms extensive exposures about 20 km further to the SSE.

Fig. 3. Field sketch of the Klevevegen profile (loc. A, for location see fig. 1). The profile is about 75 metre long.

Fig. 4. Metabasic globuliths surrounded by wildfolded migmatites in the Vålervegen quarry (loc. B, for location see fig. 1). Drawn on polaroid photos in the field.

Fig. 5. Diagrammatic sketch showing the two types of globulithic structures seen at the localities A and B.

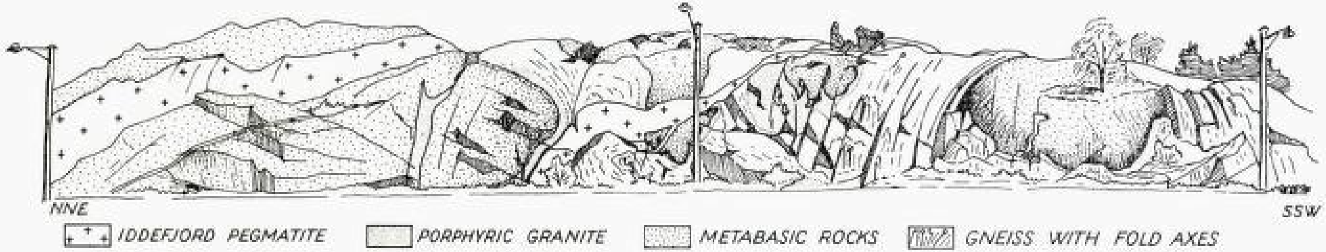

Fig. 3

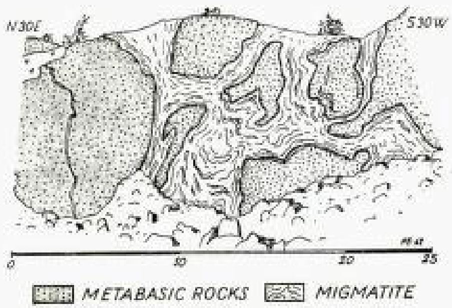

Fig. 4

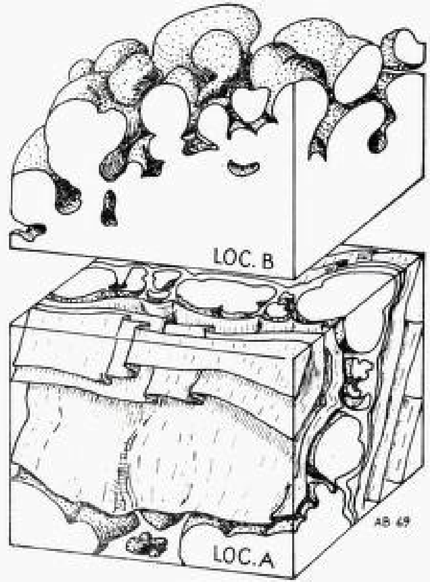

Fig. 5

In this connection it may be noted that the assumption that the globuliths intruded into a regionally preheated country rock (required to explain the anatectic contact effect of the smaller globuliths) is consistent with the updating of all basement rocks of SE Norway during the Dalslandian (Broch, 1964).

The quarry south of Vålervegen.

At the quarry south of Vålervegen, the detailed contact relations of a larger gabbro intrusion, measuring about half a kilometre across, can be studied. The intrusion has been mapped by P. Appel, who demonstrated its general discordant nature and overall semicircular outline on the map, see fig. 1.

As seen from fig. 4, the trend of the contact is, however, in detail highly irregular. The shape of the contact may be compared with that of a cauliflower, circular in general outline, but bulbous or botryoidal when studied in detail. The main gabbro body forms convex protrusions and bulbs each with its own chilled margin. Some of these marginal bulbous bodies appear to be rootless — at least no connection to main gabbro can be discerned even in the good three-dimensional exposures of the quarry.

The marginal bulbs are separated by interlobal space and pockets occupied by mixed, migmatitic gneisses displaying disharmonic folds with haphazardly oriented axes, i.e. true wild migmatites (cf. Berthelsen, Bondesen and Jensen, 1962). These gneisses carry appreciably greater amounts of hornblende and biotite and show a relatively larger grain size than the normal country rock. These features could be explained either by assuming a metasomatic contact effect from the basic magma or by postulating a basification through partial anatexis. Migmatitic veins may also occur in the marginal parts of the basic bodies trending more or lesss parallel to the contact. Such veins probably represent mobilized wall rock material introduced by means of back-veining but arranged in a contact-parallel manner because of the simultaneous contact deformation.

While the gneisses of the interlobal pockets became wildfolded due to irregular and changing compression of the mass between the growing basic bulbs, a flattening and stretching appears to have taken place around (and probably also in the marginal parts of) the convex basic lobes, because here foliated and lineated mixed gneisses and amphibolites developed.

The combined effect of the contact deformation and mobilization was the production of pseudo-concordant contacts. The relic chilled contacts observable in even satelitic bulbs, proves the primary origin (i.e. relation to the intrusive act and the solidification) of these features, which include formation of S- and B-tectonites indistinguishable from true orogenic tectonites in hand specimen.

In some of the larger marginal bulbs and in the central body of the Vålervegen gabbro, igneous banding has been noticed with steep attitudes more or less parallel to the contacts. This banding is primarily brought out by changing grain size. The apparent active pressure exerted by the individual bulbs renders it feasible that this banding signifies multiple intrusive action rather than cumulative crystal settling. The banding is cut by aplitic, acid dykes.

Summary and comparisons.

In the light of the two examples just described (see also fig. 5) it is evident that contact deformation may be extremely difficult to tell apart from superposed true orogenic (i.e. regional) deformations. This means that minor folds observed in and around globuliths should be studied in great detail and should only be used with the greatest precaution — if at all — when a structural analysis of the orogenic structure is attempted.

Since, in some parts of the Moss area, trails of globuliths (e.g. the globulith zone at Klevevegen) form the only «marker horizons» present, the recognition of the primary nature of contact deformation is of great importance for further structural work.

The occurrence of both deep-level (e.g. in the Moss area) and high-level globuliths (e.g. the Ketilidian gabbros) suggests that, during the emplacement of globuliths, the P_{H_2O} of the basic magma (? due to osmosis) underwent a rapid increase, which caused forceful swelling of the solidifying body and eventually squeezing out of not yet solidified magma in the form of satelitic bulbs or protrusions. In this connection it is interesting to note that scalloped contacts suggesting chill of basic magma against contact-anatextic acid melts appear best developed around satelitic bodies.

The recognition of a special globulithic type of intrusion in the Moss

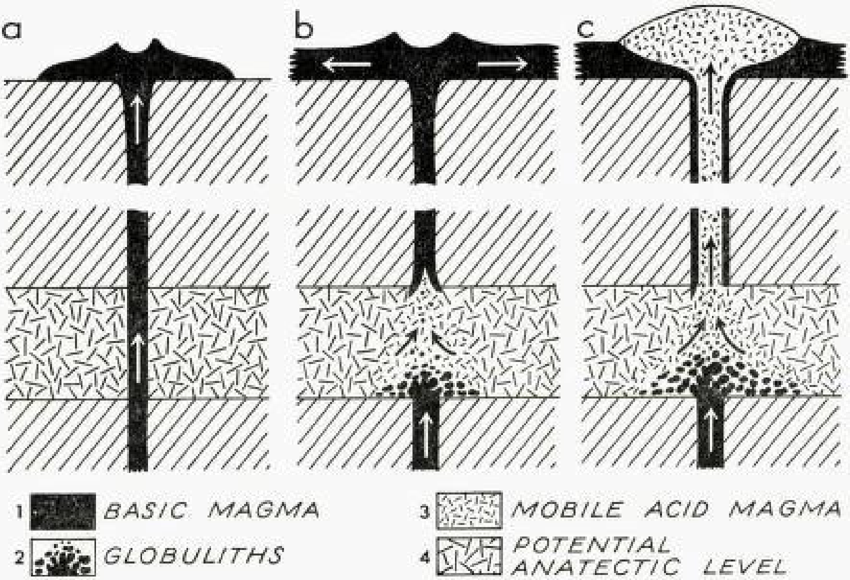

Fig. 6. Diagram showing the possible relation between deep level globulith (as in the Moss type area) and shallow level associations of basic and acid rocks. Redrawn from Blake et al. (1965) with minor changes.

area immediately raises the question whether or not this deep-level type of intrusive structure could be expected to occur in the metamorphic terrains of other regions.

It is most natural to start to look for globuliths among the basic rocks of the Bamle area SW of the Oslo graben, because for other reasons a structural continuity between this and the Moss area is to be expected. The gabbroic rocks of the Bamle area, often referred to as the hyperite group, grade from olivine gabbro and hyperites into schistose amphipolites (Bugge, 1943).

Judging from the rounded to bulbous outlines of the gabbroic bodies of, for example, the Søndeled district in the Bamle area (fig. 3 in Bugge, 1943, or fig. 2 in Barth and Bugge, 1960) and the apparent concordant contacts of the intrusions, it is tempting to suggest that these gabbroic and hyperitic bodies are globuliths. Going through Bugge's pertinent description, the author noted the following paragraph: «It is often difficult to say anything about the mechanism of intrusion, as it is impossible to know for sure to what extent the conformity has appeared during it» (Bugge, 1943, p. 40). Conceivably the idea of contact deformation was latent in Bugges's mind.

The cordierite-anthophyllite rocks occurring around the margins of the basic bodies of the Søndeled district were explained by Bugge by means of a Mg-metasomatism, the Mg having been leached from the amphibolites and gabbroic rocks. However, as recently suggested by Grant (1968), partial melting of common rocks could be a source of cordierite-anthophyllite-bearing assemblages, and since Grant explicitly refers to the Søndeled example, the author ventures to suggest that *contact* anatexis with ultimate restite formation could also explain the relative Mg-enrichment of these rocks.

The intrusive gabbro complex at Rackeby in SW Sweden (Stålhøs, 1958) also shows several features suggesting that at this locality it would be worth while testing the hypothesis formulated on basis of the Moss globuliths.

In this context, Blake, Elwell, Gibson, Skelhorn and Walker's (1965) point of view on intimate association of acid and basic magma fits like a key in the lock — if the Moss globuliths are looked upon as megapillows. The diagrammatic representation by Blake et al. of the possible relationship between, and origin of, a composite acid-basic complex of Austurhorn type is redrawn in fig. 6 with only slight modifications, the most important of which is that «Viscous acid magma» has been changed to «Potentitial anatectic level».

Following these lines of thought the Moss globuliths could be deep level analogues of the near-surface or surface examples of Austurhorn type. Some Moss globuliths, according to this scheme, would have originated in the potential anatectic level 4 of fig. 6, where the acid component will form the marginal or back-veining phase, while others examplify PT conditions corresponding to a somewhat higher level, where allochthonous anatectic material invades already solidified basic magma. The reason why several types representing different levels are represented side by side in one and the same terrain, is amongst others the wide time span covered by the invasion of basic magma.

But need globuliths or bodies suspected to be globuliths always consist of or contain basic rocks? Probably not. The shapes and contacts of several ultramafic bodies, which have usually been described as boudins without closer study, could also indicate that some ultramafic bodies were emplaced as globuliths. In surface exposures above the Kleveveg profile (Fig. 7) small basic globulithic bodies thus simulate the pattern

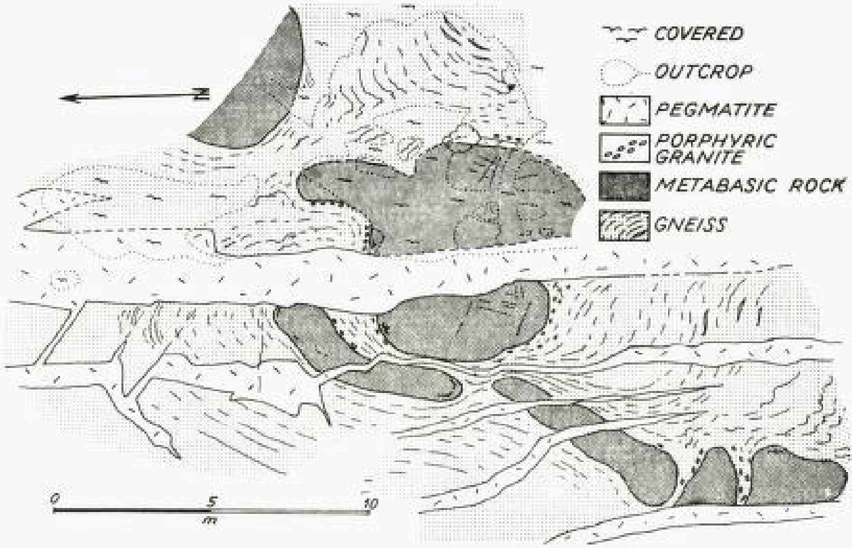

Fig. 7. Sketch map of exposures east of (and above) the Klevevegen profile (loc. A). Note how the small globuliths simulate boudin structures.

shown by great many ultramafic «boudins», and without the additional information obtainable in the vertical profils near by, these small basic globulithic structures no doubt would have passed on as boudins into the author's field notes.

However, returning once more the meaning of the German saying cited above it might in this context be relevant to recall its ironic undertone.

Coining the new term himself, the author ought to be the last to misuse it!

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Abbreviation: NGU = Norges Geologiske Undersøkelse.

Manuscript received and accepted for publication in October 1969.

AN OCCURRENCE OF UNUSUAL MINERALS AT BIDJOVAGGE, NORTHERN NORWAY

By

Carl O. Mathiesen[1])

Abstract.

A complex titanium mineral containing abundant vanadium, chromium, and iron is described. The distribution of minor cerium, yttrium and scandium characterizes the two phases of which the mineral is comprised. As associate minerals occur thortveitite, vanadio-rutile, vanadiochrome spinel, gadolinite, and euxinite. Microscopic, analytic (including microprobe), X-ray powder, and other data are given.

Introduction.

The Bidjovagge copper deposits, Pre-Cambrian in age, lie 40 km NNW of the village of Kautokeino in Finnmark, Northern Norway (Fig. 1), within a belt of predominantly volcanic, albite-rich rocks (Holmsen et al., 1957, Gjelsvik, 1957). These so-called greenstones in the vicinity of Bidjovagge are for the most part banded tuffs with associated diabasic sills. The hydrothermal chalcopyrite-pyrite-pyrrhotite-carbonate ore occurs in brecciated zones along a steep anticlinal limb consisting of albite fels, in part graphitic, and also apparently of tuffitic origin. Near one of the deposits is found a titanium mineral with complex and unusual composition. The investigation of this mineral in addition produced thortveitite, vanadiorutile, vanadiochrome spinel, gadolinite, and euxenite.

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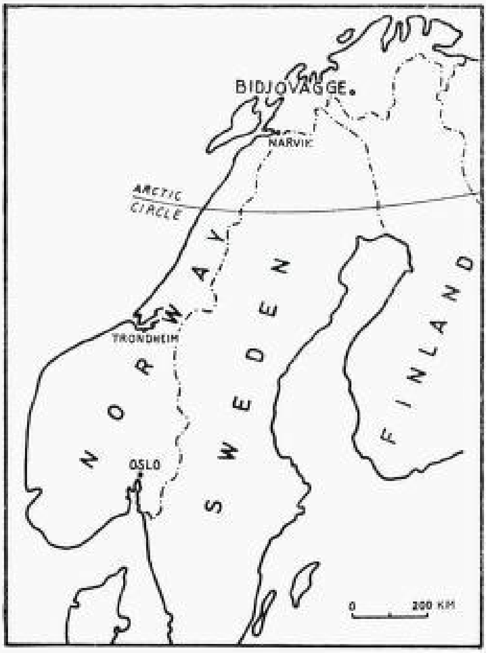

Fig. 1. Location map showing Bidjovagge.

Host rock.

The ore deposit locally known as A, from which the overburden has largely been removed, is confined to the light albite fels. Graphitic albite fels defines the hanging wall and a diabase sill the foot wall. Dividing the deposit in its upper portion lies a wedge of extremely hard, fine-grained, flint-like, light grey to reddish fels similar to the regional contact zone of the fels with the foot wall diabase. At the northern end of this wedge (local coordinates: 80N, 560E) and restricted to a few square metres are found the minerals described in this paper. The Bidjovagge area has been extensively investigated because of its ore and nowhere else there has a similar occurrence yet been observed.

Chemical analysis of the hard-fels (Fig. 2) embracing the minerals to be described shows 19.4 % Al_2O_3, 10.8 % Na_2O, and c. 70 % SiO_2, with several hundreths of a percent each of K_2O, CaO, and MgO, corresponding to 92 % albite. Differential thermal analysis shows 3½—4 % free quartz which occurs mainly as microscopic veins.

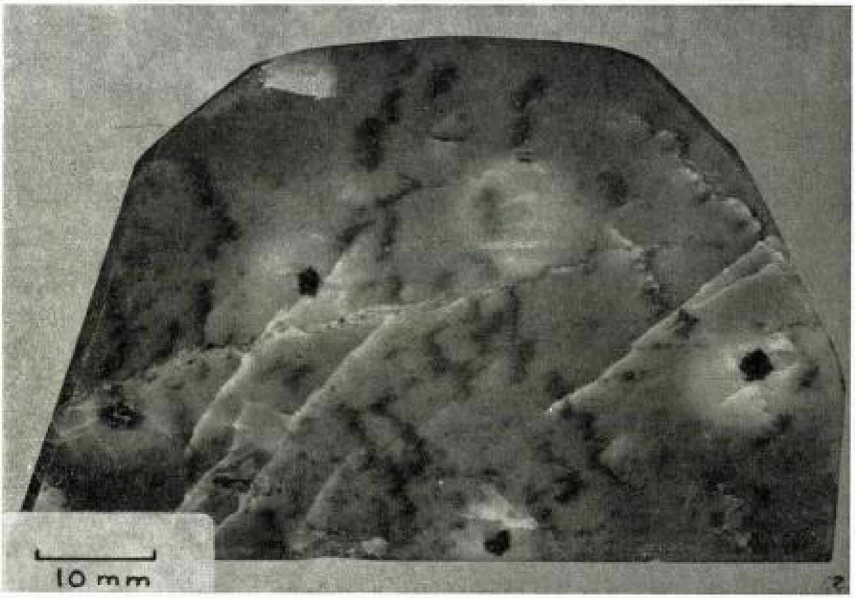

Fig. 2. Complex titanium mineral surrounded by reaction haloes in albite fels. The dark stringers are vanadio-rutile.

Complex Titanium mineral.

Dispersed through the rock at intervals of several centimetres are seen 1–2 mm grains of a black mineral surrounded by reaction haloes (Fig. 2). The mineral has a metallic-adamantine luster, gives a black-brown streak, and is brittle.

In order to obtain a pure fraction a series of steps were performed involving crushing to a few mm, concentration by Carpco magnetic separator, selection of appropriate grains by pincette, grinding, and several stages of separation by Franz magnetic separator and heavy liquids. Table 1 presents an analysis of the mineral, where Ti, V, Cr, and Ce (and the total rare earths) were determined chemically and the remainder spectographically.

Grains of the mineral are seen to often consist of a center portion surrounded by more friable material. Usually these grains are not readily dissectable, but in one case it was possible to remove much of the friable envelopment, revealing a rather well-developed apparent octahedron (Fig. 3). In polished specimens the center portion often appears quadratic. Rarely the friable adherent has grown in such a way as to give the mineral

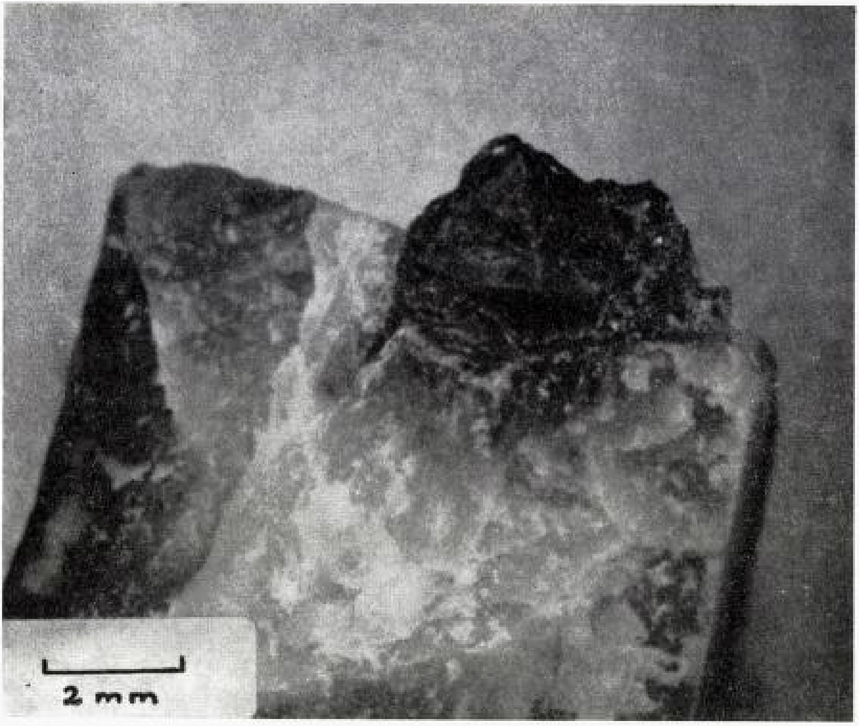

Fig. 3. Complex titanium mineral showing octahedral form.

a crosslike form (Fig. 4). Usually, however, this growth is rudimentary and deeply serrate, and completely surrounds the center portion.

Microscopically it is seen that the center portion is essentially homogeneous, bluish white in plain light, and isotropic. The surrounding material is banded parallel to the edges of the quadratic center and consists of two phases: a predominant phase creamy white in plain light and markedly anisotropic, interbanded with a phase similar to the center portion (Fig. 5). Microhardness and maximum reflectivity measurements vary somewhat because of the banding, but resemble the values for rutile, the isotropic phase being somewhat harder than the anisotropic phase.

Although in well-developed grains the center partion usually appears quadratic in section, the centers of other grains are more irregular as is the banding of the envelopment. In addition to banding parallel to the sides of the quadratic center, oblique banding (apparently twinning)

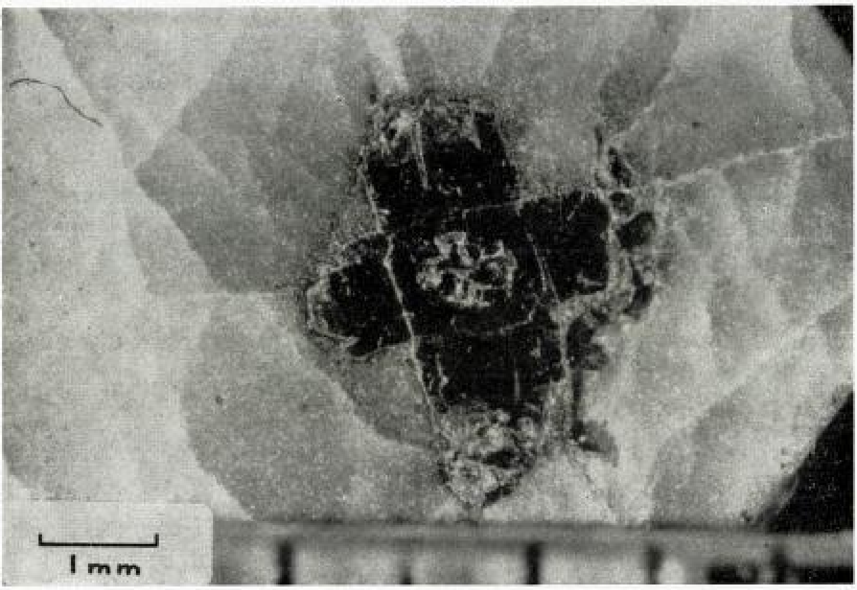

Fig. 4. Complex titanium mineral with cross-like development.

Fig. 5. Polished section of complex titanium mineral in polarized light, oil immersion.

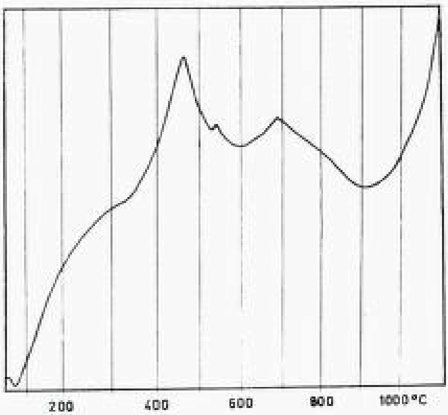

Fig. 6. Differential thermal analysis of complex titanium mineral.

may also be present, especially in the corners. Sometimes also the banding is distorted as seen in the NW and SE quadrants of Fig. 5. The banded portion of the mineral shows long traversal cracks not seen in the center portion, and the individual anisotropic bands show fine traversal cracks that do not extend into neighbouring isotropic bands. The anisotropic phase has thus apparently a greater coefficient of contraction than the isotropic phase, and it must be this phenomenon which causes the friability of the banded envelopment.

Within the mineral some inclusions of secondary rutile, quartz, chalcopyrite, marcasite, and malachite have been found. Leaching has taken place. Along the irregular periphery of the mineral a rim of leucoxene has developed. Secondary open spaces often occur between the mineral and the fels. Within the fels adjacent to the mineral hematite has accumulated.

Separation of the mineral as previously described did not provide narrow limits for specific gravity measurement. Values vary from about 4.0 to 4.2, with increase in magnetic susceptibility parallel to specific gravity.

Differential thermal analysis of the mineral shows exothermal reactions at 470° and 690°C and an endothermal reaction at 900° before sintering begins at 1000° (Fig. 6). The reaction at 470° appears to be due to the contamination of a small amount of sulphides. (Note sulphur content, Table 1.) Polished sections heated to 500° show no apparent change in comparison to unheated sections. At 700°, however, the mineral has recrystallized to a myriad of extremely tiny grains of rutile, though banding and anisotropy are still recognizable. At 900° the banding and anisotropy has completely disappeared and the recrystallization is slightly coarser. Near the periphery some grains of a grey unidentified mineral have also formed.

X-ray data on microscopically selected fractions of the mineral are presented in Table 2. The first column are the measured dÅ values for the banded phase of the mineral and the second column the dÅ values for the central isotropic phase. Subtracting the values of the second column from the first gives the values in column three which presumably represent the anisotropic phase. Identical films were obtained for each of the phases on unheated material and material heated up to 600°. At 700°, however, both phases gave a typical rutile film. (Material heated to 1300° also gave a rutile film.) This transition corresponds to the D.T.A. exothermal reaction at 690° and the observed recrystallization of heated polished specimens. Attempts to obtain X-ray single-crystal data on tiny fragments were not successful.

In order to determine the distribution of various elements in the mineral, investigation by microprobe was undertaken. Several grains were studied with very similar results. The grain described below is shown in Fig. 5. In this case the banding is well developed in one direction and poorly in the other. The probe path runs normal to the banding from near the midpoint of the grain in the quadratic center portion, through the banding, and into the surrounding fels. The distribution of six elements, V, Cr, Fe, Ti, U, and Ce, is shown in Fig. 7. The left side of the figure is a composite of oscilloscope photographs of inverse electron images and shows the probe path, with the mid-point of the grain at the top. Approximately 30 % of the path traverses the isotropic center (0—0.29 mm) before the first band is met. The distribution of V, Cr, Fe, and Ti is given in percent. As standards for Ce and U were not available at the time, the distribution of these elements is relative.

It is seen that the contents of V, Fe, and Ti show no apparent gradient across the mineral, whereas the content of Cr in the central area is approximately twice that of the banded area. None of these elements show correlation with the banding, although the distribution of V is more erratic in the banding. Ce is high in the central area, and in the banded area it slavishly follows the banding, with minimal concentration in the dark (anisotropic) bands and maximal concentration in the lighter (isotropic) bands. (In an inverse electron image the lighter the

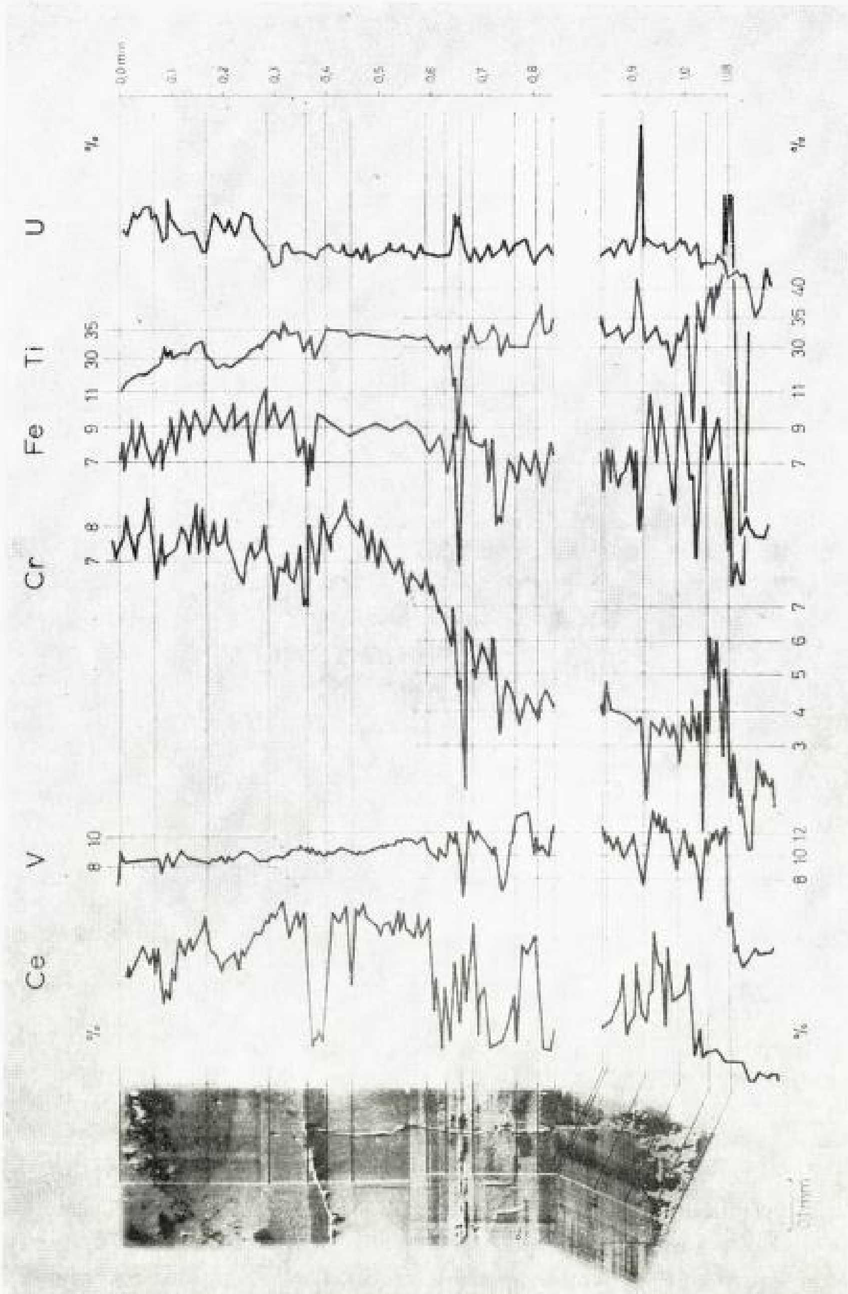

Fig. 7. Distribution of Ce, V, Cr, Fe, Ti, and U in complex titanium mineral as determined by microprobe.

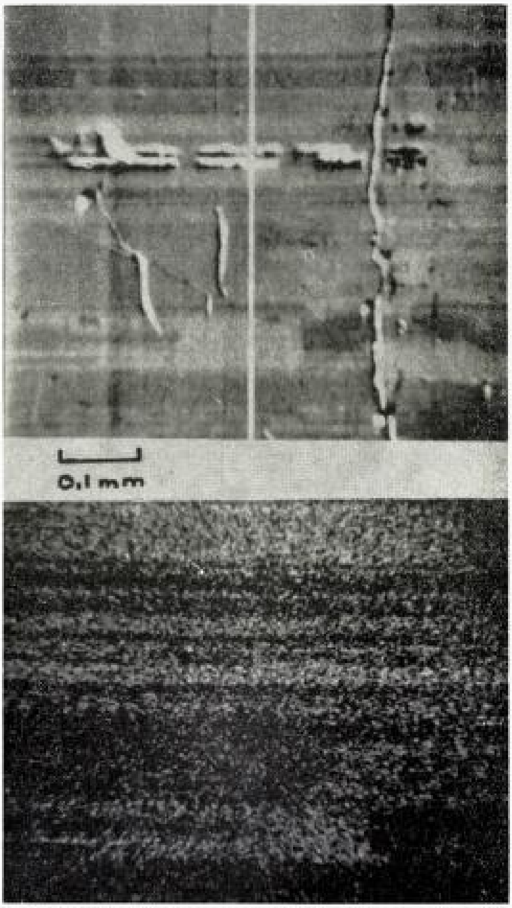

Fig. 8. Distribution of Ce in complex titanium mineral as determined by microprobe. The upper photo is an inverse electron image, the lower a Ce scan.

shade the higher the median atomic number.) Fig. 8 shows an inverse electron image of a portion of the banding (third photo from the top in Fig. 7) and a Ce scan for the same area. The distributions of Y and Sc was determined to closely parallel that of Ce.

In order to obtain quantitative values for Ce, Y, and Sc, standards were prepared using measured quantities of the oxides of these elements in $Li_2B_4O_7$ flux. Microprobe analyses were made at 0.38 and 0.43 mm (Fig. 7), in and outside the banding respectively. Reconnaissance showed these areas to be quite representative of the two phases. The results are

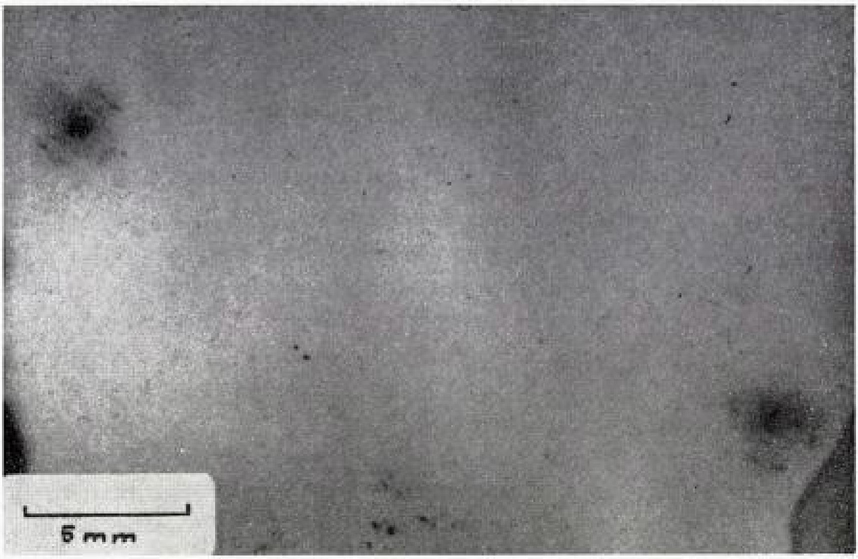

Fig. 9. Auto-radio exposure of polished fels containing two grains of complex titanium mineral.

shown in Table 3, and are compared to the general analysis taken from Table 1. The totals are low, at least in part because not all constituents were measured.

As may be seen, the values for Ce and Sc lie considerably above those obtained in the general analysis. A control spectrographic analysis of a fraction of the original material made at the Institute for Atomic Energy at Kjeller showed 1.6 % Ce, 0.23 % Y, and 0.56 % Sc (0.61 % Sc by neutron activation), and is included in Table 3 as oxides. The values for Ce and Y here obtained compare favorably with the microprobe results. A divergence, however, exists with respect to Sc, especially since Sc is also involved in the thortveitite to be described. It is possible that the content of Sc in the well-developed grains examined is not completely representative for all grains. Although the general analysis, Table 1, shows some Sr, interference prevented determination of the distribution of this element.

U is dispersed in small quantity throughout the mineral and as well occurs as discrete grains. These grains are concentrated mainly at the center of the mineral (small white spots at the top of Fig. 7 not traversed by the probe path), but also between certain bands near the edge of the mineral as at 0.92 mm. (A slow traverse speed was used to achieve the peak at 0.92 mm. Reduced speed at 0.65 mm would likely have produced a similar peak.) Fig. 9 is an auto-radio exposure on X-ray film of a surface containing two grains of the complex titanium mineral. The grain on the left is the same as that shown in Fig. 4. Due apparently to the concentration of U in the centers of grains these centers are often pitted, as seen in Fig. 3 and less clearly in Fig. 5. It was determined that Pb closely follows U of which it is likely a decay product. Th accompanied by a little U is found to be concentrated in the walls of radial cracks near the periphery of the mineral. (The walls of these cracks thus appear white in inverse electron images.) It is apparently because of the effects of U and Th that bleached haloes occur around grains of the mineral, as seen in Fig. 2.

The complex titanium mineral described above defies simple classification. It is evident that the mineral comprises two phases, an isotropic phase enriched in Ce, Y, and Sc, and an anisotropic phase containing less of these elements. The X-ray data obtained was not identifiable for either phase. It is also evident that both phases invert to rutile at about 700°C. A relationship between this mineral, especially the isotropic phase, and davidite (which is also isotropic in reflected light) may be considered. But, aside from a certain resemblance in chemical composition, the accumulated data does not convincingly indicate this to be the case. A relationship to the vanadio-rutile in the surrounding fels, described subsequently, is, however, distinct.

Thortveitite.

While examining by microprobe other grains of the complex titanium mineral for consistency of results, thortveitite (Schetelig, 1922) was discovered. It occurs as a thin discontinuous shell along the periphery of grains of the titanium mineral, and is secondary to the development of its host (Fig. 10). Since this deposition is no more than a few hundreths of a mm in thickness it was not possible to obtain material for X-ray powder determination. Table 4 presents two microprobe analyses (Analysis 1 relates to the zone seen in Fig. 10) and compares them with an analysis from Eptevann, South Norway (Marble and Glass, 1942). It is seen that the rare earth content of the Eptevann thortveitite is higher than that of the Bidjovagge variety in which the cerium lanthanides were not detectable. Vlasov (1964) cites Zr and Hf analyses

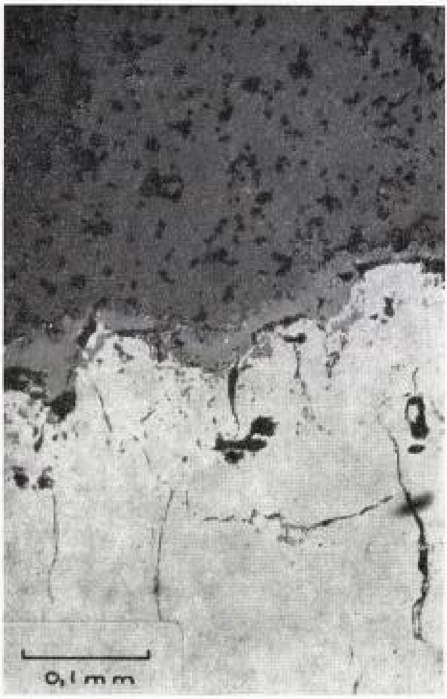

Fig. 10. Polished section of complex titanium mineral with rim of thortveitite.

from Norway and Madagascar of about 1 % for each of these elements. Neither Zr nor Hf were detectable in the Bidjovagge variety. Several tenths of a percent Zr may nevertheless be present, since the presence of Sc and Fe interferes with this determination.

Vanadio-rutile.

Within the fels containing the complex titanium mineral described above are observed in polished specimens sinuous stringers of a finely divided dark mineral (Fig. 2). From a screened fraction of the crushed fels a quantity of material was collected by pincette, which under a binocular showed no trace of the complex mineral or its reaction halo. This material was ground and subject to magnetic separation. The concentrate was finely pulverized and centrifuged in heavy liquid, sp. gr. 4. The heavy fraction, whose specific gravity was determined to be 4.3, gave an X-ray powder film typical of rutile. A combination of ordinary spectrographic and microprobe analyses gave the results shown

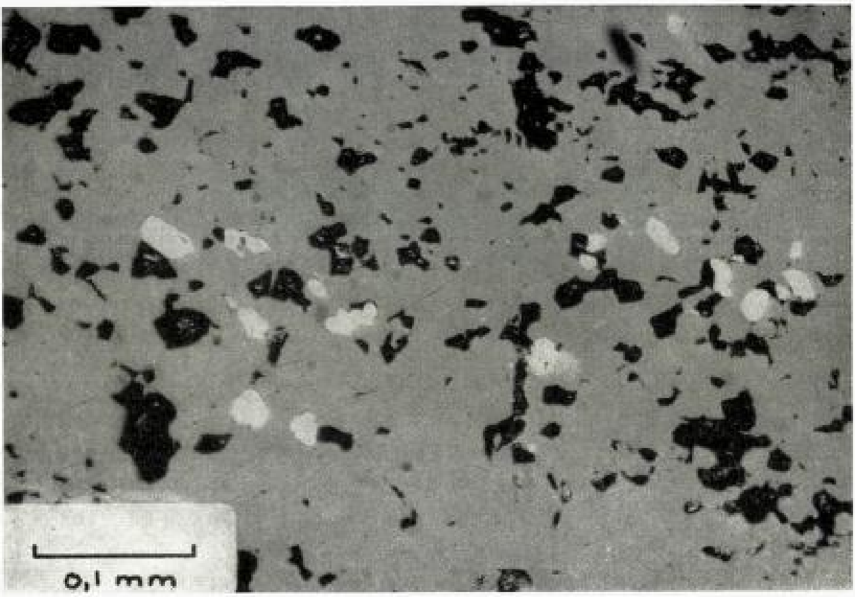

Fig. 11. Polished section of fels with vanadio-rutile.

in Table 5. (Microprobe analyses are given in parantheses.) Contamination of iron in the crushing process may explain the divergence between the spectrographic and microprobe analyses for this element.

Microscopically this mineral is seen to be completely opaque and subhedral to anhedral in form (Fig. 11). The colour of the mineral resembles closely the anisotropic bands of its complex companion, as does the anisotropy and bireflection. Under higher magnification the inner reflex characteristic of rutile is conspicuous. Minute inclusions of higher reflectivity appear to be chalcopyrite, as the spectrographically determined copper would also indicate.

As may be seen from Table 5 this rutile is of unusual composition. (Cases in which vanadium has been reported in rutile the content is never more than a few tenths of a percent.) Aside from the physical relationship between this rutile and the complex mineral described previously and the optical similarity to the anisotropic phase of that mineral, there is also a distinct chemical similarity though the contents of V, Cr, Fe, and the rare earths are considerably less.

It is observed that the rutile stringers, which are abundant in the fels, do not extend into the halo zones surrounding the complex mineral

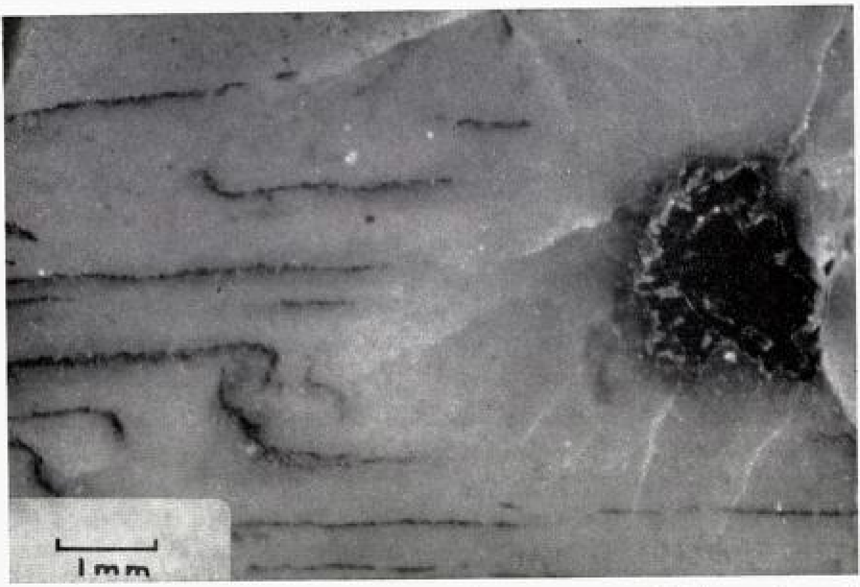

Fig. 12. Polished specimen of fels with complex titanium mineral. The vanadio-rutile stringers do not extend into the halo zone surrounding the complex mineral.

(Fig. 12). It is possible that grains of the complex were formed by the migration of ground mass rutile to certain loci, but this process would require supplementary V, Cr, Fe, rare earths and U. It seems more probable that secondary processes have removed the rutile from these radioactively affected and more porous zones as well as leaching the periphery of the complex mineral itself.

Vanadiochrome spinel.

Within the fels in the section containing the grain of the complex titanium mineral seen in Fig. 5 was found a small area with tiny (c. 0.01 mm) grains of a mineral giving the microprobe analysis shown in Table 6. Microscopically the mineral has the grayish color of magnetite, is isotropic, and exhibits edges characteristic of the isometric system. Indeed, it would have been taken for magnetite had not its chemical composition been determined. In the table Fe is calculated as Fe_3O_4 as the proportion Fe^{**}: Fe^{***} was not determinable, and V is calculated as V_2O_3. The mineral occurs in what appears to be a crack filling, together with rutile and probably albite and quartz.

The data would seem to indicate this mineral to be a variety of spinel, but no such composition has been found in the literature. The variety of magnetite, coulsonite, contains up to about 7 % V_2O_3, but with negligible Cr. Chromite contains but scant V.

Gadolinite and euxenite.

In the concentrate obtained by Carpco separation described previously a small amount of a black mineral, bottle-green on thin edges, with conchoidal fracture and vitreous luster was observed, and a number of these grains were collected by pincette. X-ray powder analysis showed the mineral to be gadolinite. Insufficient material was found for a chemical analysis, but meticulous spectrographic analysis on the small amount available provided the results shown in Table 7. The values in parentheses were determined by microprobe, Tb, Ho, Tm, Lu, and Hf being determined by interpolation. The relatively high proportion of cerium lanthanides more exactly define the mineral as a cergadolinite, with a Ce-Nd maximum.

Several dark grains with conchoidal fracture and vitreous luster, but with a deep reddish color on thin edges, were determined by X-ray powder analyses to belong to the series euxenite-blomstrandine. The amount of material available, however, did not permit closer identification.

These grains of gadolinite and euxenite (as well as apatite and antigorite) were obtained through a process designed to concentrate the complex titanium mineral. It is likely that they occur in the host rock in greater quantity than the small amount observed would indicate.

Acknowledgements.

Determinative tests were performed at the following institutions. Microprobe analyses:

Central Institute for Industrial Research, Oslo
Technical University of Norway, Trondheim

Chemical and spectrographic analyses:

Geological Survey of Norway, Trondheim
Institute for Atomic Energy, Kjeller

X-ray powder data:

Geological Survey of Norway, Trondheim
Technical University of Norway, Trondheim

In achieving the results presented in this paper a number of persons were involved and to whom thanks are due. Special acknowledgement is extended to B. Tøtdal, NTH and L. Auran, SI for microprobe tests, to J. Hysingjord and H. Hatling, NGU for X-ray powder data, to B. Berner and M. Ødegård, NGU for spectrographic analyses, and to B. Andreassen and L. Stige, NGU for chemical analyses.

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Manuscript received and accepted for publication in September 1969.

Table 1.

Analysis of complex titanium mineral. The remaining 0.1 % R.E. consist mainly of La and Yb.
Small amounts of Pr, Nd, Gd, Dy, and Er are also present.

Comp.	wt. %	Comp.	wt. %
TiO_2	54	U	0.5—2
V_2O_5	17.7	Sr	0.2
Cr_2O_3	7.4	Zn	0.00x
FeO	4.5	Ni	0.0x
Fe_2O_3	6.6	Pb	0.x
ZrO_2	0.4	Th	0.0x
Al_2O_3	1	Sc	0.0x
SiO_2	1.5	Tot. R.E.	0.91
MgO	0.5	(Ce)	(0.6)
MnO	0.06	(Y)	(0.2)
Cu	0.2		
S	1.5	SUM c.	98

Table 2.

Complex titanium mineral. X-ray powder data ,copper radiation).

Both phases		Isot. phase		Anisot. phase	
dÅ	Int.	dÅ	Int.	dÅ	Int.
3.80	6			3.80	6
3.47	10			3.47	10
2.81	6			2.81	6
2.61	7			2.61	7
2.47	6	2.46	4		
2.42	1			2.42	1
2.37	2			2.37	2
2.18	6	2.17	7		
2.14	6			2.14	6
1.89	5			1.89	5
1.80	2			1.80	2
1.68	9	1.69	10		
1.66	8			1.66	8
1.63	1			1.63	1
1.60	1			1.60	1
1.48	1			1.48	1
1.44	2			1.44	2

Table 3.

Complex titanium mineral.
Analyses of the anisotropic and isotropic phases.

Comp.	Anisot phase	Isot. phase	Gen. anal. Table 1	Anal. Kjeller
TiO_2	50.29	51.17	54	
V_2O_5	14.42	13.15	17.7	
Cr_2O_3	8.91	9.43	7.4	
Fe_2O_3	8.58	10.19	10.1	
Al_2O_3	0.60	0.86	1	
SiO_2	2.39	not anal.	1.5	
MgO	0.05	0.26	0.5	
CeO_2	0.70	4.12	0.7	2.0
Sc_2O_3	1.00	2.89	0.0x	0.9
Y_2O_3	not anal.	0.36	0.3	0.3
SUM	86.94	92.43	93.2	

Table 4.
Thortveitite. Bidjovagge analyses compared with thortveitite from Eptevann.

Comp.	Anal. 1	Anal. 2	Ave.	Eptevann
MgO	1.76	2.87	2.31	0.17
Al_2O_3	3.13	3.53	3.33	4.95
SiO_2	44.71	45.09	44.90	45.79
CaO	0.24	0.20	0.22	0.24
Sc_2O_3	39.07	38.18	38.62	34.32
TiO_2	0.68	0.85	0.76	$<$0.01
V_2O_5	1.30	1.37	1.33	
Cr_2O_3	0.37	0.83	0.60	
MnO	0.24	0.22	0.23	0.53
Fe_2O_3	4.63	6.10	5.36	2.95
Y_2O_3	2.73	2.87	2.80	
Dy_2O_3	0.18	0.20	0.19	
Er_2O_3	0.11	0.13	0.12	
Yb_2O_3	0.26	0.40	0.33	
ThO_2				0.09
ΣLa_2O_3				1.48
ΣY_2O_3				9.52
U_3O_8				0.00
Pb				0.00
ZrO_2				0.00
H_2O-				0.07
H_2O+				0.00
SUM	99.48	102.89	101.16	100.11

Table 5.
Analysis of vanadio-rutile.

Comp.	wt. %	Comp.	wt. %
TiO_2	(94.3)	U	trace
V_2O_5	3.8 (3.5)	Y	trace
Cr_2O_3	0.6 (0.4)	Sr	not detect.
Fe_2O_3	1.0 (0.2)	Zn	0.00x
FeO	$<$0.1 (chem. det.)	Ni	0.0x
Al_2O_3	$<$0.1	Pb	0.01
SiO_2	1.5	Th	0.0x
MgO	not detect.	Na	not detect.
MnO	not detect.	Nb	0.1
ZrO_2	0.3	Sc	trace
Cu	0.2	Ce	not detect.
S	not anal.	Ta	not detect.
SUM	c. 102		

Table 6.

Microprobe analysis of vanadiochrome spinel.

Comp.	wt. %
Fe_3O_4	66.28
V_2O_3	16.98
Cr_2O_3	15.09
TiO_2	0.52
Al_2O_3	1.14
SiO_2	0.19
MnO	0.x
SUM	100.2+

Table 7.

Analysis of gadolinite.

Comp.	wt. %	Comp.	wt. %
La_2O_3	0.7		
CeO_2	5.4 (2.4)		
Pr_2O_3	1.1		
Nd_2O_3	4.8		
Sm_2O_3	2.3		
Eu_2O_3	10 p. p. m.		
Gd_2O_3	2.3		
Tb_2O_3	(0.3)		
		ΣCe_2O_3	**16.7**
Y_2O_3	c. 30 (31.7)		
Dy_2O_3	1.7		
Ho_2O_3	(1.0)		
Er_2O_3	0.8		
Tm_2O_3	(0.1)		
Yb_2O_3	1.4		
Lu_2O_3	(0.5)		
HfO_2	(0.03)		
		ΣY_2O_3	35.5
		SiO_2	(22.5)
		FeO	9.5
		BeO	9.0
		CaO	(0.3)
		MnO	(0.2)
		B_2O_3	0.0x
		SUM	93.7

SUPRACRUSTAL AND INFRACRUSTAL ROCKS IN THE GNEISS REGION OF THE CALEDONIDES WEST OF BREIMSVATN[1])

By

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Contents.

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Abstract.

Two rock complexes can be distinguished in the Gneiss region west of Breimsvatn in Sogn & Fjordane, West Norway.

The upper complex is composed essentially of original supracrustal rocks: calcareous, psammitic and pelitic schists and biotite-plagioclase gneiss with significant intercalations of gneiss, meta-anorthosite, amphibolite and ultrabasite. An «eclogite-like rock» is also recorded. The lower complex is largely composed of migmatite, two-feldspar gneiss and gneiss-granite which has been mechanically transformed into augen gneiss. It is regarded as part of the infracrustal, Precambrian core of the Caledonides in West Norway. The rocks of the upper complex can be arranged in a succession of mappable units, but possible sliding tectonics and recumbent folding make stratigraphical relations still uncertain. The age of the supracrustal rocks is not clear, although it is assumed to be Eocambrian or older corresponding to Precambrian units of the central nappe region.

The contact between the supracrustal rocks and their basement is concordant, but strong differential movements shared by both complexes have taken place. Both have experienced two phases of east-west folding, the latest of which is responsible for the pattern of lithological distribution in the investigated area.

Introduction.

The Basal Gneiss problem.

The contact zone between schists on the coast and the gneisses further inland has presented intriguing problems since the early days of geological studies in West Norway (Fig. 1). Early geologists (Irgens & Hiortdahl 1864, Kjerulf 1879, Reusch 1881 and C. F. Kolderup 1923) regarded the gneiss in the east as the Archaen basement for the schists. The choice of rock types to be incorporated among the «schists» am, however, equivocal. N.-H. Kolderup (1928) reduced the area previously mapped as «schists» considerably by relating the anorthosites and their associated quartzites and mica-schists of Nordfjord and Sunnfjord to the Precambrian or possibly Eocambrian. He draw attention to the lack of evident break between the metamorphic sedimentary rocks and the gneisses: the contact was concordant and gradational like in the Oppdal—Sunndal area.

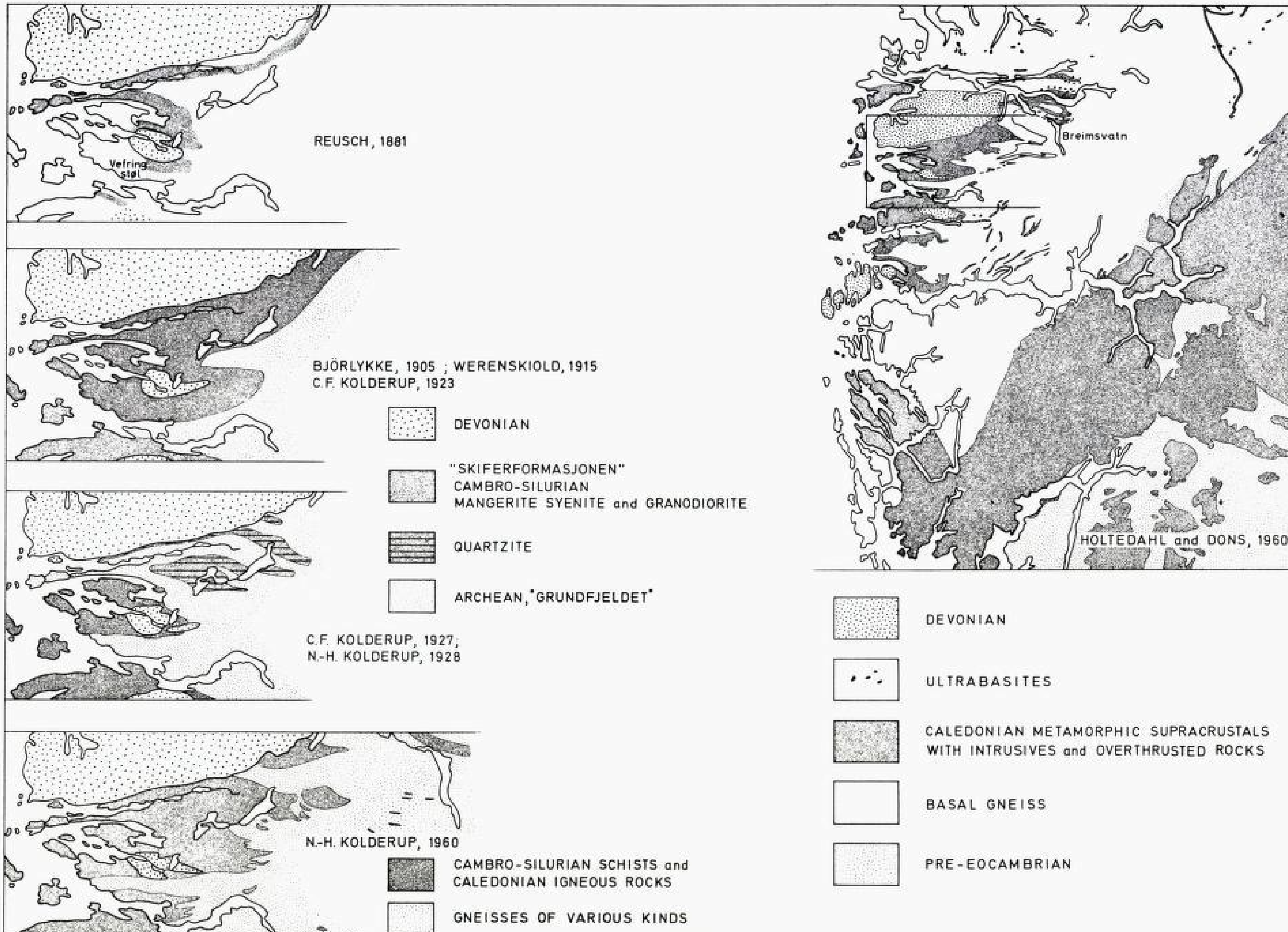

Fig. 1. Various interpretations of the contact between «schists» and the gneisses further inland in parts of West Norway. One of the localities where Reusch (1881 p. 145, 168) first described the contact between the «schists» and their Precambrian basement is indicated in the upper left insert (Vefring støl).

Results of the present mapping of the contact between infracrustals and supracrustals is shown by full and dashed line on map to the right.

The vast Gneiss region of the Norwegian Caledonides was later reinterpreted (O. Holtedahl 1936 p. 136, 1938 p. 52, 1944 map, Kolderup & Kolderup 1940 p. 129, N.-H. Kolderup 1952b p. 238): the rocks in the central zone of the orogen were «believed to have been mainly formed by high-grade metamorphism, migmatization, and granitization of pre-existing rocks (Archean, Eo-Cambrian, Cambro-Silurian)» The quote was made from the legend to the map accompanying O. Holtedahls paper from 1944. It could not be doubted that Precambrian (Archean) rocks formed part of the Gneiss region (Rosenqvist 1941 p. 25, Strand 1960 p. 232), but where and in what state?

Gjelsvik (1951 p. 36, 44) suggested that the gneisses of Sunnmøre were essentially transformed Eocambrian and Cambrian metasediments with mobilized and completely transformed Precambrian basal layers possibly uncovered in some anticlines. He inferred later (1953 p. 86, 93 and Fig. 4) that less altered Precambrian rocks might be present in the anticlines between Nordfjord and Sogn. N.-H. Kolderup (1960 p. 15) explained the gneisses of Nordfjord and Sunnfjord as products of granitization of Cambro-Silurian and Eocambrian rocks, but said that it could not be denied that also Precambrian elements might form parts of the area. A Late Precambrian eugeosynclinal stratigraphic sequence was postulated by Hernes (1965, 1967) for this and adjoining parts of the Gneiss region. Skjerlie and Kildal (personal information 1965) who have undertaken detailed geological mapping in Sunnfjord and Sogn have strongly advocated that a distinction between Caledonian supracrustals and their original Precambrian basement *could* be defined in the field, — in fact, the distinction originally made in Irgens & Hiortdahl (1864) and Reusch (1881) was valid in principle.

The structural relations of the Gneiss region were depicted by O. Holtedahl (1936, 1938, 1944 and 1952) as a system of recumbent folds. Field evidence from the Oppdal area indicated that rocks in this central part of the Caledonides had obtained a state of plasticity which made basement as well as cover deform by recumbent folding. The mobilized basement could be expected to form extended zones congruently incorporated within the supracrustals. A structure of this sort was illustrated by Muret (1960) in a significant contribution which showed piling-up of Pennine-type nappes composed essentially of Precambrian cores with thin mantles of supracrustal rocks. He claimed that evidence of an original discordant position was still discernable.

There is no reason to question O. Holtedahl's and N.-H. Kolderup's contensions that the *western* part of the Gneiss region of West Norway to a large extent is made up of altered Cambro–Ordovician and Eocambrian rocks, but the *eastern* margin has proved to be Precambrian and has been indicated as such on all geological maps (Bjørlykke 1905, Werenskiold 1915, O. Holtedahl 1944, Holtedahl & Dons 1953, revised 1960). Rekstad (1914 p. 12–14) published figures which indicated even angular unconformity below the schists, Goldschmidt (1941 p. 198) noted the presence of conglomerate above the true Precambrian basement and Landmark (1949 p. 42, 48, 57, also referred to by Strand 1960 p. 238) demonstrated that migmatitization processes in the gneisses took place in the Precambrian and that the basement was affected only by mylonitization, shearing and sheeting during Caledonian movements. Recent studies of the eastern margin of the Gneiss region (Banham & Elliott 1965, Banham 1968) confirmed the Precambrian age of the gneisses and Strand (1966, 1969 and personal information) was able to map a possible contact between the basement and younger metamorphic supracrustals at Grotli within the basal gneiss region. A true Precambrian Rb–Sr isochron age (1.000 ± 150 m.y.) was obtained by Brueckner et al. (1968) for corresponding basement gneiss at Tafjord. The complex structural relations at Grotli indicate that the basement has been deformed in Caledonian time, similar to that anticipated by O. Holtedahl (1944) on his tectonic map of the Norwegian Caledonides.

N.-H. Kolderup's maps (1928, 1960) and the most recent geological map of Norway (Holtedahl & Dons 1960) show a number of isolated outcrops of Cambro-Silurian and Eocambrian rocks in the western part of the basal gneiss region. A line drawn at the easternmost extension of these outcrops must be located near the contact between the «caledonized» Precambrian basement and the assumed Caledonian supracrustals (Wegmann 1959 p. 56, Bryhni 1966 p. 10), but its validity is dependent on the correctness of the age assignment given to these supracrustals.

The relations in the northern part of the basal gneiss region were studied by Birkeland (1958) who demonstrated that two distinct tectonic units were present in western North Trøndelag: (1) an old Precambrian basement complex and (2) a younger Caledonian system of transformed early Paleozoic geosynclinal rocks with intrusives. The younger system rests on the older with a profound unconformity. Supracrustal rocks were present in the Precambrian basement. These important results were reinterpreted (Holtedahl & Dons 1960) or overlooked in later maps or regional geological studies.

The «basal granites» of North Norway have offered the same problems as the basal gneiss of South-West Norway. Vogt (1942) considered that many, perhaps most of the granites were Precambrian and was supported by Kautsky (1946) and Gustavson (1966) who contended that the gneiss-granite in the coastal areas were thrust sheets which had been recrystallized and mobilized during the Caledonian orogeny. Rutland & Nicholson (1965) found that the Precambrian basement had rather taken part in plastic nappe tectonics comparable to that of the Pennine zone in the Alps. The main transport of the nappes was accomplished at an early stage before the main metamorphism and the slides between the nappes consequently show concordant structures on either side.

Present study.

The contact between Caledonian cover rocks and their «caledonized» basement is at present very difficult to define unambiguously in the coastal districts of West Norway, although an important attempt has been made on the MÅLØY map at scale 1 : 250 000 (Kildal 1970). Mica-schist and feldspathic quartzite in this part of the country are usually referred to as Cambro-Silurian or Eocambrian rocks. This is an assumption which should be tested — may be the metamorphic supracrustals rather belong to the Precambrian basement? The problems met with in the MÅLØY map area made it clear that detailed studies in selected areas with simpler basement/cover relations were desirable. Only by integrated interpretation of such small areas can a reliable stratigraphic-tectonic map of regional scale possibly emerge.

The present report is still at the reconnaissance scale. It gives results of a study of the contact relations between rocks of assumed supracrustal derivation and a complex of massive gneisses in an area west of Breimsvatn, Sogn og Fjordane Co. and some data on the lithologic succession in this area. Botanic studies (Nordhagen 1954) had indicated that schists and carbonate rocks were present and this was confirmed by one of the present writers (Bryhni as assistant for Prof. N.-H. Kolderup) at Blåvatn—Måsvassdalen in 1956 and Vonen—Årdalskupa in 1957. The reconnaissance indicated that the schists formed synforms above massive granitic gneisses, and the area therefore appeared promising for a closer study of general basal gneiss problems.

Fig. 2 gives the main results of the present study. Two major rock units have been distinguished:

1. An inhomogeneous complex of calcareous schist, meta-arkose, feldspathic quartzite, mica-schist and gneisses of assumed supracrustal origin. Significant amounts of meta-anorthosite, banded gneiss and amphibolite occur interlayered with these rocks, and ultrabasite and an «eclogite-like rock» do also occur.

2. A relatively homogeneous complex of massive two-feldspar gneiss, migmatite and augen gneiss.

The two units correspond respectively to the Fjordane and the Jostedal complexes of Bryhni (1966). The homogeneous complex of essentially granitic gneisses represent «infracrustal» rocks (Windley et al. 1966) which may be related to the supracrustals by sedimentary superposition, basement reactivation or by granitization/migmatization. The term «infracrustals» for these rocks is especially useful as long as the age of the supracrustals is unknown.

Lithologic succession.

Our geological mapping was performed on rather old and inaccurate maps at scale 1 : 100 000 for the major part of the about 500 km^2 area, and the tectonic or stratigraphic problems encountered in the field would certainly require more detailed studies to be properly solved. We have, however, studied some characteristic sections and established a tentative rock succession which could be a basis for later discussion of stratigraphy and structure of the area.

A description of the major rock units in the area around Storevatn follows below:

VII. *Ommedalen gneiss and meta-anorthosite.* This is structurally the uppermost unit which is exposed in the upper part of Ommedalen and in the mountain plateau south of Storeskarseggen. Massive anorthosite and schistose meta-anorthosite with layers or tectonic inclusions of amphibolite are the most common rocks together with augen gneiss.

VI. *Storeskarseggen meta-arkose.* This unit is well exposed south of Storeskarseggen where it occurs as a division several hundred meters thick of light gneiss with minor feldspathic quartzite and mica-schist. The light gneiss is essentially an epidote-muscovite-feldspar-quartz rock (minerals listed in order of encreasing amounts).

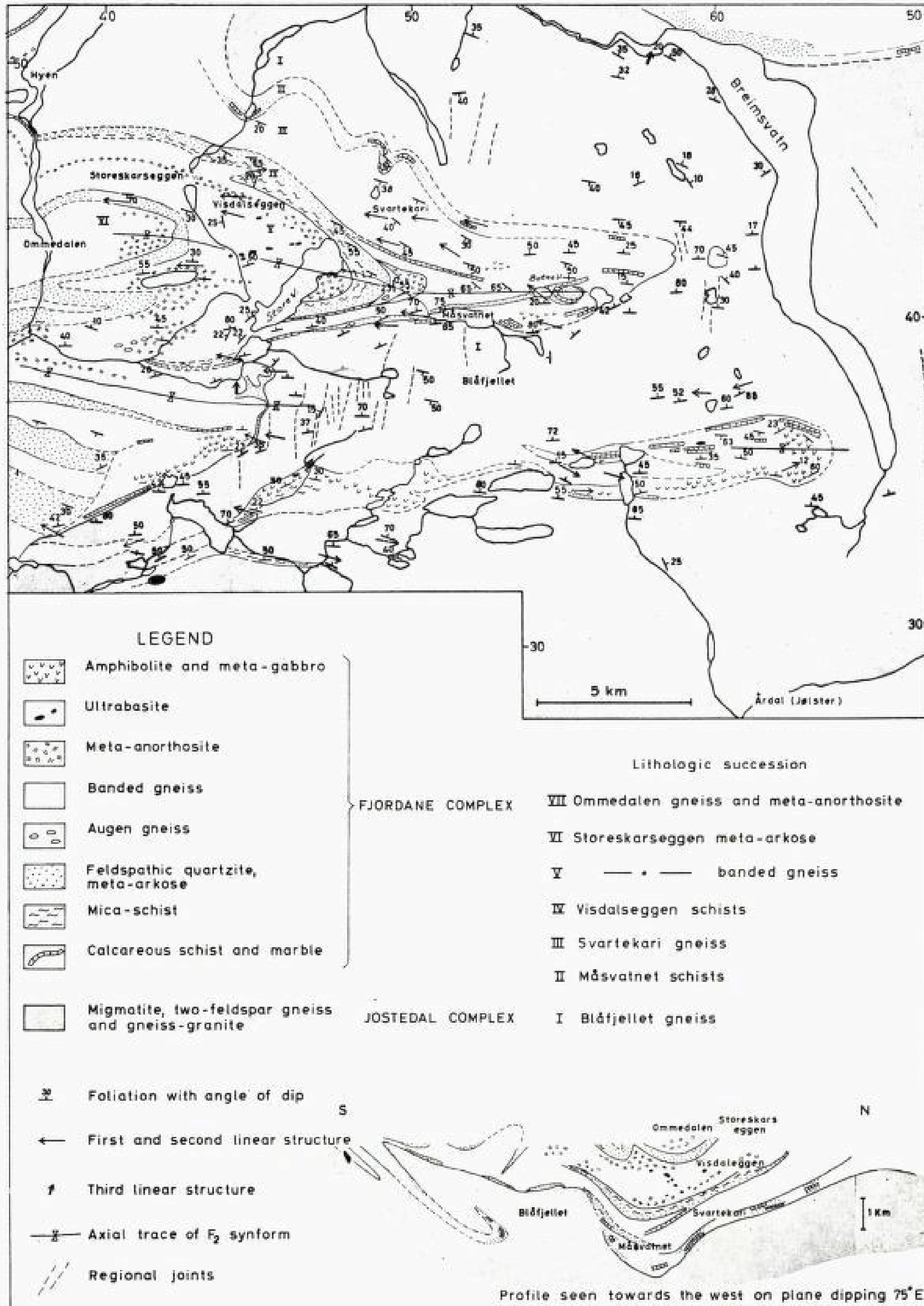

Fig. 2. Geological map, lithologic succession and tectonic profile of the area to the west of Breimsvatn. The profile is hypothetical in the southern part of the area where the fold axis has very variable trend and plunge.

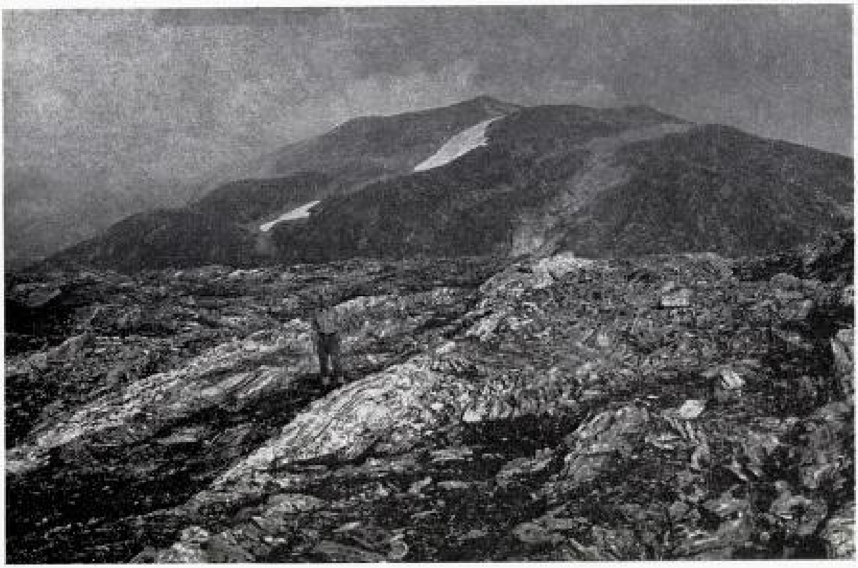

Fig. 3. Interbanding of biotite-plagioclase gneiss and meta-anorthosite. Mountain plateau south of Storeskarseggen (LP 405/455), looking east towards Visdalseggen where a larger body of white meta-anorthosite can be seen.

V. *Storeskarseggen banded gneiss.* Most of the area west of Storevatn belongs to this unit, which is several hundred meters thick. The lithology is rather varied with banded biotite-plagioclase gneiss and meta-anorthosite, big bodies of anorthositic rocks, amphibolite, augen gneiss, and other gneisses with two feldspars. Several small bodies of serpentinite have been recorded within this unit. A small occurrence of «eclogite-like» plagioclase-garnet-clinopyroxene rock from amphibolite east of Storevatn also appear to be located within this unit. Small scale banding of biotite-plagioclase gneiss, meta-anorthosite and amphibolite (with anorthositic layers less than 10 cm thick) has been recorded at the mountain south of Storeskarseggen (Fig. 3).

IV. *Visdalseggen schists.* The upper part of this unit is a characteristic garnet-quartz-mica schist with up to 15—20 % rather euhedral garnets ½—2 cm in diameter. The schist is well exposed at Visdalseggen and at the south-western part of Måsvassdalen. Layers of feldspatic quartzite 0.1—5.0 m thick are often present in this mica-schist and amphibolite is found at places. The thickness of the unit is from a few hundred metres to less than 50 metres.

Fig. 4. Svartekari gneiss with characteristic white quartz-dioritic layers and lenticles. Svartekari (LP 500/440).

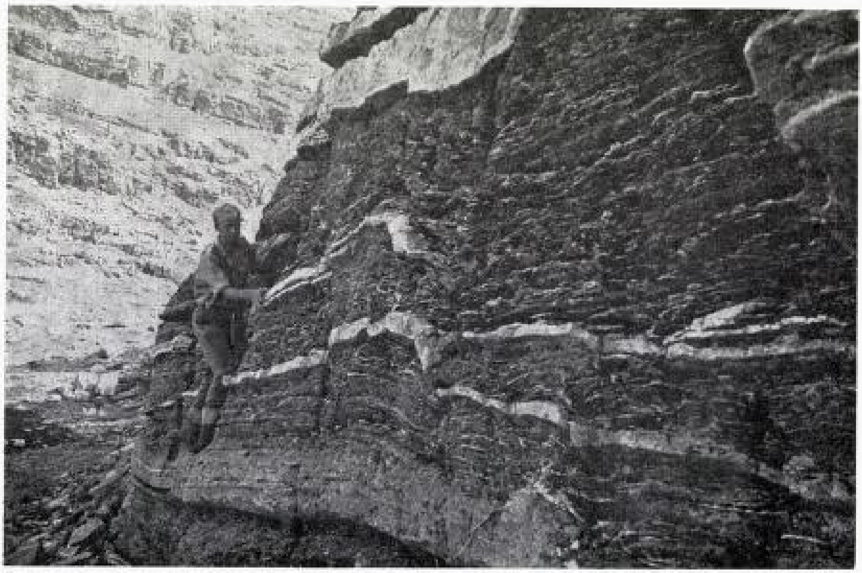

Fig. 5. Calcareous two feldspar-quartz-mica schist with layers of meta-arkose in which plagioclase is the dominant feldspar. Above Smørløysa west of Langevatn (LP 490/460).

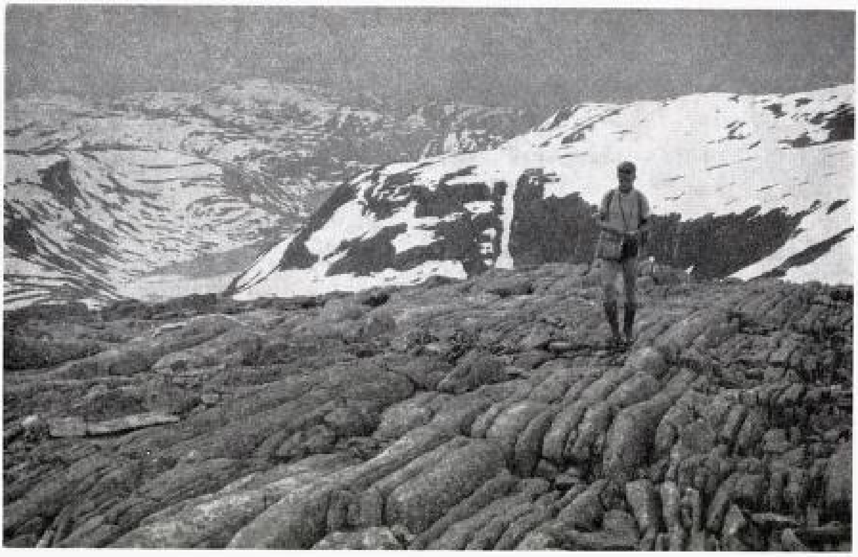

Fig. 6. Massive gneiss-granite with steep foliation (Blåfjellet gneiss). Blådalsfjell (LP 575/380).

Below the garnet-quartz-mica schist is a formation of meta-arkose 50—200 meter thick with minor feldspathic quartzite. White mica is often concentrated in layers between more massive benches in this rock.

Calcareous mica-schist with layers of impure, grey phyllonitized marble less than 1 meter thick is well exposed below meta-arkose east of Storevatn but has not been recorded elsewhere. The grey, fine-grained marble contains big books of white mica and is thus similar to marble in the Måsvatnet schists further down in the succession.

Massive augen gneiss with up to fist-size augen of potash feldspar occurs at the bottom of the Visdalseggen schists at the nose of the Måsvassdalen synform.

III. Svartekari gneiss. This is essentially a gneiss unit with banded biotite-plagioclase gneiss and biotite-quartz-two feldspar gneiss. It is present in characteristic development at the mountain Svartekari between Måsvatnet and Langedalsvatn where it has an apparent thickness of more than 1000 meters. White layers and lenticles of quartz-dioritic or of granitic composition can frequently be seen in the gneiss (Fig. 4). Pegmatitic layers and veins are especially found in the area to the east of Svartekari. Layers or isolated inclusions of amphibolite and hornblendite are rather common while quartzite and mica-schist are rare within this unit. A big body of amphibolite with relict gabbroic texture is present in Måsvassdalen.

II. Måsvatnet schists. Rocks within this unit were first referred to by Nordhagen (1954) who called attention to mica-schist and calcareous rocks at Måsvatnet. The rocks are well exposed at the south-western and at the eastern end of this lake, where they overlie the assumed basement (Blåfjell gneiss). The rocks which occur are: gneiss, amphibolite, garnet-quartz-mica schist, meta-arkose, feldspathic quartzite and calcareous mica-schist with interlayers of marble (Fig. 5). The marble is a white, pure calcite marble at Botnevatn east of Måsvatnet, but has often been mechanically transformed into a grey, fine-grained «limestone». White mica of the impure marble layers occurs as books up to several centimeters wide.

I. Blåfjellet gneiss. This unit consists of migmatites and gneisses with much potash feldspar, gneiss-granite and augen gneiss (Fig. 6). Pegmatite is rather common and often transgressive to foliation in the gneiss.

Infracrustals (Blåfjellet gneiss)

The gneiss occurring in the eastern part of our area has not previously been subjected to detailed studies. N.-H. Kolderup mentioned some related gneiss types (1928 p. 14—22 and 36) and demonstrated exposures during the Congress excursions in 1960 (Kolderup 1960). Kolderup noted that the rocks had a «sort of magmatic character» and that the gneisses were rich in potash feldspar or mica and sometimes had augen structure. The minerals reported were quartz, microcline or microperthite, albite, biotite, muscovite, garnet, epidote, sphene and black iron minerals. Myrmekite and chlorite also reported.

The area immediately west of Breimsvatn contains a variety of different rocks, among which the following structural types are prevalent:

Massive gneiss-granite.

Massive augen gneiss with abundant lensoid or folded pink augen ½—1 cm long.

Banded biotite-two feldspar gneiss with augen or lenticles of pink potash feldspar.

Migmatite with irregular coarse-grained quartzo-feldspathic veins and dykes.

The rocks are, however, rather homogeneous compared with those which occur structurally above them further to the west. The term «Blåfjellet gneiss» has been found appropriate for this unit. It is believed that the gneisses from Blåfjellet are continuous with rocks below the Jostedalen glacier to the east of our area and that they thus are representative of the infracrustal core of the Caledonides.

The massive augen gneiss of our area appear to have formed from original granitic rocks by penetrative shearing. The massive rocks have weakly developed foliation with widely spaced parting surfaces. A coarse layering or banding is, however, distinct when a cliff section is viewed at a distance. Lineation is omnipresent in the western part of the area where the gneiss has been subjected to shearing, and cross joints are very characteristic there. In fact, the infracrustal rocks in the western areas can be mapped from air photographs by their characteristic jointing.

Pegmatitic, granitic and white quartz-monzonitic layers or transecting veins are common. Amphibolitic intercalations veins occur more sparsely and have been transected by pegmatite. The most widely distributed rock is the gneiss-granite, which is essentially a magnetite-sphene-biotite-quartz-plagioclase-microperthite fels, and its sheared derivative, which is essentially a magnetite-sphene-white mica-epidote-quartz-plagioclase-microperthite augen gneiss (minerals listed in order of increasing amounts). Apatite, epidote-orthite, white mica, pyrite, and zircon are common accessory minerals and calcite, chlorite and red iron minerals occur secondarily. Fluorite and molybdenite have been found occasionally.

The most massive granitic rock is located at Førde in the south-east corner of our map. It is an unequigranular gneiss-granite or granite with 1–2 cm wide megacrysts of pink alkali feldspar which is partly intergrown with or mantled by plagioclase. Its ground-mass contains quartz and granulated plagioclase with grain-size mostly between 0.2–1.0 mm. This rock is, towards the west, progressively transformed into a foliated and lineated augen gneiss with flat or lenticular pink augen 0.5–2.0 cm long. The ground-mass is fine-grained with grains of granulated quartz and plagioclase 0.1–0.5 mm long.

Sections perpendicular to the lineation indicate that the alkali feldspar pophyroclasts have rotated relative to the matrix. The augen have remained relatively coherent during deformation while quartz and plagioclase broke down into granulated aggregates of very small grains which bend around the more resistant potash feldspar fragments.

Table I.

Modal composition of some samples from Blåfjellet gneiss. «x» indicates that the mineral is present in small amounts.

Compositions have been estimated by pointcounting thin section and stained slab of each sample.

Locality	Quartz	Alkali feldspar	Plagioclase	Myrmekite	Biotite	White mica	Chlorite	Epidote	Orthite	Sphene	Zircon	Apatite	Calcite	Black iron minerals	Red iron minerals	Pyrite
Gneiss-granite Førde, Breim, LP 66/34	24	32	33	x	7	x	x	x	x	2	x	1	x	1	—	x
Gneiss Sørsendalen, Breim, LP 58/42	31	41	18	x	5	3	x	1	x	x	x	x	—	1	x	x
Augen gneiss Blåfjellet, LP 48/37	22	41	26	x	5	x	x	4	x	x	x	x	—	2	x	x

Modal compositions of three selected rocks are given in Table I.

Quartz occurs in aggregates of lenticular grains 0.5—1.0 mm long with saccaroidal to interlobate contacts in the massive gneiss-granite and as aggregates of strongly elongated, strongly undulative grains 0.05 mm thick and 0.5 mm long with sutured boundaries in the augen gneiss.

Alkali feldspar is mostly an untwinned microperthite which at places has rods and veins less than 0.005 mm thick of fairly refrigent inclusions. The amount of these plagioclase inclusions varies from grain to grain and even within a single grain, but is usually less than a quarter of the total microperthite volume. The portion of plagioclase inclusions increases towards zones of granulated plagioclase which sometimes transects the microperthite. Undulative extinction and microcline twinning is distinct in some grains or parts of one grain. Myrmekite and aggregates of small plagioclase grains often occur in the marginal zone of the microperthite grains.

Plagioclase is obviously recrystallized in the gneiss-granite at Førde where it occurs as granular-saccharoidal aggregates of grains about 0.2 mm wide with some epidote. Plagioclase in augen gneiss have bent twin lamellae but is usually recrystallized into apparent untwinned grains. The composition of both the twinned and the recrystallized varieties of plagioclase is about An 30.

Biotite occurs as clusters of books 0.5—1.0 mm thick in the gneiss-granite and as «trains» of somewhat smaller books in very fine-grained micaceous folia in the augen gneiss. Pleochroism is: X pale yellow, Y, Z dark olive green or dark brown. Parallel intergrowths with chlorite are seen in a few grains within each thin section.

White mica is only present as tiny flakes in plagioclase aggregates in the gneiss-granite and as serictic aggregates together with biotite in augen gneiss.

Chlorite is always associated with biotite from which it appears to have formed by retrograde alteration. Pleochroism: X colourless, Y, Z green. Anomalous blue interference colours are usually seen. Many grains have acicular inclusions which probably are rutile.

Epidote minerals. Epidote occurs as tiny grains within plagioclase aggregates in the gneiss-granite and as grains 0.5—1,0 mm wide in the augen gneiss. Many grains in augen gneiss have a core of red-brown, almost isotropic orthite. Orthite sometimes occurs alone as inclusions in sphene and magnetite.

Sphene is very frequent and readily seen with the naked eye as red-brown grains 1—3 mm long. Polysynthetic twinning is often seen under the microscope.

Zircon is idiomorphic in the gneiss-granite at Førde from which it has been isolated as slender tetragonal prisms 0.2—0.3 mm long.

Magnetite is omnipresent and very characteristic for the gneiss-granite and the augen gneiss. The magnetic fraction of a crushed sample from Førde had abundant perfect octahedra 0.2 mm high which under the ore microscope were quite homogeneous.

Radiometric ages.

Two micas from pegmatite in the «Jostedalen complex» have previously yielded K/Ar ages of 590 m.y. (Ortevik, Sogn) and 582 m.y. (Loen, Nordfjord). Neumann (1960) who reported these radiometric ages assumed that they were probably lower than the true ages of rock formation because argon might have been expelled during a later period of Caledonian orogeny.

Older gneisses of Tafjord (Brueckner et al. 1968) have yielded a Rb/Sr isochron age of 1.000 (± 150) m.y. and a biotite age of 383 (± 12) m.y.

Biotite from a massive granite-gneiss at Førde south of Breimsvatn has been dated for the present study (Geochronological report No. FMK/692, 1969). A total degassing $40_{Ar}/39_{Ar}$ determination produced an apparent age of 398 (± 2) m.y. and a $40_{Ar}/39_{Ar}$ spectrum analysis gave an isochron age of 402 (± 1) m.y. Thus the last major phase of Caledonian orogeny in this area occurred around 402 m.y. and was of sufficient intensity to outgas and/or recrystallize almost all the biotite of the granite-gneiss.

It is well known that biotite ages are easily reset by later metamorphic events. It is seen clearly by the ages reported from Tafjord and it is also well demonstrated at Eidsfjord, Hardanger (Progress Report 1967 and 1968) where Precambrian rocks with very little petrographically perceivable Caledonian influence and Rb/Sr ages above 965 m.y. yielded biotite Rb/Sr and K/Ar ages of about 380 m.y. Work in progress will probably produce evidence that the infracrustal rocks around the southern end of Lake Breimsvatn are originally Precambrian like the older gneisses at Tafjord.

Supracrustals and Associated Rocks.

Cambro-Ordovician and older rocks which can be identified as original supracrustals occur extensively on the west coast of Norway (N.-H. Kolderup 1928, 1960). Various gneisses and rocks can be related to the «Anorthosite Kindred» (Kolderup & Kolderup 1940, p. 69); meta-anorthosite, amphibolite, mangerite syenite, etc., are associated with the supracrustals, and Bryhni (1966) assigned them all to a «Fjordane complex» distinct from an assumed Precambrian basal complex («Jostedal complex»).

Banded gneisses.

A distinct tectonic/stratigraphic unit with calcareous and other metamorphic sedimentary rocks (Måsvatnet schists) separates the gneisses now to be described from the infracrustal gneisses.

High content of mica — biotite and at places also muscovite — is characteristic, and glittering rock surfaces can often be seen because

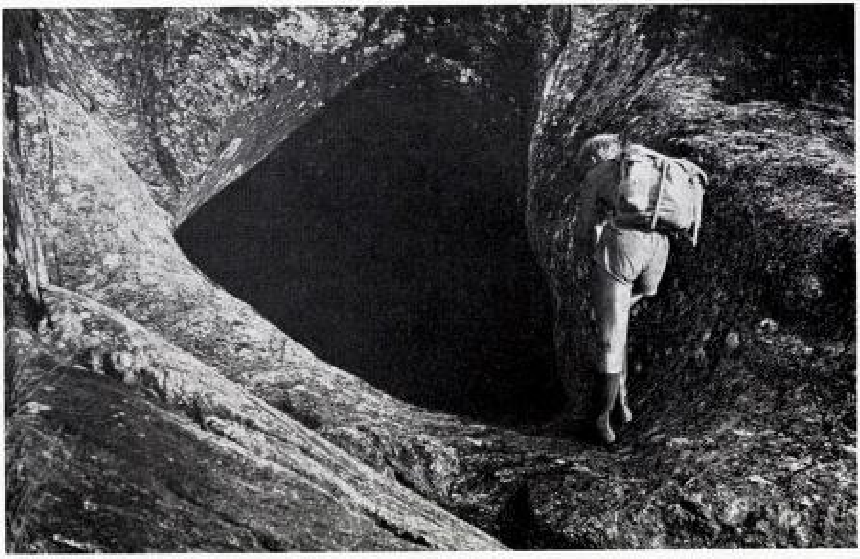

Fig. 7. One of the pot-holes carved out in massive augengneiss in Måsvassdalen (LP 490/419).

of mica-rich layers in the gneiss. The content of quartz varies much, but lenticles of pure quartz are frequently found. Many gneisses have plagioclase as the single feldspar although two-feldspar gneisses and even augen gneisses with much potash feldspar also occur.

Modal composition of some characteristic rocks is given in Table II.

Biotite-quartz-two feldspar gneiss.

Potash feldspar is present especially in the northern part of our area, in the unit termed «Svartekari gneiss». The structure of the potash feldspar bearing gneisses is usually characterized by alternating dark and light layers with white quartz-dioritic and sometimes even granitic bands often present. The amount of potash feldspar is variable, but usually small and confined to coarse layers and augen. Plagioclase is thus the dominant feldspar except in a variety of augen gneiss and in rare granitic and pegmatitic layers and veins. Augen gneiss with potash feldspar as the dominating mineral has been recorded in Måsvassdalen where it occurs in the contact zone between a feldspathic quartzite/ meta-arkose and amphibolite and on the mountain plateau south of Storeskarseggen where it is associated with meta-anorthosite. The augen

Table II.

Estimated modal composition of various banded gneisses. «x» indicates that the mineral is present in minor amounts. «—» indicates that the mineral not been observed in the thin section.

Locality	Quartz	Potash feldspar	Plagioclase	Biotite	White mica	Chlorite	Garnet	Kyanite	Epidote	Orthite	Amphibole	Scapolite	Sphene	Zircon	Apatite	Black iron minerals	Red iron minerals	Pyrite
Åstølfjellet, LP435/377	20	20	45	10	3	—	—	—	2	x	—	—	x	x	x	x	x	—
NW Svartekari, LP 485/450	25	10	40	10	10	—	—	—	5	x	—	—	x	x	x	—	x	—
Svartekari, LP 500/440	20	10	60	10	—	x	—	—	x	x	—	—	x	x	—	—	—	—
Svartekari, LP 500/440	40	—	10	30	—	—	—	—	20	x	—	—	x	x	x	—	x	x
Svartekari, LP 500/440	20	—	55	12	—	3	—	—	10	—	x	—	x	x	x	x	—	—
N. Bergstølen, LP 407/367	22	—	33	17	x	2	24	x	x	—	x	—	—	x	—	2	x	—
N. Bergstølen, LP 407/365	20	—	60	16	x	x	2	—	x	—	—	x	—	x	—	2	x	—
N. Bergstølen, LP 407/365	30	—	10	10	43	x	x	—	x	x	—	5	x	x	x	2	—	—
Myklandsdalen, LP 613/362	20	—	46	26	—	—	—	—	x	—	1	—	1	x	x	6	x	—
E. Kupevatn, LP 563/353	38	—	19	18	—	1	2	—	15	—	7	—	—	x	x	x	—	—

gneiss at Måsvassdalen is an exceptionally massive (Fig. 7) and tough rock in which fine potholes have been carved out. The banding in the most widely distributed gneiss is due to alternation of epidote-quartz-biotite and biotite-potash feldspar-quartz-plagioclase layers (minerals listed in order of increasing amounts). The content of muscovite varies within wide limits, — from highly micaceous rocks similar to mica-schists to rocks devoid of white mica.

Amphibole, apatite, chlorite, limonite, orthite, pyrite, magnetite, sphene and zircon occur as accessory or secondary minerals.

Quartz usually occurs as granular aggregates and rarely as component of myrmekite.

Plagioclase (An 20) is strongly sericitized, saussuritized and sometimes even contains biotite inclusions. It has recrystallized into grains without inclusions at places, sometimes within the same thin section where other plagioclase grains are full of inclusions.

Potash feldspar is microperthitic with tiny plagioclase-component inclusions formed as rods and as veins. Inclusions have high relief and locally zoisite indicative of an original anorthite content. Irregular inclusions of potash feldspar in plagioclase or intergrowths between the two feldspars have also been noted.

Biotite (X pale yellow, Y, Z olive green or brown) and *muscovite* are often intergrown or aggregated with epidote and sphene.

Chlorite has formed from biotite at places.

Epidote is zonar with highest interference colours at rims. A core of red-brown *orthite* is sometimes present.

Sphene sometimes occur as mantles around black iron minerals.

Zircon is present as rather round grains.

Biotite-plagioclase gneiss.

A grey gneiss with much biotite and with plagioclase as the sole feldspar is the most widely distributed rock above the infracrustal complex. It often occurs with potash feldspar bearing gneisses in the unit termed «Svartekari gneiss» but is also present in higher tectonic/stratigraphic levels interbanded with mica-schist, feldspathic quartzite, amphibolite and meta-anorthosite.

The structure of this gneiss varies from rather homogeneous and massive «dioritic» types to banded and schistose rocks, and it can sometimes be demonstrated that this variation is related to mechanical deformation and accompanying neomineralization (phyllonitization). The most massive varieties are equigranular and medium-grained while the banded and schistose types are fine-grained and granulated with extensive sericitization of feldspar and biotitized or chloritized garnet. Massive rocks are essentially biotite-quartz-plagioclase felses while the sheared rocks are essentially white mica-epidote-quartz-plagioclase gneisses and schists.

Any of the minerals amphibole, epidote, garnet, kyanite and scapolite are locally characteristic constituents. Apatite, black iron minerals, orthite, pyrite, red iron minerals, sphene, zircon and zoisite occur as accessory phases.

Quartz is strained in the most massive rocks and granulated into aggregates of smaller, unequidimensional grains with sutured outlines in the sheared rocks.

Plagioclase is a basic oligoclase to acid andesine in the most massive rocks and acid oligoclase in the schistose types. Tiny needle- or flake-like inclusions of biotite, white mica and zoisite cloud the plagioclase even in the most massive rocks. These inclusions are often concentrated in the contact zone between adjacent plagioclase grains. Granulation has started at the grain margins in the massive rocks and is complete in the sheared rocks. Twinning has been obliterated by the abundance of inclusions in plagioclase of the massive rocks, but is often distinct in some of the sheared and recrystallized varieties which have smaller unequidimensional grains of plagioclase without inclusions.

Biotite (X colourless, Y, Z brown or olive green) is fringed by small grains of amphibole in the massive rock types and intergrown with chlorite in the sheared rocks. Inclusions of zircon, black iron minerals or sagenitic rutile are sometimes present.

Amphibole is present only in the massive rocks as blue-green hornblende on the margins of biotite. More sheared or recrystallized rocks have pale-green poikiloblastic actinolitic amphibole.

Epidote of the epidote-rich rocks has vivid interference colours with highest bi-refringence at the rims. Brown orthite occurs in some specimens and clinozoisite with normal low or anomal blue interference colours is sparsely present.

Garnet has round, xenomorphic outline or has been broken up into irregular fragments. Numerous inclusions are usually present. Biotite and chlorite have formed along cracks oriented perpendicular to foliation.

Kyanite occurs in layers with much garnet as fibrous masses.

Scapolite has sometimes very irregular outlines indicating interstital growth. Inclusions of black iron minerals are common. Yellow-brown alteration products are present along the rim and on cracks in the scapolite grains.

Apatite and *sphene* occur very sparsely, and sometimes fail to show up in one thin section.

Zircon is present as round or lenticular grains mostly less than 0.2 mm wide and is surrounded by pleochroitic halo when inclosed in biotite.

Chlorite is usually of a type with high refringence, normal low or anomalous brownish interference colours, weak pleochroism: X, Y pale green, Z colourless and negative elongation possibly corresponding to prochlorite. It has inclusions of sphene. Another type, often a retrograde product of garnet, has very lov, anomalous Berlin blue interference colours.

White mica is present as tiny inclusions in plagioclase of the relatively massive rocks and as sericitic aggregates in the sheared rocks.

The two types of gneiss described in this section are related by similar appearence and occurrence in the field, similar association with amphibolite and metamorphic sedimentary rocks and by similar properties of many of their minerals as seen in the microscope.

The gneisses are in some respects like those which occur in outer Nordfjord (Bryhni 1966 p. 17—32). The biotite-plagioclase gneisses of our area are, on the other hand, also similar to metamorphic dacites which occur in thrust slices above a migmatitic basement in the Bergsdalen area (Kvale 1946a p. 47—52 etc.).

Close association with quartzitic rocks and mica-schist indicates that the gneisses may have formed by metamorphism of original sedimentary and volcanic rocks although local interlayering with anorthosite and evidence of phyllonitization from original massive rocks make this explanation difficult to apply for *all* the biotite-plagioclase gneisses. It is possible that *some* gneisses have formed from more massive but flow banded plutonic rocks which by mechanical processes and neomineralization become schistose — like what can be seen in the overthrust masses of central Norway — while *others* have been produced from dacitic and rhyodacitic volcanics or from greywacke sediments.

Mica-schist, micaceous gneisses, meta-arkose and quartzitic rocks (Meta-pelites and meta-psammites).

Mica-schist, muscovite-rich gneisses, meta-arkose and feldspathic quartzites are interbanded and grade lithologically into one another. A characteristic example of interlayering of mica-schist and feldspathic quartzite is illustrated in Fig. 8 where the quartzitic layers are 5—200 cm thick.

Tight internal folding and armouring by quartz segregated into lenticles or fold-cores have rendered some varieties of mica-schist rather tough and massive. Small rounded topographical features are characteristic for them. Carbonate is locally present as lenticles with brown colour on weathered surfaces. The mica-schist is garnetiferous at places with abundant garnets up to 1 cm wide. Feldspar locally shows up as big, round porphyroblasts and the rock then grades into augen gneiss. Pyrite and tourmaline are mesoscopically discernable in some mica-schist and micaceous gneisses.

The meta-arkoses and feldspathic quartzites are white or grey rocks with much potash feldspar and white mica. Potash feldspar is usually

Table III.

Estimated modal composition of some mica-schists, micaceous gneisses and feldspathic quartzites. «x» indicates that the mineral is present in minor amounts. «—» indicates that the mineral was not observed in thin section
[1]) *And rutile.*

Locality	Quartz	Potash feldspar	Plagioclase	Biotite	White mica	Chlorite	Garnet	Epidote	Orthite	Amphibole	Sphene	Zircon	Apatite	Calcite	Tourmaline	Black iron minerals	Red iron minerals
N. Storevatn, LP 474/433	35	—	5	x	45	—	15	—	—	—	x[1])	x	x	—	x	x	—
Årdalskupane, LP 551/360	50	—	5	5	40	—	—	x	x	—	x	x	x	x	—	x	x
S. Storeskarseggen, LP 410/450	45	15	25	x	10	x	—	3	x	—	x	x	x	—	—	2	—
NW Vaslivatn, LP 405/430	40	20	10	1	26	x	—	2	x	—	x	x	x	—	—	1	x
Årdalskupene, LP 564/350	35	20	5	x	37	x	—	x	x	—	x	x	x	—	—	3	—
Ryg støl, LP 610/365	56	29	—	—	15	—	—	—	—	—	x	x	x	x	—	—	—
Gjengedal, LP 390/415	73	21	—	—	6	—	—	x	—	x	x	x	x	—	—	x	x
Årdalskupene, LP 564/346	71	17	—	—	12	—	—	—	—	x	x	x	x	—	—	x	x

Fig. 8. Interlayering of feldspathic quartzite and mica-schist. Nesstøylen (LP 450/330).

dominating — sometimes the only feldspar present, but one sample (from Storeskarseggen meta-arkose) has more plagioclase than potash feldspar.

Apatite, black iron minerals (hematite, magnetite and ilmenite), orthite, sphene and zircon occur as accessory minerals.

Modal composition of some samples is given in Table III.

The texture of the mica-rich rocks is characterized by twisted books of white mica and by lenticular feldspar grains in a very fine-grained matrix of granular-sutured quartz. The quartzitic rocks are also usually very fine-grained with unequal-dimensional grains of strongly sutured quartz elongated in the foliation surface.

Potash feldspar has sometimes microcline «grill» but is mostly without twinning. Grains with irregularly distributed veins of plagioclase have been noted in micaceous gneiss but feldspar is usually non-perthitic. Potash feldspar of some quartzitic rocks is «dusty» due to very small brown inclusions.

Plagioclase is usually untwinned and never contain inclusions of potash feldspar. Albite has been noted in one thin section and basic oligoclase in another.

White mica often forms books up to 5 mm wide which may be externally granulated into sericite. Mantling by fine-grained biotite has been noted in a non-feldspathic calcite-bearing mica-schist. White mica of the quartzitic rocks has a green tint indicative of high iron content.

Biotite is always present in much lower amounts than white mica. Pleochroism is:

X colourless, Y, Z dark green, olive green or dark brown. Chloritization of biotite has been noted in one sample.

Epidote-minerals. Brown, zonar orthite is the only epidote-mineral in some of the mica-schists. It is then sometimes intergrown with black iron minerals. Epidote-rich rocks have orthite only as cores within highly birefrigent epidote.

Sphene sometimes forms big porphyroblasts, but is usually present as small isolated grains or clusters of small grains intimately associated with black iron minerals.

The black iron minerals are in some cases magnetite, in other ilmenite and hematite. A polished sample shows the latter two to be finely intergrown.

Calcareous mica-schist and marble.

The calcareous rocks in the area are usually easily recognized by the rusty weathered surface, cavernous weathering and by the natural rock-gardens with Silene acaulis, Dryas octopetala and various types of Saxifraga (Nordhagen 1954).

The main rock type is a coarse-grained calcite-quartz-mica schist with layers less than 1 meter thick of white marble or grey, impure, fine-grained «limestone». Potash feldspar as well as plagioclase occur in some varieties of the schist, and quartzo-feldspathic rocks are locally interlayered with it. Lenticles of quartz and of calcite are frequent.

The combined thickness of the calcareous rocks is usually less than 20 meters and often as little as 2 meters although thickness above 50 metres are reached at Botnevatn where tectonic thickening has taken place.

White marble from the west end of Måsvatnet is rather coarse-grained calcite aggregate with only a few grains of quartz and graphite. An impure «limestone» from the SE side of Storevatn is gray on fresh surface, contains white mica in books up to 1 cm wide but is otherwise very fine-grained. It contains about 60 % calcite, 20 % quartz, 15 % muscovite and 5 % plagioclase and is microscopically much sheared with folded lamination, twisted books of mica and granulated grains of quartz and plagioclase in a very fine-grained matrix of calcite, black iron minerals, sericite, sphene and orthite. The rock gives a false appearance of low metamorphic grade because of the grain-size and gray colour, but it is rather an impure marble which has been granulated.

A layer of calcareous mica-schist in feldspathic quartzite at Sletteheia west of our area has about 50 % quartz, 25 % plagioclase, 15 % calcite and 10 % muscovite. The quartz is present as segregations or as granulated aggregates. Apatite, green biotite, chlorite, black iron minerals, limonitized pyrite and sphene occur as accessory minerals. The high content of plagioclase in this rock is remarkable, and indicates that there may be a lithological transition between impure marble and quartzo-feldspathic gneisses. Local presence of plagioclase and potash feldspar in some calcite-quartz-mica schists and interlayering of such schist with quartzo-feldspathic rocks are, in fact, indicative of an original sedimentary environment where calcareous and arkosic psammites were deposited together.

Ultrabasites.

Ultrabasites occur within the series of assumed supracrustal rocks. They are developed as serpentinized dunite, serpentinite, chlorite schist and as talc schist. Only the biggest body, at Nes in Naustedalen, has been mentioned in the literature previously (Kolderup 1928) but the ultrabasites in our area appear to be related to those described from an adjoining area to the north (Kolderup 1952a). The ultrabasites west of Storevatn form an aligned association in micaceous gneisses with anorthosite. The biggest body is about 200 m long and 75 m wide, while the smallest are only 5—15 m thick schistose bodies interlayered with the micaceous gneiss.

Asbestos occurs in irregular veins and talc is concentrated in layers up to 1 m thick.

The two bodies in upper Naustedalen are developed as serpentinite schist on the contact between amphibolite and micaceous gneiss or between calcareous mica-schist and feldspathic quartzite.

It is interesting that the ultrabasites are associated with amphibolite and anorthosite. The whole rock complex has been too much altered to give any indication about the original relations between these rocks, but the fact that similar rocks occur together in vast areas north of our area (Bryhni 1966 p. 48) points towards a genetic connection, possibly involving magmatic fractionation, flow differentiation or differential melting of an igneous basic body.

«Eclogite-like rock» and other basic inclusions.

A massive rock with garnets in a pale-green, fine-grained clinopyroxene matrix has been recorded from an ultrabasic part of amphibolite east of Storevatn. The garnets ($n = 1.742$) are marginally symplectitized and granulated while clinopyroxene occurs both as strained porphyroclasts and as fine-grained, granulated matrix, none of which appear to be symplectitized. Acid plagioclase (An 30) and quartz occur

Fig. 9. Hornblenditic inclusions in Svartekari gneiss (LP 515/422).

rather subordinately as fine-grained, granular aggregates while rutile is a conspicuous accessory mineral.

The rock is «eclogite-like» in hand specimen and also microscopically similar to an eclogite except for the presence of minor plagioclase. It is possible that plagioclase has formed by recrystallization of a meta-eclogite, but it is more likely that the rock was metamorphosed in the boundary field between gabbro and eclogite, i.e. in the granulite stability region.

The two-feldspar gneisses east of Svartekari contain a variety of basic inclusions, which occur as isolated bodies, trains of small bodies or as layers, — much in the same way as the eclogites and associated rocks do in outer Nordfjord (Fig. 9). Many of the basic bodies consist essentially of coarse-grained hornblende, while others are garnet-hornblendites, garnet-epidote-hornblende felses and garnet amphibolites. A sample selected for microscopic examination was essentially a garnet-epidote-hornblende rock with minor muscovite, biotite, plagioclase, chlorite and rutile. Symplectites were not observed, and we have thus no proof that the basic inclusions can be related to the eclogites.

Amphibolites.

Amphibolites are very common associates of the micaceous gneisses and -schists. Relict gabbroic texture has anly been preserved in a large body in Måsvassdalen. All other amphibolites occur as layers with apparent thicknesses from less than one meter to at least 400 meters. Like other rocks, they are tightly similar-folded with axial planes almost parallell to lithologic boundaries. The highest figure of «layer thickness» is therefore due to tectonic thickening. Thin layers of amphibolite have often been deformed into boudins. Layers with rusty surface colour due to oxidized pyrite are very characteristic for the amphibolitic rocks. Several mineralogical and structural varieties occur: The most common type is a quartz-epidote amphibolite, but quartz-garnet-epidote amphibolites, biotite-amphibolites and white plagioclase-rich schists frequently occur as bands. Gneiss and white quartz monzonite may also occur as layers within amphibolitic rocks.

Only two specimens of quartz-garnet-epidote amphibolites have been studied in thin section, both representative of the south-eastern part of our area. They have granoblastic texture in plagioclase-rich or quartz-rich aggregates. Plagioclase has composition An 14, is untwinned and sometimes contains lenticular inclusions of quartz. Amphibole has pleochroism: X pale yellow, Y and Z blue-green, $2V_x = 78°$, $Z/C = 20°$.

Epidote, garnet and quartz make up about 15 %, 10 % and 10 % respectively of the bulk.

Biotite, apatite and zircon occur as accessory minerals.

The occurrence of amphibolite as widely extended layers in micaceous rocks and their association with other supracrustal rocks indicate that they might be transformed basalts or other basic rocks originally formed at the surface. In some cases this might be true, but local preservation of gabbroic texture and association with anorthosite rather suggest that the amphibolite was produced from original igneous plutonic rocks.

Meta-anorthosites.

White, altered anorthosites occur intimately associated with amphibolitic rocks and with biotite-plagioclase gneisses. Augen gneiss is sometimes also associated with anorthositic rocks. Some of the gneisses are certainly related to the «Anorthosite Kindred» as defined by Kolderup and Kolderup (1940).

The Storeskarseggen gneiss unit west of Storevatn contains relict of only slightly altered anorthosites. The plagioclase of these rather massive rocks has locally retained its lilac colour and lustre on cleavage surfaces. Most anorthosites have been altered, however, to schistose, tightly shearfolded, very fine-grained rocks. The mineral association found in these rocks are:

plagioclase-zoisite-muscovite-chlorite

plagioclase-epidote-muscovite-biotite (amphibole-garnet).

The plagioclase is an andesine, — sometimes basic andesine (An 45), Bright-green chromian muscovite occurs in pockets and veins. Amphibolites are often associated with the altered anorthosites. Amphibolites in the south-eastern part of our area contain white rocks with less obvious relationship to the «Anorthosite kindred». The mineral associations found in these white bands are:

plagioclase-zoisite-actinolite-prochlorite

plagioclase-epidote-muscovite-biotite.

The first association was found in a specimen sampled in Årdalskaret east of indre Kupevatn. It has a very fine-grained granoblastic-sutured mosaic of plagioclase between extremely well oriented tablets of zoisite and amphibole. Bigger grains of plagioclase are twinned and externally granulated into very fine-grained mosaic of largely untwinned plagioclase.

The other association was found in a specimen sampled at Skjorta east of Myklandsdalen. This rock is also finegrained, but the texture in aggregates of plagioclase (An 40) is granoblastic-saccharoidal. The plagioclase aggregates are surrounded by epidote and muscovite in a felted mass of sericite and pale-brown biotite. Books of muscovite are often twisted and externally granulated into sericite. Inverse zoning is often seen in plagioclase.

The thin, white layers in amphibolite are composed of essentially andesinic plagioclase and epidote minerals and should be classified with the meta-anorthosites encountered further west and northwest in our area.

Presence of white meta-anorthositic layers in amphibolite and interlayering of gneiss and anorthositic rocks must be considered in any theory for the formation of anorthosites in our area. Such field relations are best explained by calling for an original heterogeneous igneous plutonic complex where each lithological unit became widely extended by original flow layering or secondary mechanical deformation with neomineralization.

Structure.

Contact between lithologic units.

Contacts between lithologic units are sometimes difficult to define because shearing and recrystallization in the contact zone has rendered the rocks on each side similar to each other. This difficulty is marked especially in the most important of them all: the contact between the infracrustal (Blåfjellet gneiss) and the supracrustal complexes (Måsvatnet schists). An old observation by Reusch (1881 p. 167) had shown that the «schists» contained a variety of rocks: light schists, mica-rich schists, augen gneiss and hornblende-rich schists while the «Archean» basement contained gneiss-granite and gneiss. We think that the *heterogeneous character* is characteristic for the rocks we have related to the supracrustals and that this is of diagnostic significance for the Fjordane complex in contrast to the monotony of the rocks of the Jostedal complex.

It is possible that the contact regions locally are influenced by «granitization» processes as inferred by Kolderup (1960) for the north side of Breimsvatn where gneisses and quartz schists have veins and apophyses of granitic substance. This possibility cannot be ruled out for the north-west shore of Breimsvatn either where granitic and pegmatitic bodies frequently occur as layers, dykes or as the filling of tensional fissures. We have, however, mapped this area as part of the infracrustals because we think that such structures are best explained by partial melting processes or by segregations in the infracrustal complex.

Growth of feldspar porphyroblasts has locally taken place in the supracrustals immediately overlying the contact. A feldspathic quartzite at Kupevatn has augen of red microcline and is even interlayered with augen gneiss west of the same lake. Such relations might be explained by smallscale «granitization» processes, but could just as well be explained by compositional variations in the original psammitic and pelitic sediments. The most interesting rock with respect to possible «granitization» processes in our area is probably an augen gneiss which occurs between our Svartekari and Visdalseggen units in the closure of the Måsvassdalen synform. The big posttectonic feldspar augen of this rock appear to have formed in the sheared, possibly granulated, contact zone of two major lithologic units in a low-pressure regime.

Acute shear folds and tectonically elongated pegmatite lenticles are locally characteristic for the infracrustal gneisses near the contact towards supracrustals. The contact zone is often crushed and a mylonitic banding has at places almost obliterated the original lithological banding of original psammitic rocks. Individual lithological layers have often been sheared out in the contact zone between major units. This is well displayed in the Måsvatnet schists where for example feldspathic quartzite/meta-arkose alone or feldspathic quartzite/meta-arkose underlain by garnetiferous mica-schist and carbonate schist occur above the contact. The thickness of this feldspathic quartzite/meta-arkose formation varies from virtually nothing to several hundred meters along the contact between Vonen and Måsvatnet. We think that the main shearing in the contact zone took place during or before the first discernable folding of the rock body, although some later movements have produced mylonitic rocks and deformed feldspar porphyroblasts at places.

Planar and linear structures.

Foliation is in some rocks very regular without apparent folding, but close inspection will often indicate the presence of very tight shear folds or isolated fold-hinges with axial planes subparallel to lithological banding and formational contacts.

Most rocks exhibit a penetrative lineation which is defined by preferred orientation of minerals. Infracrustal gneisses have pretectonic augen aligned with their longest dimension parallel to axes of tight shear folds. Mica-schists and other rocks have lineation defined by rods of quartz and small-scale crenulations.

Foliation, penetrative lineation and tight folding with axial-planes subparallel to foliation are all related to the first folding, F_1, in the area. Strong transposition must have taken place during this folding, and the thickness of a layer or rock unit is therefore unrelated to original stratigraphical thickness.

Axial planes of the first folds are folded in similar manner as the foliation, and curved axial planes can often be seen (Fig. 10). This is indicative of nonplane cylindroidal folding or due to a second folding episode, F_2, comparable to a superposed folding further to the west (Bryhni 1962 p. 336). Linear structures related to F_1 and F_2 both trend EW to ESE-WNW (Fig. 11). The plunge in the northern and major part of our area is towards the west while the plunge in the southern part is towards the east as well. NS joints in the Blåfjell gneiss area dip

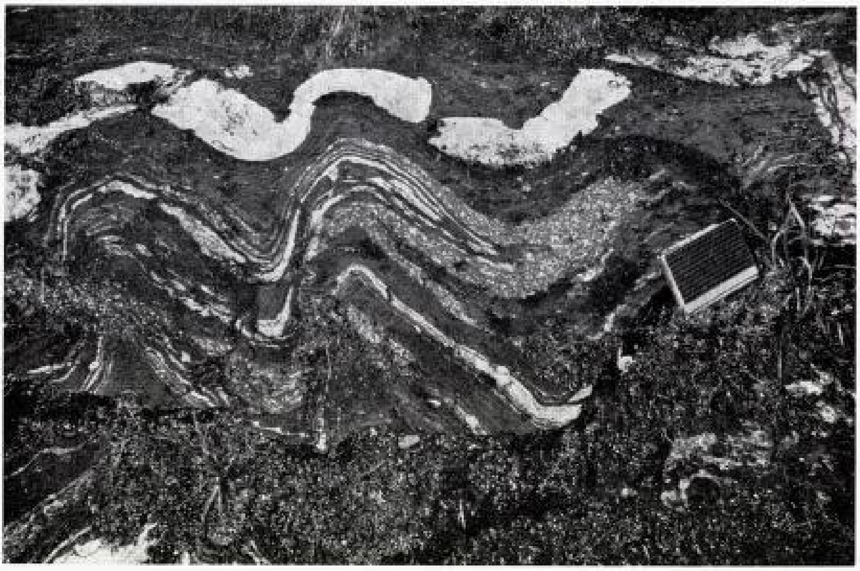

Fig. 10. Fold with curved axial plane. West side of Myklandsdalen near Rygg st. (LP 375/605). Interpreted as F_1-folds folded by F_2.

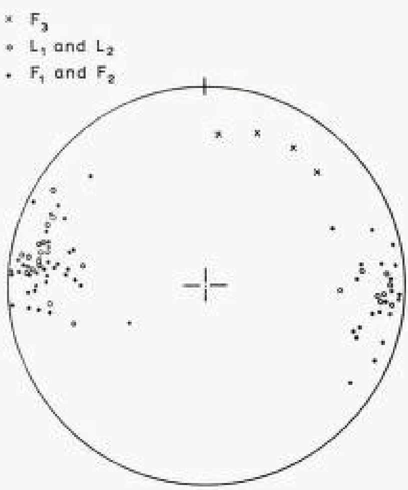

Fig. 11. Projection of linear structures in the area. Lower hemisphere, Schmidt net.

consistently to the east corresponding to their position as cross fractures for west plunging linear structures.

A few minor, open folds with axes trending NS to NNE-SSW and fracture cleavage with corresponding strike have been noted at places (e.g. at tributary to Slettelva, 700 m S of Dalevatnet). The folds have affected a lineation parallel to F_2 in quartz lenticles and must therefore be termed F_3. The fact that the first and second linear structures in the southern part of our area plunge both to the west and to the east can be attributed to this third folding.

Regional structure.

The geologic map and profile show a major synform at Måsvatnet which is bounded to the south by a major antiform. It is believed that the Måsvatnet synform is an F_2 structure comparable with the four «synclines» described by Kolderup from West Norway (1960 p. 8—13) and our supracrustal rocks can, in fact, be followed continuously with similar lithologic succession into his syncline IV north of Breimsvatn.

The supracrustals and the rocks associated with them have been folded in the same general way as the infracrustals, but have experienced more irregular folding with shorter wave-length as if decollement has taken place. The psammitic rocks in the Måsvatnet unit appear mesoscopically to strike perpendicularly to the NS-trending contact south of Dalevatn as if a discontinuity were present, but the relations have proved to be due to the west plunge of the fold axis and to the fact that the supracrustal rocks are deformed with tight folding while the infracrustal gneiss has been laid into broad, open undulations.

Discussion.

We think that the infracrustal rocks west of Breimsvatn are continuous with rocks which occur below the Jostedalen glacier and with rocks of established Precambrian age at the south-eastern margin of the Gneiss region. Infracrustal and supracrustal rocks of our area are, however, interfolded in such a way that age relations are not evident. The assumed basement shares all essential structures with its cover and the present conformity appears to be due to a deformation shared by both complexes.

The main deformation of the rock body in our area took place along EW to ESE-WNW axes. The linear structures formed during the first folding may have been rotated by the second folding, but we have no evidence that they originally diverged much from the axes of second folding. Nor do we have any evidence that the two folding episodes were separated by any significant time interval. We therefore provisionally suggest that the first and second linear structures are due to continued folding on essentially the same EW to ESE-WNW fold axis whereby initial axial planes become eventually refolded.

EW fold axes were found to be characteristic for the Precambrian basement at the south-eastern margin of the Gneiss region (Banham 1968), and one could argue that the supracrustals west of Breimsvatn were Precambrian as well. One has to remember, however, that EW folding is characteristic for large areas of the Norwegian Caledonides (Strand 1949) and has even affected rocks which are generally regarded as Eocambrian rocks and Cambro-Silurian rocks.

The deformation responsible for the second folding in our area probably left its imprint in the radiometric age of 402 m.y. and can perhaps be related to a late EW folding recorded by Banham & Elliot (1965) from the south-eastern margin of the Gneiss region.

The supracrustals west of Breimsvatn might well have a Precambrian depositional age. Birkeland (1958) has described metamorphic supracrustal rocks from a more northern part of the Gneiss area which he thinks were originally Precambrian, and Kvale (1946 b) has suggested that the metamorphic supracrustal rocks he encountered in Bergsdalen and which he related to the Precambrian «Telemark suite» have, in fact, a wide distribution in West Norway. The association of metamorphic psammitic rocks with amphibolite and biotite-plagioclase rocks in our area is not unlike the quartz-schist/metabasalt/metadacite association of Bergsdalen. The correspondence can be extended to include anorthositic rocks because such occur within the upper Bergsdalen nappe (Kvale 1960 p. 191). A possible argument against the assignment of the supracrustals and their associated rocks west of Breimsvatn with rocks of Bergsdalen and the «Telemark suite» is that the psammitic rocks are different. The Bergsdalen and Telemark metamorphic psammites are often ortho-quartzites and quartz schists with a very low content of feldspar while the psammitic types west of Breimsvatn are very feldspathic. The feldspathic quartzites and meta-arkoses west of Breimsvatn are in this respect more closely related to the Eocambrian of Norway (Kolderup 1928 p. 69–70), or to the particular feldspar-rich psammites of the Bandak group in the Precambrian Telemark suite (Bugge 1931 p.

162). Metamorphic supracrustal rocks identical with those which occur in our area are present in Oppdal, Nordmøre (Tingvoll group of Hernes 1965, 1967), Grotli (Strand 1966, 1969) and in the upper Gudbrandsdalen area (Strand 1964). The thick amphibolitic intercalations which occur west of Breimsvatn can then perhaps be correlated with the basic rocks present in flagstones and mica-schist of the Sethø complex at Oppdal (H. Holtedahl 1960, p. 13—17) which sometimes have formed from intrusive sills (Holmsen 1955, p. 138—139) and to gabbro and basalt associated with sparagmites of the Kvitvola nappe (Bjørlykke 1965).

Anorthositic rocks are generally regarded as Precambrian, and we have no reason not to accept this age for the metamorphic anorthosites west of Breimsvatn. At least three different mechanisms may account for the presence of Anorthosite Kindred rocks in our area:

1. Tectonic emplacement like in the central nappe region. A such mode of origin was suggested by Kolderup (1960 p. 11) and it is quite possible that nappes with Anorthosite Kindred rocks in West Norway and the central nappe region have a related tectonic position on each side of the Jostedalen culmination.

2. Intrusion of the Anorthosite Kindred rocks into the supracrustal series, differentiation by settling and later by flow layering to form extensive layers or bodies of contrasting composition.

3. Metasomatic derivation from original sedimentary rocks (Hietanen 1963, Korshinskij 1965).

Our data do not permit further eluciation of the problem of anorthosite formation, but we think that the area west of Breimsvatn lends support to the two first alternatives, and the true story of the formation of Anorthosite Kindred rocks in our area might involve both of them. The lithologic sequence given in Fig. 2 for the Fjordane complex is scarcely a normal stratigraphic succession. The Ommedalen and the Storeskarseggen units contain Anorthosite Kindred rocks similar to those which elsewhere in the Caledonides occur in thrust position. The wide extension of these units, their varied lithology with gneisses, anorthositic rocks, amphibolite and ultrabasite indicates that one is here dealing with plutonic rocks which first become intruded into metasediments, then strongly differentiated and deformed together with the enclosing rocks. Original intrusives would be phyllonitized or broken up into lenticular bodies congruently folded with the enclosing rocks. The mesoscopic style of folding indicates that recumbent F_1 folds are present. It is tempting, therefore, to relate our two anorthosite-bearing lithologic units with the two layers of anorthosite at Gloppen just to the north of our area and join Kolderup (1960, p. 13) in the interpretation of the area as composed of nappes. The Måsvatnet, Visdalseggen and Storeskarseggen schists and meta-arkose contain rather similar rocks and we think *that the Svartekari, Storeskarseggen and Ommedalen gneiss and/or anorthosite units are likely to form three individual thrust nappes. The schists and meta-arkose may then be the Precambrian or Eocambrian mantles between Precambrian fold cores.* Extension of our studies into areas to the west and more detailed mapping would be necessary to solve the structural problems, but we think that our tectonic profile (Fig. 2) indicates that large nappe structures of the type which was postulated by O. Holtedahl (1936) are likely to occur in the area west of Breimsvatn.

The existence of a complicated nappe system within the area west of Breimsvatn makes it plausible that rocks of very different ages were present. True Cambro-Ordovician rocks occur far to the west of our area, but are to our knowledge everywhere bounded by tectonic contacts towards lithologies which can be compared with the supracrustals west of Breimsvatn. We therefore would regard the rocks west of Breimsvatn as Precambrian or Eocambrian until more evidence is available to alter this opinion.

The rocks in our area are obviously polymetamorphic. Their grade of metamorpism corresponds generally to the almandine amphibolite facies although the presence of metamorphic varieties of the Anorthosite Kindred rocks and of Plag-Ga-Clpx «eclogite-like rocks» indicate that granulite facies has been attained at places. Later retrogression has caused saussuritization of plagioclase in gneiss, conversion of anorthosite into schistose meta-anorthosite and phyllonization of various rock types.

Acknowledgements.

One of the authors (I.B.) received funds from the Geological Survey of Norway (NGU) for his field work while the other (E.G.) was sponsored by the Norwegian Research Council for Science and the Humanities (NAVF). The Norwegian Research Council for Science and the Humanities (NAVF) made it possible for us to obtain radiometric age determinations which were carried out by the firm F. M. Consultants Ltd., Herne Bay, Kent, England. Dr. Frank Fitch of the latter firm has given us valuable information on the K/Ar dating of our rocks.

The present study is a result of research in which many have given inspiration and valuable suggestions. Foremost among these are Professor Niels-Henrik Kolderup, Professor Trygve Strand, Ellen S. Kildal and Finn J. Skjerlie who all have worked in the same general area or have been engaged in the same problem as we in the Gneiss region.

We are indebted also to Eli Knap for drawing figures and to Reidar Nordquist and Bjarne Berge for making the thin sections and polished sections.

Special thanks are due to Dr. W. L. Griffin and Professor D. H. Green who corrected the English of this paper.

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ABBREVIATIONS:

N.G.T.: Norsk Geologisk Tidsskrift.

N.G.U.: Norges Geologiske Undersøkelse.

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Manuscript received 30. December, 1969.

AN OCCURRENCE OF NATURALLY LEAD-POISONED SOIL AT KASTAD NEAR GJØVIK, NORWAY

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Abstract.

A 100 m^2 area with exceptionally sparse vegetation in the forest at Kastad, 6.5 km north of Gjøvik, Norway, is described. The area bears no signs of human activity. Determination of heavy metals in the soils of the area shows no really abnormal contents of V, Mn, Co, Ni, Cu, Mo, Ag, and Zn, but a mean concentration of 4.7 % Pb in top soil dry matter. There is a clear correlation between sparsity of vegetation and lead content in the soil, and it has been concluded that the area is lead poisoned.

Most plants normally growing in the area were not found on the lead-rich soils, although some registered species can tolerate considerable lead in the substratum. The vegetation growing on the lead rich soil has an abnormally high lead content, and might therefore have an injurious effect on fauna.

It is presumed that the lead in the soil is accumulated from natural solutions produced during chemical weathering of a galena-bearing quartzite. Registration of areas with sparse or selective vegetation might be used as a tool in lead prospecting.

Introduction.

Nearly ten years ago a program was established in Norway by the National Forest Survey, the Agricultural College, and the Geological Survey involving the investigation of forest soils for the combined purpose of general soil research and geochemical prospecting (Låg 1962, p. 111). In conjunction with its normal inventory work the Forest Survey has collected humus samples from systematically distributed plots in Nord-Trøndelag, Oppland, and Buskerud. These humus samples

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have been analysed chemically and spectrographically at the Agricultural College and the Geological Survey.

In addition to this regional work detailed investigations have been carried out, especially of methodological character. One such investigation includes a closer study of an approximately 0,1 km^2 tract in the Gjøvik district, where lead anomalies occur in the humus (Hvatum 1965). The presence of lead mineralization in the bedrock had earlier been reported (e.g. Føyn 1954, and Eriksen 1962). During a reconnaissance by the present authors on the 18th of August 1967 we observed a small, stony, barren area where sparsity of vegetation led us to suspect the presence of poisoned soil. Chemical analyses of soil and vegetation samples showed this to be the case.

Description of the area.

This stony, barren area lies in generally forested terrain on the Kastad farm (owned by Sverre Braastad) in Gjøvik township, ca. 6.5 km north of the town Gjøvik (Fig. 1). The area is fan-shaped with irregular borders, ca. 20 m long and up to 10 m wide, thus covering approximately

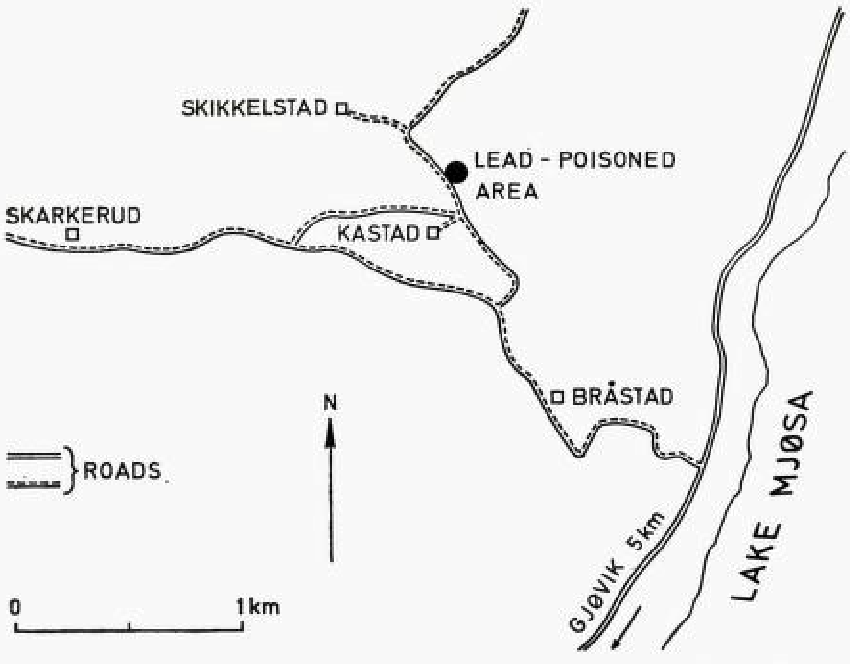

Fig. 1. Location of a lead poisoned, barren area at Kastad, near Gjøvik, Norway.

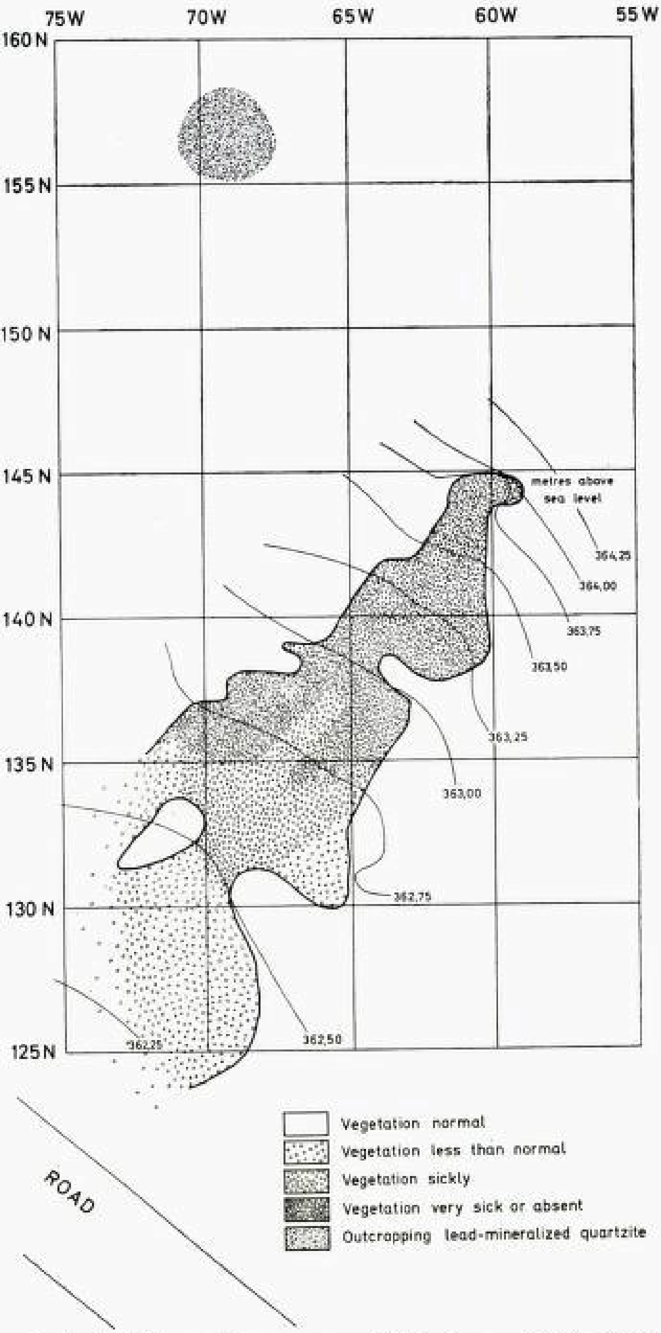

Fig. 2. A lead poisoned, barren area at Kastad, near Gjøvik, Norway.

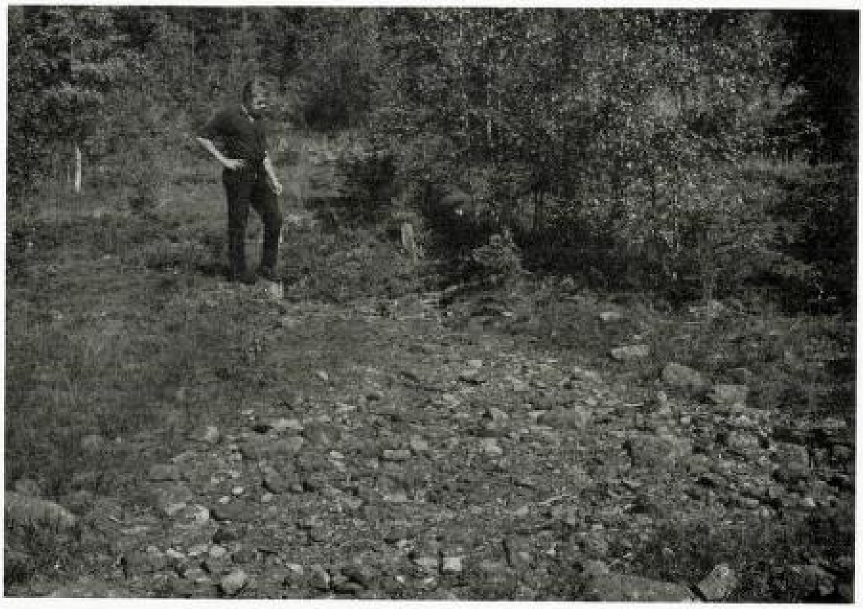

Fig. 3. A lead poisoned, barren area at Kastad, near Gjøvik, Norway.

100 m^2. It forms a slight depression in a weakly sloping field, which rises a little more abruptly NE of the narrow end of the fan (Fig. 2). As the soil between the surface stones bears little or no vegetation, the area appears in summer as a stony, dark patch in the normal green environment (Fig. 3).

Geologically the area belongs to the southeast edge of the Caledonian mountain chain. Cambro-Silurian sediments resting on the Sub-Cambrian peneplain have been overthrust during the Caledonian orogeny by Eo-Cambrian and Cambro-Silurian rocks, producing an imbricate structure with beds of varying composition and age (Skjeseth 1963). Lead mineralization occurs as finely distributed galena with accessory sphalerite and pyrite predominantly in Eo-Cambrian quartzites (e.g. Bjørlykke 1967, with bibliography). Similar mineralization is exploited in Sweden and described among others by Grip (1960, 1967).

Fine sandy morainic material in the order of a meter or so covers much of the bedrock in the Kastad district. The soils surrounding the small barren area of study have podzol profiles. Within the area higher vegetation is patchy, scant or absent. Where the vegetation is scant, weak podzolization with 0—5 cm of bleached horizon occurs. Where vegetation is absent, no real soil development is found. Iron precipitation is weak in the area's narrowest portion (144 N – 61 W), while in the middle of the area (139 N – 64 W) it extends to a depth of ca. 50 cm, and in the widest portion (129 N – 70 W) somewhat deeper. The humus layer varies in thickness from one to six centimeters.

Investigation of the soil.

Within the investigated 1 km^2 tract 90 soil samples were taken in 1966–67 from plots spaced at intervals of 50 m E-W and 25 m N-S. The sampling was carried out in conformity with the procedure used by the Forest Survey (Låg 1968, p. 337). At each plot ten sub-samples were taken within a circular radius of 5.64 m and combined into one sample for analysis.

In 1967 soil samples from the barren area were collected from 35 points at 1–2 m intervals. At each point two samples were taken, one at a depth of 2–4 cm, the other at 5–20 cm. In 1968 three additional samples were taken at a depth of 50 cm (Fig. 4, points B, C, and D). All samples were analysed chemically and spectrographically, and the results are presented in Figures 4 and 5, and in Table 1, 2, and 3.

Investigation of the vegetation.

A study of the distribution of vegetation in a number of 100 m^2 areas within the 1 km^2 tract was made in the summer of 1966. Registrations of the main plant species were made by two students, Kjell Ivar Flatberg and Asbjørn Moen. The plant species found in one such area lying 30 m north of the barren area are listed in Table 4. In the autumn of 1967 samples were taken of the vegetation of the barren area and grouped according to degree of apparent health. Where sufficient substance was available, the samples were analysed chemically (Table 3). Vegetation samples from the barren area were also collected in the autumn of 1968. Determination of the occurring species was made by Professor Eilif Dahl of the Agricultural College of Norway (Table 4).

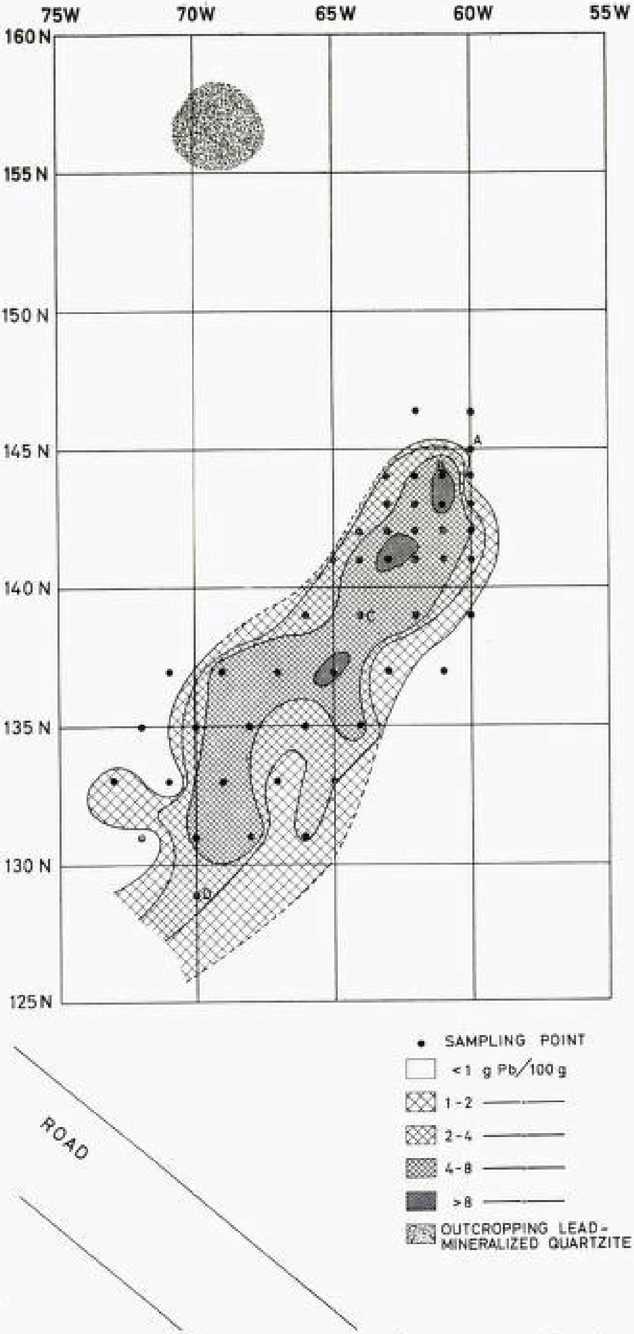

Fig. 4. Lead in top soil (depth: 2—4 cm) in a barren area at Kastad, near Gjøvik, Norway.

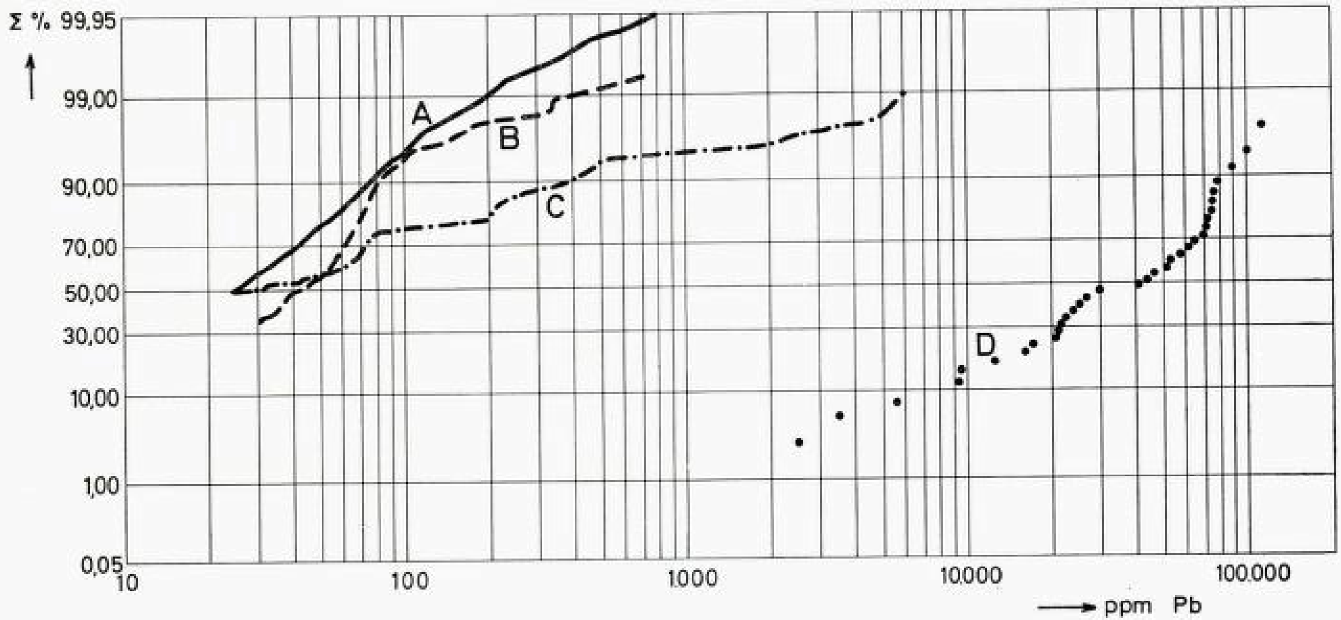

Fig. 5. Frequency distribution of total lead in top soil (mainly raw humus).

A. 2086 samples from Oppland county, Norway (40 000 km²).

B. 168 samples from Vardal, Oppland county (100 km²).

C. 90 samples from Kastad, Vardal (0.1 km²).

D. 35 samples from a barren, lead poisoned area at Kastad (100 m²).

Analytical methods.

All samples were first dried in warm air (ca. 80 °C). The soil samples were then screened using 2 mm aluminium sieves, and 10 g of each fine fraction was ashed at 430 °C. The vegetation samples were similarly ashed whole.

After pulverization in an alumina mill, the ash was analysed semi-quantitatively for V, Mn, Co, Ni, Cu, Mo, Ag, and Pb by optical spectrography (arc excitation of the ash mixed with carbon, visual evaluation of the spectra). The metal content in the ash is given in ppm using the concentration scale 1, 3, 10, 30, 100 etc.

Zn was determined directly in the ash by X-ray fluorescence. In addition all samples from the barren area which gave Pb-values higher than 30 ppm were re-analysed by X-ray fluorescence using Bi as reference element.

pH was determined by glass electrode in soil-water suspensions of unashed samples, the ratio soil/water being 1/2.5.

Discussion.

The frequency distribution of Pb in the top soil (35 samples) of the barren area is compared in Figure 5 with the distribution of Pb in the soils of the surrounding district. The soil samples from the barren area have a lead content which is a) ca. 1600 times as high as that of systematically collected soil samples from Oppland and Buskerud, and b) ca. 1000 times as high as that of the soil samples from the 1 km^2 tract surrounding the barren area.

The total content of lead in the upper 25 cm of the soil over the 100 m^2 barren area is at least one ton. There are no signs of human activity which might account for this. High natural lead content in soils, up to 20 000 ppm, occurring under similar geological conditions has been reported by Brotzen (1967, p. 211).

It appears that the Cu and Ni contents in the soil samples while varying but slightly, show a positive correlation to Pb (Table 1). The concentrations of Cu and Ni, as well as V, Mn, Co, and Ag are not abnormal (e.g. Mitchell 1955, pp. 262—263; Vinogradov 1959; Hawkes and Webb 1962, pp. 363—377; Hvatum 1963; Bølviken 1967, p. 230).

Table 3 shows a clear relationship between the health of the vegetation and the lead content of the substratum. It can therefore be concluded that the sparsity and effected state of the vegetation in the area is a consequence of lead poisoning.

Poisoning of vegetation due to high lead content in soil has been observed elsewhere (e.g. Mudge et al. 1968, p. 28).

As seen in table 5 only two species, *Deschampsia flexuosa* and *Cladonia chlorophaea*, occur in all health classes, regardless of the Pb content of the soil. The species *Betula pubescens*, *Picea abies*, *Pteridium aquilinum*, *Vaccinium myrtillus*, *V. vitis-idaea*, *Agrostis tenuis*, *Polytrichum juniperinum*, and *Webera nutans* together with various *Bryophyta* species can also tolerate a wide range of lead concentrations in substratum. *Agrostis tenuis* and some other grasses have previously been reported as tolerant of high lead concentrations, the tolerance probably being genetically controlled. (Bradshaw 1952; Gregory and Bradshaw 1965; Jowett 1958, 1959, 1964; Wilkins 1957, 1960). Under the present conditions the following species seem to be especially competitive on lead rich soil (Pb > 1 per cent): *Linnaea borealis*, *Agrostis sp.*, *Cladonia coccifera v.pl.*, *C. deformis*, *C. mitis*, *C. ochrochlora*, *C. verticillata*, *C. pyxidata* and other species of *Cladonia*, as well as *Peltigera aphthosa* and various green and bluegreen algaes.

It is well known that lead is poisonous for animals and human beings, even in rather small concentrations. According to Bersin (1963, p. 417) dry plant material used as fodder normally contains 0.1—1 ppm Pb, however plants growing on lead rich soil may contain ten times as much or more.

Our investigation (Table 3) shows that apparently normal vegetation from the poisonous area has a lead content in dry matter in the order of 1000 ppm, and sickly vegetation 3000 ppm (maximum 9000 ppm). Evidently such concentrations of lead must also affect the fauna.

According to table 3 the ash content of the vegetation samples is 5—10 per cent, indicating that the plant material analysed is contaminated by soil. The vegetation itself might therefore contain less lead than shown in the table. However, this does not alter the general conclusions drawn. The plant material eaten by animals must also to some extent be contaminated by soil.

It is of interest to note that natural local concentrations of lead in soil and plants can far surpass that caused by artificial contamination shown by Rühling and Tyler (1968).

The manner in which the lead is bound in the soil of this area has not yet been established. However, the following observations have been made:

a) An attempt to identify lead minerals in the two richest samples (ca. 11 % Pb) under the binocular or by X-ray diffraction on untreated material was not successful. The predominance of the lead probably does not exist in crystalline form.
b) Tests indicate that the lead content in the soil samples is highest in the finest fractions. Till samples separated by heavy liquid (sp.wt. 2.8) showed similar lead content in the light fraction as in the heavy fraction. It seems probable therefore that the lead in the samples is at least in part bound to the surface of soil particles.
c) During ashing of litter from the samples richest in lead it was possible to see small beads of lead being «sweated out». A considerable amount of the lead must be bound to organic matter.

The vertical distribution of lead and pH in the soil of the barren area (Table 2) discredits an upward transport of lead as a dominating mechanism for the enrichment of lead near the surface.

A 5 m^2 outcrop of lead-mineralized quartzite occurs approximately 15 m from the upper end of the poisoned area. (Similar mineralization perhaps also occur over a larger area). Lead is thought to be leached out of the galena-bearing quartzite and transported through the soil in hydrous solution. During wet periods some of this water emerges in the upper portion of the fan (pt A, Fig. 4). Upon reaching the surface the solution loses lead to the soil. The lead accumulates in the soil, which becomes increasingly toxic to the vegetation. As the toxic soil impedes the normal growth of vegetation, the area becomes less protected from erosion. A slight depression is formed, which tends to channel the drainage, causing an increase in the poisoning effect. It is likely that the position, shape, and size of the poisoned area has varied somewhat during postglacial time. Among other factors variations in climatic condition may have led to changes in the degree of poisoning. However, one must assume that the effects of lead poisoning have characterized the area throughout the postglacial period.

Conclusion.

Investigation of these phases of lead geochemistry will be continued. Upon the bases of the work done, however, it seems possible to draw the following tentative conclusions:

a) Lead minerals in quartzite (mainly galena) are decomposed by chemical weathering. The products enter into hydrous solution and are transported away.
b) From this solution lead is reprecipitated in the soil, mainly in the upper humusrich layer. In the course of postglacial time (less than 10 000 years) concentrations in the order of up to 10 % Pb have accumulated in the soil.
c) Such concentration of lead in the soil has resulted in a poisoning of the vegetation. There is a great variance in the ability of different plants to tolerate high lead concentrations.
d) Since lead is toxic to many animal organisms, even in small amounts, such concentrations as found in this case may have an injurious effect on fauna.
e) Localization of areas with a conspicuous sparse or selective distribution of plant species may be of assistance in prospecting for lead ores.

Sammendrag.

En forekomst av blyforgiftet naturlig jordsmonn i Vardal.

I skogen ved Kastad i Vardal, ca. 6,5 km nord for Gjøvik by, er det funnet et nesten vegetasjonsfritt felt på ca. 100 m^2. Feltet er uten synlige tegn på kunstige inngrep. Like i nærheten finnes i fast fjell en blotning av kvartsitt med finfordelt blyglans. Blymineraliseringen hører til fjellkjederandens forekomster. Liknende mineralisering er utnyttet økonomisk i Sverige (Laisvall og Vassbo).

En mistanke om at den sparsomme vegetasjon skyldes naturlig blyforgiftning, ble bekreftet ved kjemiske analyser av jord og vegetasjonsprøver. 35 jordprøver fra rotsonen (5—20 cm dybde) hadde et gjennomsnittsinnhold på 2,4 % Pb, og 35 prøver fra de øverste jordlag (2—4 cm dybde) et gjennomsnittsinnhold på 4,7 % Pb med maksimalinnhold i enkeltprøver på over 10 %. De andre elementer det ble analysert på (V, Mn, Co, Ni, Cu, Mo, Ag, Zn), viste konsentrasjoner som ikke avviker særlig fra normalverdier for jordprøver. Fordelingen av Pb og pH i jordsmonnet tilsier at oppkonsentreringen av bly i de øverste jordlag ikke kan skyldes transport oppover i profilet. Sannsynligvis er blyet blitt frigjort fra den nærliggende blyholdige kvartsitt ved kjemisk forvitring, og deretter i fuktige perioder transportert i vannoppløsning til jordoverflaten i feltets øvre parti (se fig. 2), der det er blitt avgitt, særlig til det humus-holdige sjikt. Blyet er akkumulert i jordsmonnet, som etter hvert har fått så høyt innhold at det er blitt giftig for vegetasjonen.

Vegetasjonsprøver fra det forgiftede feltet viste et gjennomsnittlig blyinnhold på 0,22 % med maksimalt innhold i enkelte prøver på 0,9 %. Blyinnholdet i plantene økte i takt med blyinnholdet i det jordsmonn plantene vokste på.

Planteartenes evne til å tåle høye blykonsentrasjoner i jordsmonnet varierer sterkt. To arter, smyle *(Deschampsia flexuosa)* og laven *Cladonia chlorophaea* ble funnet på selv de mest blyrike deler av det forgiftede feltet, foruten under mer normale forhold i nærheten. I alt 19 arter ble påvist på jord med mer enn 2 % bly i rotsonen, mens det under tilsynelatende normale forhold på tilgrensede felter ble påvist 63 arter.

Undersøkelser over prosesser som fører til opphoping av bly i jordsmonn og vegetasjon vil bli fortsatt. Foreløpig er det konkludert med følgende:

a) Krystallinske blyforbindelser i kvartsitter (vesentlig blyglans) nedbrytes kjemisk og forvitringsproduktene transporteres bort i vandig løsning.
b) Fra den naturlige blyløsning er blyet avsatt i jordsmonnet, i størst mengde i det øverste humusholdige sjiktet. I løpet av postglacial tid (mindre enn 10 000 år) er det blitt anriket bly i jordsmonn til konsentrasjoner opp til størrelsesorden 10 %.
c) Naturlige, sekundære blyanrikninger i jordsmonnet har ført til forgiftning av vegetasjonen. Det er stor forskjell på ulike planters evne til å tåle store blykonsentrasjoner.
d) Da bly er giftig for dyreorganismer, selv i små mengder, må konsentrasjoner som påvist i dette tilfelle ventes å ha sterk innflytelse på den stedlige fauna.
e) Lokalisering av vegetasjonsfrie felter og felter med spesielt selektert vegetasjon kan brukes som hjelpemidler ved leting etter blymalm.

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Table 1

Heavy metals in dry matter of soil from a barren area at Kastad near Gjøvik, Norway.

		Ash %	V ppm	Mn ppm	Co ppm	Ni ppm	Cu ppm	Mo ppm	Ag ppm	Pb %	Zn ppm
Depth 2—4 cm, n = 35	m	50.2	19	177	4	4	51	0.8	<3	4.7	163
	s	13.0	8	86	2	2	58	0.9		3.3	75
Depth 5—20 cm, n = 35	m	91.6	39	340	7	4	35	1.0	<3	2.4	355
	s	3.2	14	225	4	2	24	0.7		1.2	138

m: Arithmetic mean $\Sigma\frac{x}{n}$; s: Standard deviation $\sqrt{\frac{\Sigma (x-m)^2}{n-1}}$

Table 2

Lead content and pH of soil dry matter from different depths in a barren area at Kastad, near Gjøvik, Norway. Sample location shown in fig. 4.

Sample location			cm depth	% ash	% Pb	pH
145 N	60 W	(A)	2—4	65.8	0.3	4.5
145 N	60 W	(A)	5—20	91.8	0.7	5.0
144 N	61 W	(B)	2—4	46.6	11.6	5.4
144 N	61 W	(B)	5—20	83.2	4.4	5.5
144 N	61 W	(B)	50	96.1	1.8	5.3
139 N	64 W	(C)	2—4	62.4	6.7	5.6
139 N	64 W	(C)	5—10	90.8	3.8	5.6
139 N	64 W	(C)	50	99.2	0.6	5.9
129 N	70 W	(D)	2—4	55.3	2.5	5.4
129 N	70 W	(D)	5—20	96.7	1.1	5.5
129 N	70 W	(D)	50	98.2	0.2	5.8

Table 3

Ash and heavy metal content of dry material of soil and vegetation from a barren area at Kastad near Gjøvik, Norway.

Type of samples	Class of Vegetation	Number of samples	Ash		V		Mn		Co		Ni		Cu		Mo		Ag	Pb		Zn	
			per cent		ppm		ppm		ppm		ppm		ppm		ppm		ppm	per cent		ppm	
			m	s_m	m	s_m	m	s_m	m	s_m	m	s_m	m	s_m	m	s_m	m	m	s_m	m	s_m
Vegetation	III Apparently normal	6	5.8	0.6	1.3	0.1	100	13	0.2	0.1	0.5	0.1	1.4	0.1	0.1	0.1	< 3	0.11	0.07	66	10
	IV Subnormal	7	8.5	1.0	2.2	0.1	107	20	0.6	0.1	0.8	0.1	2.9	0.9	0.1	0.1	< 3	0.28	0.12	99	25
	V Sickly	2	9.5	2.4	2.8	0.7	80	38	0.6	0.1	0.9	0.2	2.8	0.7	0.1	0.1	< 3	0.35	0.12	86	26
Soil at 2—4 cm depth.	III Apparently normal	9	42.3	5.3	17	3	218	35	3.0	0.5	3.5	0.4	13	3	0.6	0.2	< 3	1.69	0.72	139	15
	IV Subnormal	10	50.2	5.1	17	3	222	46	3.6	0.4	3.5	0.4	21	4	0.6	0.2	< 3	2.23	0.31	200	29
	V Sickly	10	50.3	5.5	21	3	206	36	4.8	0.8	4.7	0.8	23	3	0.6	0.2	< 3	2.87	0.53	188	34
	VI Sick or absent	16	54.9	2.7	19	2	172	18	4.4	0.7	4.7	0.4	90	17	1.2	0.3	< 3	7.43	0.57	168	14
Soil at 5—20 cm depth.	III Apparently normal	9	93.1	1.0	46	5	279	84	7.3	2.7	2.3	0.5	19	6	1.1	0.2	< 3	1.29	0.35	300	78
	IV Subnormal	10	91.9	0.7	42	5	311	79	4.9	0.7	0.9	0.0	32	8	1.0	0.2	< 3	1.99	0.39	365	49
	V Sickly	11	90.8	1.1	37	4	464	67	6.8	0.6	3.1	0.5	26	3	1.0	0.2	< 3	2.32	0.35	350	43
	VI Sick or absent	16	90.7	0.7	34	3	279	38	6.2	0.5	5.2	0.5	56	5	1.2	0.2	< 3	3.09	0.17	430	23

m: Arithmetic mean $\Sigma \frac{x}{n}$; s_m: Standarddeviation of the mean $\sqrt{\frac{\Sigma (x-m)^2}{n(n-1)}}$

Table 4

Plant registrations around and within a barren area at Kastad, near Gjøvik, Norway.

	Vegetation normal to sick					
	I	II	III	IV	V	VI
Betula pubescens	x	x			x	
Juniperus communis	x					
Picea abies	x	x	x		x	
Pinus silvestris		x				
Populus tremula		x				
Sorbus aucuparia	x	x				
Alchemilla sp.	x					
Anemone nemorosa	x					
Chamaenerion angustifolium	x					
Euphrasia frigida	x					
Galeopsis sp.	x					
Hieracium silvaticum coll.	x					
Linnaea borealis					x	
Lycopodium annotinum	x					
Maianthemum bifolium	x					
Melampyrum pratense	x					
Melampyrum silvaticum	x					
Oxalis acetosella	x					
Potentilla erecta	x	x				
Pteridium aquilinum		x	x	x		
Ranunculus acris	x					
Rumex acetosa	x					
Rumex acetosella	x		x			
Rubus idaeus	x	x				
Solidago virgaurea	x	x				
Trientalis europaea	x					
Vaccinium myrtillus	x	x	x	x	x	
Vaccinium vitis-idaea	x	x	x	x	x	
Veronica chamaedrys	x					
Veronica officinalis	x	x				
Viola riviniana	x					
Agrostis canina v. fascicularis (?)					x	
Agrostis tenuis		x	x	x	x	
Anthoxanthum odoratum	x					
Carex leporina	x					
Carex pallescens	x					
Carex pilulifera	x					
Deschampsia caespitosa	x					
Deschampsia flexuosa	x	x	x	x	x	x
Festuca ovina	x					
Festuca rubra	x					

	Vegetation normal to sick					
	I	II	III	IV	V	VI
Luzula multiflora	x					
Luzula pilosa	x	x	x			
Phleum commutatum	x					
Poa pratensis	x					
Barbilophozia lycopodioides	x	x				
Brachythecium sp.		x				
Bryophyta sp.		x	x	x		
Dicranum bergeri	x					
Dicranum scoparium			x			
Dicranum undulatum		x				
Dicranum sp.	x	x				
Hylocomium splendens	x	x				
Hylocomium squarrosum				x		
Pleurozium Schreberi	x	x	x	x		
Polytrichum juniperinum		x	x	x	x	
Polytrichum piliferum			x			
Polytrichum strictum			x	x		
Polytrichum sp.	x					
Ptilium crista-castrensis	x	x				
Webera nutans		x			x	
Webera sp.				x		
Cetraria islandica		x	x			
Cladonia arbuscula	x	x				
Cladonia chlorophaea		x	x	x	x	x
Cladonia coccifera v. pleurota			x	x	x	
Cladonia cornuta	x					
Cladonia crispata	x					
Cladonia deformis					x	
Cladonia gracilis		x		x	x	
Cladonia mitis			x	x	x	x
Cladonia ochrochlora					x	
Cladonia pyxidata						x
Cladonia rangiferina		x	x			
Cladonia silvatica coll.	x	x				
Cladonia verticillata			x		x	
Cladonia sp.				x		x
Peltigera aphthosa					x	
Algae sp.						x

I 1. July 1966. 100 m^2 approx. 30 m north of the barren area.
II 5. October 1968. Around the barren area. Normal vegetation.
III 5. October 1968. Within the barren area. Apparently normal vegetation.
IV 5. October 1968. Within the barren area. Less than normal vegetation.
V 5. October 1968. Within the barren area. Sickly vegetation.
VI 5. October 1968. Within the barren area. Vegetation very sick or absent.

Table 5

Occurrence of plant species in relation to the health of the vegetation and the lead content of substratum.

Health classes, normal to sick (see tables 3 and 4) Average lead content in the root zone of the soil (per cent)	II <1.0	III 1.3	IV 2.0	V 2.3	VI 3.1	Number of species
Deschampsia flexuosa, Cladonia chlorophaea	—	—	—	—	—	2
Betula pubescens, Webera nutans	—			—		2
Picea abies	—	—		—		1
Vaccinium myrtillus, V. vitis-idaea, Agrostis tenuis, Polytrichum juniperinum	—	—	—	—		4
Pteridium aquilinum	—	—	—			1
Cladonia coccifera v. pleurota		—	—	—		1
Cladonia mitis		—		—		1
Cladonia verticillata		—	—	—	—	1
Cladonia sp.			—	—	—	1
Linnaea borealis, Agrostis sp., Cladonia deformis, C. ochrochlora, Peltigera aphtosa				—		5
Cladonia pyxidata					—	1

INVESTIGATIONS AT THE SOUTH-WESTERN BORDER OF THE SPARAGMITE BASIN (GAUSDAL VESTFJELL AND FÅBERG VESTFJELL), SOUTHERN NORWAY

by

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Contents.

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With 22 illustrations in the text.

Abstract.

Descriptions are given of lithologies deposited in the original Sparagmite Basin. These rocks can be classified into a parautochthonous older part and an allochthonous younger unit. Most of the formations of the Sparagmite Group can be traced from the lake Mjøsa, where they have their typical development, to the Gausdal area without any great changes of the lithology.

The Biskopåsen Conglomerate (earlier Biri Conglomerate) is found in four separate bodies (three of them lens-shaped) indicating deposition in deltaes. The Biri Limestone in Gausdal is a facies variant of the Biri Limestone at Mjøsa. Among the carbonate deposits in Gausdal there are sandstones, conglomerates (with boulders of carbonate rocks, granites and quartzites), and beds with ooliths. The Moelv Tillite in Vestre Gausdal is developed as a shale conglomerate with a matrix consisting of graded bedded silt and clay. The boulders consist mostly of grey granite, light quartzites and the underlying carbonate rocks.

In the Gausdal area the deposits within the Sparagmite Group indicate a transport of the material from a westerly point, and that most of the clastic material was weathered before deposition. Some of the formations display rapid changes of sedimentary facies.

The structures have disclosed that the youngest rigid members behaved differently from the older members during the Caledonian orogeny, the former being derived allochthonously from the parautochthonous and folded older members. The rigidity of the south-western border of the basin resulted in a dragging of the fold axes, and to the west even a sheltering effect where folds die out.

Introduction.

Location of the area.

The area studied is located between 61° 0′ and 61° 15′ N. Lat. and 0° 30′ and 0° 46′ Long. West of Oslo. The maps which were used are:

Synnfjell 1 : 100 000 Gradteigen F 31 aust.

Deutsche Heereskarte, Norwegen 1 : 50 000, F 31 0 Synnfjell.

Norway—Norge 1 : 50 000, Sheet 1717 II Ams series M 7 II Synnfjell. In addition to these, preliminary copies of a new 1 : 50 000 map being prepared by the Geographical Survey of Norway were also used. From 1960 aerial photographs of the eastern part of the area were available, and the map Fig. 2 is based on them.

Aim and means of study.

The purpose of the investigations was to determine firstly the position of the western and south-western border of the Sparagmite Basin within the Vestre Gausdal area, and secondly the influence of the border on the sediments deposited in the basin and on the tectonic development of the area.

The rocks of the area are mainly coarse-grained ranging from sandstones to conglomerates; generally they are rudites with angular fragments (sparagmites). Sedimentary structures are few. As most of the rocks are coarse and display marked lateral variations, it is difficult to choose representative samples. Nevertheless, many specimens were collected and the minerals and textures seen in thin-sections have been described.

Of the tectonic structures observed in the field lineations are the most numerous and tend to be prominent in limestones and shales. Folds are rarely seen. All measurements are based on the 400° scale, and planes of bedding and joints have been projected on to the upper hemisphere of a Schmidt net. As the bedrock is poorly exposed, the geological maps have entailed a good deal of interpretation.

Previous investigations of the area.

The first description of the area was that of Hiorthøy in 1785. He found that the rocks were usually grey in colour and rich in quartz. In 1826 Keilhau estimated the position of the southern border of the Sparagmite Basin from west to east not far from where it is now suggested to lie.

Kjerulf (1857) mentioned a limestone occurring at Bratland and Forsetseter which could be traced to Biri. In 1873 he described a peculiar limestone and grey sparagmite together with a coarse conglomerate. These lithologies could be traced from Mjøsa to Gausdal. He found that above the conglomerate in Gausdal there is a younger limestone at a higher level than the one at Biri. In 1879 Kjerulf described a section from the Herfjell area (Fig. 1, E 3):

Glistening shale with lumps of quartz

Blue-quartz-sandstone and
limestone alternating 700′
together with other quartzites

Conglomerate with fragments of granite (grey granite),
fine-grained grey sparagmite and shale, limestone like the Glomstad-limestone.

Bjørlykke (1891) described the Biskopåsen Conglomerate as grey granite boulders in a matrix of grey sparagmite. He remarked on the different varieties of limestone and estimated their total thickness as ca. 120 m (in Roppa). Above the limestone he found 6 m of dark shale and 12 m of quartzite: these are overlain by a graptolite-bearing shale.

Münster (1891) described the limestone from Roppa to Slåtbakken seter (Dekken seter on the map Fig. 1, J 4). He suggested that this was the Biri Limestone.

Bjørlykke (1893) introduced the name «The Sparagmite Formation» for those deposits older than the Biri Limestone. He followed the limestone from the west slope of Vestre Gausdal under Herfjell to Biri. The upper boundary of the Biri Limestone he described as a calcareous sparagmite with fragments of granite and quartz. He found that the limestone was overlain by a shale containing fragments 3—9 cm in diameter. Above this shale is a thinner shale which is in turn overlain by a 5—50 m thick quartzite. Bjørlykke asserted that the secondary cleavage is steeper than the primary one of the area.

Törnebohm (1893) presented a thorough description of the limestone in Gausdal. Under the Biri Limestone he found a coarse conglomerate extending from Biskopåsen at Mjøsa to Hindalssjøen (Fig. 1, F 4). He established that the boundary between the conglomerate and the Biri Limestone was of some importance — since the two formations were deposited under essentially different conditions. Furthermore he found limestone at several levels. The limestones he found drew an arched boundary. Inside the boundary he thought there were uniform grey sparagmites and shales, whereas outside there were various other rock types.

Törnebohm (1896) repeated his earlier (1893) views in considering the geology of this area but did, however, add that the coarse sparagmites show rapid changes of sedimentary facies, and that rocks of similar lithology are to be found at all horizons. He also mentioned that the

Biskopåsen Conglomerate (earlier Biri Conglomerate) could be followed from Fåvang to Skjellbreidvatn (Fig. 1, G 5).

Münster (1900) referred to all limestones in the area by the name Biri Limestone. He noticed that the limestone is overlain by a grey shale containing fragments, especially of granite, and he found that the younger sparagmite could be mistaken for the Biskopåsen Conglomerate.

Münster travelled from Mjøsa to Herfjell and Verskei (Fig. 1, D 2) without finding what he called «the younger formation» (at Ringsaker he measured it as being 350 m in thickness). At Herfjell he found that the shale containing fragments was overlain by a shale without fragments, and above the shales was a «sparagmite-sandstone». He thought the Biri Limestone had its maximum thickness to the north-west, and that the younger sparagmite in the north-west is very thin, increasing in thickness towards the south-east. He supposed that calcareous arkoses within the younger sparagmites at Ringsaker and Biri were equivalents of calcareous arkoses in Gausdal. Thus he concluded that the conglomerate at Moelven railway-station (the Moelv Tillite) corresponds to the shale conglomerate at Nyseter and Herfjell (Fig. 1, E 3).

Bjørlykke (1905) drew the southern and western border for the older dark sparagmite (with conglomerate and limestone) along the valley Vismunddalen (Fig. 1, I 6) and the mountain Herfjell to the hill Evenvoldkampen (between Vestre Gausdal and Østre Gausdal). He remarked that here there were various fold systems, and thrusts with imbricate structure.

Holtedahl (1920) studied the limestones of south-eastern Norway and found that the Biri Limestone at Brattland (Fig. 1, A 3) is a very pure carbonate deposit.

The present investigations in this area were started in 1959 and continued until 1962.

Outline of the geology.

The Brøttum Sandstone and Shale is the oldest formation within the area. It can be traced from Mjøsa without any lithological changes. Upwards it grades into the overlying Biskopåsen Conglomerate. The Biskopåsen Conglomerate is found in four separate bodies; the uppermost parts of the conglomerate are calcareous.

Within this area the Biri Limestone is a facies variant of the Biri Limestone at Mjøsa. It may be divided into four interdigitating members:

Nyseter Limestone	ca.	0—20	m
Fjello Sandstone	»	0—200	»
Vismund Limestone	»	0—20	»
Åltjerna Limestone	»	0—15	»

The Åltjerna Limestone is a black limestone and shale alternating with compact black limestone with white calcite veins. It is this part of the formation that most closely resembles the Biri Limestone at the type locality. The Vismund Limestone is a coarse calcareous arkose with gravel conglomerates, and displays graded bedding. The Fjello Sandstone is either a coarse arkose or a medium-grained feldspathic sandstone. It resembles the Ring Sandstone at Mjøsa, which it is a part of. The Nyseter Limestone resembles the Vismund Limestone, but upwards it is enriched in carbonate and contains oolitic layers in its uppermost parts.

Within the area there is a sharp boundary between the Biri Limestone and the Moelv Tillite. This is exposed at Nyseter (Fig. 1, E 3), south of Herfjell (Fig. 1, E 3) and in the stream Briggebekken (Fig. 1, D 4). Up to the present time these are the only places known in Southern Norway where the tillite rests on limestone in an autochthonous position.

The Moelv Tillite in this area is a boulder clay. The clay matrix has varves which bend conformably around the boulders. The boulders measure up to 1 m and comprise fine-grained quartzites, coarse grey granite, fine-grained light granite, gneisses, mylonites, limestone, pegmatite-quartz, shale fragments and metaamphibolites. The tillite grades upwards into the overlying Ekre Shale by a decrease in the boulder content. The Ekre Shale, in this area, is a varved shale which grades upwards into the overlying Vangsås Formation, the transition marked by an increasing number of sandstone beds.

The Vangsås Formation comprises the members Vardal Sandstone and Ringsaker Quartzite. Since they are difficult to separate in the field, a hypothetical boundary has been drawn on the maps and sections. The Vangsås Formation can be followed from Mjøsa to Gausdal without facies variations. In Vestre Gausdal it grades into the overlying Cambrian shales; the transitional zone is ca. 1 m in thickness.

It is supposed that the Sparagmite Basin, as a separate basin of sedimentation, was brought to an end during the lower Cambrian transgression.

Thin-sections show that most of the clastic feldspar in all formations is weathered, a feature suggesting that material supplied to the Sparagmite Basin was derived from weathered Precambrian rocks. Some of the weathering may, however, be post-depositional. Fjello Sandstone and younger deposits have clastic grains of dark quartz and feldspar. All rocks in this area are metamorphosed to the lower part of Eskola's greenschist facies.

During the Caledonian orogeny the rocks deposited in the Sparagmite Basin were compressed in a southerly direction. The rigid Vangsås Formation (earlier the Quartz-Sandstone Formation) was thrust out of the basin while the older formations were folded up against the south-western border. The folds formed in connection with the «Quartz-Sandstone nappe» (Skjeseth, 1963) die out westwards, possibly because the deposits in the west were sheltered by the rigid block forming the western border of the Sparagmite Basin. A change of trend of fold axes may be due to a dragging effect along the western margin. The joints may be either tension joints perpendicular to the fold axes or post-Caledonian fissures of north-south trend.

Acknowledgements.

This paper is a part of a cand. real. thesis presented at the University of Oslo. No later investigations have been added. The field work for the thesis formed part of the regional mapping of late Precambrian and Eocambrian formations in Southern Norway started by the Geological Survey of Norway (NGU) under the leadership of Professor S. Skjeseth. NGU kindly defrayed the field expenses.

The author is grateful for all kinds of help received during the field work and the preparation of this manuscript. In particular, sincere thanks are due to Professor S. Skjeseth and State Geologist P. Holmsen for their critical reading of the manuscript, and to Professor O. Holtedahl and Dr. P. Jørgensen for helpful discussions.

I. Dillan has drawn all the illustrations, and O. Brynildsrud, M. Moen and H. Gjølme have made the photographs. Dr. D. Roberts has kindly corrected the English manuscript.

Stratigraphy and Petrography.

In Gausdal and Fåberg Vestfjell area have the rocks belonging to the Sparagmite Group (Late Precambrian and Eocambrian) been divided into several lithostratigraphical units. The stratigraphical terminology used is the one proposed by Henningsmoen (1955).

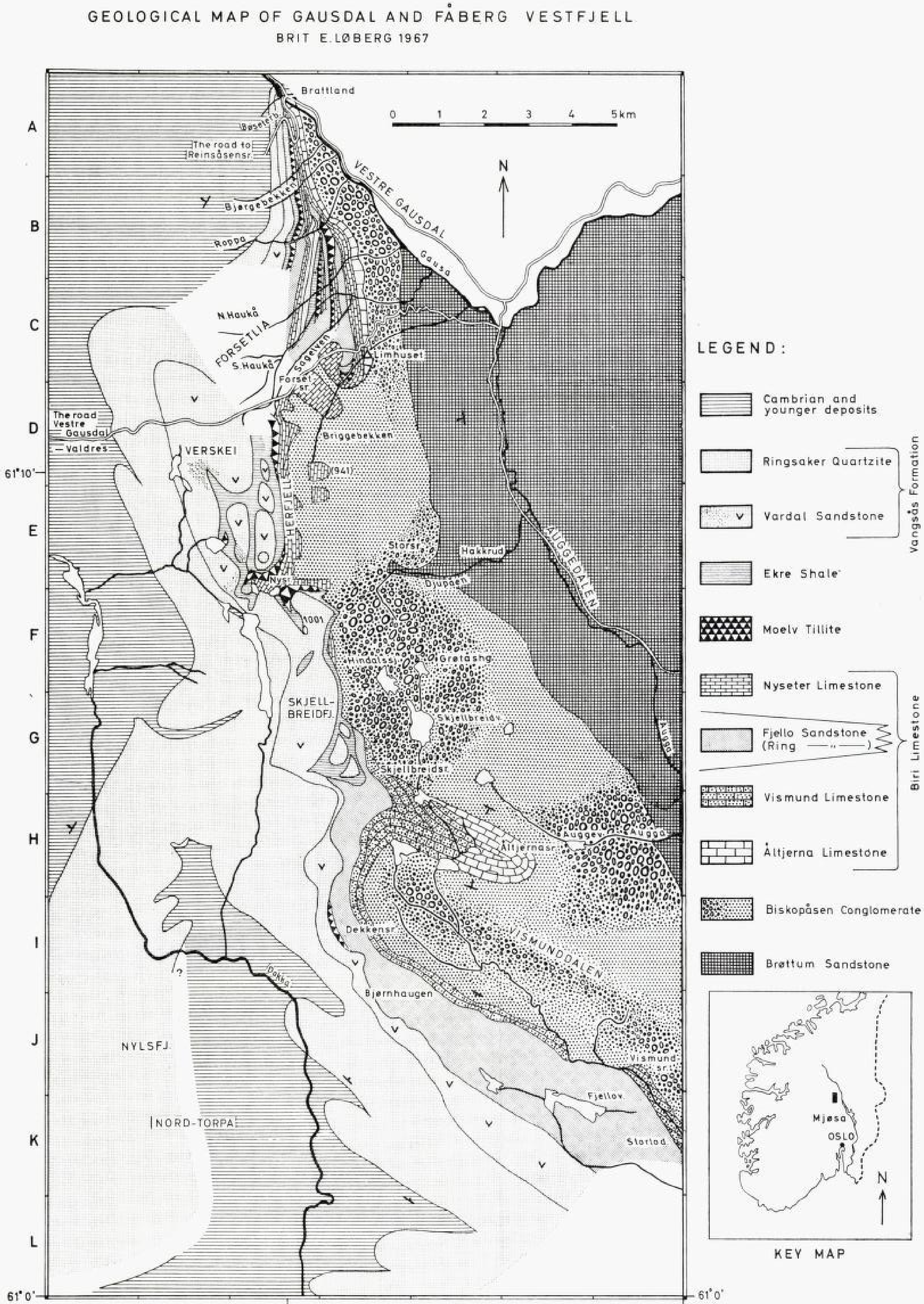

Fig. 1. Geological map of Vestre Gausdal and Fåberg Vestfjell.

Some of the classical names for formations of the Sparagmite Group have recently been changed (Bjørlykke, Englund and Kirkhusmo, 1967) and here the revised names will be used.

Nomenclature for latest Precambrian and Eocambrian formations in S. Norway:

Vogt (1924), Holtedahl (1960) and Skjeseth (1963)	Bjørlykke, Englund and Kirkhusmo (1967)	This paper
Ringsaker Quartzite } Quartz Vardal Sparagmite } Sandstone	Ringsaker Quartzite } Vangsås Vardal Sandstone } Formation	{ Ringsaker Quartzite { Vardal Sandstone
Ekre Shale	Ekre Shale	Ekre Shale
Moelv Conglomerate (= Moelv Tillite	Moelv Tillite	Moelv Tillite
Moelv Sparagmite	Ring Formation	Biri Shale
		Fjello Sandstone > and
Biri Shale and Limestone	Biri Shale	Limestone
Biri Conglomerate	Biskopås Conglomerate } and	Biskopåsen Conglomerate
Brøttum Shale and Limestone	Limestone	Brøttum Sandstone and Shale
Brøttum Sparagmite	Brøttum Formation	
Elstad Sparagmite	Elstad Formation	

Brøttum Sandstone and Shale

The Brøttum Sandstone and Shale is the oldest formation in the area. The formation consists of closely alternating dark feldspatic greywackes and dark shales, the thickness of each layer varying from a few centimetres to several metres. The formation is only found in the eastern part of the area, where it dips westwards under younger strata. The base is not exposed.

In the river Søndre Haukå (Fig. 1, C 3-4) the formation gradually changes upwards to a coarse conglomerate (the Biskopåsen Conglomerate). The transitional zone is more than 100 m in thickness.

As the section shows (Fig. 3) the Brøttum Sandstone and Shale at the south-western border of the basin is apparently 785 m thick, discounting possible disturbances. In the southern part of the Sparagmite Basin, Skjeseth (1963) estimated the thickness of the formation to be 1000–1500 m.

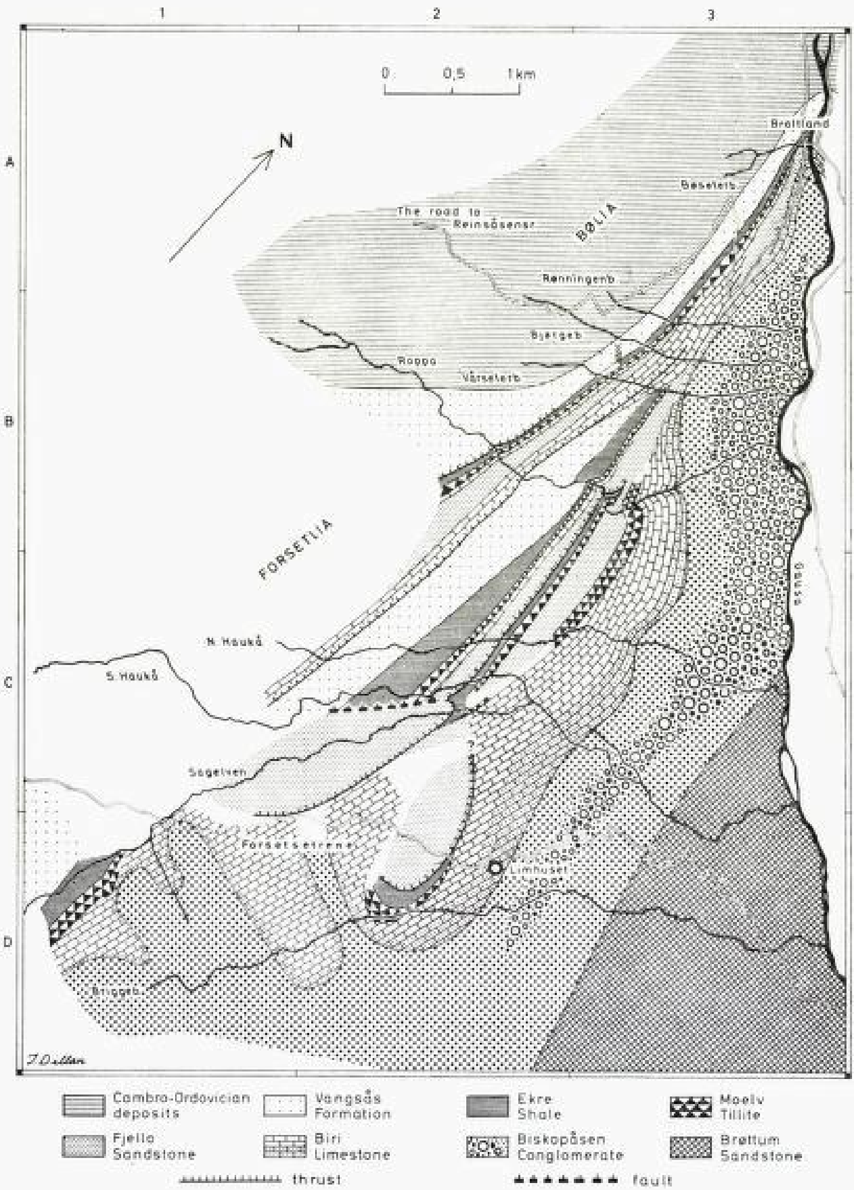

Fig. 2. Geological map of the western slope of Vestre Gausdal.

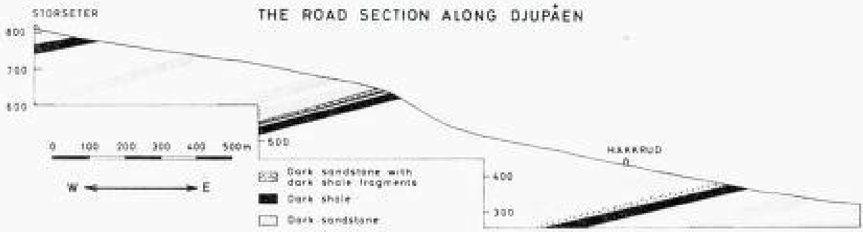

Fig. 3. Section along the river Djupåen. The heights are in metres above sea level.

The grain size of the Brøttum Sandstone ranges from that of the matrix (less than 0.05 mm) to ca. 1 cm in diameter. Grains smaller than 1 mm are angular, while those of 1 mm or more are angular to subrounded and are fractured. Open fractures are filled with matrix or recrystallized quartz. The clastic material is dispersed in a dark matrix consisting of sericite, chlorite, fine-grained quartz, small pyrite crystals and finely disseminated graphite. The rock fragments are quartzite, shale and granite. Most of the clastic material consists of grains of quartz with strongly undulose extinction, and feldspar (mainly K-feldspar) moderately to strongly sericitized. All feldspar material is coloured light brown. Plagioclase is albite and Na-rich oligoclase. Clastic grains of biotite are less than 0.5 mm in size. Their colour varies from tan to light reddish brown, and they are bent so producing a wavy extinction.

From the character of the Brøttum Sandstone and Shale within the area, it is clear that the material was transported only a short distance before its subaqueous deposition. It evidently came from an area where quartzite, shale and granite were undergoing erosion.

Biskopåsen Conglomerate.

The Brøttum Sandstone and Shale gradually passes up into the overlying Biskopåsen Conglomerate. No sharp boundary has been found. In the transitional zone in Søndre Haukå, the Brøttum Sandstone contains intercalations of conglomerate beds. The Biskopåsen Conglomerate forms the floor of the valley Vismunddalen (Fig. 1, I-J 5-7), and the western slopes of the valleys Auggedalen (Fig. 1, E-H 4-7) and Vestre Gausdal (Fig. 1, A-C 3-4).

Within the area the Biskopåsen Conglomerate crops out in four separate bodies as shown on the map (Fig. 1). These bodies are connected to one another by dark arkoses which grade laterally into the conglomerates. The matrix of the conglomerates is a coarse-grained dark arkose, freequently containing some calcite.

Three of the conglomerate bodies exhibit a lenticular outcrop pattern. The one at Augga (Fig. 1, H 7) is ca. 100 m in maximum thickness, the one at Skjellbreidvatn (Fig. 1, F-G 4-5) more than 150 m thick, and the one in Vestre Gausdal (Fig. 1, B 4) at least 170 m in thickness.

In Vismunddalen (Fig. 1, I 6), although good exposures are lacking a zone of coarse conglomerate can be followed along the length of the valley. In the coarser parts the long axes of the boulders reach a maximum of 25 cm, decreasing to less than 4 cm 1 km to the north and south. The boulders are well rounded and consist of light quartzite, pegmatite quartz and grey granite.

At Augga (Fig. 1, H 7) the conglomerate is well exposed in the river where it is ca. 100 m in thickness. The boulders are 20–30 cm across and well rounded. They consist of grey granite, light quartzite and gneiss. Upwards the grain size decreases, and the pebble material then consists mainly of light quartzite. The coarsest part of the conglomerate at Augga is exposed near the lake Auggevatn (Fig. 1, H 6), in the river Augga and along the road that runs parallel to the river.

Southwards from the lake Auggevatn the conglomerate thins out and the boulders (mainly of light quartzite) diminish in size and are scattered in a matrix which gradually becomes more fine-grained. The southern margin of the Augga conglomerate body is situated approximately two kilometres south of lake Auggevatn.

North of Auggevatn abundant erratics (2–5 m^3 in size) of conglomerate with well rounded boulders 15 cm in diameter suggest that the conglomerate at Augga extends northwards for at least one kilometre. Westwards the conglomerate is overlain by a dark arkose which, although devoid of boulders, contains some scattered fragments (1.5 cm) of white quartzite. This dark arkose is the youngest exposed stratum of the formation at Augga.

A plateau ca. 900 m a.s.l. surrounds the lake Skjellbreidvatn (Fig. 1, G 5). Here the Biskopåsen Conglomerate is extencively exposed on all hills and in all the river beds. The matrix of the conglomerate in this particular body contains more calcite than at any other locality within the area. This calcite is thought to be primarily of clastic origin, probably derived from a source outside the area and possibly from the «Brøttum Limestone» between the Brøttum Sandstone and Shale and the Biskopåsen Conglomerate (Skjeseth, 1963, p. 28).

The oldest part of the Biskopåsen Conglomerate at Skjellbreidvatn is found at the top of the steep slope west of the valley Auggedalen and on the east slope of the hill Grøtåshaugen (Fig. 1, F 5). There it is a dark arkose with fragments of light quartzite, similar to the upper layer of the conglomerate at Augga.

The size and number of boulders increase westwards. In a zone between Grøtåshaugen and the lake Hindalssjøen (Fig. 1, F 5) the maximum long axes of the boulders measure 30—40 cm. The thickness of the conglomerate and the size and number of the boulders decrease northwards and southwards from Hindalssjøen. West of Hindalssjøen lengths of boulders rarely exceed 10 cm. Further west the size and number of the boulders decrease, and the uppermost part of the deposit is represented by a bluish-grey calcareous arkose.

In its coarsest parts, the conglomerate contains boulders which consist almost exclusively of a coarse-grained grey granite, sometimes porphyric. The matrix is a coarse calcareous gravel with angular pebbles. Outside the very coarsest zone, the conglomerate contains a larger number of boulders of gneiss and quartzite. Pebbles in the fine-grained conglomerate at the margins of this body consist almost solely of quartzite.

In V e s t r e G a u s d a l the Biskopåsen Conglomerate is best studied along the 50 m deep canyon of the river Roppa (Fig. 1, B 3-4). A section is drawn along the edge of the canyon (see Fig. 4). The oldest part of the conglomerate in the Vestre Gausdal area is found in the river Søndre Haukå (Fig. 1, C 3-4) where the Brøttum Sandstone and Shale gradually passes up into the conglomerate. In the lower and upper beds of the conglomerate the boulders are scattered, while in the central parts of the body boulders are observed to be tightly packed and their maximum lengths frequently exceed one metre: the boulders are elongated but well rounded. In the coarsest part of the conglomerate the boulders consist almost entirely of grey granite. Where the conglomerate is moderately coarse, the boulders comprise grey granite, gneisses, light quartzite, dark sandstone and dark quartzite. In the most fine-grained horizons near the top of the deposit, the pebbles consist of light quartz and quartzite. Imbrication of the boulders indicates a transport of material from the west or north-west. The uppermost layer of the conglomerate in Vestre Gausdal is a bluish-grey arkose which gradually passes up into a calcareous arkose and the overlying Biri Limestone.

The matrix of the Biskopåsen Conglomerate in Vestre Gausdal has been studied in thin-section. The clastic material is the same as that of the Brøttum Sandstone and Shale, but the ratio clastic material: fine-grained matrix is higher for the conglomerate.

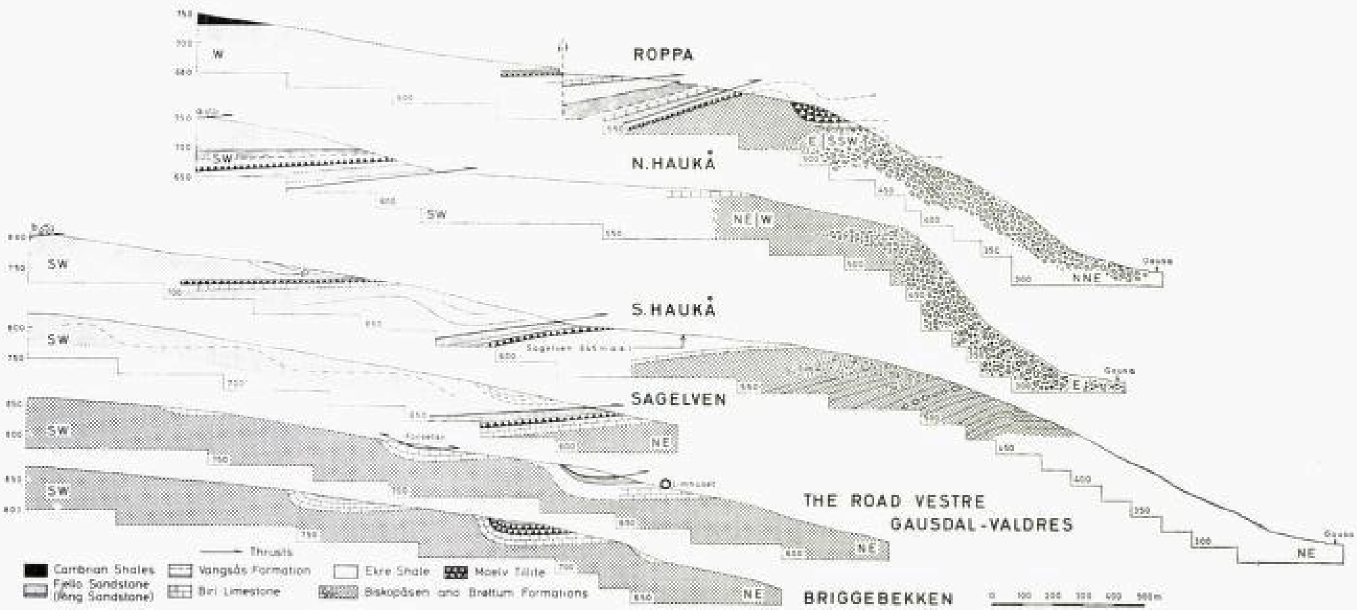

Fig. 4. Sections in Forsetlia, a and b are blocks of limestone. The vertical scale is twice the horizontal scale, the heights given as metres above sea level. (See also Fig. 21).

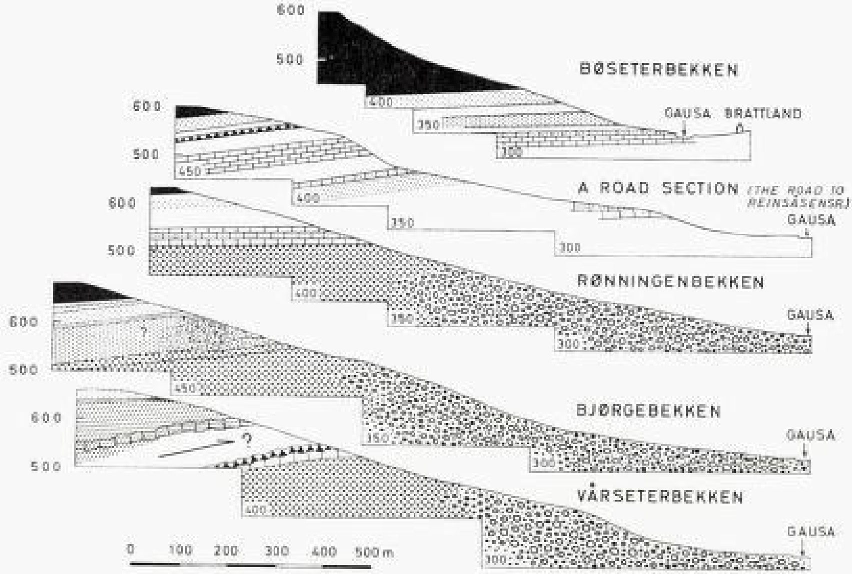

Fig. 5. Sections in Bølia; horizontal and vertical scales equal. (See also Fig. 21).

Fractures in the clastic grains are filled with recrystallized quartz or sericite and occasionally by chlorite. The chlorite, which sometimes penetrates clastic grains of quartz and feldspar, is greyish green (weak pleochroism) with an anomalous blue interference colour and positive elongation.

In consideration of sedimentation conditions and environment, it is likely that the conglomerate bodies were deposited as river deltaes along the south-western coast of the Sparagmite Basin. Whether the conglomerates are of exactly the same age is not known, though as they are all younger than the Brøttum Sandstone and Shale and older than the Biri Limestone, they must have been deposited at approximately the same time. With regard to the changing lithological character of the sediments, from greywackes of the Brøttum Sandstone and Shale to the conglomerates and arkoses of the Biskopåsen Formation it would appear that crustal movements must have occurred around that time. These lithological changes can be studied along the western, south-western and southern borders of the Sparagmite Basin. The coarseness and roundness of the hard boulder material of the Biskopåsen Conglomerate indicates that the gradients of rivers increased and that the material was transported over considerable distances. This favours the idea either that the crustal movements took place outside the basin of deposition, or that a general uplift of the area possibly occurred rather than an actual subsidence of the basin itself. Gradually the deposition of cobbles was followed by pebbles, then gravel and arkoses, and in turn there followed a period with sedimentation of the Biri Limestone.

Biri Limestone.

Above the Biskopåsen Conglomerate there are carbonate deposits which are partly arkosic. This formation, the Biri Limestone, is exposed from Storlodammen (Fig. 1, K 7) in the south to Brattland (Fig. 1, A 3) in the north. The transition from the Biskopåsen Conglomerate to the Biri Limestone in this area is best seen at the mountain Herfjell (Fig. 1, E 3) and in the stream Bjørgebekken (Fig. 1, B 3).

As mentioned earlier (pp 161 and 165), the Biri Limestone of the Vestre Gausdal area differs from the Biri Limestone at the type locality at Mjøsa. In Vestre Gausdal it can be divided into four members:

Nyseter Limestone	ca.	0— 20	m	thick
Fjello Sandstone	»	0—200	»	»
Vismund Limestone	»	0— 20	»	»
Åltjerna Limestone	»	0— 15	»	»

To a large extent these members interdigitate and no sharp boundaries have been observed between them.

Åltjerna Limestone.

The Åltjerna Limestone is the oldest member and the one that mostly resembles the Biri Limestone from the type locality. It can be traced from the type locality to the Vestre Gausdal area. At Storlodammen (Fig. 1, K 7) and Vismundseter (Fig. 1, J 7) the Åltjerna Limestone is a sequence of alternating dark limestones and shales, the layers 2—5 mm in thickness. At Åltjerna setrene (Fig. 1, H 6) and in Vestre Gausdal it is a dark limestone with white calcite veins. In Vestre Gausdal it is overlain by an interbanded grey limestone and shale alternating rapidly in 2—5 mm thick layers. By the confluence of the rivers Sagelven and S. Haukå (Fig. 1, C 4) the interbanded grey limestone and shale is resedimented, a current from the west has moved the limestone fragments a few centimetres.

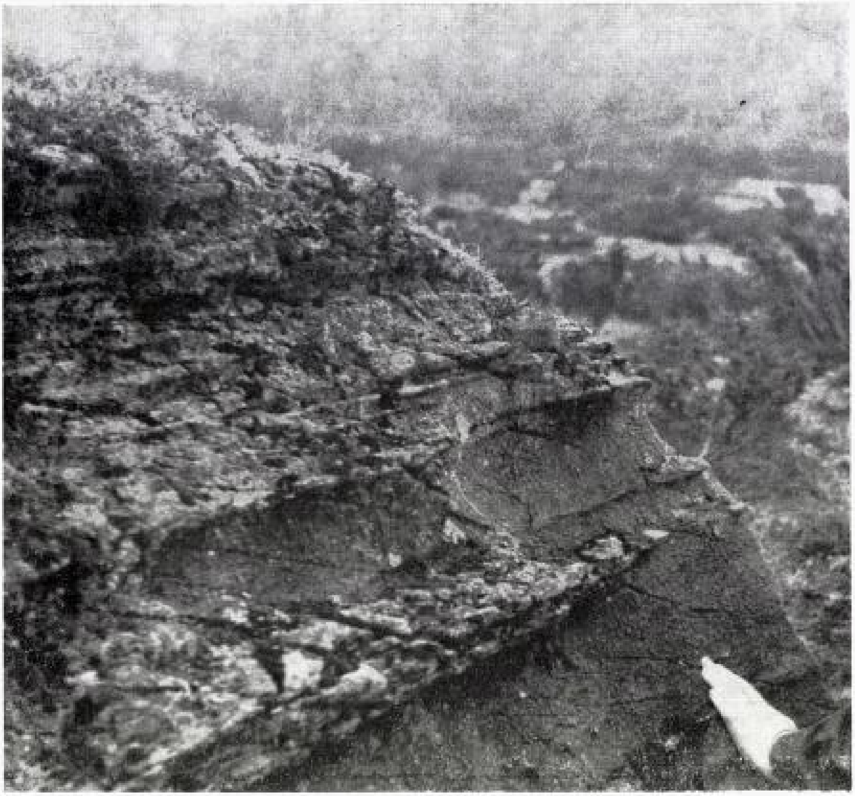

Fig. 6 A. Nyseter Limestone from Nyseter.

Vismund Limestone.

In the valley Vismunddalen the Vismund Limestone is exposed in the lowest parts of the surrounding hills. It is evidently younger than the Åltjerna Limestone though the actual contact is unexposed. The Vismund Limestone contains an admixture of coarse clastic material, and so is partly a calcareous arkose which sometimes displays graded bedding. Upwards it gradually passes into the overlying Fjello Sandstone.

Fjello Sandstone.

The formation Ring Sandstone at Mjøsa can be traced into this area. It thins out westwards and then has limestones both below and above it; in Vestre Gausdal, this sandstone is a member (the Fjello Sandstone)

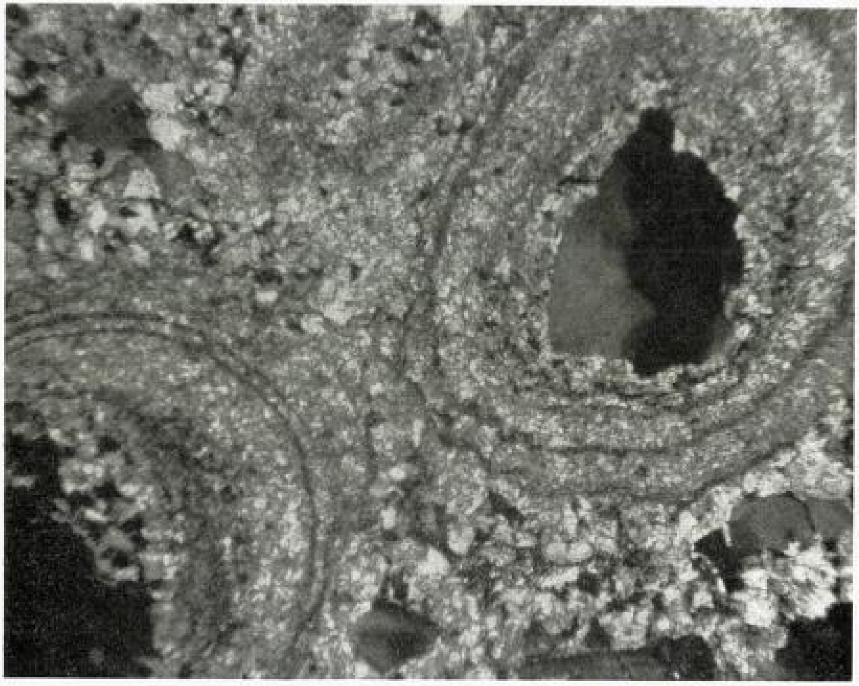

Fig. 6 B. Oolitic texture of the rock at Fig. 6 A, crossed nicols. (Enlarged ca. 50x).

of the Biri Limestone formation. In the southern part of the area, the Fjello Sandstone (Ring Sandstone) is a thick deposit of bluish-grey arkose which is very coarse and partly calcareous. At the lake Fjellovatnet (Fig. 1, K 6) it is ca. 200 m thick but thins out north-westwards to ca. 100 m on the hill Bjørnhaugen (Fig. 1, J 4). The uppermost layers are not exposed.

The Fjello Sandstone in the northern part of the area is mostly a medium-grained dark feldspathic sandstone which rarely exceeds 10 m in thickness. Similar rock types are very common among older and younger formations, but the Fjello Sandstone can be identified in particular by its upward enrichment in carbonate.

Nyseter Limestone.

The Nyseter Limestone crops out from the eastern slope of the mountain Skjellbreidfjell (Fig. 1, F 4) in the south, to west of the farm Brattland (Fig. 1, A 3) in the north. It does not differ from the Vismund

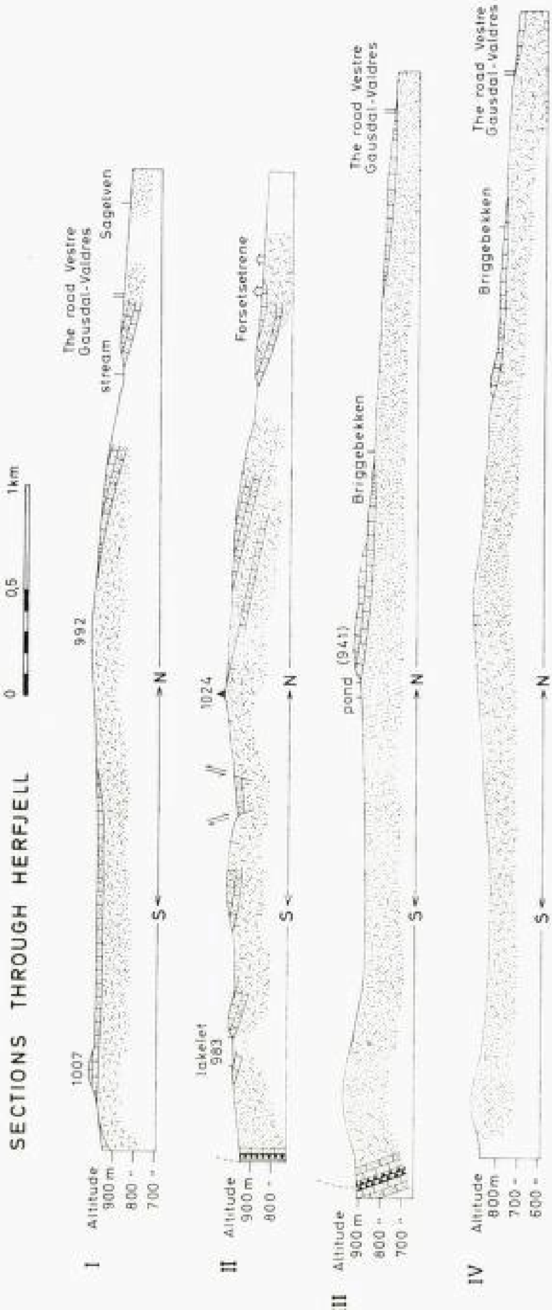

Fig. 7 A. North-south sections across the mountain Herfjell. Ca. 400 m between the sections. For legend, see Fig. 7 B, p. 178.

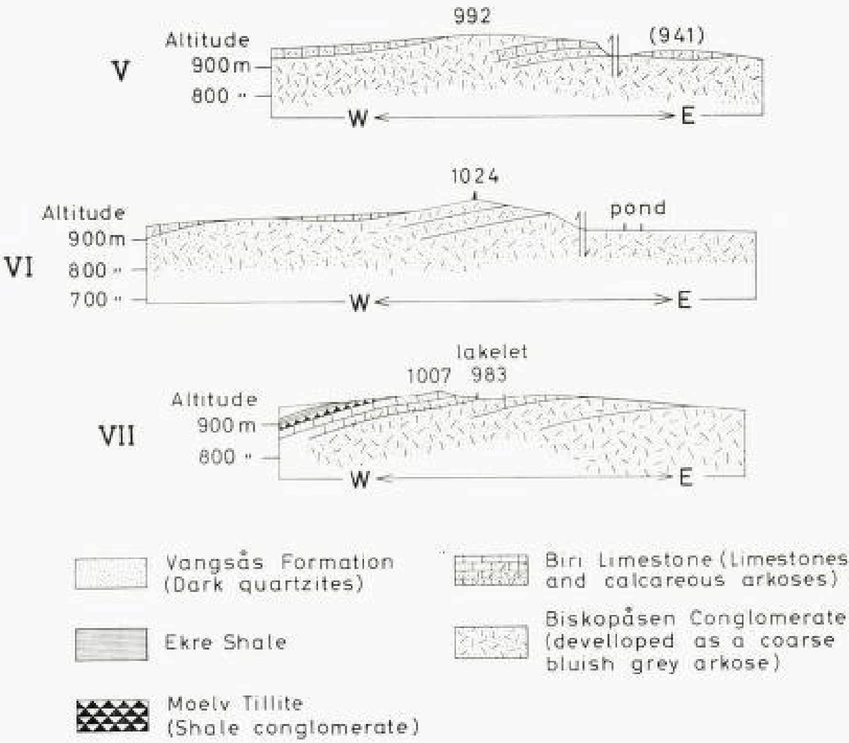

Fig. 7 B. East-west sections across the mountain Herfjell. Ca. 0,5 km between V and VI, and ca. 1,5 km between VI and VII.

Limestone, other than in its stratigraphical position. The Vismund Limestone rests on Åltjerna Limestone and grades upwards into the Fjello Sandstone, while the Nyseter Limestone rests on arkoses (the top layer of the Biskopåsen Conglomerate) and is overlain by the Moelv Tillite. The Nyseter Limestone is the most exhaustively studied member of the Biri Formation within the area. It is well exposed at Nyseter and Herfjell (Fig. 1, E 3) where it has its maximal thickness and shows a good deal of facies variations. At these particular localities no other members of the Biri Formation are present.

The rocks of Herfjell represent a transitional zone between the Biskopåsen Conglomerate and the overlying Biri Limestone. The transitional zone from the arkose at the top of the Biskopåsen Conglomerate to the Nyseter Limestone is ca. 100 m in thickness; the Nyseter Limestone is

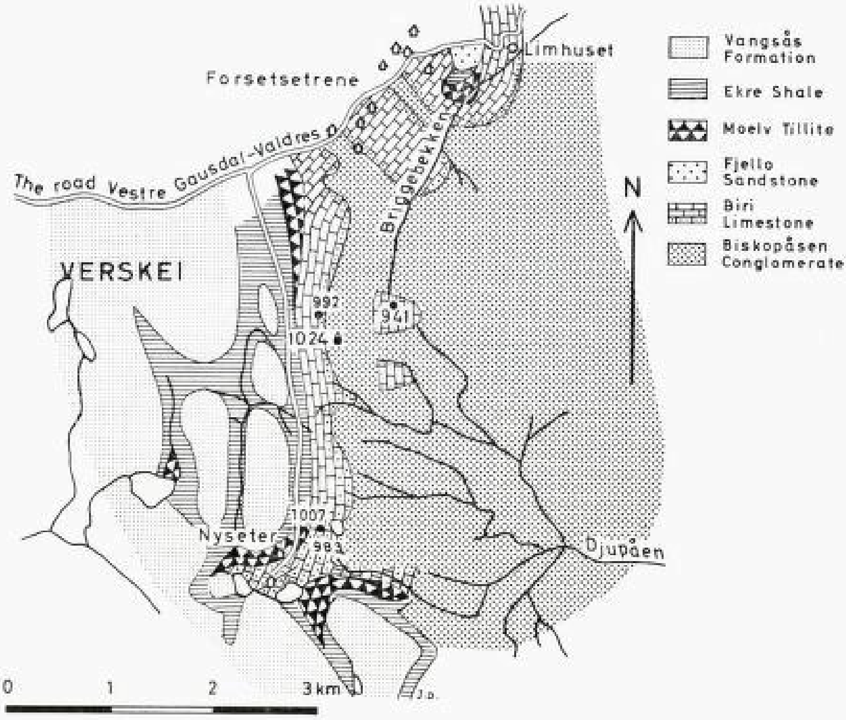

Fig. 7 C. Geological map of the Herfjell area.

ca. 20 m thick. Because of the transition it is difficult to estimate or measure the thickness with any accuracy. The local stratigraphy is sketched in Figs. 7 and 8. Upwards (northwards and westwards on the map, Fig. 1), the rocks at Herfjell are gradually enriched in limestone beds 2—3 cm thick with intercalated 0.2—2 m thick arkose units. Limestone layers become more and more numerous higher up in the sequence,

Fig. 8 A. Light bluish-grey silty limestone interbanded with a conglomeratic limestone in the south wall of the hill (941) east of Herfjell.

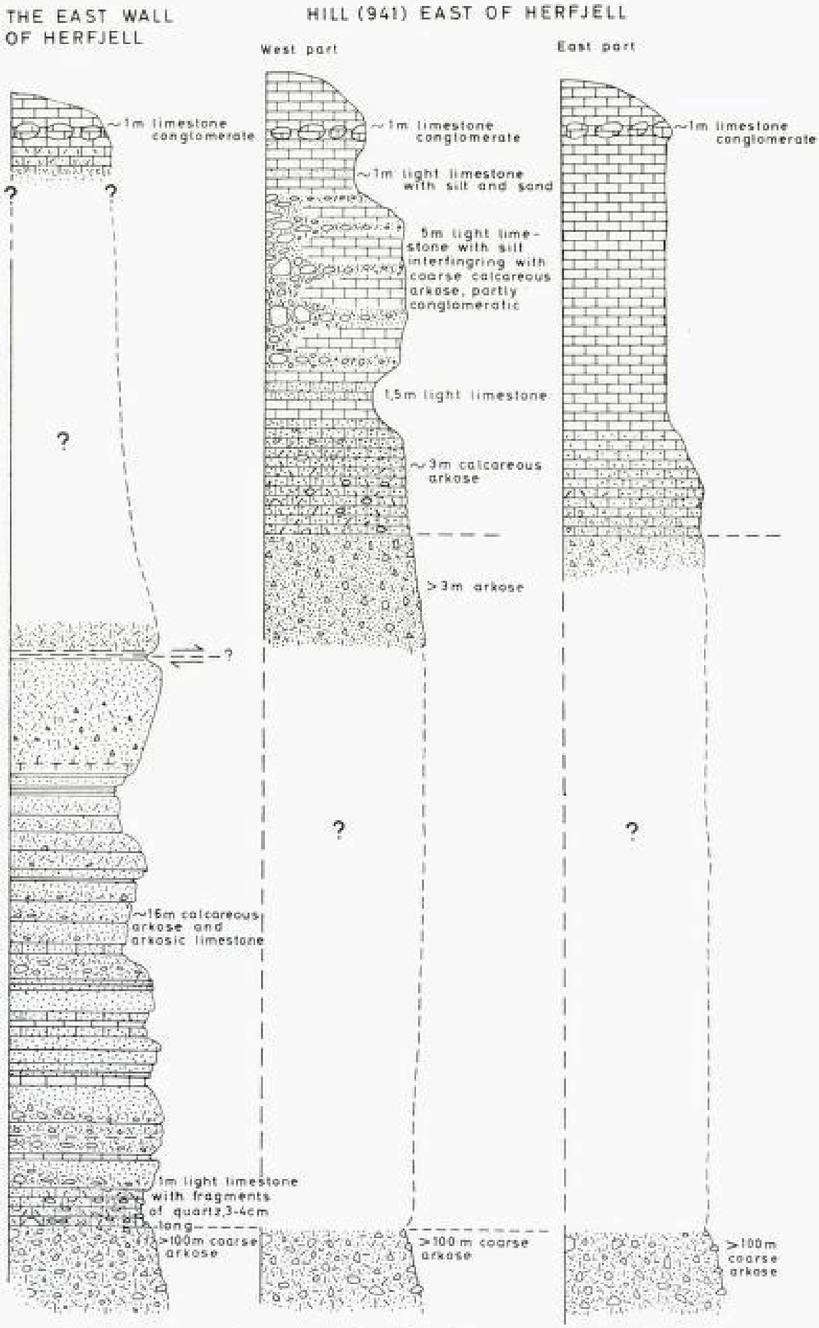

Fig. 8 B. Detailed sections through calcareous rocks at Herfjell.

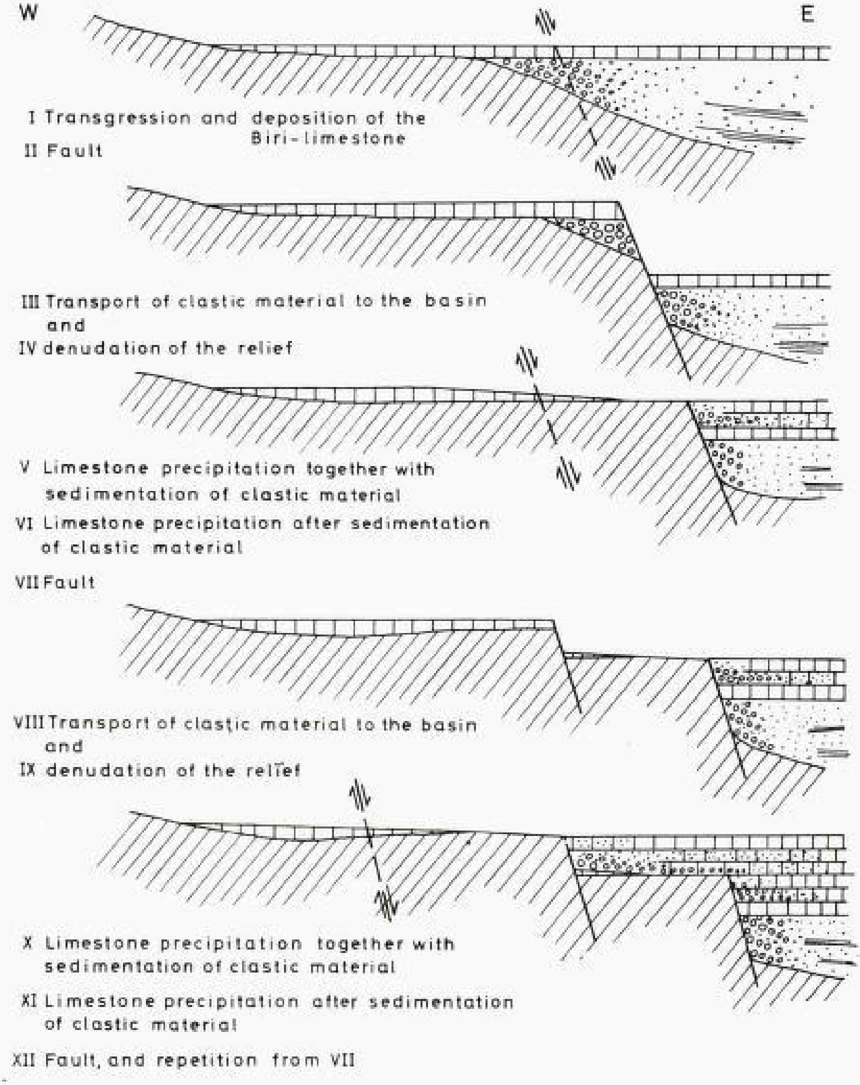

Fig. 9. A sketch of the south-west border of the Sparagmite Basin as it may have been during the time of Biri Limestone deposition.

and at the same time the matrix of the arkose interbeds is enriched in carbonate. A limestone-conglomerate occurs near the top of this Nyseter Limestone (Fig. 8 B).

The uppermost layers show graded bedding. Near the top the calcareous arkoses have an oolitic texture (Fig. 6). These layers interdigitate irregularly with a light silty limestone, ca. 2 m thick: this limestone marks the top of the Nyseter Limestone.

Thin-sections of dark limestones show a very fine-grained matrix of calcite and graphite with dispersed fragments of quartz and feldspar, usually ca. 0.1 mm in diameter. The fragments are corroded by the calcite of the matrix and have lobate outlines. The quartz grains show irregular extinction. Most of the K-feldspar grains are microcline, but some of them are too small to show any visible microcline or perthite textures. Plagioclase has not been observed. Crystals of vein calcite occur in all sizes up to ca. 1 mm. Secondary quartz or feldspar is not observed.

It is likely that parts of the Biri Limestone in this area are the temporal equivalents of the Ring Sandstone, especially as the Biri Limestone contains arkosic layers (the Fjello Sandstone). On the other hand, the Fjello Sandstone may be regarded as a near-shore facies of the Biri Limestone as indicated in Fig. 9. The variation in facies and thickness of the Biri Limestone may partly be due to irregularities of the sea bottom ensuing the deposition of the Biskopåsen Conglomerate.

The most obvious near-shore facies of the Biri Limestone within the area is found near the confluence of the rivers Sagelven and Søndre Haukå (Fig. 1, C 4 or Fig. 2, C 2-3). Here the original limestone and shale has been broken and resedimented, and the fragments moved but a few centimetres indicating currents from the west. The shore-line at the time of deposition of the Biri Limestone must have been situated a little to the west of this locality.

The oolitic layers at the top of the Nyseter Limestone also point for deposition near the shore. Illing (1954, pp 35—44) found that ooliths are inorganic growths formed in warm marine waters washing along shores with carbonate sands. He defines ooliths as grains of different origin surrounded by one or more concentric layers of fine-grained carbonate. According to this definition and origin the Nyseter Limestone with ooliths may have been a gravel in the littoral zone, and it may indicate where the shore of the basin was located during the deposition of this particular formation.

Moelv Tillite.

Within this area the Moelv Tillite is a boulder-bearing shale (Fig. 10). It is found from Nyseter (Fig. 1, E 3) in the south to Bølia (Fig. 1, B 3) in the north. The shale matrix is dark grey and usually shows a greenish

Fig. 10. Moelv Tillite in the river Sagelven (Fig. 2, C 2).

grey weathering surface. West of Dekken seter (Fig. 1, I 4) there are some erratics, > 1 m^3, of a coarse conglomerate with angular boulders; these are also interpreted as Moelv Tillite.

The tillite rests on the Nyseter Limestone, and this boundary can be seen at Nyseter (Fig. 1, E 3), south of Herfjell (Fig. 1, E 3), and in the stream Briggebekken (Fig. 1, D 4). These localities are (so far) the only ones known in Southern Norway where the tillite is resting on limestone in an autochthonous position. The contact itself is sharp. The tillite also occurs allochthonously within the area (Figs. 4, 5 and 21). In such cases it is mostly crushed or mylonitized, but still recognizible.

The thickness of the formation varies from ca. 30 m at Roppa (Fig. 1, B 4) to 2.5 m in the upper thrust sheet at Roppa and Bjørgebekken (Fig. 1, B 3). In all probability the variation in thickness is partly due to tectonic deformation.

The tillite is markedly polymict, the boulders and pebbles varying in size from 1 m down to the fine grains of the matrix. The matrix of the deposit displays a varved structure and consists mainly of laminated silt, each lamina ca. 1 mm thick.

Throughout the area the most common material of the boulders and blocks are fine-grained light quartzites, grey granite, and a fine-grained light granite. Boulders of gneisses, limestones and mylonites are frequently seen, but those of pegmatite quartz, shale fragments and meta-amphibolites are rare. On the northern slope of Herfjell (Fig. 1, D 3) one pebble of red granite has been found in the tillite, while at Nyseter (Fig. 1, F 3) a 30 cm long boulder of Nyseter Limestone with graded bedding was deposited, and as this particular rock is too brittle to have been transported over long distances without disintegrating, its presence as a sizable boulder at Nyseter would appear to confirm the local derivation.

In thin-sections the laminae of the matrix are seen to bend conformably around the clastic fragments. The fine-grained matrix consists of scattered chlorites, sericite, granulated quartz and small pyrite crystals. Most clastic grains consist of quartz and quartzite of variable size and roundness. All quartz displays irregular extinction. Rock fragments are of the same material as the pebbles, boulders and blocks. Many of the feldspar grains are light brown or mottled brown and some of them are strongly sericitized and they are often corroded by carbonate. Many of the plagioclase (albite) grains have bent twin lamellae. Clastic grains of brown biotite (partly altered to chlorite), muscovite, phlogopite, zircon and ore have been observed but these are rare (Fig. 11).

As shown in Fig. 11, in thin-section the rock looks like a greywacke. According to Pettijohn (1957, pp 272—273) pure tillites will resemble greywackes, having a matrix rich in chlorites and micas.

The unsorted character of the boulders together with the finely stratified matrix indicate that the boulder clay within the area has been deposited under water. Kulling (1938, pp 292—296) found that unsorted angular material dispersed in a laminated matrix where the layers bend conformably around the clastic fragments is indicative of a glacial origin.

Holmsen (1954, pp 105—119) described facies variations of the Moelv Tillite. He interpreted the type found at Moelv (Holtedahl, 1922, pp 168—172) as a near-shore facies, and the boulder-bearing shale as having been deposited farther from the coast, the boulders being ice-rafted. Earlier, Münster (1901, pp 18—19) had compared these deposits, and thought that the cleavage in the tillite was formed during the regional metamorphism of the area. Englund (1966, pp 77—79) described a boulder-bearing shale at Fåvang (ca. 20 km north of this area) in a similar stratigraphic position, and he maintained that was of glacial origin.

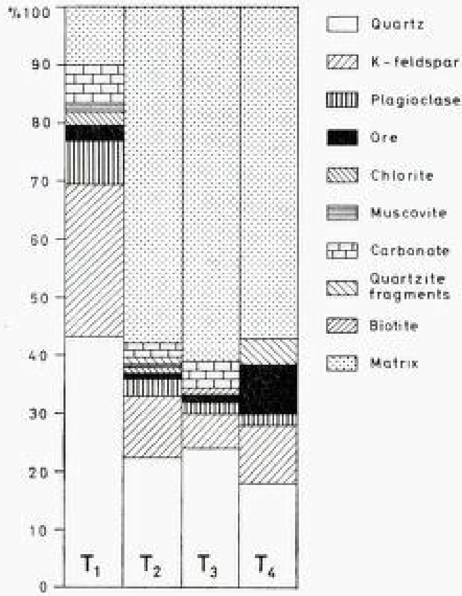

Fig. 11. Modal analyses of Moelv Tillite from the area.

T_1 = Light variety with sand and silt in matrix, from Nyseter.
T_2 = Dark shaly variety from lower thrust sheet in Roppa.
T_3 = Light shaly variety from 2 km north-west of Nyseter.
T_4 = Dark brown, strongly tectonized variety from 1 km south-east of Nyseter.

As the matrix of the tillite west of Herfjell consists mainly of silt, this suggests that the boulder clay here was deposited nearer to the shore than the equivalent deposits north and east of Herfjell. The boulder of Nyseter Limestone found in the tillite at Nyseter indicates that the top of the Biri Formation was eroded. Its occurence also suggests that the border of the Sparagmite Basin was situated not far west from Herfjell during the time of deposition of the Moelv Tillite.

The contact between the Nyseter Limestone and the Moelv Tillite can be correlated with a sudden change in climate at the same time as areas west of the Sparagmite Basin were being uplifted. The nature of the boulders of the tillite provides evidence that the area west of the Sparagmite Basin consisted mainly of quartzites and grey granites at the time of deposition of the Moelv Tillite. The quartzites found in the tillite resemble the rocks described as quartzites and quartzitic gneisses by Bugge (1939, pp 20—29) from the areas Hemsedal and Gol (ca. 80 km south-west of this area). Grey granite occurring as boulder material is similar to the granites described by Smithson (1963, pp 36—37) from Hedal (ca. 70 km SSW of this area). The Precambrian areas that supplied material to the Sparagmite Basin are to-day covered by younger deposits, and it is difficult to locate the source of these particular boulders. However, some of the rocks are thought to resemble lithologies from the Precambrian in Valdres (Strand, 1943). (Valdres is ca. 70 km west of this area.)

Ekre Shale.

The Ekre Shale is commonly a grey or greenish grey silty shale (Fig. 12) similar to the matrix of the tillite. It is exposed from the southern part of Skjellbreidfjell (Fig. 1, G 4) to Bølia (Fig. 1, A-B 3).

No sharp sedimentary contact has been seen, either to the tillite below or to the formation above the Ekre Shale. The thickness of the Ekre Shale varies from ca. 40 m at the northern part of Skjellbreidfjell where it is relatively undisturbed to ca. 0.5 m at the southern part of Skjellbreidfjell where it is squeezed between allochthonous units: on the map, Fig. 1, the Ekre-Shale is drawn with exaggerated thickness. The silt content of the Ekre Shale decreases northwards as does the silt content of the matrix of the tillite.

Thin-sections of this lithology show that it is a laminated clay-silt rock. Each layer is 0.3—4.0 mm thick, those more than 1 mm in thickness sometimes showing graded bedding. The lighter layers consist of fine-grained sericite and chlorite with some scattered clastic grains of angular quartz up to 0.05 mm in diameter. There are also some 0.5 mm long flakes of bent muscovite and 0.2 mm long flakes of biotite. In the light layers the ratio of clastic grains : matrix varies from 1 : 5 at Skjellbreidfjell to 1 : 8 at Roppa (Fig. 1, B 3). The dark layers consist mainly of dark clastic material packed in a graphite-rich matrix. Most of the clastic grains are ca. 0.1 mm in diameter, but 0.2 mm grains are also present. The quartz and K-feldspar are dark because of numerous inclusions. The quartz displays undulose extinction, and the K-feldspar is a «patchy» perthite. Plagioclase is rare, only two small grains having been observed. Although clastic muscovite and biotite are present, they appear more frequently in the lighter layers. Biotite is partly altered to chlorite. Haematite and limonite are found as inclusions, 0.1 mm or smaller, in biotite-chlorite individuels. In the darker layers the clastic grains : matrix ratio varies from 3 : 2 at Skjellbreidfjell to 1 : 2 at Roppa.

Kuenen and Migliorini (1950, p. 119) maintained that floculation of lutite in marine waters will make it settle together with the silt so that varves cannot be developed. When the ice that deposited the tillite melted, it is possible that big rivers from the melting glaciers ran into

Fig. 12. Ekre Shale from 1 km ESE of Nyseter (¼ size).

the Sparagmite Basin and so made the water near the mouths of the rivers brackish or fresh. The turbulent water that brought material into the basin at the time of deposition of the Ekre Shale was probably so cold and heavy that it sank on entering the basin. Its load is thought to have been deposited under quiet conditions in brackish water with the concomitant development of graded bedding.

K. Bjørlykke (1963, pp 90–92) described the Ekre Shale from the eastern border of the Sparagmite Basin, and Englund (1964, pp 108–111) has presented a description of this same formation occurring at Fåvang (ca. 20 km north of this area). Neither of them, at that time, found laminated rocks in those areas. K. Bjørlykke suggested that the constituents of the shale were derived from rocks rich in clay minerals. The lack of lamination may imply deposition in saline waters further away from the river estuaries. Later, however, both K. Bjørlykke (1966 p. 32) and Englund (1966) reported the localized occurrence of a varve-like lamination in their areas.

The average varve within the area is ca. 1 mm thick. If the varves represent seasonal deposits and the thickness of the Ekre Shale at Skjellbreidfjell (ca. 40 m) is close to the true thickness, the Ekre Shale of the area was deposited probably during a period of ca. 40 000 years. The presence of clastic grains of dark quartz and feldspar similar to those found in the Fjello Sandstone indicate that the same source rocks were being eroded over a long period of time.

As the Moelv Tillite is a glacial deposit, the oldest part of the Ekre Shale is a late-glacial to post-glacial sediment. The boulder material of the tillite gradually decreases, a feature possibly signifying a gradual withdrawal of the ice. If the water of the Sparagmite Basin was saline at that time, it was almost certainly warmer than that of the rivers which flowed into it.

The Vangsås Formation.

The Vangsås Formation comprises the two members Vardal Sandstone and Ringsaker Quartzite. The older member the Vardal Sandstone, gradually passes into the Ringsaker Quartzite with decreasing grain-size and feldspar content; the border as indicated on the maps and sections is therefore transitional.

Vardal Sandstone.

The Vardal Sandstone is mostly a dark arkose with a tectonic contact to the Ekre Shale, but on the west side of Skjellbreidfjell (Fig. 1, F 3) a sedimentary boundary can be observed. At this boundary there are alternating layers of shale and arkose, the arkosic beds thickening upwards while the layers of shale diminish in thickness and number. The Vardal Sandstone has its maximum thickness (ca. 100 m) in the south of the area; this decreases northwards to ca. 5 m in the stream Bøseterbekken (Fig. 1, A 3).

The Vardal Sandstone is well exposed in all hills west and south of Vismunddalen. The mountain Skjellbreidfjell is built up of Vardal Sandstone lying in both autochthonous and allochthonous positions. On the west side of the hill Verskei (Fig. 1, D 2) the Vardal Sandstone is developed as a conglomerate with pebbles of white quartzite and a matrix of dark arkose. No imbrication has been observed. The weathered surface of the conglomerate is reminiscent of the Fjello Sandstone. A more finegrained variety of the conglomerate is found in some of the hills west of Nyseter (Fig. 1, E 3) and in the river Søndre Haukå (Fig. 2, C 2). In Søndre Haukå there is a 20 cm wide zone of coarse arkose (7 mm-size

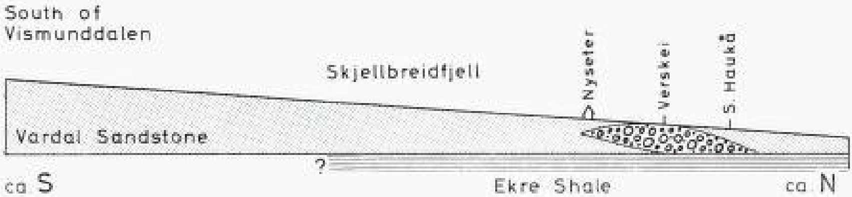

Fig. 13. Facies variations in the Vardal Sandstone.

grains) between the Ekre Shale and the ordinary Vardal Sandstone. Fig. 13 illustrates a possible explanation of the field relationships and facies variations of the Vardal Sandstone.

Microscopic examination shows that in this area the Vardal Sandstone is an arkose which grades upwards into a feldspathic sandstone. The upper part of the feldspathic sandstone is recrystallized. The lower part of the arkose has clastic grains surrounded by some matrix, while the upper part consists of tightly packed clastic grains with little or no matrix, the grains, however, being partly crushed. The clastic material consists mainly of quartz and feldspar, the size of the grains decreasing upwards. The grains are subrounded to well rounded. Quartz shows irregular extinction, and in the crushed individuals this is strongly undulose. The feldspar is mainly microcline and perthites; plagioclase is uncommon — only albite is present. The feldspars are usually light brown or mottled brown. Clastic muscovite is present, but biotite has not been found. Accessory minerals are zircon, haematite and leucoxene. The matrix consists of sericite, fine-grained quartz and some chlorite.

Although the Vardal Sandstone may resemble a fine-grained variety of the Fjello Sandstone, it has less plagioclase and also lacks biotite. Moreover the grain-size and roundness indicate that the material of the Vardal Sandstone has been transported over longer distances (or for a longer time) than that of the Fjello Sandstone.

The conglomerate at Verskei may be interpreted as a lensshaped body thinning out northwards and southwards as indicated in Fig. 13. It has not been found east of Verskei. The character of this conglomerate suggests rather near-shore conditions with a transport of material from a westerly source.

Vogt (1923, pp 327—329) described the Vardal Sandstone from Mjøsa (ca. 40 km south-east of this area). He assumed that the deposition was caused by a rapid uplift of the areas surrounding the basin. This uplift, or uplifts, is thought to have continued irregularly and as a result there are several zones of coarse-grained beds rich in feldspar within the profile. Vogt suggested that the upward decrease in feldspar content corresponded with gradually quieter conditions: the land-mass supplying material to the Vardal Sandstone was gradually being worn down and the transport velocity of the material decreased so that most of the feldspar disintegrated before deposition. This is in agreement with the present writer's observations.

Skjeseth (1963, pp 32—34) described the lower part of the Vangsås Formation in the Mjøsa district as a grey arkose which gradually changes upwards to a white quartzite. Layers of Vardal Sandstone in the Ekre Shale mark the transition from Ekre Shale to Vardal Sandstone. He described a conglomerate occurring in the middle of the deposit along the southern border of the Sparagmite Basin: it is characterised, by white quartz pebbles in a grey sandstone matrix. Skjeseth's descriptions might well be applied to the observations from this area.

The varying thickness of the Vardal Sandstone from north to south in the area can be understood in terms of transport of material from south to north, but then it becomes difficult to explain the even grain-size and the conglomerate at Verskei. As the grain-size, roundness and feldspar content is quite constant throughout the area, it is therefore more likely that the material was brought in from a westerly direction. As indicated in Fig. 13, the sedimentation of the Vardal Sandstone may have lasted for a longer time in the south than in the north of the area.

The rocks that supplied material for the deposition of the Vardal Sandstone were most certainly the same as those which gave material to the older formations: acidic rocks with dark quartz and feldspar, and light quartzites that yielded pebbles for the conglomerate at Verskei. The western border of the basin at that time was probably situated near the area of deposition of this conglomerate.

Ringsaker Quartzite.

The Ringsaker Quartzite within the area is mostly a dark grey to black feldspathic sandstone. This sandstone gradually passes into the overlying Cambrian shales through a transitional zone ca. 1 m thick (Fig. 14). This boundary is seen in the stream Bjørgebekken (Fig. 1, B 3). At other places within the area the boundary has been disturbed by tectonic movements. The Ringsaker Quartzite is found from Nord-Torpa (Fig. 1, K 2) in the south to Vestre Gausdal (Fig. 1, A-B 3) in the north, homogeneously developed, though its thickness (as with the Vardal Sandstone) decreases northwards. At Nylsfjell (Fig. 1, J 1) the thickness is more than 65 m, south of Vismunddalen (Fig. 1, K 5-6) more than 50 m, and at Brattland (Fig. 1, A 3) ca. 10 m.

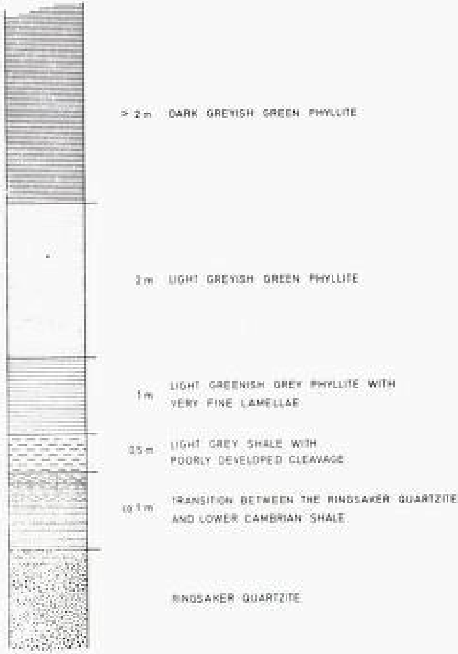

Fig. 14. The transitional zone from Ringsaker Quartzite to the Cambrian shales as developed at Bjørgebekken (Fig. 2, B 3).

In Bjørgebekken (Fig. 2, B 3) there are cyclothems ca. 1,5–2 m thick; the bottom layer of each cyclothem may be conglomeratic. The conglomerates are 1 cm thick and consist almost exclusively of well-rounded grains of dark quartz 7–10 mm in diameter. The thickness and grain-size of the cyclothems decrease upwards in the profile at the same time as thin shale lamellae increase in number and thickness. At the top of the member there is a transitional zone, as shown in Fig. 14, wherein the sandstone layers rarely exceed 2–3 mm in thickness.

Specimens of the Ringsaker Quartzite studied in thin section are found to be feldspathic sandstones and orthiquartzites. Although they are recrystallized, the primary outlines of the clastic grains are still visible under parallel nicols. The grains are closely packed and there is very little matrix. The matrix consists of fine-grained quartz, sericite, finely disseminated graphite and small crystals of pyrite. The clastic grains are usually ca. 1 mm in diameter and well rounded in the lower part of the member, decreasing upwards to ca. 0.05 mm and subrounded to angular near the top. Clastic material consists of quartz and feldspar; other minerals have not been observed. Most of the quartz and feldspar is dark on account of numerous small inclusions. The quartz displays a strongly undulose extinction. Feldspar is light brown or mottled brown. The K-feldspars are microcline and perthites; plagioclase is rare, but where present is albite.

Thin-section studies show that the Vardal Sandstone carries haematite while the Ringsaker Quartzite contains pyrite. This may indicate that the Vardal Sandstone was deposited in an oxidizing milieu and the Ringsaker Quartzite in a reducing environment within the area: it is possible, however, that the pyrite may have been introduced diagenetically.

As the Ringsaker Quartzite in this area is fine-to-medium-grained, this would suggest that quieter conditions of sedimentation prevailed at this time than during the deposition of the Vardal-Sandstone. Vogt (1923, pp 329—331) asserted that deposition of a pure quartzite represents a greater span of time than does an equivalent thickness of arkose. He found that the Ringsaker Quartzite at Ringsaker (ca. 40 km southeast of this area) is 240 m thick, thinning out remarkably towards the north. This led him to suggest that the land adjacent to the basin was uplifted from north to south. Oftedahl, however, (1947, pp 167—168) proposed that the Ringsaker Quartzite was thin in the middle of the basin, and that might be the reason why it is thin in Vestre Gausdal and thicker further south. This agrees with the recent observations.

Skjeseth (1963, p 34) found Scolithus in the Ringsaker Quartzite at Mjøsa indicating near-shore conditions there. No criteria have been found within the area that can help prove that the Ringsaker Quartzite is a near-shore deposit, but as it is possible to follow this member as a continuous unit from the Mjøsa districts into the present area, it is tempting to think that it has been developed under rather similar conditions.

As the Ringsaker Quartzite within the area becomes finer grained upwards in the profile at the same time as it is intercalated with shale lamellae of increasing thickness, it would seem as if it was being deposited gradually further away from the coast; alternatively, the sea may have transgressed to the south and west beyond the border of the original Sparagmite Basin. Quite possibly the thick layers of quartzite in the south-western part of the area mark the position of the border of the Sparagmite Basin during the deposition of most of the Ringsaker Quartzite. The rocks which provided material for the Ringsaker Quartzite were most likely the same as those that supplied material to the Vardal Sandstone. The upward enrichment of quartz within the profile may be due to a longer transport of the material before its actual deposition.

Cambro-Ordovician Shales.

In this area the Ringsaker Quartzite is everywhere overlain by at least 0.5 m of light grey shale (Fig. 14). It is mainly slightly phyllitic, but the cleavage is only poorly developed.

Dapples, Krumbein and Sloss (1938, p. 1932) described shales from stable shelf areas as having certain significant characteristics that suggest their correspondance to quartz sandstone, and that they represent the fine-grained equivalent of the latter. They differ more in grain-size than in mineralogy; however, the clay minerals settle with the same velocity as the very fine quartz grains. Such very fine silt-shales are uniform over great areas and show a gradual transition to sandstones without abrupt contacts. The colour is often greenish grey though this is not diagnostic. This description suits rather well the lowest part of the Cambrian shales within the area, and strengthens the supposition that the Ringsaker Quartzite is closely related to the overlying Cambrian shale.

The Cambro-Ordovician deposits and their fossils from the area have been described thoroughly by Bjørlykke (1891, pp 1—10) and Münster (1891, pp 22—34). Clearly the transgression which occurred in Lower Cambrian times extended far outside the present area.

Brief comments on the sediments.

From the examination of thin-sections of lithologies from several of the formations it can be seen that most of the clastic feldspar is weathered. The feldspar is usually brown or mottled brown in colour, regardless of grain-size. Holtedahl and Schetelig (1923, p. 16) described weathered Precambrian overlain by Cambrian basal conglomerate at Brandbukampen (ca. 80 km south of this area), and they found that this weathered Precambrian continues northwards. Weathered Precambrian most probably existed close to the Sparagmite Basin and supplied material for the deposition of the sedimentary rocks which occur there today. However, some of the weathering and alteration of the feldspars may be a post-depositional phenomenon.

Clastic grains of dark quartz and feldspars are found in the younger formations, their darkness caused by the presence of numerous small inclusions. These have not been studied, but a thorough investigation could probably throw light on the provenance of the Ringsaker Quartzite and other dark quartzites occurring along the Caledonian orogen.

All clastic quartz observed in thin-sections has an irregular extinction, as well as often being crushed. This indicates that the rocks of the area have been involved in tectonic movements. The presence of chlorites and altered biotite points to a low grade metamorphism, most probably the lower part of Eskola's greenschist facies. James (1955, pp 1455–1488) has referred to this part of the greenschist facies as the chlorite zone. In sedimentary rocks with chlorite, sericite, carbonate, clay material and clastic mica this zone is signified by the presence of a cleavage parallel to the bedding.

A feature of particular interest in this area is the occurrence of limestone overlain by tillite. The contact is abrupt. Similar boundary relationships are found at other places in the world and they have been interpreted as indicating a sudden change of climate.

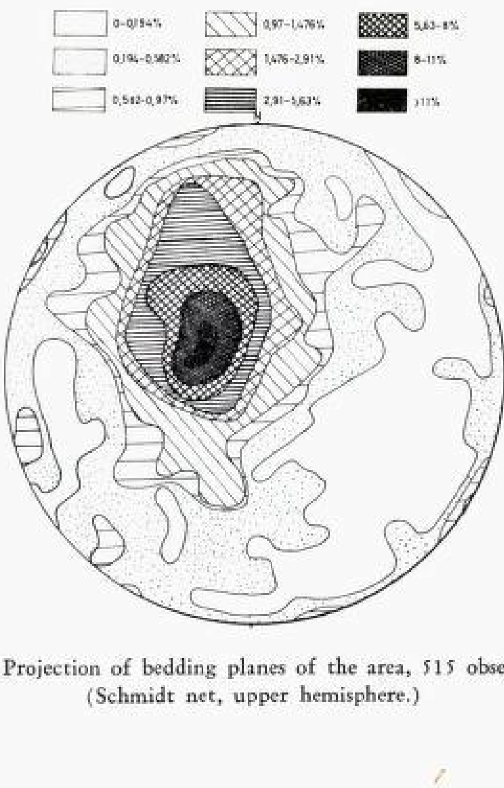

Fig. 15. Projection of bedding planes of the area, 515 observations. (Schmidt net, upper hemisphere.)

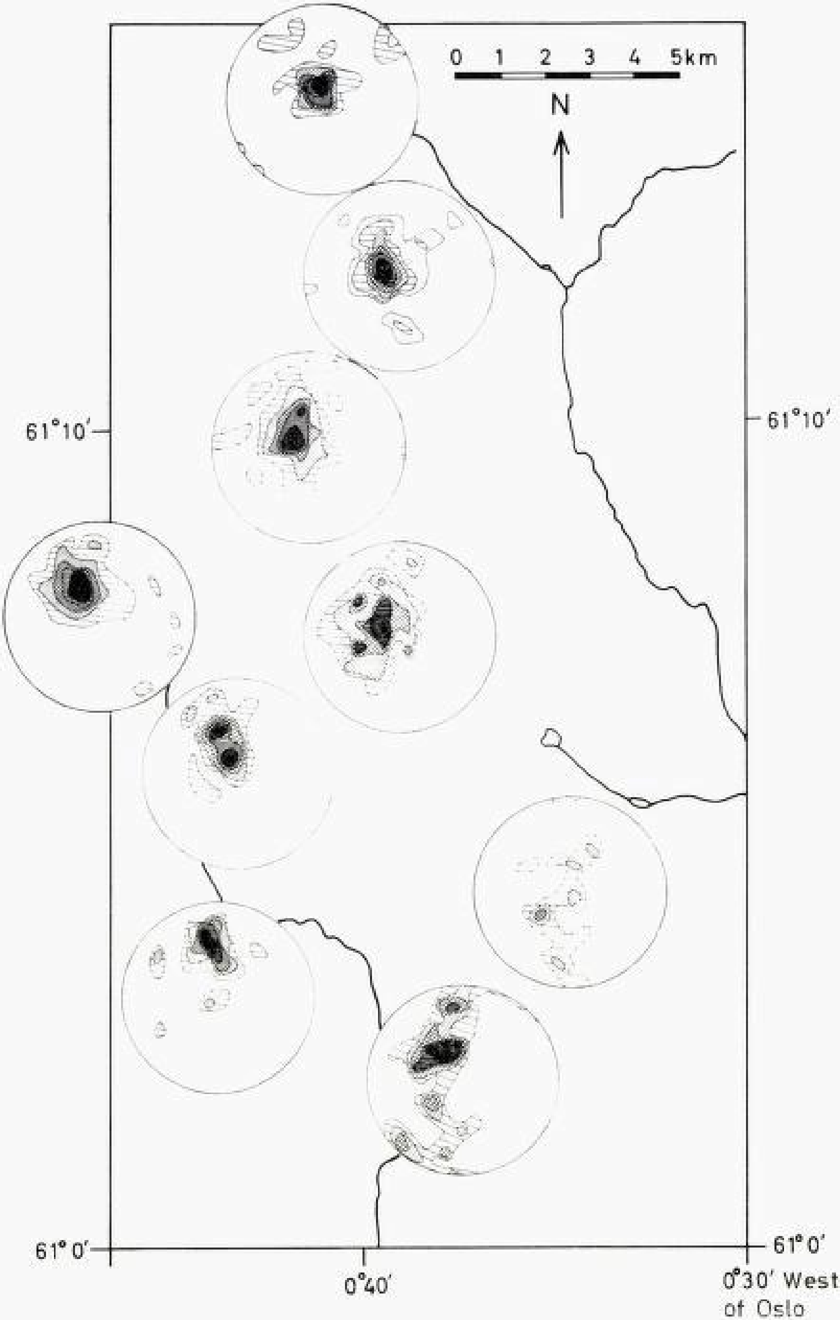

Fig. 16. Bedding within the area (Schmidt net, upper hemisphere).

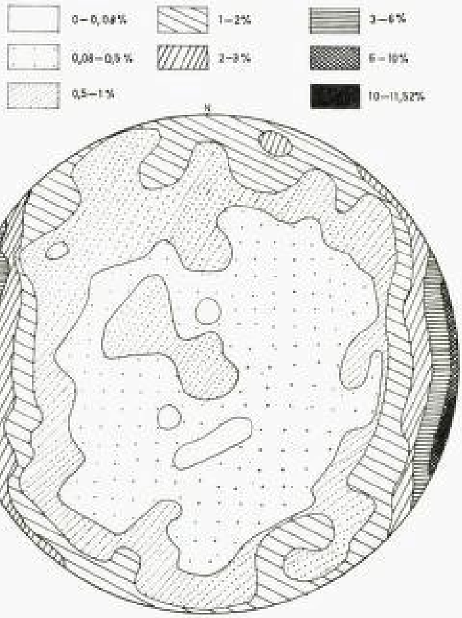

Fig. 17. Projection of faults and joints from the area, 1230 observations (Schmidt net, upper hemisphere).

Structure.

Regional tectonics.

During the Caledonian orogeny, the rocks which now occur in this area were transported towards the south-east from their original site of deposition. The rocks of the Sparagmite Basin were moved southwards against the border of the basin and partly beyond it. In this account only the main structural features of the area will be described.

Bedding commonly dips gently to the north-west (Figs. 15 and 16). Folds of any size are rarely seen in the field, but a lineation is conspicuous in conglomerates, limestones and shales (Figs. 19 and 20). The joints of the area are mainly vertical and show a north-south trend (Figs. 17 and 18); gently dipping joints may be interpreted as minor thrusts. Faults can be demonstrated within the area but are rare (Fig. 20). From a tectonic structural standpoint three groups of rocks can be recognized within the area:

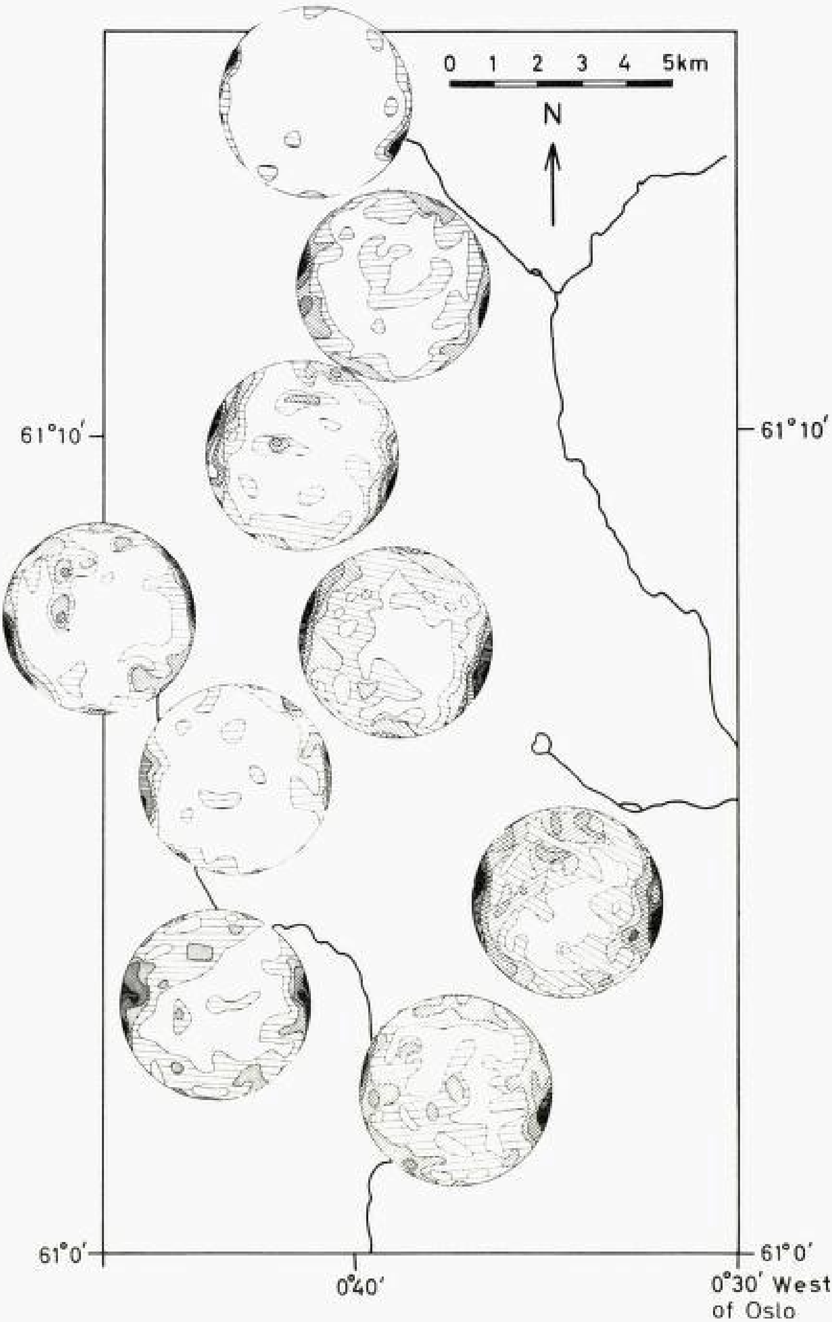

Fig. 18. Joints and faults of the area (Schmidt net, upper hemisphere).

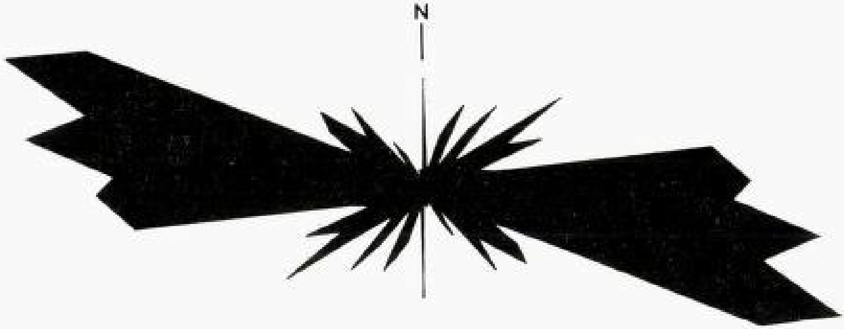

Fig. 19. Orientation of quartz-grains and conglomerate boulders and pebbles. 215 observations.

Allochthonous rocks
Folded rocks
Autochthonous rocks.

Allochthonous rocks (thrusts).

In the south-eastern part of the area the Vangsås Formation has been thrust southwards out of the Sparagmite Basin and inverted, such that parts of it rest on Cambrian shales (Fig. 1, J-L 3-6) and constitute a continuation of the Ringsaker Inversion (Skjeseth, 1963).

The mountain Skjellbreidfjell (Fig. 1, G 4) is mainly built up by Vardal Sandstone and Ekre Shale in imbricated thrusts (Fig. 20).

The maps (Figs. 2 and 20) and sections (Figs. 4, 5 and 21) show that some thrusts are present in Forsetlia (Fig. 1, B-C 3-4): the rocks are very little folded. The thrust planes, where exposed, are smooth and the contact to the underlying rocks is sharp. The underlying rock may be crushed, as is the tillite at Roppa, but as the Ekre Shale and Cambrian shales mainly acted as lubricants, rocks near the thrust planes are only mylonitized close to the contacts if tectonized at all.

Folded rocks.

Most of the rocks of the Sparagmite Group within this area are folded to some extent, though as indicated in Fig. 22 the intensity of the folding increases southwards. Folded sequences are best exposed along the river Roppa and on the mountain Herfjell.

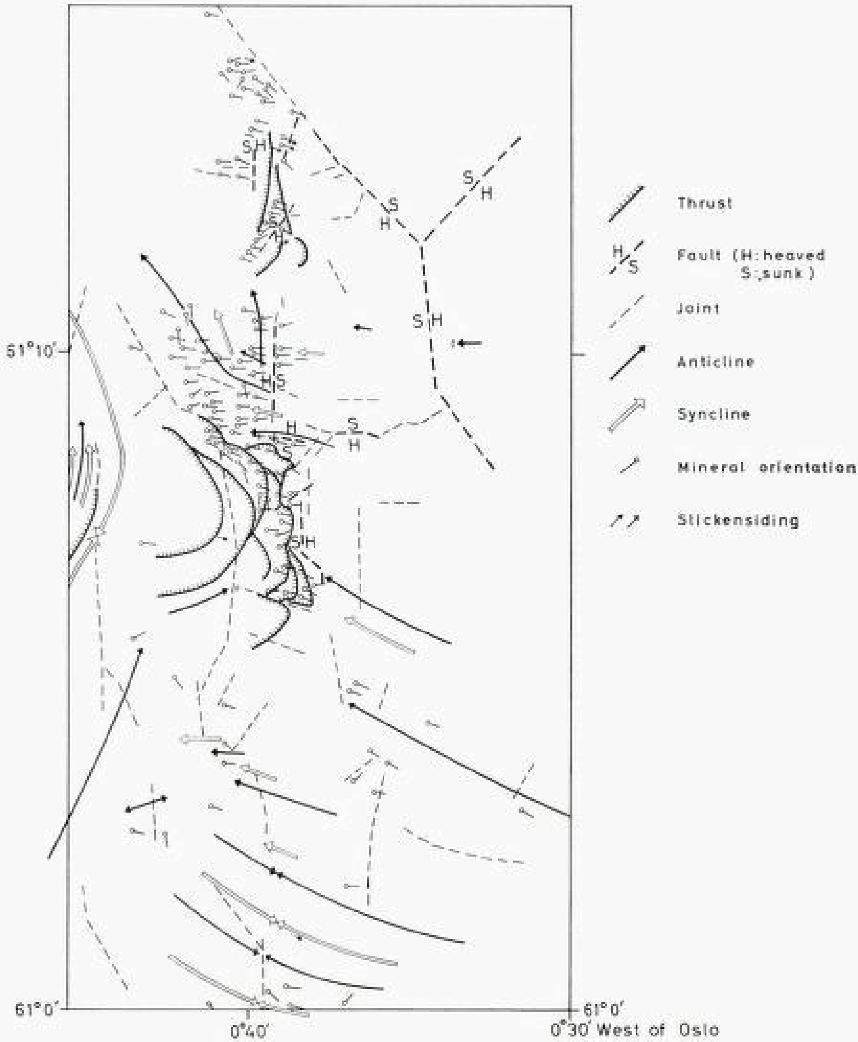

Fig. 20. Structural map of Vestre Gausdal and Fåberg Vestfjell.

The Roppa section.

The river Roppa cuts through both folded and thrusted rocks as shown in Fig. 4. In the lower part of the river gently folded rocks can be observed in the walls of a 50 m deep canyon; the fold axes are ca. north-south and the axial plane dips steeply to the west. The wavelength

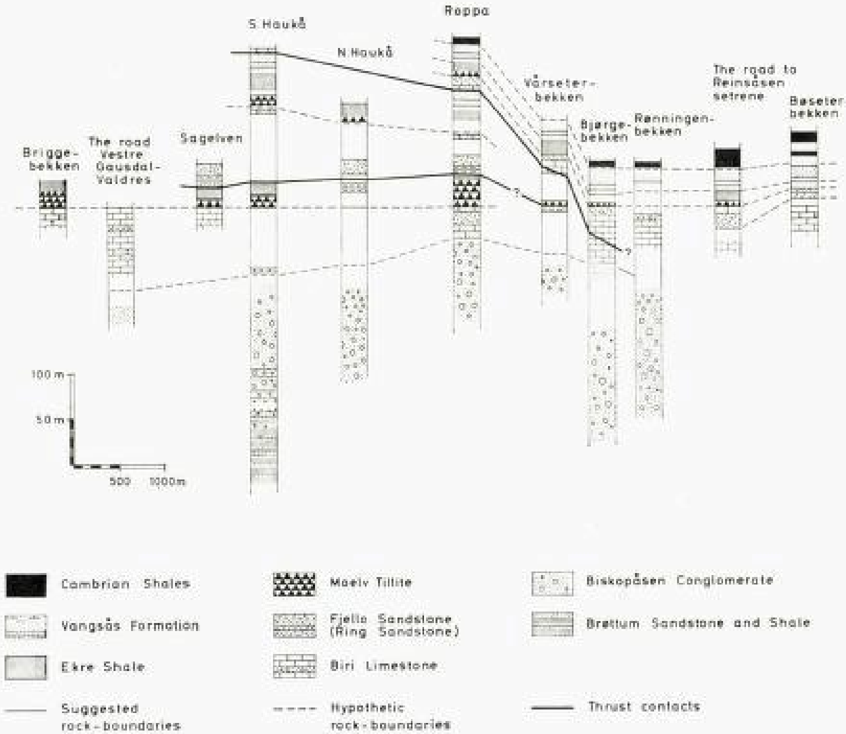

Fig. 21. Sections in Vestre Gausdal (Bølia and Forsetlia).
(See also Fig. 2, 4 and 5.)

is ca. 50 m and the amplitude ca. 10 m. Even the Biskopåsen Conglomerate is folded at this locality, the longest axes of the pebbles and boulders trending east-west (Fig. 19). As indicated in Fig. 4, the Biskopåsen Conglomerate has suffered from tectonic deformation at some levels. Immediately above the Biskopåsen Conglomerate the Biri Limestone is quite undisturbed, but higher up in the section it is folded and thrusted.

Where the Roppa shows a sharp change of direction (Fig. 2, B 3), the Moelv Tillite occurs within a syncline. The thickness of the tillite, here ca. 30 m, has almost certainly been exaggerated by tectonic movements. West of the tillite is an anticline in the core of which is the arkosic top layer of the Biskopåsen Conglomerate (as shown in Fig. 4). The lower thrust sheet in the Roppa valley rests on the western limb of this anticline. This sheet moved forwards over a layer of Ekre Shale and Moelv Tillite. Although these have been thinned to ca. 5 m, they are still recognizible. The thrust sheet itself is ca. 50 m thick and is comprised of strata from the Fjello Sandstone to the Vangsås Formation (Fig. 21).

The upper thrust sheet in this Roppa area rests directly upon the lower one. It evidently moved from west to east on Biri Limestone. The thrust sheet is ca. 50 m thick and comprises strata from the Biri Limestone to the Cambro-Ordivician shales. The folding possibly took place before the thrusting at Roppa, as the thrust sheets seem to stop against the fold.

The Herfjell anticlinorium.

As indicated in Figs. 7 and 22, the mountain Herfjell lies within an anticlinorium. Strata in the south-eastern part of the mountain are inverted, the fold axes dipping at 10° to the WNW. Both the axes of small-scale folds and a mineral lineation follows this direction. Fracture cleavage dips to the north (Fig. 6 A).

With the exception of parts of Nord-Torpa, in the Vangsås Formation (Fig. 1, J-L 47), Herfjell is the only locality within the area where rocks are found in an inverted position. The bedding attitude changes from nearly horizontal at Nyseter through vertical 0.5 km east of Nyseter to inversion 1 km east of Nyseter where the fold is overturned, its southern limb dipping steeply to the north (Fig. 7 A).

It is evident that a strong frictional dragging effect occurred along the western border of the Sparagmite Basin when the rocks of the Sparagmite Group were folded and pushed southwards. The calcareous rocks of Herfjell are less competent than the corresponding rocks further south in the area, and this may be the reason why they were inverted in Herfjell.

The fold causing the inversion in the southern part of Herfjell shows a change of axial trend from a WNW-direction in Herfjell to a more NW-direction in the hill Verskei (Fig. 1, D 2) where the fold dies out (Fig. 20).

The folded area south of Herfjell.

South of Herfjell the lithological sequence in the Sparagmite Basin seems to be gently folded. On the plateau west of Auggedalen (Fig. 1, F-G 4-5) no folds are seen; however, the Biri Limestone at Skjellbreidseter (Fig. 1, G 4) has boudinage structures indicating a compression from the north, the boudins being aligned east-west.

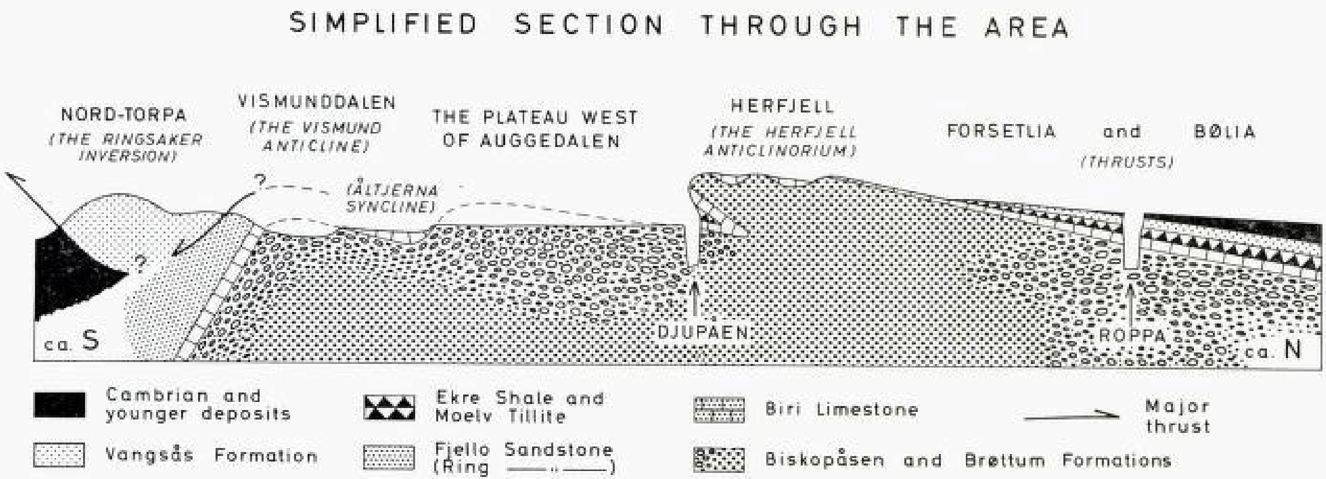

Fig. 22. A simplified section from ca. NNW to ca. SSE through the area.

The flat area around the Åltjerna setrene (Fig. 1, H 5) displays a gentle synclinal structure, the Åltjerna syncline.

An anticline (the Vismund anticline) is exposed in the flat bottom of the valley Vismunddalen (Fig. 1, I-J 5-7). To the east the anticline is well developed with a steep southern limb, while to the west the anticlinal structure dies out. The Vismund anticline is a western continuation of the Moelv anticline (Skjeseth, 1963, pp 89—91).

As indicated in Fig. 22, all the folds south of Herfjell are asymmetrical with southern limbs dipping more steeply than the northern ones. The axes dip gently to NW (Fig. 20).

Autochthonous rocks.

Strata to the north of Roppa are less disturbed than in any other part of the area. Folds here are gentle and, as seen from Fig. 16, the bedding is nearly flat. The thrusts found in the Roppa valley die out northwards, as shown on the maps. Joints are uncommon. Their main trend is NNE-SSW as shown in Fig. 18: the lower part of the river Roppa follows this direction. The ground north of the Roppa valley is that part of the area where fossils (in Cambro-Ordovician strata) and primary sedimentary boundaries are best preserved.

Synopsis of the tectonic history.

The tectonic history of the area may be summarized as follows:

1) The Vangsås Formation, with overlying Cambrian shales, was thrust out of the Sparagmite Basin.
2) The older formations were compressed and folded against the south-western border of the basin, the fold axes being approximately parallel to the border.
3) Thrust sheets were transported towards the east (e.g., Roppa).
4) Faulting took place and joints were developed; some faults, however, were initiated contemporaneously with the folding, and some of the joints are also thought to have developed at that time.
5) Some of the vertical joints of north-south trend may be of post-Caledonian age.

Within the area there are many structural problems as yet unsolved: here only the main features have been outlined.

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Abbreviations:

GFF : Geologiska Föreningen i Stockholms Förhandlingar. Stockholm.
J. Geol.: Journal of Geology. Chicago.
NGT : Norsk Geologisk Tidsskrift. (Kristiania) Oslo.
NGU : Norges Geologiske Undersøkelse. (Christiania) (Kristiania) Oslo.
SGU : Sveriges Geologiska Undersökning. Stockholm.

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First manuscript received in November 1969, revised manuscript received March 1970.

GEOLOGICAL INVESTIGATION IN THE BYGLANDSFJORDEN–GYVATN AREA

THE PRECAMBRIAN ROCKS OF THE TELEMARK AREA IN SOUTH CENTRAL NORWAY. NO IX

by

Juliusz H. Teisseyre[1])

ABSTRACT

The area consists of Precambrian migmatitic gneisses in which a group of banded gneisses (migmatitic gneisses, sensu stricto) and a group of granite gneisses have been distinguished. Two stages of folding are readily ditinguishable, and it is possible to find some traces of an even older structure (F 1). The F 1 structures were refolded by a system of NW-SE folds (F 2) and later by a system of E-W folds (F 3). The second period of folding produced the most prominent structures, while the third episode is only of minor importance.

INTRODUCTION

The mapped area lies in the north-western part of Aust Agder county in the Setesdal valley, S. Norway (see Location Map Fig. 12). The northern boundary of the area is at 58°40′15″ N and the southern boundary at 58°40′00″ N; in the east the area is bounded by Byglandsfjord lake and in the west by the Skjerka river and Gyvatn lake. The area covered is about 73 km^2.

The Setesdal valley is the major topographical feature; it is U-shaped and, in the area under consideration, about 1.5–3 km wide. Gyvatn

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lake and Sandvatn lake lie in a second U-shaped valley to the west, while the central part of the area is an elevated lake plateau with several ranges of hills rising to 100–150 m.

The area consists exclusively of migmatitic gneisses partially covered by thin layers of Quaternary deposits. All rocks in the area completely recrystallized. For this reason it is impossible to introduce even an approximate stratigraphic sequence.

The area under consideration has not previously been investigated geologically. It forms part of the great Precambrian shield of Southern Norway and belongs to the so called Telemark–Rogaland Region. The mining district of Evje to the south has been described by Barth (1945, 1947) and Bjørlykke (1947).

The present field investigations included geological mapping on a scale of 1 : 25 000, structural measurements, and some megascopic observations of rocks and ore minerals.

This preliminary report contains a description of the rocks and a summary of the structural investigations.

DESCRIPTION OF ROCKS

It seems practical to subdivide all of the migmatic gneisses into two main groups: granite gneisses and banded gneisses. This structural classification is based mainly on the works of Mehnert (1962 p. 97–111) and Polovinkina (1966). Rock descriptions are based largely on megascopic examination and staining techniques (using sodium cobaltinitrite) have been employed to facilitate mineral identification.

Granite gneisses.

Rocks belonging to this group are widespread throughout most of the studied area, and there are several structural types of the granite gneisses.

Homophanic granite gneisses. The term «homophanic texture» was introduced by J. J. Sederholm for those granite gneisses which show random orientation of mineral grains and which are structurally similar to intrusive granites. The present author chooses to use this term instead of the name granite to stress the metamorphic origin of these rocks. Bands, veins, nests, or even completely irregular bodies of homophanic granite gneiss are common in all parts of the area, but

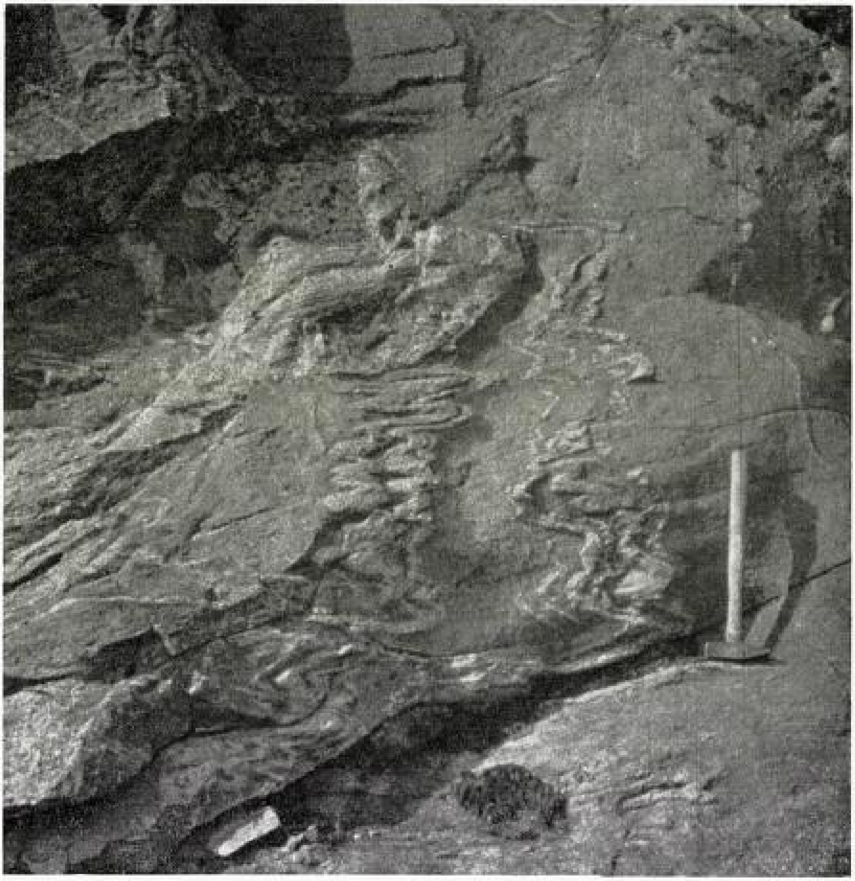

Fig. 1. Mafic inclusion with ptygmatic folding within granite gneisses. About 1.2 km SE of Frøyrak.

bodies large enough to be shown on the present map scale are rather rare. The homophanic granite gneisses are found mainly in the vicinity of Bø. These rocks are fine- to medium-grained, pink, pale rose, or greyish white in colour. The chief mafic constituent is biotite while green to brown-green hornblende plays a minor role. Felsic constituents of these rocks exceed 75 % by volume. Quartz grain boundaries vary from straight to irregular. Undulose extinction is a characteristic feature of the quartz. Plagioclase is an intermediate or basic oligoclase up to andesine. Porphyroblasts of this mineral (up to 6 mm) are fairly uncommon. Sericitization of plagioclases may be present either in the cores or along cleavages. K-feldspars frequently display microcline twinning; microperthitic as well as myrmekitic textures are common. Sphene, magnetite, and zircon are the most notable accessory minerals.

Stained slabs show that quartz and plagioclase are randomly dispersed while K-feldspar is often concentrated in nests and shows a distinct tendency to form porphyroblasts up to 5—7 cm across, and usually displaying tabular habit. They usually contain small inclusions of plagioclase and myrmekite, and commonly show a megascopic perthitic structure. Where K-feldspar porphyroblasts become common, the rock passes into porphyroblastic granite gneiss. The boundary between equigranular and porphyroblastic varieties is always gradational. The porphyroblastic granite gneisses are closely related to the augen gneisses. Both varieties are rare as large rock masses, but small bodies are common in several parts of the area.

Spotted granite gneiss («Forellen-migmatit» or «granite gneiss with Forellen-migmatische struktur», Angel & Staber 1937) contain numerous elongated patches of mafic minerals (mostly hornblende) in a completely homophanic rock matrix, the patches forming a pattern similar to that on the skin of a trout. This type of rock is probably transitional between the homophanic granite gneiss and the laminated granite gneiss. In certain cases where the mafic patches are especially elongated, the rocks display a more or less distinct small scale foliation.

Laminated granite gneisses are widespread. Compared to the homophanic gneisses described above, the laminated gneisses are richer in plagioclase and mafic constituents, especially in biotite which is concentrated along the subparallel laminae. Where the laminae become

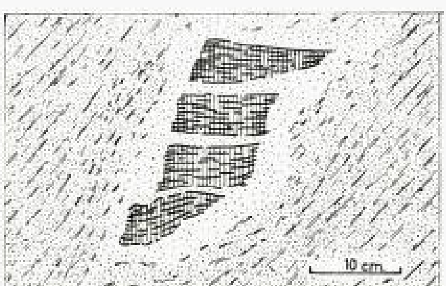

Fig. 2. Mafic inclusion within laminated granite gneiss. The body is broken along structural planes and rimmed by a leucocratic zone. From the ridge between the unnamed lakes at elevations marked 574 m and 593 m.

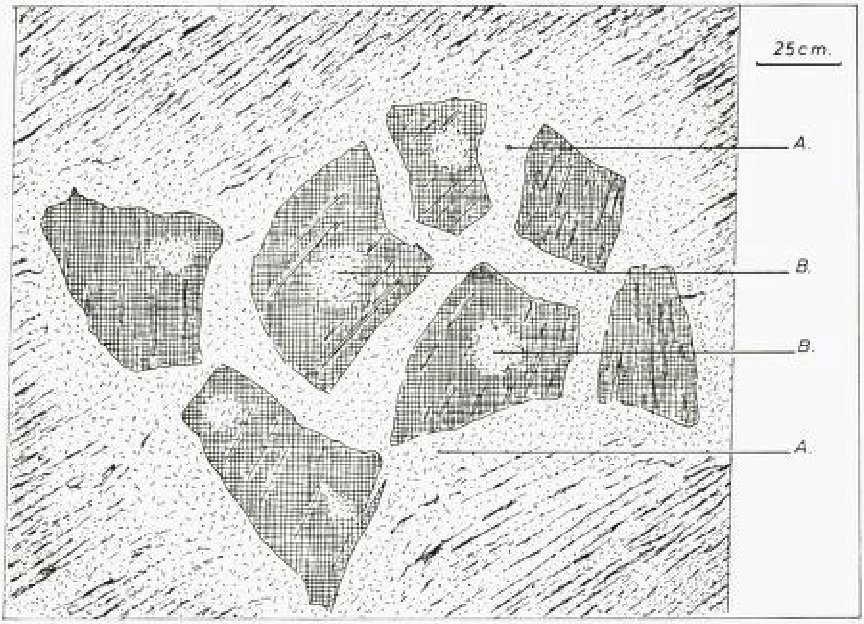

Fig. 3. An agmatite formed by mafic bodies within distinctly laminated granite gneiss. The dark bodies show rather sharp boundaries against the surrounding pegmatite (A), while pegmatite nests (B) inside the dark bodies have gradational boundaries. Roadside exposure 2.2 km SE of Bø.

wider the rock passes into s t r i p e d g n e i s s and finally into banded gneiss.

Irregularly shaped, widely spaced dark bodies are common among all varieties of granite gneiss, and it is possible to map zones especially rich in such bodies. The petrographic character of the dark bodies is rather diverse. Most abundant are amphibolites and monomineralic biotite rocks. The last remaining traces of these mafic bodies are skialiths and nebulitic structures in granite gneiss. Granitized mafic inclusions commonly display a distinct folding of ptygmatic character (see Fig. 1), while less strongly granitized bodies were apparently more rigid and formed breccia or agmatite structures (see Figs. 2, 3, 4). Around the dark bodies there are often light aureoles, sometimes pegmatitic. Pegmatites are also commonly present inside the mafic bodies as veins or nests, Figs. 2, 5. Where dark bodies within granite gneisses are numerous and display regular subparallel arrangement, the rock may grade into banded gneiss.

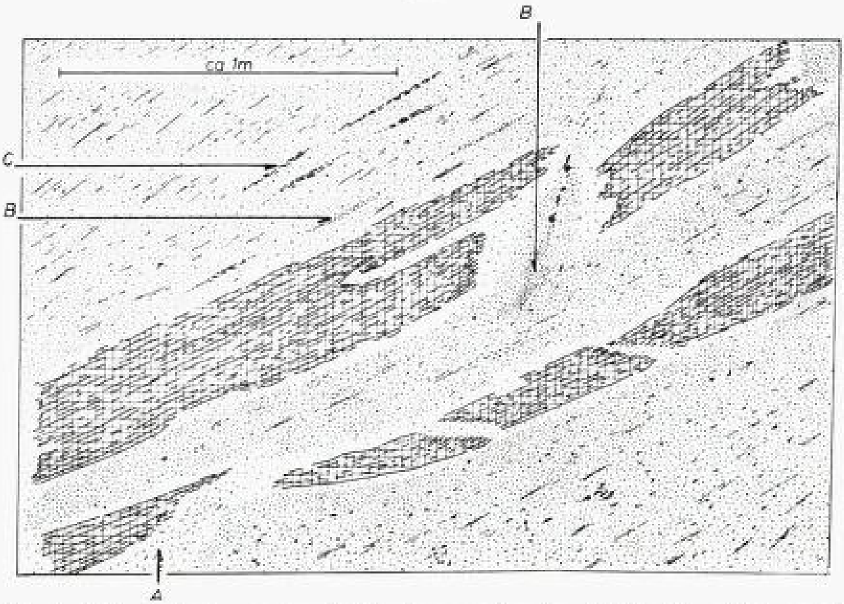

Fig. 4. Mafic inclusions surrounded by leucocratic rims (sahlbands) within granite gneiss. (A) Pegmatite nest with indistinct boundaries. (B) Nebulitic structure formed by dark minerals. (C) Schlieren structure (compare diagram Fig. 8 B). Roadside exposure about 0.4 km to the north of Sines Peninsula.

Fig. 5. Net of pegmatite veins in a dark band of banded gneisses, (light band to the right). Underwater exposure occurring in the southern part of the unnamed lake (565 m).

Banded gneisses.

The banded gneisses are less widespread, tending to predominate in the western and south-western part of the area. Commonly there is no sharp boundary between these groups, and all transitions are possible.

Banded gneisses comprise alternating dark and light coloured bands varying in thickness from several centimetres to several metres. Petrographically, the rocks within the light bands may be similar to rocks of the homophanic granite gneisses, the laminated granite gneisses or (in some areas) the pegmatites.

The rock material in the dark bands is much more diversified. The most strongly granitized members are light- to dark-grey gneisses which are dioritic in character. Their structure ranges from gneissic to almost completely homophanic; the colour is dark grey, much darker than the granite gneisses. Plagioclase is rather abundant but quartz is scarce.

The most widespread type of mafic layers is essentially a finely laminated grey gneiss. These fine-grained plagioclase gneisses commonly show what German geologists have called «Perlen-gneiss-structur». Some rather broad and uniform bands of perlen gneiss extending over distance of some hundred metres have been indicated on the accompanying map. These rocks are usually rich in hornblende and contain relics of amphibolites.

Felsic constituents are quartz and plagioclase (andesine), with K-feldspar in small amounts. There are two variants, one with biotite, the other with hornblende as the main mafic mineral. Biotite-rich types often contain cordierite and secondary muscovite. Hornblende-rich types may gradually pass into an amphibolite by a decrease in quartz content.

Laminae up to 1 cm thick or larger bodies of amphibolite are widespread, but most of them are too small to be shown on the map. These rocks vary from hornblendites to rocks rich in plagioclase. The first type consists almost exclusively of hornblende ($\alpha =$ yellowish green, $\beta =$ green, $\gamma =$ bluish green) with magnetite as an important constituent mineral. Apatite and sphene are common accessories.

Amphibolites are much more common. Hornblende and plagioclase (andesine) constitute the bulk of all specimens examined. The hornblende of these rocks displays various colours from bluish green to brownish green. Sphene, magnetite and apatite are commonly present while quartz and biotite occur in variable amounts. Almost all of the amphibole rocks show distinct traces of biotitization of hornblende. This process is related to a rather strong granitization of the whole area. Biotitization of amphibolites is clearly visible in the vicinity of point 690 m on the map (to the north of Åmdal). This extensive amphibolite is composed of laminae of hornblende with layers of biotite and feldspar. Some parts of the body are almost completely biotitized. There are also numerous nests of epidote which probably originated during the biotitization of the amphibolite.

Pegmatite, aplite and silexite.

Numerous veins, nests or completely irregular bodies of pegmatite, aplite, and silexite cut all other rock types; Figs. 1, 2, 3, 4, 5, 6, 7.

Pegmatites are associated in particular with granite gneisses, but are also developed within mafic inclusions in granite gneisses and within dark layers of banded gneisses including smaller or larger amphibolite bodies. As mentioned above, pegmatites in some places form aureoles around mafic bodies.

Pegmatites rich in K-feldspar are associated with granite gneisses in which they often form veins and nests. When associated with mafic bodies the pegmatites are often pure quartz-plagioclase rocks, commonly forming veins (see Fig. 5), although nests without sharp boundaries can be seen in places. Dark minerals in the pegmatites include both biotite and hornblende. A description of the ore minerals is presented later.

Fine-grained pegmatites with few mafic minerals may grade into very fine-grained aplites. Numerous veins of aplitic character have been observed. These always form small bodies which mostly accompany the granite gneisses. In contrast to the pegmatites, the aplites contain no ore minerals.

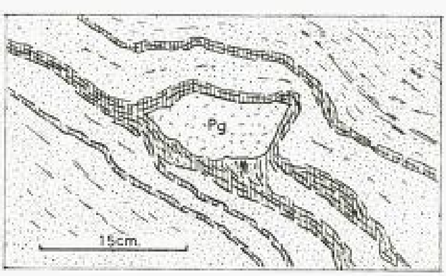

Fig. 6. An irregular pegmatitic body (Pg) within the banded gneisses. From the roadside exposure on the north-east side of the Daleåna river.

With decreasing feldspar content both pegmatites and aplites grade into silexites. Zones, bands and dykes rich in quartz are common. They are grey or bluish grey in colour, and are commonly parallel to the lamination of the surrounding rock. They may however cut across structural planes. A few silexites contain ore minerals.

Ores.

Ores occurring within the area studied are mostly associated with pegmatites. Sulphide-bearing amphibolites have been found only in one road outcrop near Vik creek.

Among the ore-bearing pegmatites, magnetite-bearing rocks are the most widespread. They are common on the east slopes of Vik heii and along the shore of Byglandsfjord lake to the north of Vik creek. In several cases sulphides accompany magnetite. Sulphide-bearing pegmatites are also common. Such rocks are exposed along the road south-east of Tveitå farm. At these same localities hematite-bearing pegmatites have been found. It is possible that the hematite is a product of oxidation of magnetite.

Fragments of rocks containing molybdenite were removed from the cellar of a building at Dale farm during excavation.

The most important occurrences of ore minerals are shown on the accompanying map (Fig. 12).

STRUCTURAL INVESTIGATIONS

Introduction.

The character of the following mesoscopic structures have been studied and measured systematically: folds, faults, veins, several types of «S» surfaces including planes of rock laminations, joint planes, slickenside surfaces. All structural measurements are given in terms of the 360° scale.

The gneisses of the Byglandsfjord—Gyvatn area have undergone polyphase deformation. Two distinct episodes of folding are present, designated F 2 and F 3 together with probably the earliest fold phase F 1, which is of a relict character.

A gneissose structure in granite gneisses and a banded structure in banded gneisses are found to be the dominant planar structures in the area under investigation. The former is a kind of small scale lamination and comprises almost infinite repetition of laminae enriched in biotite. This structure has been developed during the F 2 phase of folding. Micas however have recrystallized mimetically after the earlier fabric, most probably after initial lamination of the supracrust transposed during the F 1 phase of folding. The banded structure is comprised of alternating maphic and granitoid bands of varying thickness from centimetres up to metres and even dozens of metres. This feature is probably of complex origin, probably due to an incomplete metamorphic segregation of the main rock-forming minerals. Moreover some local intrusions of pegmatites subparallel to the foliation of the host rocks are often visible. Regular banded gneisses with straight mafic layers extending over long distances, may suggest that the banding is partly inherited from the bedding of sediments as well as the initial banding of ignous rocks.

Apart from the gneissosity and banded structure, the joint systems are the most pronounced planar structures in the rocks of this area. They have most probably originated after the main metamorphic stage and for that reason will be elaborated separately in a later paper.

Minor folds and related structures.

A. First generation of minor folds — F 1.

Minor folds of F 1 age have been observed almost exclusively in the mafic layers of the banded gneisses. They fold the banding which, in most cases, seems to be inherited from the initial bedding or banding of the rocks. These first generation minor folds are tight or isoclinal. Minor cracks and thrust surfaces commonly accompany these folds (see Figs. 7, 9). Pegmatites have very often intruded the banded gneisses along these thrust planes or cut obliquely across the limbs of folds (see Fig. 7).

Axial directions of the F 1 folds are strongly superimposed by later deformations (F 2 and F 3 phases of folding). The structures which belong to this folding phase are presented in a diagram (see Fig. 11 a).

B. Second generation of minor folds — F 2 and related structures.

Minor folds.

The minor structures associated with the second episode of folding are abundant in the whole area between Byglandsfjorden and Gyvatn. The folds belonging to this phase have deformed earlier minor folds.

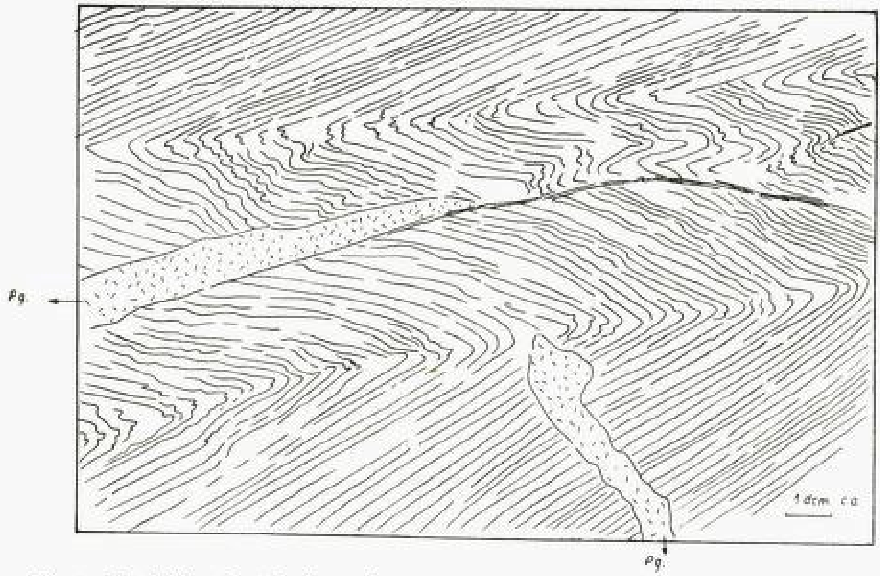

Fig. 7. F 1 fold and axial planar shear. (Pg) pegmatites emplaced along shear surface and across lamination of the rock. About 1.5 km WNW of Sandvikfjellet (721 m).

They may be distinguished from the F 1 folds by their structural style. Second generation minor folds usually show evidence of plastic deformation, the ptygmatic type of folding being common. The fold structures are seldom associated with cracks and thrust planes (see Fig. 1). They are however quite variable, this variations mainly dependent on the (1) physical properties of rock (structure texture and mineral composition) (2) local tectonic environment. It must be stressed that the great mechanical differences between mafic paleosom and leucocratic neosom caused some pecularities in the development of F 2 structures. In places rich in mafic bodies, agmatitic and breccia like structures are clearly associated with the second generation of folds.

Breccia and agmatitic structures.

Some amphibolitic bands and layers (very often completely biotitized) have been fractured due to movement of much more plastic, strongly granitized light-coloured rock material. Amphibolitic bodies have apparently been pulled apart along structural planes (Fig. 3) or along shear planes oblique to the lamination of the rocks (Fig. 4). The second case probably resulted from a stretching of the amhpibolitic bodies while tension acted in the plane of lamination.

The distribution of the mafic bodies is readily studied in the extensive road exposures about 0.4 km to the north of Sines point (Fig. 4). Plots of the orientation of the lamination in the main rock types at this locality are given in Fig. 8 B.

In the area south-west of Hommeknuten several «boudins» of amphibolite surrounded by light-coloured rocks have been observed. In some places the amphibolite layers have apparently been pulled apart into isolated «boudins»[1]) while in others they display a typical necking with alternate thinning and thickening. The amphibolite layers are up to several metres thick and boudinage is developed extensively. These boudins are observed in the limbs of major folds most probably connected with the second episode of folding.

Another type of breccia-like relationship has been observed in the extensive road-cuts about 2 km south-west of Bø. Several dark masses are broken into smaller bodies (from several centimetres to several decimetres in diameter) and dispersed within the granite gneisses, often being surrounded by pegmatite. These fragments were apparently broken along more or less irregular surfaces which usually have no systematic relationship to the internal lamination of the mafic bodies. Later those bodies have been dispersed and rotated. The exposures did not permit measurement for structural analysis. The author believes that these true agmatitic structures are also related to the second generation folds which are clearly evident in nearby crags.

It is believed that where the dark bodies show little evidence of granitization and are relatively rigid, they were dispersed and rotated discordantly within the granitic migma. More strongly granitized bodies show a deformation concordant with that of the surrounding granite gneisses.

Ptygmatic folds and their relationships to agmatitic structures.

The relationship between agmatitic structures and ptygmatic folds is seen in the extensive road-side exposure about 1.2 km to the south-west of Frøyrak. There are several types of mafic enclosures in the granite gneisses. In some parts of this exposure, the rocks display distinct

[1]) Similar structures have been described as so called «Zerrungstexturen» by Mehnert (1962).

«Schollentextures» (term used by K. R. Mehnert 1962). Blocks («Schollen») of mafic rock have been broken and then dispersed in the strongly migmatized and plastic granitic «leucosome». These bodies display various stages of granitization. Less altered by the metasomatic and migmatitic processes are the amphibolites and even monomineralic biotitic rocks. These show evidence of having been rigid bodies during deformation since they are broken up into polygonal blocks with rather sharp boundaries. In the next stage of granitization dark enclosures commonly display venitic structure and are strongly folded (see Fig. 1), rather than broken. In some places it is possible to find some ptygmatic folds which appear as structures resulting from the strong superposition of two or more folding episodes. According to the opinion of several authors such type of ptygmatic folds may be formed even in one folding phase (see H. Ramberg experiments, 1959). The interrelation between axes of ptygmatic folds within the mafic enclosures and the deformation of the surrounding granite gneiss is visible in some places (Fig. 8 A). As the rock becomes more granitic, the outlines of the mafic bodies become indistinct and they may then pass into «schlieren» and finally into nebulitic structures. While granitized xenoliths in several places show random orientation in comparison with the structure of the surrounding granite gneisses, schlieren and nebulitic bodies are more or less concordant. This is clearly visible on the structural diagram of fold axes and poles to lamination for Frøyrak exposures as seen in Fig. 8 A.

This concordance of fold axes in the mafic bodies isolated within granite gneiss may be explained in one of two ways:

1. folding has been generated and controlled by similar forces in all the mafic bodies in spite of the fact that they are isolated from each other by more plastic granitic material;
2. granitization is younger than the folding of the mafic rocks.

The latter point of view is hardly tenable in this case. It is thought that migmatitization processes are syn- or late-kinematic with respect to the F 2 phase respectively and have thus probably been accompanied by tectonic movements.

C. Third generation of minor folds – F 3.

Minor folds of the third generation commonly show evidence of plastic deformation, but tight folds and folds of ptygmatic type do not occur among structures of this episode. Folds of the third generation

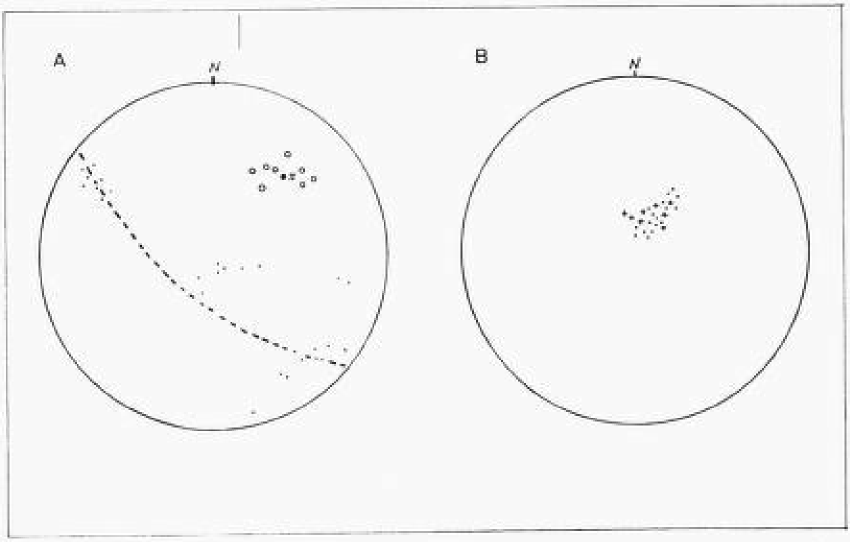

Fig. 8 A. Fold axes (small circles) in venite and poles to lamination planes (dots) in granite gneiss. Pole of girdle marked by black dot. (Lower hemisphere of Wulff net). Exposure 1.2 km SW of Frøyrak (See Fig. 1).

Fig. 8 B. Poles to lamination planes in granite gneisses (dots) and in dark inclusion (crosses). Roadcut about 0.4 km N of Sines peninsula (see Fig. 4).

Fig. 9. Fold structure (probably F 1 fold) dislocated by crestal fracture in banded gneisses, small crag in south Aneknuten (577 m).

are mediumsize structures (several centimetres up to decimetres in amplitude), open in style and symmetrical. Axial direction is constant, E-W up to ESE-WNW. These structures are noticeable fewer in number than F 2 and even F 1. The F 3 structures seem to be of lesser significance than the earlier folding of the F 1 and F 2 phases. The third folding phase has probably been post-migmatitic; however there was probably some pegmatite emplacement during this stage.

D. *Structural control of pegmatitic and aplitic veins and bodies.*

All the rocks in the whole area studied are cut by pegmatitic veins. Many are subparallell to the banding in the host rocks.

In several places «concordant» veins of pegmatites, aplites and silexites pass into discordant veins. The discordant veins have often been intensively folded probably due to differential movement along structural planes (Fig. 10). Small pegmatitic bands in the banded gneisses are in some localities squeezed into boudins.

Some peculiar structures have been observed in certain parts of the banded gneisses, and it is almost impossible to describe them in the conventional terminology of structural geology. They consist of irregular pegmatite bodies surrounded by irregularly folded, dark layers of banded gneiss (Fig. 6). The relationship between pegmatites and agmatitic structures has been described above.

Summing up the above described features we can say that pegmatites were formed mainly during the F 2 tectonism and are strongly involved with this deformation. Some pegmatitic bodies are probably younger.

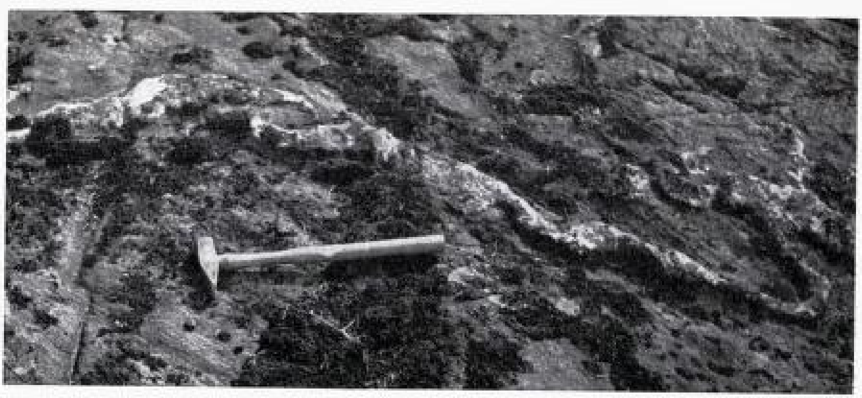

Fig. 10. Pegmatite vein in the banded gneisses, folded probably due to differential movement along lamination planes.

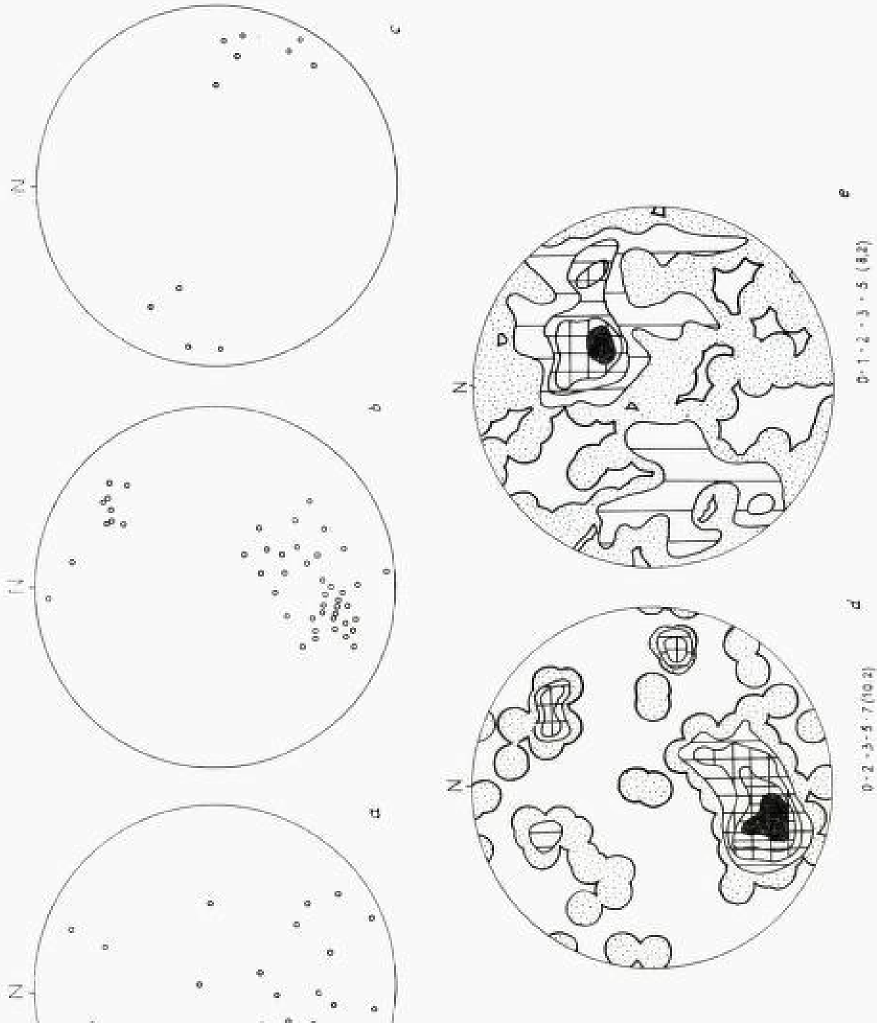

Fig. 11. Structural diagrams from the Byglandsfjorden—Gyvatn area: (a) F 1 fold axes, (b) F 2 fold axes, (c) F 3 fold axes, (d) contour diagram of the fold axes F 1 + F 2 + F 3, (e) contour diagram of the poles of lamination.

Geometric analysis of mesostructures.

During the field work 83 measurements of axes of minor folds and 379 measurements of foliation were made together with a few hundred measurements of joint surfaces. Measurements of joints have not been analysed as yet.

The diagram of F 1 folds (see Fig. 11 a) shows such a strong dispersion of axial directions that it is impossible to speak of a dominant F 1 trend. This dispersion is a consequence of refolding, involving early minor folds, during F 2 and F 3 phases.

The diagram of F 2 fold axes (see Fig. 11 b) shows a much clearer pattern. Most of the F 2 minor fold axes plunge toward SSW, S and SSE at an intermediate angle. A lesser concentration of plots around NNE is also clearly visible. This pattern of distribution of fold axes may be explained as an effect of superposition of the F 2 structures by the later F 3 phase.

The diagram of F 3 folds (see Fig. 11 c) shows a few measurements distributed in the E-ESE and W-WNW directions. All the folds belonging to the F 3 phase have either horizontal or gently plunging axes.

A contour diagram (Fig. 11 d) for the whole area shows two main trends of fold axes. The first system is strongly dispersed and trends

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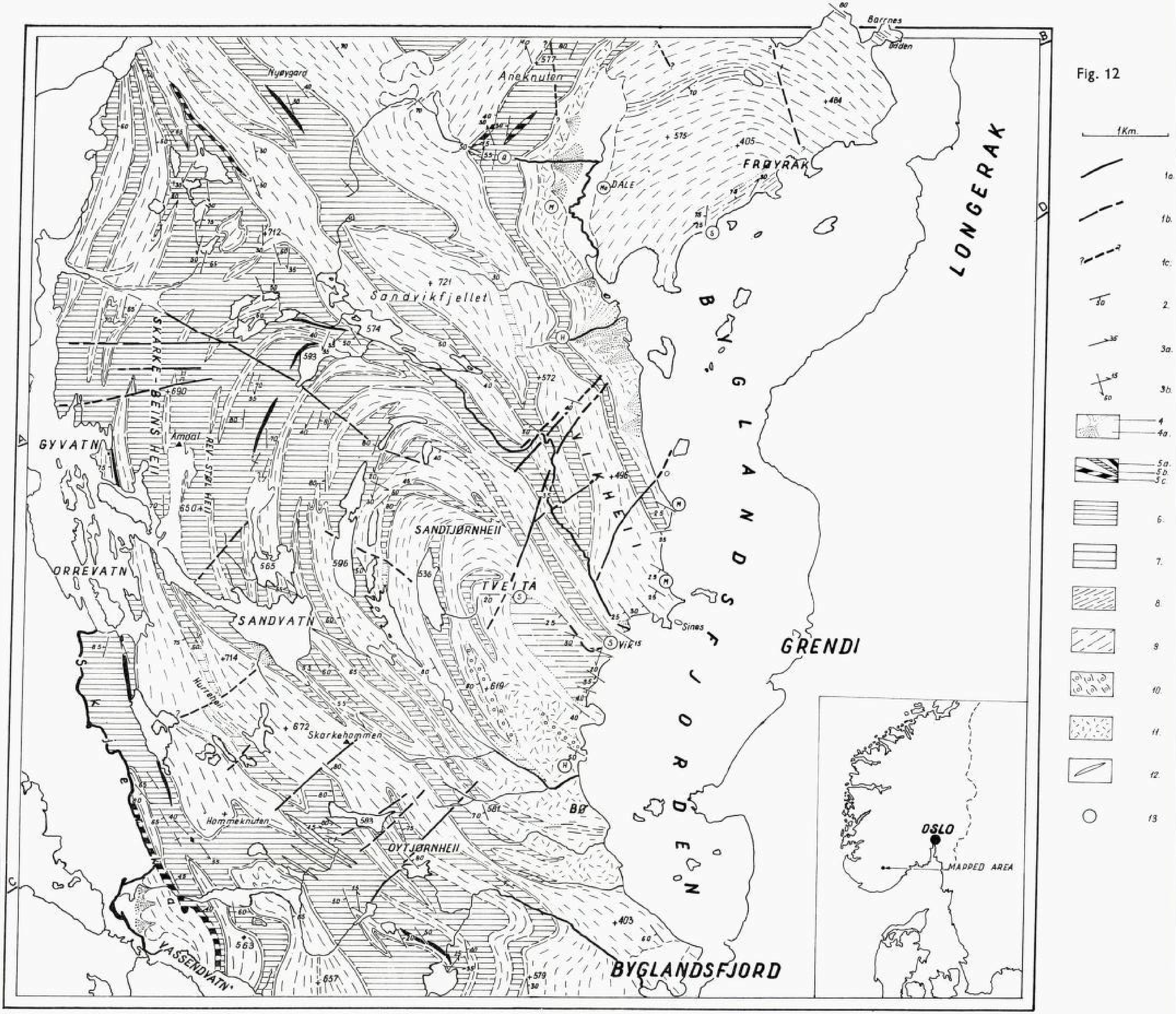
Fig. 12. Geological map of the area between Byglandsfjorden and Gyvatn (Setesdal, S. Norway).

1. Faults. a. definite, b. approximate, c. inferred or concealed.
2. Strike and dip of metamorphic foliation or banding.
3. a. Direction and plunge of minor fold axes or lineation.
   b. Axial directions of cross folds.
4. Quaternary deposits (undifferentiated glacial and fluvioglacial deposits, alluvium. 4a. = talus).
5. a. Amphibolites, b. biotitized amphibolites, c. grey plagioclase gneisses with numerous amphibolite intercalations.
6. Banded gneiss.
7. Light coloured variety of banded gneiss, gradational into granite gneiss.
8. Granite gneiss rich in dark bodies, gradational into banded gneiss.
9. Granite gneiss with laminated, striped or «Forellen-migmatite» structure.
10. Granite gneiss with porphyroblastic texture grading into augen gneisses.
11. Granite gneiss with homophanic and equigranular structure.
12. Larger pegmatite bodies.
13. Mineral occurrences: Mo = Molybdenite, M = Magnetite, H = Hematite, S = Sulphides, Q = Rose quartz.

Fig. 12



LONGERAK
BYGLANDSFJORDEN
GRENDI
BYGLANDSFJORD
FRØYRAK
DALE
Barrnes
Aneknuten
Nyøygard
Sandvikfjellet
VIKHEII
SANDTJØRNHEII
TVEITA
Sines
Vik
BØ
SKARKE-BEINS HEII
REV-STØL HEII
Amdal
GYVATN
ORREVATN
SANDVATN
Hurreheii
Skarkehommen
Hommeknuten
OYTJØRNHEII
VASSENDVATN
OSLO
MAPPED AREA
1 Km.
1a.
1b.
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5b.
5c.
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7.
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13.

NNE-SSW to NW-SE with a 10,2 % maximum to the SSW and two weak sub-maxima plunging NNW and NE. The second is an E-W fold system showing a weak maximum plunging gently towards ESE. It should be stressed that fold axes belonging to the E-W system plunge gently in both directions and show little dispersion. The «first system» described above is most probably complex, it consists of F 1 structures rotated during the second period of folding (F 2) together with F 2 folds. The «second E-W system» is that of the F 3 fold episode.

Diagram Fig. 11 b showing poles to lamination surfaces presents an unclear picture, because of the superimposed folding (F 2 and F 3), but it is possible to trace the girdle related to the folding about axes of the F 2 fold system. The other system of folds F 3 (W-E) has probably produced some of the minima on this girdle.

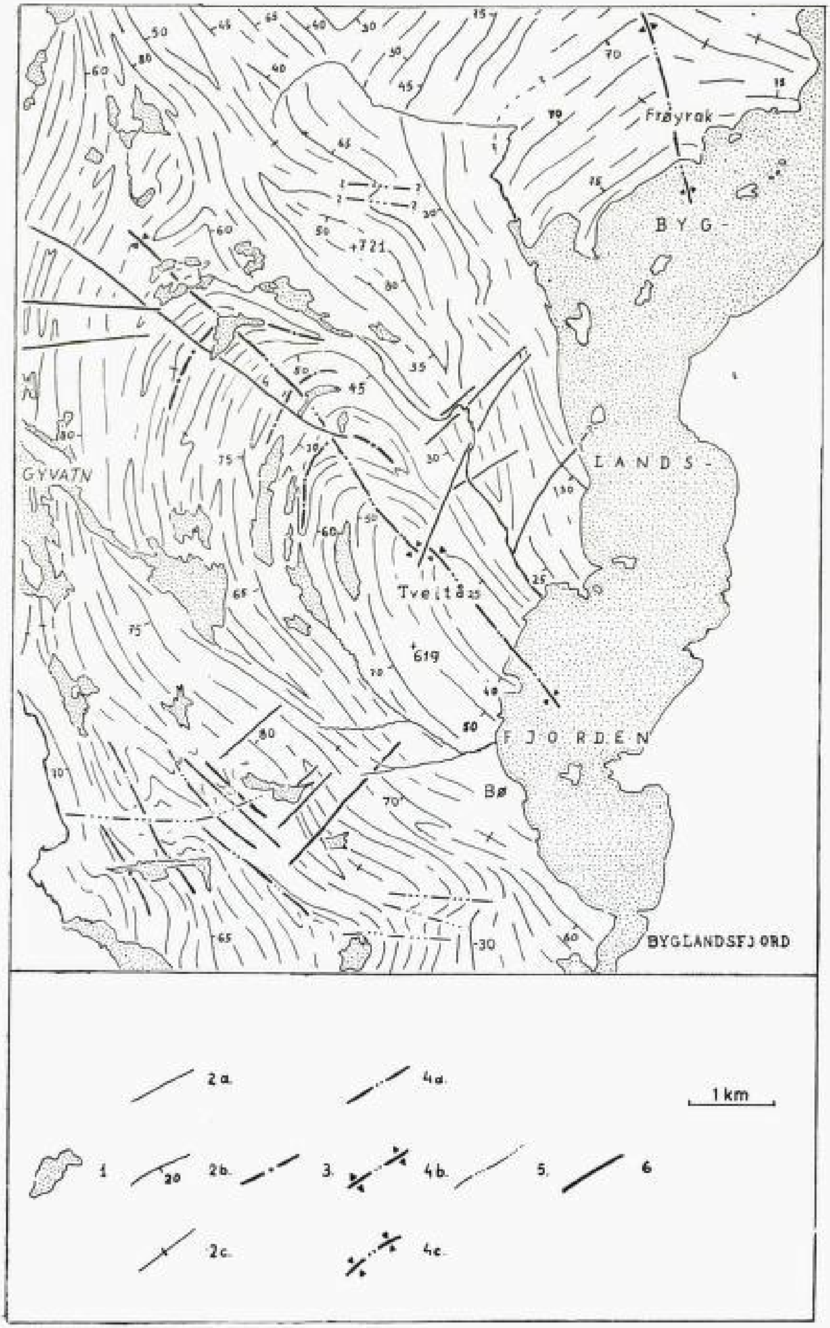
The major structures.

All three periods of folding may be traced on a larger scale by analysis of the outcrop pattern. The minor F 1 fold structures can be traced only as relics. They are best seen in the area to the north-west of the farm Tveitå.

The F 2 structure are dominating in the whole area. They have been studied by a statistical analysis of minor structures as well as a study of the outcrop pattern on the geological map (see Figs. 8, 11, 12, 13). From this the following conclusions have been reached.

The north-eastern part of the area includes a large portion of a big fold structure which appears to be antiformal. This structure is here referred to as the Frøyrak antiform. Observations of minor structures in this domain show that fold axes plunge towards north-north-east, north and north-west at low or intermediate angles. Weak concentrations of plots of fold axes and the rather random orientation of lamination surfaces may be caused by the superposition of F 2 folds and subsequent folding and faulting on earlier F 1 structures. Unfortunately, it is very hard to discern between effects of F 1 and F 2 episodes of deformation especially in the core of the antiform.

The Tveitå synform is situated in the central part of the area. The north-east limb of this synform dips 25°—40° to the south-west, and poles of lamination surfaces along this limb form a distinct concentration on the structural diagram (Fig. 11 e). The western limb of the synform, which has steeper dips, is marked in the diagram by a less distinct


GYVATN
Frøyrak
BYG-
LANDS-
FJORDEN
BYGLANDSFJORD
Tveitå
Bø
+721
619
1 km
1
2a.
2b.
2c.
3.
4a.
4b.
4c.
5.
6

but clearly visible concentration of plots. The axial place trace of the Tveitå synform can be followed in the narrow zone where the strike direction turns from NNW-SSE to N-S and where the dip angle increases from ca. 25°—40° to 60°—90° (see Fig. 9). Observations from the Tveitå synform show a distinct concentration of axes of minor folds plunging 25°—60° toward the SSE, S and SW with a distinct maximum at SSW on the contour diagram (Fig. 11 d). The axis of the synform plunges approximately SSW at an angle ranging between 35° and 50°, while the trace of the axial plane of the fold trends NW-SE. These data are confirmed by the structural diagram as well as by analysis of the outcrop pattern (Figs. 11, 12, 13).

Both the Frøyrak antiform and the Tveitå synform were probably formed during the second period of folding (F 2). E-W trending folds (F 3 structures) are also present, but they don't play any major role. The two cross sections (Fig. 14) as well as the structural map (Fig. 13) may clarify some of the interrelations between the two major structures.

The southern area which includes the slopes of Hommeknuten, Hurrhei, Oytjørnheii and Svaen up to the Skjerka river as well as northern shores of Vassendvatn, seems to be more complicated, and it is very hard to distinguish clearly any major structures. The NW-SE trending fold axes are strongly dispersed, while the E-W or F 3 folds play a more important role than in the northern areas.

A great dispersion of fold axes belonging to the NW-SE system (F 2) supports the theory that in this southern area F 1 and F 2 fold structures have been strongly affected by superposed F 3 folding or faulting or both.

←

Fig. 13. Geological structures of the area between Byglandsfjorden and Gyvatn (Setesdal, S. Norway).

1. Lakes.
2. a. Trace of metamorphic foliation (dip undefined).
   b. Trace of metamorphic foliation (dip defined by figure).
   c. Trace of metamorphic foliation (vertical).
3. Approximate axial plane trace of major F 1 folds.
4. a. Axial plane trace of F 2 folds.
   b. Axial plane trace of Frøyrak antiform (F 2 fold).
   c. Axial plane trace of Tveitå synform (F 2 fold).
5. Axial plane trace of F 3 folds.
6. Faults.

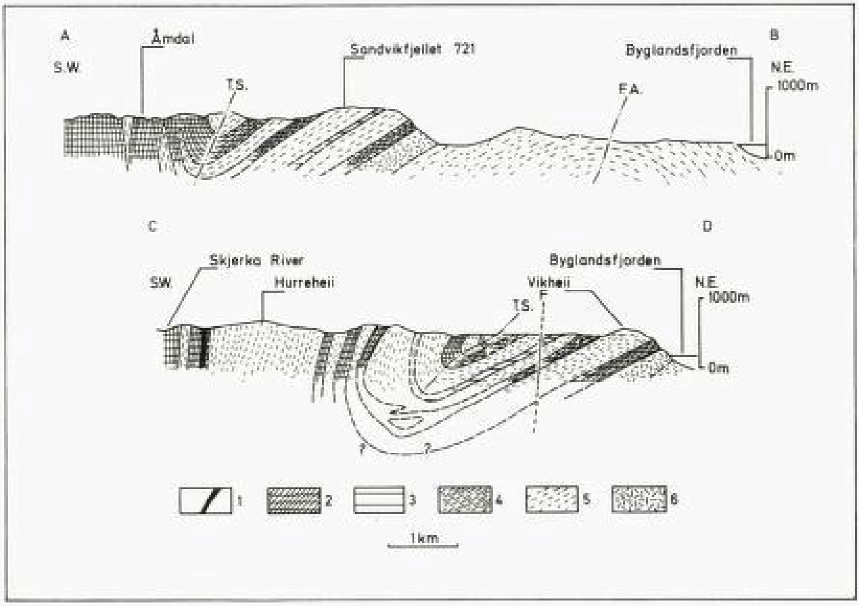


Fig. 14. Cross sections. A—B in the northern part of the area, C—D in the southern part. F.A. = axial plane in the Frøyrak antiform. T.S. = axial plane of the Tveitå synform. F = fault.

1. Amphibolites.
2. Banded gneisses.
3. Light coloured variety of banded gneisses gradational to granite gneisses.
4. Granite gneisses with numerous dark bodies gradational to banded gneisses.
5. Granite gneiss with laminated or striped structure.
6. Homophanic granite gneisses.

Faults.

The most prominent fault direction is NE-SW as seen in the area of Vik heii and Øytjørnheii. Another system trending E-W to ESE-WNW is present in the area of Skarkehommen heii and Revstøl heii. The region to the north of Frøyrak and Dale seems to be affected mostly by faults trending N-S to ESE-WSW.

It is hard to establish even approximately the character of the fault structures. Some faults seem to be younger than the F 3 structures, and these are probably of Caledonian age.

SUMMARY OF THE STRUCTURAL AND METAMORPHIC DEVELOPMENT

Even though examination of the collected material is not yet complete, sufficient data are available to permit a determination of the relationship between tectonic and metamorphic events affecting the rocks of Byglandsfjorden–Gyvatn area. The structural and metamorphic development of this area can be outlined and summarized as follows:

A) The deposition of the sediments. Due to very strong metamorphism it is almost impossible to reconstruct their initial character. These sediments probably consisted of graywackes polymictic sandstones and arkoses while shales and carbonatic rocks have played a subordinate role or even were absent. Within the whole area studied neither kizingite gneisses nor rocks of the marble – lime-silicate kindred have been found.

B) Intrusions or extrusions of basic or ultra basic rocks (now amphibolites, plagioclase gneisses and hornblendites). It also seems probable that some of these rocks may be much younger. R. K. O'Nions, R. D. Morton and H. Baadsgard (1969) have recently reported a synkinematic intrusion of gabbroic sheets from the Bamble Region.

C) The first period of folding (F 1). Some minor folds and some macrostructures belonging to this folding phase are traceable. It is quite impossible to reconstruct any trend of these structures due to the very strong superposition of later deformations. The earliest phase of regional metamorphism – of probable greenschist facies – accompanied this period of folding. The F 1 period of folding may be tentatively regarded as an equivalent of the oldest (NW-SE) folding system reported by Barth (1960) from the Evje–Ivland region.

D) Post F 1 granitisation and anatexis coeval with the stage of maximum of metamorphic grade. Migmatic gneisses of venitic type originated during this stage. These processes continued into the F 2 kinematic phase, and most of the pegmatites are of this age.

E) Late-kinematic phase of F 3. High metamorphic conditions continued up to this stage of deformation, rocks remaining in semiplastic condition.

F) Faulting and jointing which probably developed at least in part, during the Caledonian movements.

Some of the problems concerning the structural and metamorphic development of the Byglandsfjorden–Gyvatn area are solved only tentatively — others still remain unclear. More detailed mapping may help with elucidating some of the structural problems, especially those concerning relationships of the F 1 structures to later deformations. Further petrographical and petrochemical investigations may throw more light on some of the petrogenetical problems.

ACKNOWLEDGEMENTS

Field work was sponsored by the «Telemark Project», directed by Mr. Johannes A. Dons, Head Curator of the Mineralogical Geological Museum of Oslo.

The manuscript for this paper was prepared during the author's two months Norwegian State fellowship at the Mineralogical Geological Museum of Oslo, and completed with some petrographic studies carried out at the Petrological Laboratory, Geological Institute of the Polish Academy of Sciences.

The writer is particularly indebted to Mr. Johannes A. Dons for providing facilities for field work as well as for his kind help in later studies at the Mineralogical Geological Museum in Oslo. Much assistance has also been received from all members of the staff of the museum. The writer wishes to express his gratitude to Professor W. Romey of Syracuse University U.S.A. for reading and criticising an early draft of this paper. He would also like to thank Professor K. Smulikowski for helpful discussion and criticism as well as other members of the staff of the Geological Institute, Polish Academy of Sciences. The author is grateful to Prof. T. F. W. Barth and Dr. D. Roberts as well as to Mr. J. A. Dons for helpful criticism of the final draft of this paper.

The writer is also grateful to Mrs. D. Labecka for drafting the main map and to Miss B. Maurtiz for help with the photographic work. Special thanks must also go to Mr. Torleif Bø for his help and hospitality during the course of the fieldwork.

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First manuscript received January 1970, revised manuscript received in April 1970.

CORRELATION BETWEEN THE LATE PRECAMBRIAN OLDER SANDSTONE SERIES OF THE VARANGERFJORD AND TANAFJORD AREAS

PRELIMINARY REPORT

by

Signe-Line Røe[1])

Abstract.

Sedimentary rocks belonging to the late Precambrian Older Sandstone Series (overlain unconformably by the Eo-Cambrian—Lower Cambrian Vestertana Group) are described from the Vadsø—Komagelven area on the northern side of Varangerfjord. The c. 1300 m thick Varangerfjord Older Sandstone Series succession is divided into twelve informal members.

A correlation of the upper c. 1000 m of this succession with the late Precambrian Tanafjord Group is then presented.

Finally the regional discordance present between the Older Sandstone Series and the Vestertana Group is briefly mentioned.

Introduction.

The area investigated is situated on the northern side of Varangerfjord, East Finnmark, between longitudes 29° 38′ and 30° 20′ east and latitudes 70° 04′ and 70° 18′ north.

For mapping purposes, 1 : 50 000 AMS topographical maps were used as aerial photographs were not available. The mapped area covers approximately 450 sq. km. Good exposure is, in general, restricted to the coastal section and to the inland valleys.

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Previous to this investigation the area had not been mapped geologically. The geological map presented with this paper does, however, include a tract of ground mapped by Siedlecka & Siedlecki in 1968, that of the Kjøltindene–Grythaugen–Holmfjellet area. (Fig. 1 and map, fig. 3).

A brief description of the rocks occurring along the coastal section of the present area was given by Holtedahl (1918). In 1969 Hobday, Geddes and Reading (Oxford University) studied these same sediments, this work forming a part of a recent publication (Banks et al. 1970).

The purpose of the present paper is to make a lithostratigraphical comparison between the late Precambrian rocks (The Older Sandstone Series) of the Tanafjord area and part of the succession within the investigated area: this is based on six weeks field work during the summer of 1969.

Geological background.

From a geological point of view, the present area forms part of the Tanafjord–Varangerfjord region (Siedlecka & Siedlecki, 1967), which is underlain by sediments of late Precambrian and Eo-Cambrian – Lower Cambrian age. (The find of *Platysolenites antiquissimus* (Føyn 1967) has established that the youngest formation on the Varanger Peninsula is of Lower Cambrian age). To the south, this sedimentary succession unconformably overlies the crystalline Precambrian basement. To the north-east, the Trollfjord–Komagelv thrust fault forms the boundary of the Barents Sea Region (Fig. 1, also Siedlecka & Siedlecki 1967).

The rocks of the Tanafjord–Varangerfjord region have been divided into two parts: –

2. Vestertana Group (Reading 1965) c. 1450 m – sandstones and shales with two tillite formations at the base.
1. Older Sandstone Series (Føyn 1937) c. 1300 m (in the Tanafjord area) – sandstones and shales with dolomite in the upper part.

The late Precambrian rocks or the Older Sandstone Series, consisting of sandstones and shales with dolomites occurring in the upper part, is overlain with a slight angular unconformity (Holtedahl 1918, Føyn

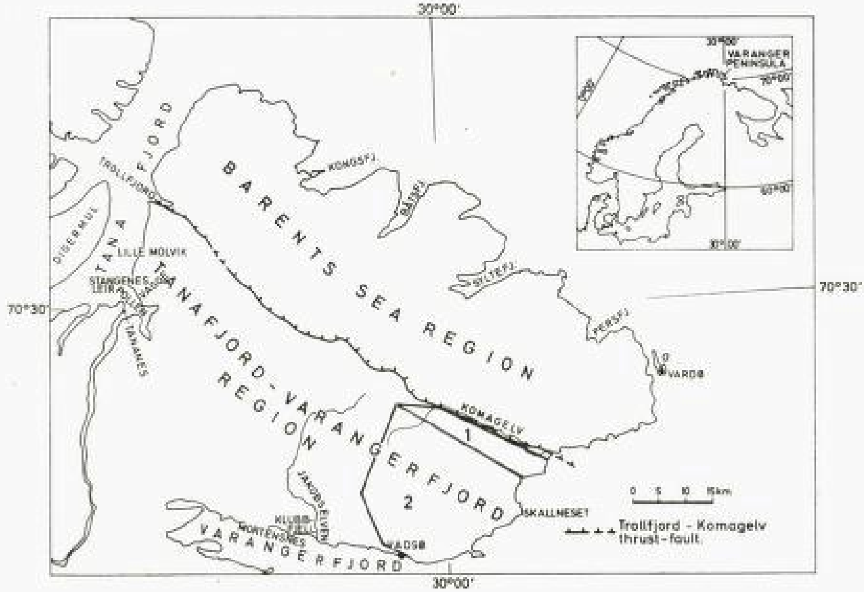

Fig. 1. Map showing the situation of the Vadsø—Komagelven area.
1. Mapped by A. Siedlecka and S. Siedlecki. (1968)
2. Mapped by the present author. (1969)

1937) by the Eo-Cambrian – Lower Cambrian tillite-bearing sandstone and shale succession, the Vestertana Group.

Detailed stratigraphic studies on the sediments of the Older Sandstone Series have previously been restricted to two coastal areas; the Tanafjord area (Føyn 1937, Siedlecka & Siedlecki 1970), and the inner part of the Varangerfjord area (Holtedahl 1918, Rosendahl 1931, Bjørlykke 1967, Banks et al. 1970). Siedlecka & Siedlecki (1970) traced the units of «The Older Sandstone Series on the Tanafjord» (Føyn 1937) inland to the Kjøltindene and Grythaugen areas (Fig. 3), and have proposed a detailed lithostratigraphic classification for this succession.

Stratigraphy.

Mapping by Siedlecka & Siedlecki in 1968 established that rocks constituting part of the Vestertana Group occur in the areas south of Kjøltindene and south-east of Skallelvskaret. In the investigated area these rocks unconformably overlie an approximately 1300 m thick sedimentary sequence belonging to the Older Sandstone Series.

Older Sandstone Series.

The Older Sandstone Series (Føyn 1937) on the northern side of the Varangerfjord consists exclusively of clastic sedimentary rocks. Unconformities appear to be absent within the succession.

Banks et al. (1970) divided the part of the sequence appearing between Mortensnes and Skallneset (Fig. 1) into six informal members; in the present paper a different scheme is proposed, the three youngest members of Banks et al. (1970) being sub-divided into a total of six members. In addition, three still younger members constituting this same succession have been introduced. Thus, the present author proposes a subdivision of the entire Older Sandstone Series succession on the northern side of Varangerfjord into twelve mappable units (members I to XII), with an estimated thickness of c. 1300 m[1]) (Fig. 2). The interrelationship of the two subdivisions is as follows:

Banks et al. (1970)	Present author
	member XII
	member XI
	member X
Upper Sandstone Member	members IV to IX
Upper Siltstone Member	
Middle Sandstone Member	
Middle Siltstone Member	member III
Lower Sandstone Member	member II
Lower Siltstone Member	member I
base unknown	

The three lowest members of the sequence (I Lower Siltstone member, 20 m; II Lower Sandstone member, 112 m; III Middle Siltstone member, 50 m) crop out along the coast of Varangerfjord between Mortensnes and Vadsø. These have not been studied by the author, but detailed descriptions are to be found in the paper by Banks et al. (1970).

IV. *Grey sandstone member, 40 m.*

Only the upper few metres of the Grey sandstone member is exposed in the south-western part of the investigated area. This member is

1) The estimates are uncertain, mainly due to the inaccuracy of the topographic maps and to the paucity of outcrops.

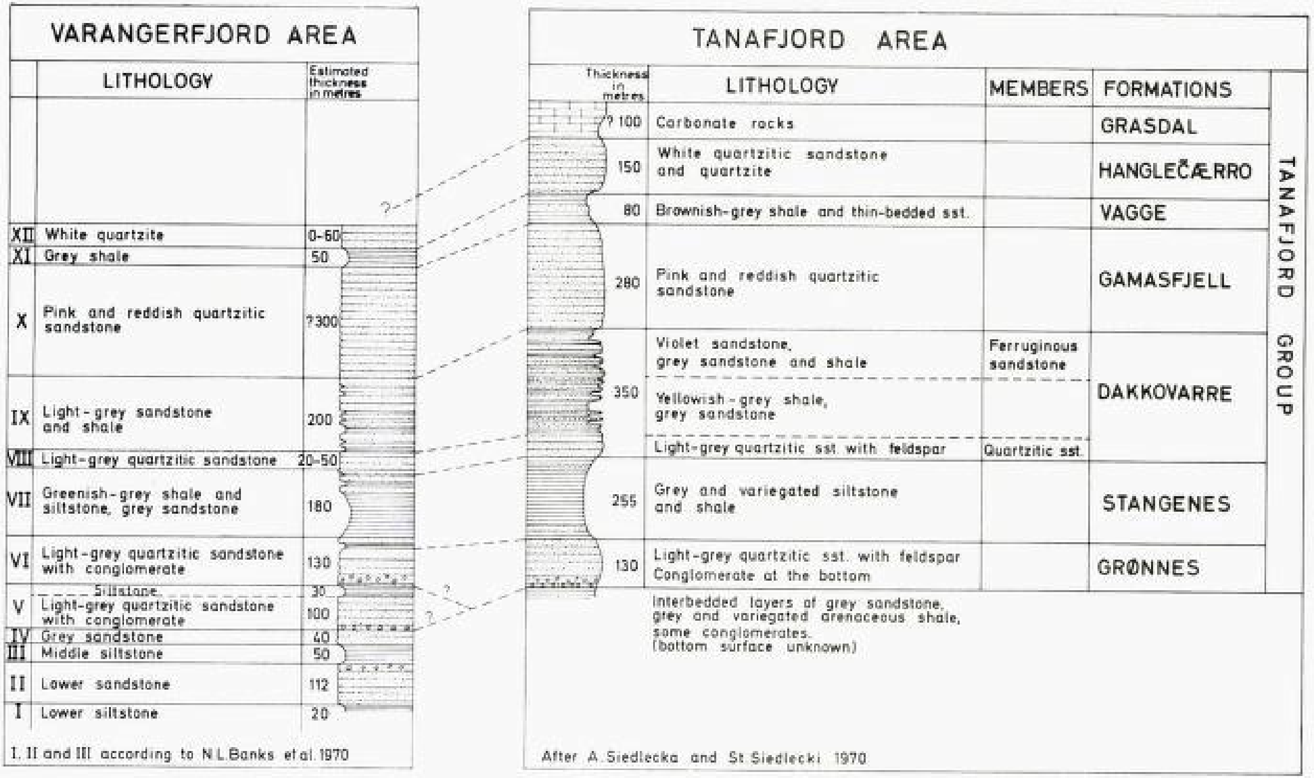

Fig. 2. Lithostratigraphic correlation between the late Precambrian sedimentary rocks of the Varangerfjord and Tanafjord areas.

assumed to correspond to the lower part of the Middle Sandstone Member of Banks et al. (1970).

The lithology, as seen in the limited exposure, is a medium-grained grey to light-grey sandstone. On weathering surfaces the colour varies from yellow to red.

V. *Light-grey quartzitic sandstone member with conglomerate, upper part siltstone, 130 m.*

A continuous section through this member occurs along the coast between Kiby and St. Ekkerø (Fig. 3). The member can be subdivided into two lithologically different parts.

1) The lower c. 100 m consist mainly of medium-grained light-grey quartzitic sandstone which tends to be friable on weathered surfaces. Fine- to coarse-grained (partly conglomeratic) feldspathic sandstone occurs subordinately within this unit. East of Gulneselven (Fig. 3), green mud-clasts measuring up to 50 cm are present in these sandstones. A notable feature of the sandstones are pyrite concretions with diameters of about 1 cm. Near the base of this lower part of the member lenticular layers of quartzite conglomerate appear, these being interbedded with sandstone. The thickness of the layers varies from a few centimetres up to 2 metres. The matrix of the conglomerate varies from quartzitic to feldspathic sandstone. The pebbles are subrounded to rounded up to 4 cm across. A few pebbles of red jasper have been observed. This conglomerate horizon has been traced discontinuously from Melkevarden to Tomaselven and is thought to be an equivalent of the conglomerates mentioned by Holtedahl (1918) from Klubbviken and Klubbfjell (Fig. 1).

2) The upper part of member V, which corresponds to the Upper Siltstone member of Banks et al. (1970), is exposed on St. Ekkerø.

It consists of light-grey siltstones interbedded with fine-grained grey sandstones. These grade upwards into siltstones and medium-grained light-grey sandstones. On weathering surfaces these rocks have a characteristic brown colour. Inland, rocks of the upper unit have been found only as loose fragments on the ground. The unit seems to thin out towards the west, this conclusion being supported by the observation that south of Sildstadhaugen the thickness of the rocks occurring between the more resistant sandstones above and below is about 10 metres.

The Upper Sandstone Member of Banks et al. (1970) has been divided by the present author into four separate members, each of which has a distinctive lithological character and can be mapped from the coast inland towards the west.

VI. *Light-grey quartzitic sandstone member, with conglomerate, 130 m.*

Along the coast the lower part of the light-grey sandstone member crops out on St. Ekkerø, and the upper part at and to the north of Krampenes.

Near the base, above c. 4 m of light-grey sandstone with siltstone partings, there are conglomerate layers which are quite similar to the conglomerates described in member V, except that here no jasper pebbles have been seen. The conglomerates have been traced westwards to Tomaselven, appearing inland mainly as a belt of loose quartz pebbles on the ground. The conglomerates are overlain by light-grey medium-grained quartzitic sandstone which is the predominant lithology of this member.

The upper c. 20 metres of the member consist of fine-grained laminated sandstones, massive sandstones and some shales.

VII. *Greenish-grey shale and siltstone, grey sandstone member. 180 m.*

These rocks have been called «green coloured arenaceous shales» by Holtedahl (1918). Along the coast, good sections through this member appear between Krampenes and Urneset. The lower boundary of the member is drawn above the uppermost massive quartzitic sandstone of member VI.

Sandy and clayey shales with interbedded grey siltstones and sandstones constitute the lower part. The shales are mostly greenish-grey with thin red-violet laminae. In some horizons the red-violet colour is predominant. Upwards, the shales pass gradually into a heterogeneous succession of thick-bedded light-grey sandstones and shales, the sandstones dominating in the upper part.

It should be mentioned here that, in the author's opinion, the Upper Siltstone member of Banks et al. occurring in the upper part of the Jakobselv valley just west of the present area is in fact the equivalent of this member (VII).

VIII. *Light-grey quartzitic sandstone member, (20—50) m.*

The resistant nature of this member of the sequence compared with the rocks above and below facilitates its mapping over large areas. The thickness is variable; from c. 20 m to c. 50 m.

The typical lithology is a light-grey to white quartzitic sandstone. A transition is present between these sandstones and the greenish-grey shales and siltstones with grey and light-grey sandstone of the preceding member, and the lower boundary has been placed below the lowermost observed bed of white quartzite.

IX. *Light-grey sandstone and shale member. 200 m.*

The lower part of this member is best exposed in a river section on the plateau above the valley west of Nattfjellet. The upper part is best studied along the coast and in the area north of Vasavannet.

Lithologically this member is heterogeneous, the rocks comprising light-grey quartzitic sandstones interbedded with brownish-grey laminated sandstones, green shales and light-coloured siltstones. Red thin-bedded sandstones occur interbedded in the upper part of the member, while towards the top massive light-grey sandstones gradually predominate.

A feature common to the sandstone members described so far are the ferruginous spots seen on weathered surfaces. The size of these spots is usually less than 2 mm, although in the Light-grey sandstone and shale member (IX) spots up to 5 mm across have been measured.

X. *Pink and reddish quartzitic sandstone member. 300 m.*

This member of the sequence is the most extensive of all members occurring in the Vadsø—Komagelven area. Because of lack of exposure in the area between Skallneset and Skallelven, only 2 metres of these sandstones can be observed along the road north of Skallneset.

The lithology of this member, in contrast to the underlying rocks, is quite homogeneous. The sandstones, which are medium-grained and massive, are notable for their pink and red coloration. The dark red colour is seen partly as bands paralleling the bedding planes and partly as irregular patches.

A transitional lithology separates this sandstone unit from the underlying member, and the lower boundary has been drawn where the sandstone first exhibits a distinct pink colour.

XI. *Grey shale member, 50 m.*

These rocks appear in the area north of Falkefjellet and in Skallelven south-east of Skallelvskaret (Fig. 3). South-west of Skallelven their extent is uncertain on account of the widespread cover of Quaternary deposits. Only the lowest three metres of these rocks are exposed in the investigated area, but north of Falkefjellet the position of the member can be followed quite easily by observing the disintegrated fragments on the ground surface and by noting its occurrence between two resistant sandstone members.

The main lithology is a greyish-green silty shale, which turns brown on weathering. The shales display both parallel and cross lamination. In the lowest three metres intercalations of light-grey thin-bedded sandstones and siltstones are present.

XII. *White quartzite member (0–60) m.*

This is the youngest member of the Older Sandstone Series in the investigated area. Its outcrop extent is restricted to the area south-east of Kjøltindene; it is found to wedge out towards the south-west.

The characteristic lithology is a medium-grained massive quartzite. Although the colour is mostly white, some quartzite horizons are observed to be reddish-grey.

Vestertana Group.

In the Vadsø–Komagelven area a part of the Vestertana Group unconformably overlies the Older Sandstone Series. In the north, the sequence rests on the White quartzite member (XII), in the south on the Grey shale member (XI). North of Skallelvskaret the Older Sandstone Series is overlain by a c. 2 m thick sedimentary breccia with angular quartzite fragments up to 10 cm across in a ferruginous sandy matrix. The quartzite fragments are obviously derived from the White quartzite member (XII); an erosional unconformity is thus indicated. The breccia is interpreted as being the basal horizon of the Vestertana Group (at this particular locality), a formation of which – the Stappogiedde Formation (Reading 1965) – crops out nearby.

In the present area the Vestertana Group consists of the Upper Tillite Formation and the lower and middle parts of the Stappogiedde Formation (Quartzitic sandstone member and the Blue-green and red-violet shale member, Reading 1965). The Lower Tillite Formation and the

Nyborg Formation are absent in this area. The tillite, thought to be the Upper Tillite (Siedlecki, personal communication), is exposed in Skallelven south-east of Skallelvskaret. The lateral extent of the tillite towards the south-west is unknown, due to lack of exposure. To the east, a tectonically strongly disturbed part of the Upper Tillite crops out north of Holmfjellet (Siedlecki, personal communication). North of Falkefjellet, the Stappogiedde Formation rests directly on the Older Sandstone Series, except at the locality north of Skallelvskaret where the thin basal breccia is present.

Correlation with the Tanafjord area.

Problems concerning the relationship between the Older Sandstone Series of the Tanafjord and the Varangerfjord areas have been discussed by Holtedahl (1918, 1932), Rosendahl (1931, 1945) and Bjørlykke (1967). Rosendahl (1945) correlated the sandstones and shales in the Mortensnes—Klubbfjell—Vadsø area (Fig. 1) with the sequence underlying the «Vagge quartzite» (Føyn 1937) of the «Older Sandstone Series on the Tanafjord». This view was supported by Bjørlykke (1967) who suggested that about 500 metres of the sequence appearing in the Tanafjord area had been eroded (or perhaps not deposited) before the deposition of the tillites in the Varangerfjord area.

Field work by the present author has indicated that Føyn's units from the Tanafjord area, with the exception of the upper carbonate rock unit, also occur north of Varangerfjord with little or no change in lithology or thickness. Furthermore, to the west of the Vadsø area, the tillites rest on rocks older than those described by Føyn (1937), and possibly on the oldest unit of this author, from the Tanafjord area.

Siedlecka & Siedlecki (1970) have proposed a formal lithostratigraphical classification for Føyn's «Older Sandstone Series on the Tanafjord», based on detailed studies of these units in the Vagge and Trollfjord Profiles (Fig. 1), and on the tracing of these units inland to the Kjøltindene—Grytehaugen areas. The complete succession has been named the Tanafjord Group and is divided into seven formations. The base of the Group is drawn below the Grønnes Formation (Fig. 2, also Siedlecka & Siedlecki, 1970) which is correlated with Føyn's quartzitic sandstone (zone d) from the Tananes section. In the description which follows, Siedlecka & Siedlecki's (1970) nomenclature has been adopted.

The middle and upper parts of the Older Sandstone Series of the

Varangerfjord area show great similarities, both stratigraphical and lithological, with the Tanafjord Group (excluding the Grasdal Formation, Fig. 2). A feature common to the two successions is the thick sequence of light-coloured quartzitic, partly ferruginous sandstones, interbedded with shales and siltstones. Whereas shales and siltstones are common in the lower part, sandstones tend to dominate higher up.

The placing of boundaries within the Varangerfjord succession has been influenced, to some extent, by the difficulties encountered during mapping in the peninsula interior, where generally only the sandstones protrude sporadically through the superficial cover.

For this reason, and because the thicknesses mentioned are uncertain estimates only, it is considered premature to draw any general conclusions concerning the apparent differences in thickness between the informal members of the Varangerfjord succession and the formations of the Tanafjord Group.

The youngest formation known within the Tanafjord Group – the Grasdal Formation – is not present in the Vadsø–Komagelven area. Here, the Vestertana Group rests unconformably on the members XII (White quartzite member) and XI (Grey shale member).

XII. *White quartzite member.* During mapping in the Kjøltindene –Grytehaugen area in 1968, Siedlecka & Siedlecki established that the White quartzite member could be correlated with the 150 m thick Hanglecærro Formation (Fig. 2). In the area investigated by the author the thickness is about 50 m, thinning out towards the south. South-east of Skallelvskaret this member is absent.

The apparent lack of facies change between this member of the Varangerfjord succession and the equivalent formation in the Tanafjord sequence, and the occurrence of angular fragments of white quartzite in the overlying breccia (north of Skallelvskaret) indicate that the thinning out of the White quartzite member was the result of erosion.

XI. *Grey shale member.* This member is correlated with the Vagge Formation (Vagge shale of Føyn, 1937) which consists of shales and thin-bedded sandstones. The correlation is based on the stratigraphic position of these sediments between two characteristic sandstone members. Detailed lithological differences or similarities between this member and the Vagge Formation have not been observed on account of the poverty of exposure.

X. *Pink and reddish quartzitic sandstone member.* The characteristic feature of this member is its colour and homogeneity. This, mostly pink, medium-grained sandstone is lithologically identical to the Gamasfjell Formation and must be regarded as a marker horizon for the present correlation.

IX. *Light-grey sandstone and shale member.* Within the Dakkovarre Formation two informal members have been distinguished by Siedlecka & Siedlecki (1970). The lower Quartzitic sandstone member is a homogeneous quartzitic sandstone containing feldspar, the colour being mainly light grey and white. The middle part of the formation consists of interbedded sandstones and shales, present in nearly equal amounts. Red-violet sandstones are characteristic for the upper Ferruginous sandstone member. These are interbedded with light-grey feldspathic and quartzitic sandstones with brown ferruginous spots, and shales.

The Light grey sandstone and shale member in the Vadsø—Komagelven area underlies the Pink and reddish quartzitic sandstone (member X). The similar lithology and the gradual transition in colour between the upper part of the member IX and member X is taken to indicate a gradual and slow change of sedimentary environment.

The correlation between member IX and the middle and upper part of the Dakkovarre Formation is based on lithological similarities and on the continuity of deposition between members IX and X evidenced by the transition noted above. The heterogeneity of the rocks, with the cyclic alternation of sandstones and shales, is a feature common to the correlated strata. Sandstones seem to be more dominant in the Light-grey sandstone and shale member than in the Dakkovarre Formation.

Red-violet colours, like the ones prevailing in the upper Ferruginous sandstone member of the Dakkovarre Formation, have not been observed in the present area. However, the Light-grey sandstone and shale member (IX) in the Vadsø—Komagelven area is ferruginous as shown by the occurrence of brownish-black spots in the sandstones and by the occasional appearance of a rusty red coloration, presumably of secondary origin.

VIII. *Light-grey quartzitic sandstone member.* The Light-grey quartzitic sandstone member (VIII) and the upper few tens of metres of light-grey sandstone of member VII may be correlated with the Quartzitic sandstone member of the Dakkovarre formation.

A distinguishing feature common to the two correlated sequences is their marked resistance to weathering as shown by conspicuous white sandstone ridges.

VII. *Greenish-grey shale and siltstone, grey sandstone member.* The uppermost c. 20 m of member VI and the Greenish-grey shale and siltstone, grey sandstone member, excluding a few tens of metres at the top (cf. preceding paragraph), is correlated with the Stangenes Formation.

The main lithology of this rock assemblage, and of the Stangenes Formation, is partly clayey and partly sandy shale, usually greenish-grey and containing thin red laminae. Siltstones and thin-bedded sandstones occur interbedded. In the present area sandstones are more frequent in the upper part as compared with the Stangenes Formation.

VI. *Light-grey quartzitic sandstone member with conglomerate.*

V. *Light-grey quartzitic sandstone member with conglomerate, upper part siltstone.* Because of lithological similarities between these two members and the absence of detailed description of the several hundred metres thick rock sequence underlying the Tanafjord Group east of Lille Molvik (Fig. 1, also see Siedlecka & Siedlecki 1970), it is uncertain whether the Grønnes Formation should be correlated only with member VI (excluding the uppermost 20 metres) or whether the correlation should also include member V.

The Grønnes Formation consists of light-grey to white fine- to coarse-grained quartzitic sandstone, with distinctive quartzite conglomerate layers at the base. Occasionally jasper pebbles may be found in the conglomerate. No silty beds have been described from the Grønnes Formation.

Points favouring a correlation between the Grønnes Formation and member VI only may be summarised as follows:

1) The only conglomerate horizons described from the Grønnes Formation occur at or near its base.
2) Concretions of pyrite occur in member V of the present area, and also in the rocks underlying the Grønnes Formation in the Tanafjord area (Siedlecki, personal communication); no such concretions have been found either in member VI or within the Grønnes Formation.
3) The siltstone unit occurring in the upper part of member V has no equivalent within the Grønnes Formation.
4) The estimated thickness of member VI corresponds well with that of the correlated formation in Tanafjord.

Arguments in favour of a correlation of the Grønnes Formation with both the members VI and V may be outlined as follows: –

1) No thick units of light-grey quartzitic sandstone are mentioned in the brief descriptions of lithologies occurring below the Tanafjord Group (Siedlecka & Siedlecki, 1970).
2) The heterogeneity of grain size seen within the Grønnes Formation is also a characteristic feature of member V.
   The quartzitic sandstones of member VI are rather homogeneous from a textural point of view.
3) A few jasper pebbles are seen in the conglomerates near the base of member V, but have not been found in the conglomerates of member VI.
4) The siltstone unit in the upper part of member V seems to thin out towards the west, a feature suggestive of non-deposition of this lithology in the Tanafjord area.

On the basis of this comparison of part of the Varangerfjord sequence with the Tanafjord Group («The Older Sandstone Series on the Tanafjord», Føyn 1937) it is proposed that the present members XII to VI (± V) be assigned to the Tanafjord Group.

If the correlation outlined above is valid, there would appear to be certain minor facies differences between corresponding units within the middle and lower parts of the two sequences. Member IX and the upper part of member VII, for example, are more arenaceous than their Tanafjord equivalents. Members IX and VII also seem to be thinner.

Whether or not there are significant facies differences between the Grønnes Formation and members VI ± V of the present area must remain an open question until a more detailed description of the succession below the Grønnes Formation in the Tanafjord area is forthcoming. Only then will it be possible to decide whether or not member V should be included in the Tanafjord Group.

West of Vadsø, in the Mortensnes area, the tillite unconformably overlies the Older Sandstone Series (Holtedahl 1918); towards the east it lies on progressively younger members of the sequence. At Mortensnes the Lower Tillite (the Smalfjord Tillite of Bjørlykke et al., 1967) rests directly on the Lower Siltstone member (Reading, NGU report, 1969), which is the oldest known member of the Older Sandstone Series on the northern side of Varangerfjord. Thus, 1300 m of this sequence is absent in the Mortensnes area. In Skallelven, the Upper Tillite overlies the Grey Shale member XI; from the observed stratigraphic relationships of the tillite to the subjacent rocks at these two localities, the angular unconformity can be calculated at 1–2 degrees.

From his studies in the Tanafjord area, Føyn (1937) showed that to the south and south-west the tillites overlie progressively older formations of the Tanafjord Group. The lowest unit immediately underlying the tillite in this area is the upper part of the Dakkovarre Formation (zone m, Føyn), at the head of Tanafjord. The maximum estimated angular unconformity in the Tanafjord area is 2 degrees (Føyn 1937).

Assuming an erosional explanation for the unconformity established in these areas (Holtedahl 1918, Føyn 1937, Bjørlykke 1967, and the present paper), the stratigraphic relationships of the tillites in the Tanafjord and Varangerfjord areas would suggest that uplift occurred in and to the south-south-west of the area underlain by the Older Sandstone Series prior to the deposition of the tillite.

Acknowledgements.

The writer is deeply grateful to Drs. Anna Siedlecka and Stanislaw Siedlecki for introducing her to the problems on Varangerhalvøya, and for subsequent encouragement and many helpful discussions. Norges geologiske undersøkelse, of whose mapping programme on Varangerhalvøya this work forms a part, has kindly provided field equipment and defrayed the field expenses. Various members of staff at Norges geologiske undersøkelse are thanked for their assistance in the preparation of this paper. In particular, the writer is indebted to Statsgeologene Tore Torske and Dr. David Roberts for their constructive criticism and correction of the English text. The writer also wishes to express her gratitude to Hr. Bjørn Mathisen of Vadsø for his generosity, hospitality and assistance during the course of the fieldwork.

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GEOLOGICAL MAP OF THE

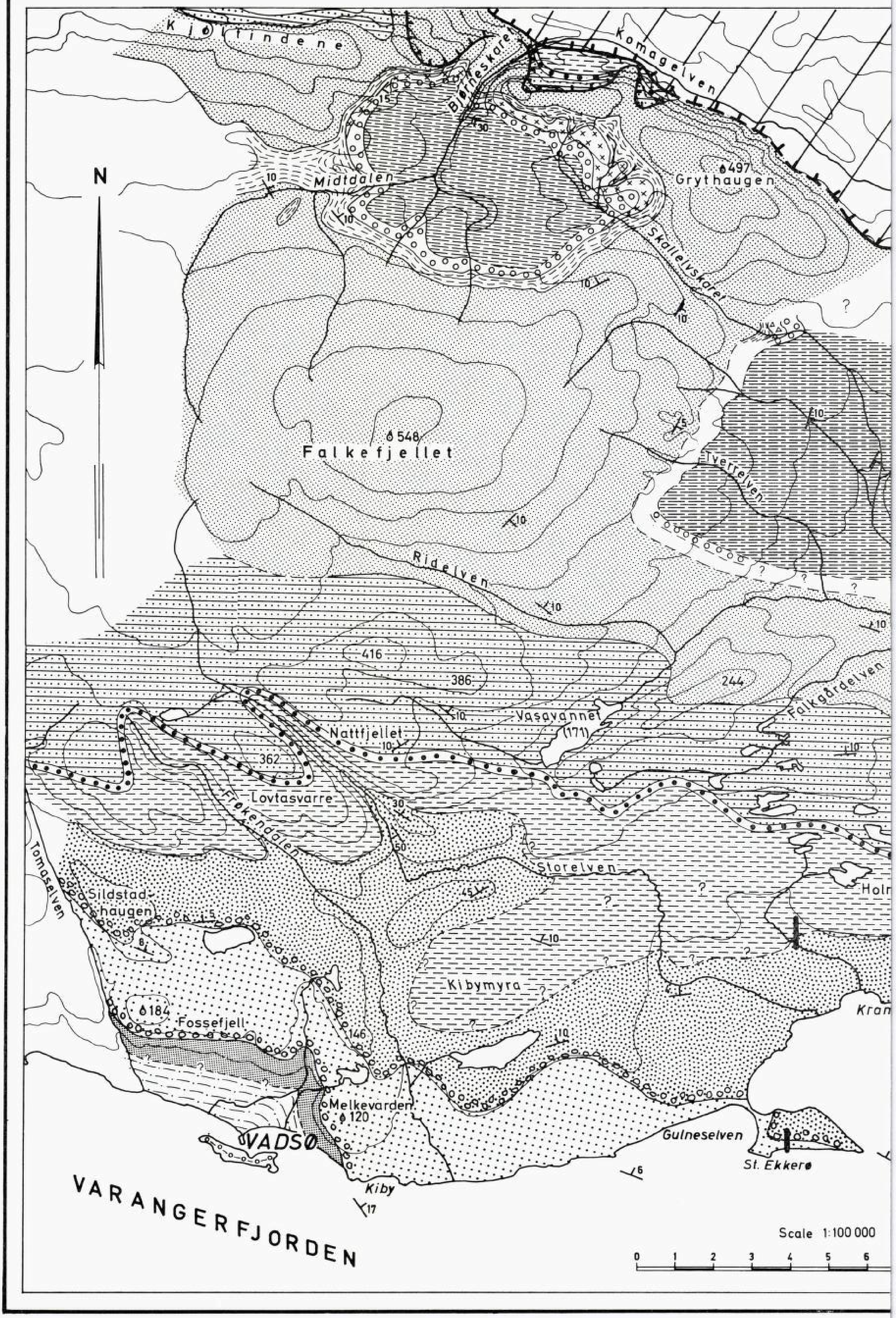
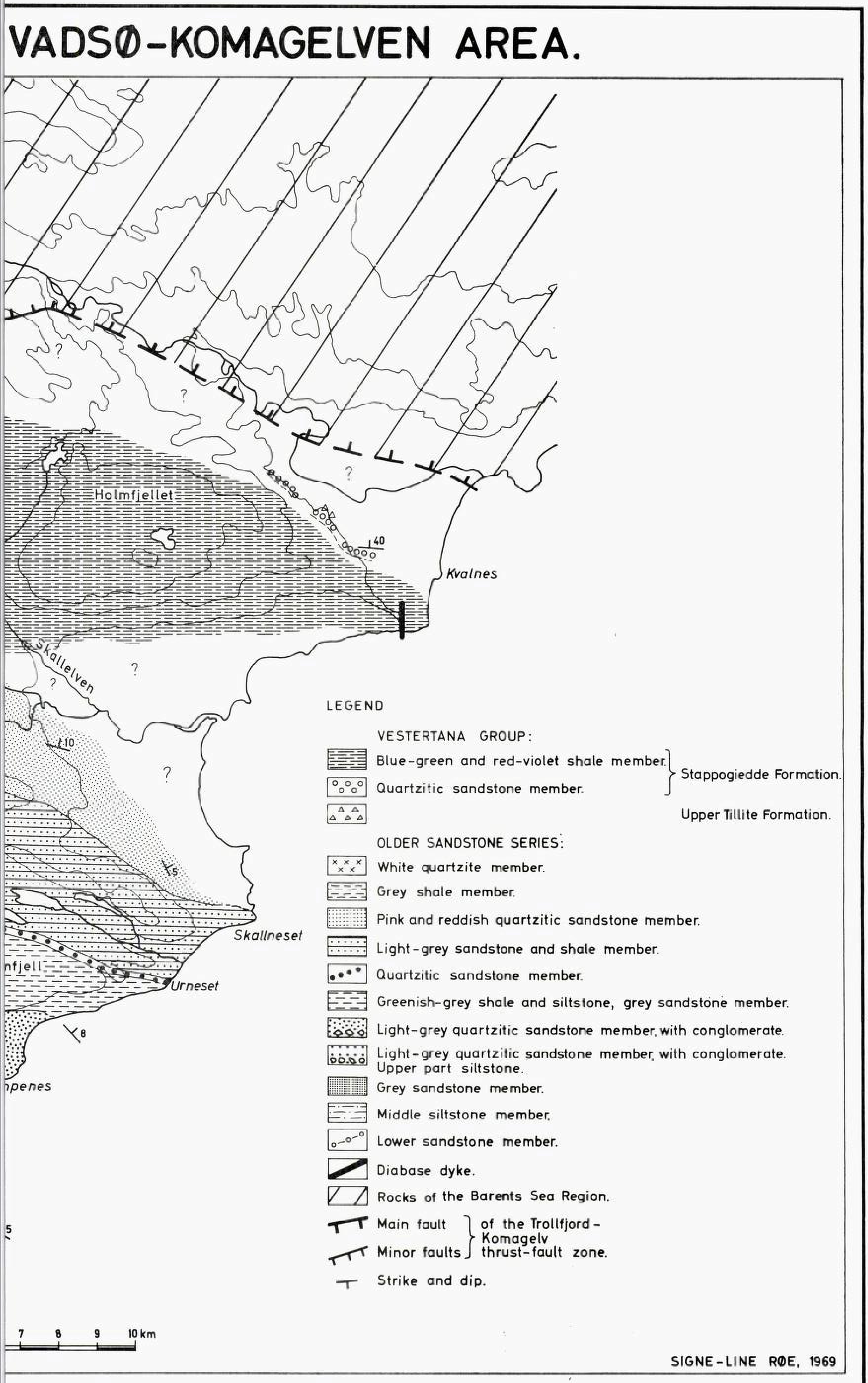


Fig. 3.


VADSØ-KOMAGELVEN AREA.
Holmfjellet
Kvalnes
Skallelven
Skallneset
Urneset
nfjell
penes
LEGEND
VESTERTANA GROUP:
Blue-green and red-violet shale member.
Quartzitic sandstone member.
Stappogiedde Formation.
Upper Tillite Formation.
OLDER SANDSTONE SERIES:
White quartzite member.
Grey shale member.
Pink and reddish quartzitic sandstone member.
Light-grey sandstone and shale member.
Quartzitic sandstone member.
Greenish-grey shale and siltstone, grey sandstone member.
Light-grey quartzitic sandstone member, with conglomerate.
Light-grey quartzitic sandstone member, with conglomerate. Upper part siltstone.
Grey sandstone member.
Middle siltstone member.
Lower sandstone member.
Diabase dyke.
Rocks of the Barents Sea Region.
Main fault
Minor faults
of the Trollfjord-Komagelv thrust-fault zone.
Strike and dip.
7 8 9 10 km
SIGNE-LINE RØE, 1969

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Manuscript received in April 1970.

EINERGREIN AV PREBOREAL ALDER FUNNET I ISRANDAVSETNING I EIDFJORD, VEST-NORGE

av

Noralf Rye[1])

Abstract.

In 1969 a branch of juniper *(Juniperus communis)* was found in a sand-pit on the ice marginal delta in Eidfjord, West-Norway. Radiocarbon dating gave an age of 9680 ± 90 years B. P., which means that the delta is of Preboreal age or younger.

Den store israndavsetningen i Eidfjord, Hardanger (Fig. 1 og 2), er oppbygd i en periode da isfronten lå ved nedre ende av Eidfjordvatnets overfordypede bekken. Store mengder materiale som breelvene førte med seg ble akkumulert i havet foran breens front. Etter hvert bygde det seg opp et isranddelta som overveiende består av glasifluvial sand og grus.

Eidfjord-avsetningen består nå av flere terrasser. Den mektigste ligger på Hæreid på dalens NØ-side (jfr. fig. 2). Denne terrassen er bredest i sin proksimale del, og fyller her mesteparten av dalbunnen. Framover mot havet smalner terrassen noe av. Overflaten skråner mer eller mindre jevnt utover mot elva, fra ca. 112 m o.h. i nærheten av fjellsiden i terrassens proksimale parti, til 100—102 m o.h. ved kanten av rasskråningen. Det er også et mindre fall i overflatens høyde distalt. På motsatt dalside fins det en rekke lavere terrassenivåer, fra ca. 90 m o.h. og ned til ca. 7 m o.h. Disse tolkes som erosjonsflater, og dermed regnes terrassene som rester av glasifluviale masser som fylte dalen til Hæreidterrassens nivå.

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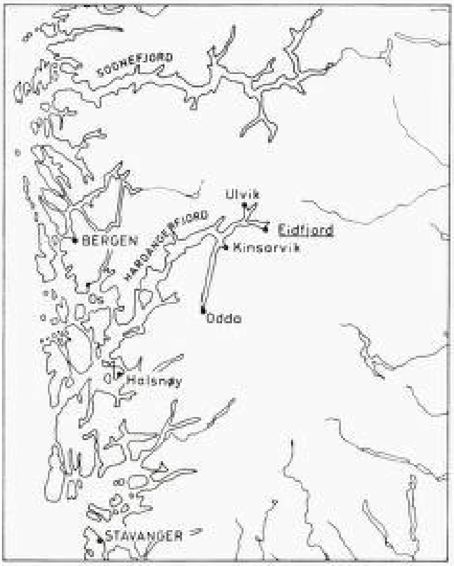

Fig. 1. Lokalitetskart.
Location map.

Ved A/S Hardanger Cementvarefabrikks sandtak (Fig. 3, jfr. også fig. 2), ble det høsten 1969 funnet ei ca. 1 m lang grein med største diameter 7–8 cm (Fig. 4). Cand real. Dagfinn Moe ved Botanisk Museum, Universitetet i Bergen, har bestemt greina til å være einer (*Juniperus communis*). Denne er C-14 datert (T - 886) ved Laboratoriet for Radiologisk Datering, Norges Tekniske Høgskole, og alderen er oppgitt til 9680 ± 90 år. B. P. (1950). Einerens alder tilsvarer dermed relativt sen Preboreal tid, dersom periodens varighet regnes fra 8300–7500 B. C. Sandtaket der greina ble funnet ligger altså i en erosjonsterrasse. Overflaten er ca. 23 m o.h., mens greina befant seg ca. 5 m o.h. dvs. omlag 18 m nede i terrassen (Fig. 3). Strukturelt er det her utpreget foreset-lagning som faller i dalens lengderetning (mot NV). Snittet som er avbildet på fig. 3 ligger omlag på tvers av dalen og viser temmelig nær horisontal lagning, men med et mindre fall fra elva. Greina lå parallelt med lagene, og befant seg ved spaden på bildet. Da greina ble funnet var den hel, men den ble knekt av gravemaskinen. Som det går fram av fig. 3 og 5 er greina meget godt oppbevart, og det skyldes hovedsakelig at den i hele sin lengde var innkapslet i ei hylse av sand og grus som var sammenkittet av jernoksyd. I de mørke lagene

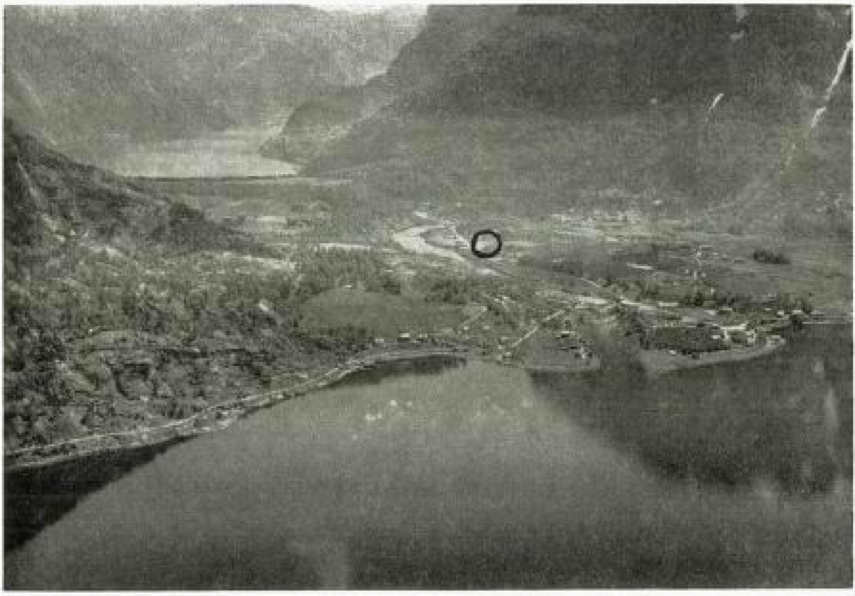

Fig. 2. Israndavsetningen i Eidfjord sett mot SØ. Det oppdemte Eidfjord-vatnet (19 m o.h.) i bakgrunnen. Den høyeste terrassen, Hæreid, ligger rett foran vatnet. Omlag midt på bildet befinner det sandtaket seg (innsirklet) der greina ble funnet. (Foto: Widerøe).

The ice marginal delta in Eidfjord, seen from NW. The Eidfjord lake in the background. In the center of the picture (circle) lies the sand-pit where the branch was found (Photo: Widerøe).

ved spaden på fig. 3 er sand- og gruskornene dekket av et rødbrunt belegg, og det var i et slikt lag den innkapsla greina befant seg. Kornfordelingsanalyse av prøve fra dette laget viser Md = 1,95 mm og So = log Q_{75}/Q_{25} = 0,40. Belegget ble fullstendig løst ved syrebehandling og sammensetningen går fram av følgende tabell, prøve nr. 1. I tabellen er også tatt med resultatet av tilsvarende analyse av selve hylsa som greina var innkapslet i, prøve nr. 2. Denne prøven viser relativt stor konsentrasjon av Fe_2O_3. Analysene er utført ved Kjemisk laboratorium, Geologisk avdeling, NGU.

Prøve	Belegget i %	Beleggets sammensetning i %				
nr.	av prøven	Fe_2O_3	CaO	MgO	MnO	CO_2
1	2,3	63,8	13,4	21,9	0,9	—
2	25,9	98,2	spor	0,6	0,3	0,9

Fig. 3. Sandtaket der greina ble funnet.
The sand-pit where the branch was found.

Sedimentasjonen av materiale der greina befant seg må betraktes som et ledd i oppbyggingen av israndavsetningen, dvs. primæravsetningen. Einerens alder angir derfor maksimumsalderen på israndavsetningen. Deltaet var bare omlag halvveis utbygd da greina ble sedimentert, men vi vet ikke hvor lang tid som gikk med til hele oppbyggingen. Bare avsetningens proksimale del kan være noe eldre enn eineren. Likevel må vi anta at israndavsetningen i Eidfjord ikke er eldre enn fra relativt sen Preboreal tid. Dette er i samsvar med Anundsens (Anundsen og Simonsen, 1967) konklusjon når det gjelder alderen på Eidfjord-avsetningen. Dateringen strider heller ikke mot den oppfatning at Ra-framstøtet har gått over Bergen—Os området (Holtedahl, 1964), og at dette framstøtet er representert av endemorener så langt ute i Hardangerfjorden som f.eks. på Halsnøy (Holtedahl, 1967 og Undås, 1963).

Funnet av trerester i Eidfjord-avsetningen *betinger* ikke noe breframstøt, iallfall ikke noe betydelig framstøt. Greina kan tenkes å ha drevet inn fra sjøen, selv om det må ha vært relativt sterk strøm utover der greina ble begravd. Det kan videre være mulighet for at den har

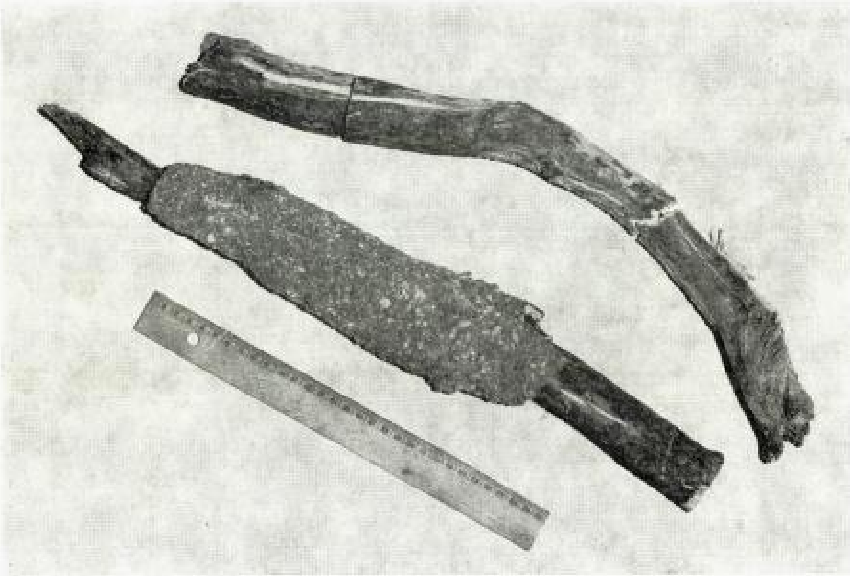

Fig. 4. Einer-greina fra Eidfjord. Bildet viser også litt av den hylsa av grus og sand som greina var innkapslet i.
The branch of juniper. The branch was capsuled by a covering crust of gravel and sand, cemented by iron-oxide, etc.

rast ned fra dalsiden, eller den kan ha blitt ført med isen ved et mindre framstøt. Det fins mange eksempler på vegetasjon tett inn til isfronten, også av einer som klimatisk sett regnes som et forholdsvis indifferent treslag. Dette eksemplaret er imidlertid så kraftig og har så regelmessige årringer at det kanskje er tvilsomt om det kan ha vokst med isen i umiddelbar nærhet.

Anundsen og Simonsen korrelerer Eidfjord-avsetningen med randmorener i fjellområdene omkring, og setter den således i sammenheng med et generelt breframstøt i disse trakter. Havnivået under breframstøtet var etter deres framstilling ca. 120 m over nåværende havstand. Hovedbegrunnelsen for dette synes å være at det i Ulvik (jfr. fig. 1) er marine terrasser over 120 m o.h. Etter mitt skjønn er det lite rimelig med et så høyt havnivå da Eidfjord-avsetningen ble oppbygd. Når et isranddelta av slike dimensjoner (1,5 km langt) og med slik materialsammensetning bygges ut, må det i sitt proksimale parti nå omlag til havnivå før det bygges så langt utover. Dersom havstanden under oppbyggingen har vært ca. 120 m høyere enn nå, må det bety at den høyeste

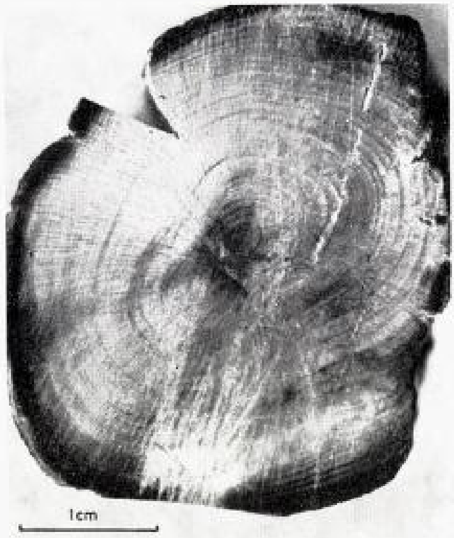

Fig. 5. Tverrsnitt av greina. Før den tørket var største tverrsnitt 7—8 cm. Årringene viser en alder på ca. 45 år.
The branch in section. The yearly rings shows an age of ca. 45 years.

terrassen (Hæreid-terrassen) senere er blitt kraftig erodert i sitt proksimale parti. Noen tegn på en slik utvikling er ikke funnet. Tvertimot ser det ut til at avsetningen delvis er bygd opp over havets overflate som et sandurdelta, og at den marine grense ikke kan være vesentlig høyere enn 100 m o.h. Dette skulle da omlag tilsvare havnivået i Eidfjord for maksimalt 9680 ± 90 år B.P. De høytliggende terrassene i Ulvik må etter dette være eldre enn israndavsetningen i Eidfjord.

Når det gjelder andre større israndavsetninger ved dalmunninger i indre Hardanger, f.eks. i Kinsarvik og Odda, vet vi foreløpig ikke om noen av disse i tid kan korreleres med Eidfjord-avsetningen. Nærmere kjennskap til marine grenser og isobasenes forløp kan gi oss sikrere holdepunkter her.

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Manuskriptet mottatt i april 1970.

UNDERSØKELSER AV STEINFRAKSJONENS RUNDINGSGRAD I GLASIGENE JORDARTER

av

Ole Fredrik Bergersen[1])

Abstract.

On the degree of roundness of the stone fraction in glacigene deposits.

The main bulk of the stone material in Norway has undergone only one or a few transport phases. A classification has been made for the deposits in the interior of Eastern Norway on the basis of the degree of roundness of pebbles (2—6 cm), and the petrographical composition. The degree of roundness has been found to be dependent on mode and length of transport.

Ordinary till contains less than 10 % rounded particles. Glaciofluvial material contains less than 20 % of rounded material after a supra- or en-glacial transport. Subglacial and normal fluvial transport rapidly increases the degree of roundness. In these deposits more than 50 % of the particles are rounded.

Innledning.

Rundingsgraden av forskjellige kornfraksjoner i kvartæravleiringer er lite undersøkt. Rundingsanalyser er derfor en lite brukt metode i kvartærgeologiske undersøkelser.

Jeg har foretatt en del rundingsanalyser på fraksjonen småstein (mellomakse 2—6 cm) i morene-, glasifluviale- og fluviale avsetninger i Gausdals- og Gudbrandsdalsområdet (fig. 1). Området ligger hovedsakelig innenfor sparagmittformasjonene, og sedimentære bergarter, særlig sandsteiner og kvartsitter, dominerer i steinfraksjonen. Gabbroide, kaledonske skyvedekkebergarter, her kalt jotunbergarter, opptrer også

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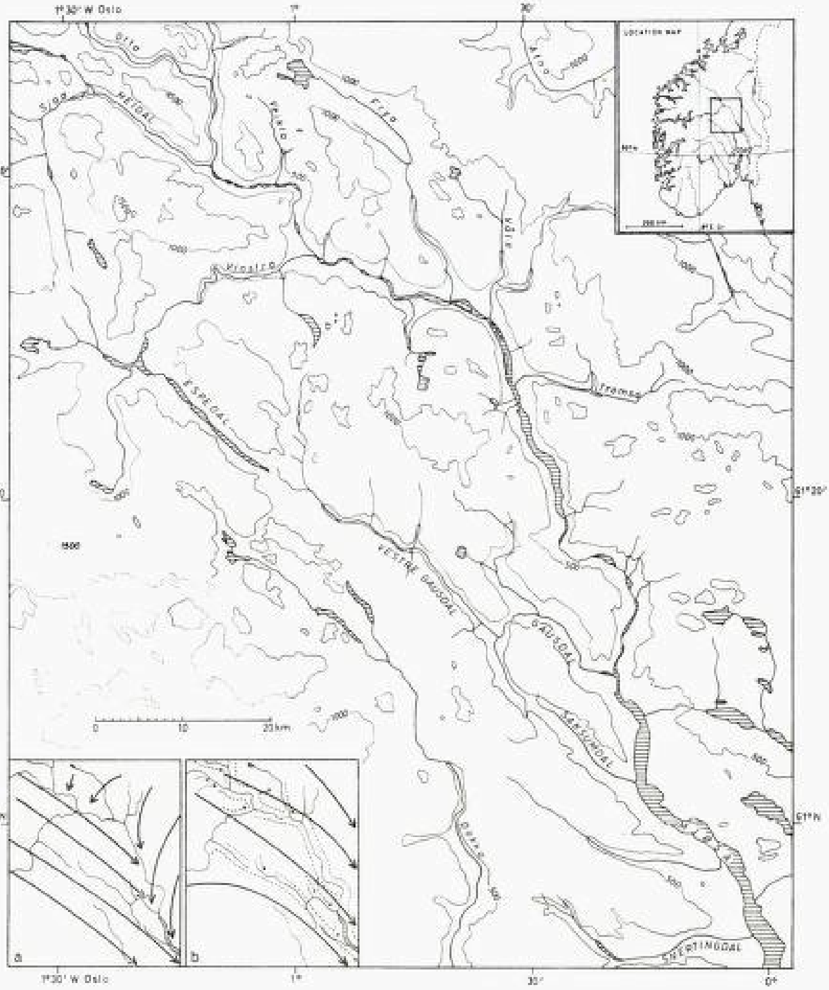


Fig. 1. Oversiktskart med hoveddrag av topografi og isbevegelser innen det undersøkte området. Isbevegelsene er rekonstruert i samarbeid med cand. mag. Kari Garnes.

a) Eldre isbevegelser.
b) Yngre og yngste isbevegelser (prikket).

Key map with main topographic features and ice flow directions.
The ise flow directions have been reconstructed in cooperation with cand. mag. Kari Garnes.

a) Older ice.
b) Younger and youngest ice (dots).

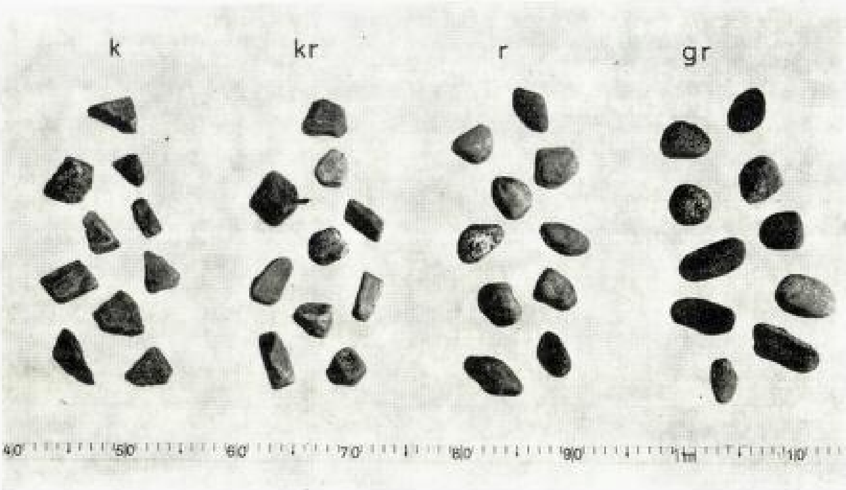


Fig. 2. Standardsett for rundingsklasser: kantet (k), kantrundet (kr), rundet (r), godt rundet (gr).

Kantet: Mer enn halvparten av hjørner og kanter er skarpe, uregelmessig overflate.
Kantrundet: Mer enn halvparten av kantene er slitte, men fortsatt tydelige.
Rundet: Kantene sees bare delvis. Steinen er konveks, med ovalt eller rundt omriss i minst ett plan. Glatt overflate, men ikke uten uregelmessigheter.
Godt rundet: Glatt overflate. Steinen er regelmessig konveks med tydelig ovalt eller rundt omriss i minst to plan.

Standard samples of the four classes of roundness: angular (k), abraded angles (kr), rounded (r), well rounded (gr).

Angular: more than the half of all points and edges are sharp, the surfaces are uneven.

Abraded angles: more than the half of all points and edges are worn, but still well defined.

Rounded: Few edges are well defined. The particle is convex with oval or circular silhouette along at least one axis. Smooth surface, but not without irregularity.
Well rounded: Smooth surface. The particle is clearly convex with oval or circular silhouette along at least two axes.

i betydelig antall. Disse har spilt en stor rolle for undersøkelsene fordi de er lette å skille fra lavmetamorfe sedimentære bergarter.

Jotunbergartene er tilført undersøkelsesområdet fra Jotunheimen. Transporten, som har vært glasial og/eller fluvial, har helt overveiende fulgt hoveddalførene (fig. 1).

Til analysene er brukt en meget enkel metodikk: Stein er inndelt i 4 klasser etter visuelle kriterier (kantet, kantrundet, rundet og godt rundet) (fig. 2).

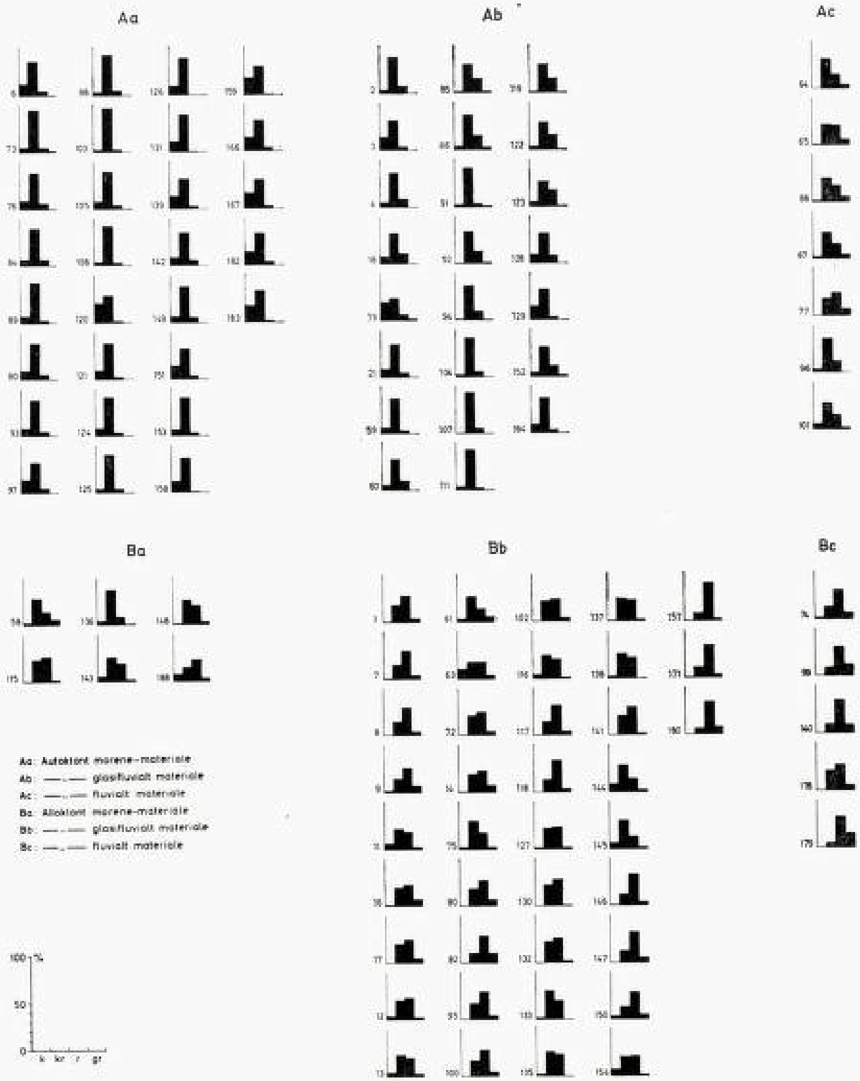
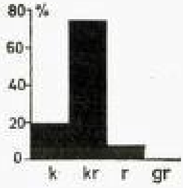


Fig. 3. Morfogrammer klassifisert etter materialets sammensetning og genesis.
Morphograms classified according to composition and origin of material.

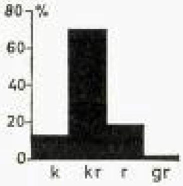
Metodikken er stort sett den samme som Reichelt (1961) har anvendt i Tyskland, og som forfatteren også tidligere har benyttet (Bergersen, 1964).

Til de fleste analyser er det plukket ut 100 stein tilfeldig.



Aa : MORENEMATERIALE

Bunn- og ablasjonsmorener på vidder, i sidedaler, og delvis i hoveddaler.

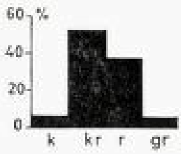


Ab : GLASIFLUVIALT MATERIALE

Glasifluviale avleiringer utenom avleiringer langs hoveddaler. Vesentlig supra- og englasial vanntransport.

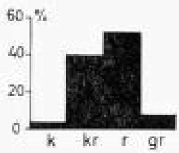
Ac : FLUVIALT MATERIALE

Resente fluviale vifter ut for munningen av sidedaler.



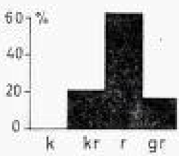
Ba : MORENEMATERIALE

Bunnmorener langs hoveddaler. Sammensatt materiale.



Bb : GLASIFLUVIALT MATERIALE

Avleiringer etter drenering langs dalene. Subglacial eller subaeril vanntransport.



Bc : FLUVIALT MATERIALE

Resente fluviale avleiringer langs hoveddaler.

Fig. 4. Gjennomsnittsmorfogrammer for forskjellige jordarter, klassifisert etter sammensetning og genesis.

Average morphograms for different types of material, classified according to composition and origin.

Det viser seg at løsavleiringer innen det undersøkte området, særlig morenematerialet, på grunnlag av steintellinger og rundingsanalyser naturlig kan to-deles:

a) Avleiringer med nær utelukkende korttransportert steinmateriale. Disse er kalt *autoktone* avleiringer.
b) Avleiringer med et betydelig innhold av langtransportert stein. Disse er kalt *alloktone* avleiringer.

Figur 3 er en sammenstilling av rundingsanalyser fra forskjellige jordarter, tegnet som morfogrammer. Klassifikasjonen av jordartene er altså foretatt, først og fremst, etter innholdet av langtransportert stein, det vil si på grunnlag av steintellinger. Som praktisk mål på dette er særlig benyttet innholdet av jotunbergarter.

Autoktone- og alloktone avleiringer er videre inndelt etter genesis i morene-, glasifluviale- og fluviale avleiringer.

Morenemateriale.

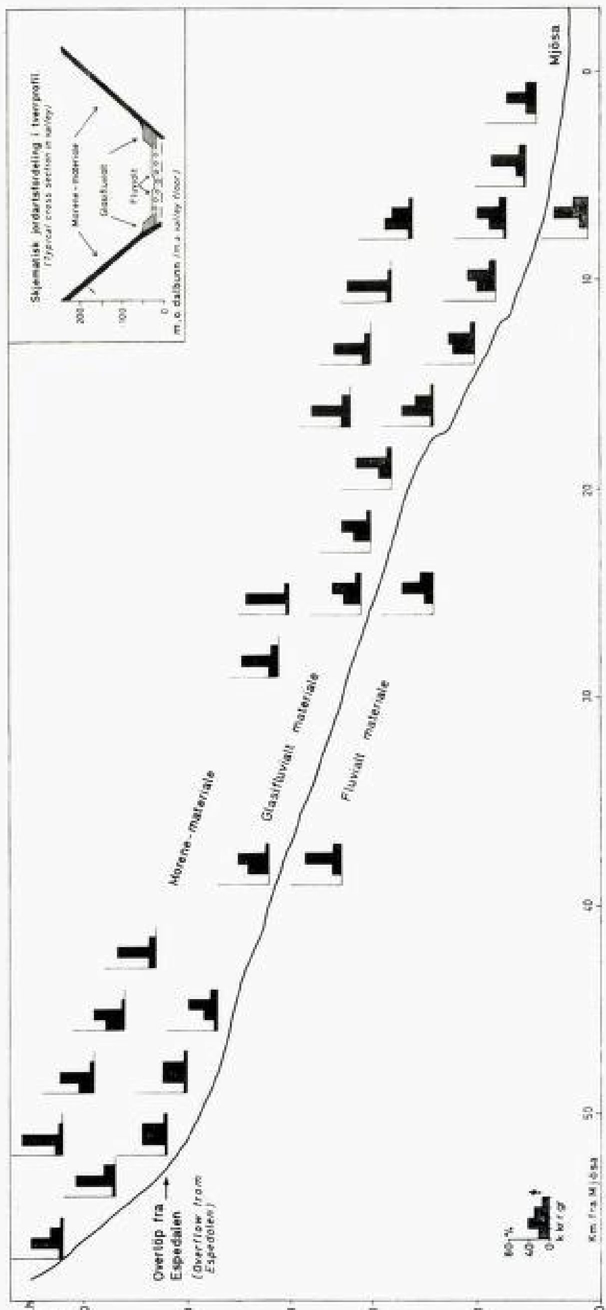
Morenetype Aa domineres av kantrundet materiale. Rundet stein utgjør under 10 % av antallet. Typen representerer det alminnelige morenemateriale innenfor det undersøkte området, og omfatter både bunn- og ablasjonsmorener.

Den andre typen morenemateriale, Ba, har et sammensatt morfogram med alle rundingsklasser representert. Rundet steinmateriale utgjør ca. 40 % i gjennomsnitt (fig. 3 og 4). Dette morenematerialet er bare funnet i hoveddalene, helst i dalbunnene, men også som støtsidemorener et stykke oppover dalsidene. En betydelig del av steinmaterialet har åpenbart flere transportfaser bak seg, med minst én glasifluvial (fluvial) fortid. Det rundete materialet er plukket opp av breer som har gått fram over eldre avsetninger i dalbunnen.

Flere ting, bl. a. mange mammutfunn, peker mot at de nevnte submoreneavleiringer — og dermed steinenes runding kan stamme fra en interstadial tid. Forfatteren leder for tiden undersøkelser av slike sedimenter i Gudbrandsdalen.

Glasifluvialt materiale.

Også for glasifluvialt materiale gjør det seg gjeldende en to-deling: *Type Ab*, autoktont glasifluvium, domineres av kantrundet steinmateriale, mens rundet ikke utgjør over 20 %. Som det framgår av fig. 3 og 4, skiller ikke dette seg mye fra det autoktone morenematerialet, som



Skjematisk jordartsfordeling i tverrprofil
(Typical cross section in valley)
Morene-materiale
Glasifluvialt
Fluvialt
m.o dalbunn (m.a valley floor)
Morene-materiale
Glasifluvialt materiale
Fluvialt materiale
Overløp fra Espedalen
(Overflow from Espedalen)
Mjøsa
m.o.h.
600
500
400
300
200
0
10
20
30
40
50
Km fra Mjøsa

det stammer fra. Typen omfatter alle glasifluviale avleiringer som ikke er avsatt av hoveddreneringer *langs* hoveddaler.

Type Bb, derimot, har omtrent 50 % rundet materiale. Dette synes å være alminnelig for avleiringer langs hoveddreneringskanaler under isavsmeltningen.

Steinmaterialets transport.

Den store forskjellen på de to typer glasifluvialt materiale mener jeg særlig skyldes *transportmåten,* i mindre grad *transportlengden.*

Så lenge materialet transporteres glasifluvialt supra- eller englasialt, kan det synes som om steinmaterialet ikke oppnår å bli rundet selv når transportlengden er flere ti-kilometre. Det antas at den vesentligste transport for materialet i mindre eskers, kames, vifter o. l. har foregått supra- og/eller englasialt. Også når det gjelder langtransporterte stein i morenemateriale er det mulig at en vesentlig del av transporten er foregått glasifluvialt på denne måten, selv om ikke materialet er blitt rundet.

Subglasial- og alminnelig elvetransport, derimot, runder materialet meget hurtig. Etter mindre enn 10 km er halvparten funnet rundet under antatt subglasial drenering. Dette støttes av morfogrammene for Ac, som er fra resente vifter med kort transport. Her er det, sammenliknet med type Ab, en klar forskyvning mot rundet.

At *transportmåten* er avgjørende for rundingsgraden synes å framgå av figur 5, som viser et lengdesnitt langs Gausdal, og hvor det er inn-

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◄

Fig. 5. Lengeprofil av Gausa med morfogrammer av morene-, glasifluvialt- og fluvialt materiale langs dalføret: Morenematerialet domineres av kantrundet stein, antall runde stein holder seg konstant: 10—20 %. Morenelokaliteten ytterst i dalen representerer alloktont morenemateriale. Glasifluvialt materiale øker sin rundingsgrad nedover dalen, men tilskudd fra morenematerialet gjør at det er minst like mye kantrundet, som godt rundet materiale, også i dalmunningen. Fluvialt materiale er klart bedre rundet enn tilsvarende glasifluvialt materiale.

Longitudinal profile of the Gausa river with morphograms of the till-, glaciofluvial-, and fluvial material. The till chiefly consists of material with abraded angles, the amount of rounded particles stays constant between 10 to 20 %. The till locality near the valley mouth consists of allochtonous till. Glaciofluvial material increase in degree of roundness with increasing transport length, but addition of till keeps the content of material with abraded angles at least as high as the content of well rounded material. This is also the case in the valley mouth. Fluvial material is distinctly more rounded than similar glaciofluvial material.

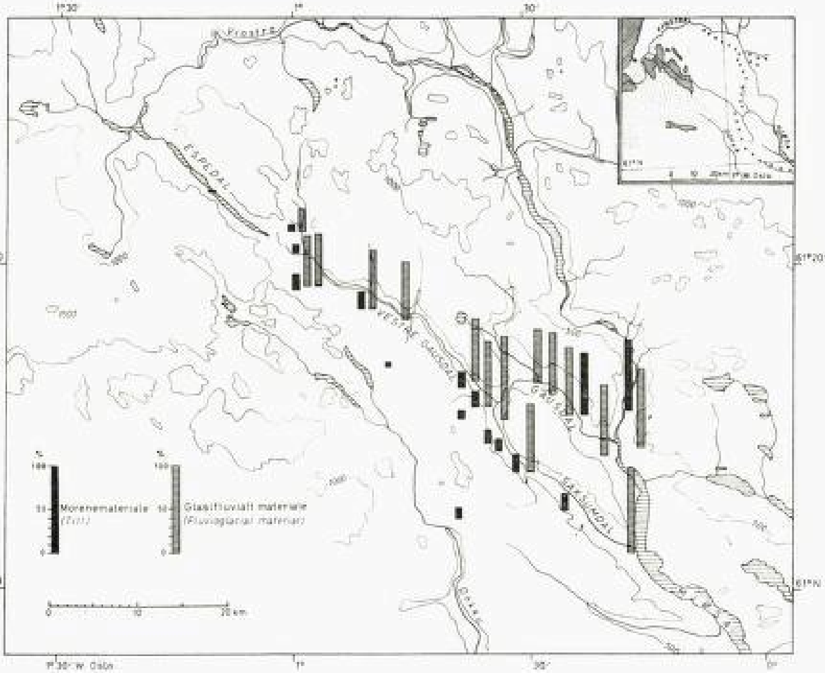

Fig. 6. Prosent rundet (rundet pluss godt rundet) jotunbergarter i morene- og glasifluviale avleiringer. Geologisk oversiktskart som viser jotunbergartenes mulige opphavssted. Skravert: Jotunbergarter.

Prikker: Valdressparagmitt, og andre sedimentære skyvedekkebergarter, som kan forveksles med jotunbergarter.

Ringer: Biskopås- (Biri-) konglomeratet, som inneholder boller av bl.a. jotunbergarter.

I det alminnelige morenematerialet har jotunbergartene lav rundingsprosent. Inneholder moreneavleiringer jotunbergarter som er rundet, tolkes avleiringene som sammensatt av materiale med flere transportfaser bak seg.

Percentage of rounded plus well rounded particles of Jotun rock types in till and glaciofluvial deposits. Geological map shows possible sources of Jotun rock types. Hatched: Jotun rock types.

Dots: Valdres sparagmite and other sedimentary nappe rocks that can be mistakingly classified as Jotun rocks.

Rings: Biskopås (Biri-) conglomerate which contains particles of Jotun rock types.

In the normal till deposits the Jotun rock types have a low degree of rounding. Till deposits with rounded Jotun rock particles are interpreted as compound deposits with several phases of transport.

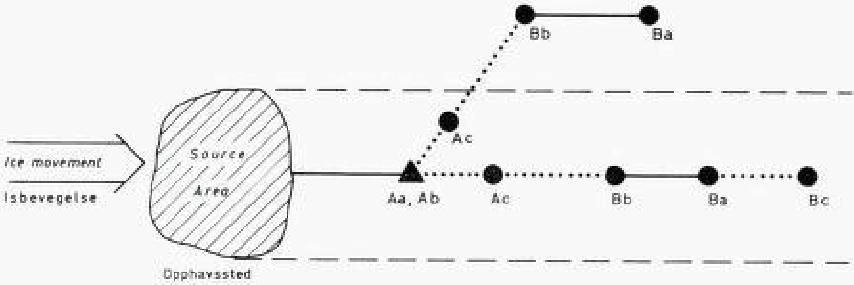

Fig. 7. Skjematisk fremstilling av steinmaterialets transporthistorie, og sammenhengen mellom autoktont- og alloktont materiale: Fra opphavsstedet føres løsrevet materiale gjennom første transportfase fram til morfogramtype Aa (autoktont morenemateriale) eller til type Ab (autoktont glasifluvialt materiale). Mindre enn 20 % av steinmaterialet er rundet. Transporten skjer ved istransport, eller ved englasial/supraglasial smeltevanntransport. Slik transport er vist ved heltrukkete linjer. Dette materialet vil ved kort elvetransport omlagres til type Ac (autoktont fluvialt materiale), eller gjennom glasifluvial transport langs dalbunnen (subglasial/subaeril) drenering til type Bb (alloktont glasifluvialt materiale). Begge deler gir avleiringer hvor mer enn 50 % av steinmaterialet er rundet. Slik transport er stiplet. Hvis dette materialet blir flyttet av en senere isbevegelse, dannes morenetype Ba (alloktont morenemateriale). En påfølgende glasifluvial eller fluvial omleiring vil gi henholdsvis morfogramtyper Bb og Bc. Figuren illustrerer også hvorfor bergarter fra et bestemt sted nesten alltid (mer enn 80 % av antallet), er rundet når det opptrer utenfor sektoren for isbevegelsesretningen: Disse bergartene har minst én glasifluvial (fluvial) fortid, før de er innbakt i morenemateriale.

Schematic presentation of transport history and development of autochtonous and allochtonous material. Freshly eroded debris from the source area gives deposits having morphograms of the type Aa (autochtonous till) or Ab (autochtonous glaciofluvial material) showing less than 20 % of rounded particles. Transporting agent is ice or englacial or supraglacial meltwater. This transport is shown in unbroken lines. The material will, after a short transport by river, yield deposits with morphograms of the type Ac (autochtonous fluvial material). If the transport is glaciofluvial along the valley floor (subglacial/suaearial) the result will be deposits of the type Bb (allochtonous glaciofluvial material). Both types yields deposits where more than 50 % of the material is rounded. This transport is shown in broken lines. Where this material has later been transported by ice, it gives till deposits of the type Ba (allochtonous till). A later glaciofluvial or fluvial transport/deposition will give morphograms of the type Bb and Bc. The figure also illustrates why rocks from a given locality in nearly all cases (more than 80 %) are rounded when found outside the sector of ice flow. This material has undergone at least one glaciofluvial/fluvial transport before being caught by ice and deposited in till.

tegnet en del morfogrammer for morene-, glasifluvialt- og fluvialt materiale. Man ser at morenematerialets morfogrammer forandres lite nedover dalen. Selv om flere av analysene er tatt nær dalbunnen utgjør rundet materiale under 20 %. Ett unntak er lokaliteten ytterst i dalen og som representerer type Ba.

Glasifluvialt materiale langs dalbunnen rundes altså raskt. Men på grunn av tilblandingen fra sidene, i første rekke fra morenemateriale, utgjøre kantrundet overalt en betydelig del, minst 20 %.

Figur 6 viser *jotunbergarters* rundingsprosent i henholdsvis morene- og glasifluvialt materiale. Rundingsprosent er summen av prosent rundet og godt rundet. Her får man bekreftet at det er *transportmåten* som har størst betydning for morfogrammet: Det glasifluviale materialet langs Gausdals bunn rundes raskt, mens runde jotunbergarter i moreneavleiringer utgjør mindre enn 20 % av de samlete jotunbergarter. Dette gjelder for materialet som er avleiret under de siste isbevegelser. Kommer man utenfor sektoren for isbevegelser fra opphavsstedet for bergartene, er mer enn 80 % av jotunbergartene rundet, noe som er i overensstemmelse med en sammensatt transporthistorie, med minst én glasifluvial (fluvial) fortid.

På fig. 7 er sammenhengen mellom transporthistorie og rundingsgrad vist skjematisk.

Konklusjon.

De forskjellige glasigene jordarter er funnet å ha *karakteristiske morfogrammer*, helt vesentlig bestemt av *transportmåte* og *transportlengde*.

Rundingsanalyser gir derfor viktige opplysninger om transporthistorien for de enkelte stein, og dermed også opplysninger om avleiringers genesis.

Dersom en større eller mindre del av steinene har flere transportfaser bak seg, blir morfogrammene vanskeligere å tyde, men også ved slike avleiringer mener jeg analyser av rundingsgraden er et betydningsfullt hjelpemiddel.

Litteratur.

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Manuskriptet mottatt i april 1970.

ISOTOPIC EVIDENCE ON THE AGE OF THE TRYSIL PORPHYRIES AND GRANITES IN EASTERN HEDMARK, NORWAY

by

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Abstract.

Rocks from the (sub-Jotnian) acidic plutonic and volcanic basement complexes in the Trysil area, eastern Hedmark, yield a Rb—Sr isochron age of 1541 ± 69 million years. This agrees within the limits of error with the isochrom age of 1590 ± 65 million years determined for the Dala porphyries and granites in Dalarna, Sweden, which are the continuation of the acidic igneous complexes in the Trysil area. The sub-Jotnian acidic magmatism in the eastern Hedmark—Dalarna region can thus be dated at 1570 ± 40 million years ago, i.e. some 100 million years younger than the termination of the Svecofennian orogeny. (Ages computed with $\lambda = 1.47 \times 10^{-11}$ yr^{-1}; errors with 95 % confidence level). Chemically, this magmatism is characterized by a granitic to alkali granitic and alkali syenitic composition.

The Trysil area has also been affected by a tectonothermal event in Sveconorwegian time, about 925 million years ago, as evidenced by the Rb—Sr and K—Ar ages of separated biotites.

Introduction.

Studies on the geology of the Trysil area in eastern Hedmark have been published by Schiøtz (1903), Reusch (1914), Holmsen (1915), Holtedahl (1921), Dons (1960) and Holmsen et al. (1966). The Quaternary deposits were mapped by Holmsen (1958, 1960). A geological sketch map of the area is shown in Fig. 1 (mainly after the Geologisk Kart over Norge, 1960).

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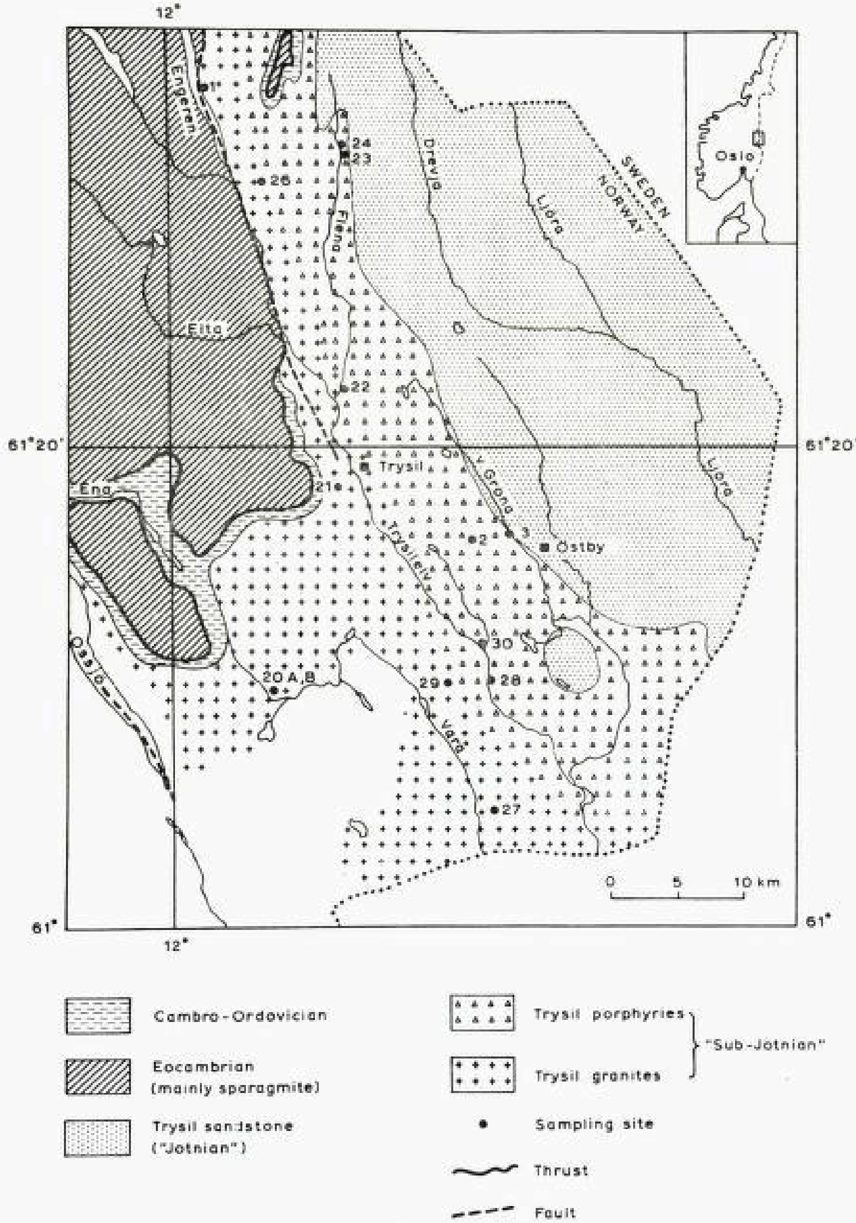

Fig. 1. Geological sketch map of the Trysil area, eastern Hedmark (mainly after the Geologisk Kart over Norge, 1960), showing the locations of the samples investigated. The numbers 1—30 correspond to the sample numbers 67 Hed 1 — 68 Hed 30 (see also Table 1).

The sub-Jotnian basement is made of acidic volcanic and plutonic rocks, i.e. the «Trysil porphyries» in the eastern part and the «Trysil granites» in the western part of the map area. All basement rocks have been tectonized, occasionally showing a gneissic appearance. The boundary between porphyries and granites is of tectonic origin (strike N-S, with steep dips); the porphyry series has moved downwards with regard to the granites (Holtedahl, 1921). Reusch (1914) already suggested a co-magmatic origin for the plutonic and volcanic complexes.

Towards the Swedish border, the porphyries are overlain by the Jotnian Trysil sandstone (mostly eastward dipping, with increasing deformation to the West). In the western part of the map area thick series of «Eocambrian» rocks (mainly sparagmites) and Cambro-Ordovician deposits occur.

The Trysil granites, porphyries and sandstone continue into Dalarna, Sweden, where they are designated as Dala granites, porphyries and sandstone, respectively. According to Hjelmqvist (1966), the Dala granites range in composition from true granitic and granodioritic rocks to quartz syenites, while the Dala porphyries mostly have an ignimbritic character. For these plutonic and volcanic rocks likewise a co-magmatic origin was postulated (Rutten, 1966).

The present study reports the results of a Rb–Sr isochron study of 14 whole-rock samples from the Trysil area (seven granites and syenites, one aplitic vein and six porphyries). Also, biotites separated from two granites were dated according to the Rb–Sr and K–Ar methods. The sampling sites are shown in Fig. 1 and listed in Table 1.

Plutonic and volcanic rocks investigated.

The plutonic rocks are coarse-grained, usually with reddish or greenish colours. Occasionally, aplitic veins can be observed. The volcanic rocks are fine-grained and mostly porphyric, usually in greyish, greenish or reddish colours. Major element compositions of the samples investigated have been analyzed by X-ray fluorescence spectrometry, using G-2 as external reference standard (Table 2). The chemistry points to a magmatism of granitic to alkali granitic and alkali syenitic composition.

With the exception of 68 Hed 29, all rocks contain quartz, microcline-perthite and albite as main components. These minerals also make up the phenocrysts in the porphyric volcanics. The granite 68 Hed 29 bears oligoclase (An_{18}) instead of albite. The plagioclase feldspars are

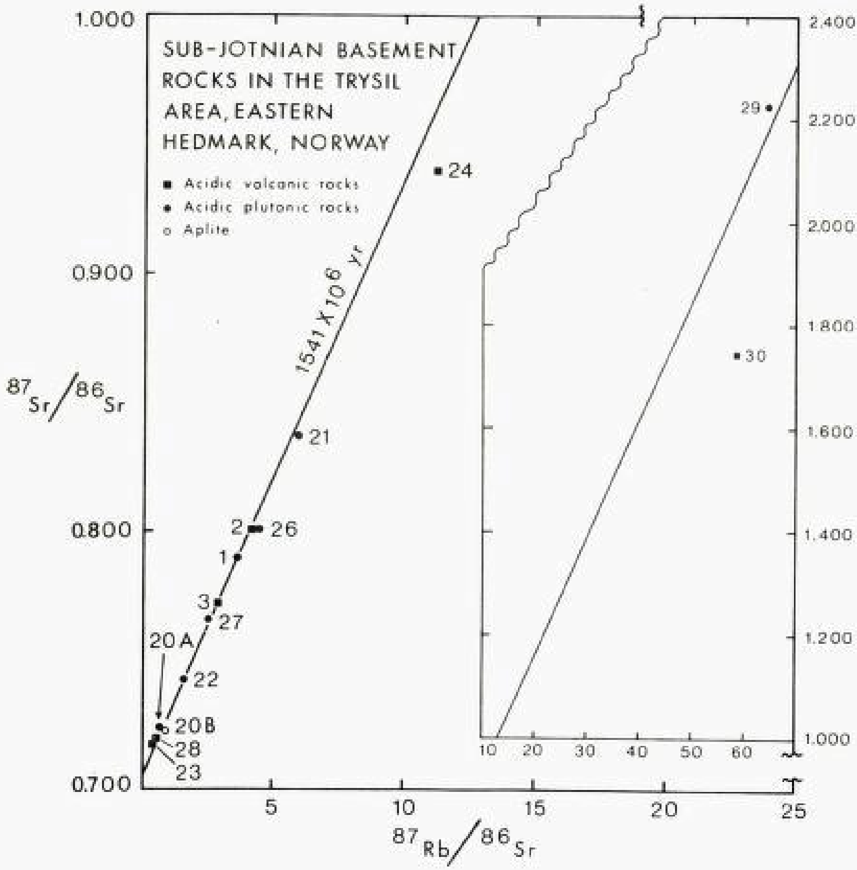

Fig. 2. $^{87}Sr/^{86}Sr$ *versus* $^{87}Rb/^{86}Sr$ *plot of the whole-rock samples from the Trysil area. The numbers 1—30 correspond to the sample numbers 67 Hed 1 — 68 Hed 30. Data points 24 and 30 were omitted from the isochron calculation.*

often filled with sericite and epidote. Biotite (partly chloritized) and colourless mica occur in varying amounts in all samples. Ferrohastingsite and actinolitic hornblende are major constituents in the granite 68 Hed 21 and the trachyandesite 68 Hed 23, respectively. Stilpnomelane is present in varying amounts in the samples 68 Hed 21, 22, 26 and 29; the occurrence of this mineral seems to be restricted to zones of shearing and mylonitization. Wide-spread accessories are titanite, zircon, apatite, fluorite, orthite, calcite, hematite and opaque ore. Especially in the plutonic rocks titanite often occurs in conspicuous crystals, easily recognizable in hand specimen, while the zircon likewise forms fairly large crystals (up to 0.15 mm in length).

Effects of shearing and mylonitization are shown by four samples, three (68 Hed 22, 26 and 29) from the boundary zone between the plutonic and volcanic complexes, and sample 68 Hed 21 probably from a NNW–SSE trending fault zone. Stilpnomelane has exclusively been found in these rocks.

Experimental procedures.

Splits of crushed and pulverized whole-rock samples were analyzed for their Rb and Sr contents by X-ray fluorescence spectrometry; five rocks have also been measured by stable isotope dilution. Separated biotites were analyzed by stable isotope dilution only. The isotope dilution measurements were made with spikes enriched in ^{87}Rb and ^{84}Sr, respectively. For the isotope measurements a 20 cm, 60° mass-spectrometer with digital output was used, utilizing thermal ionization and multiplier detection. A single Ta filament source was employed for all measurements. Except for the biotites, ^{87}Sr/^{86}Sr ratios were measured directly on unspiked strontium; whenever isotope dilution analyses were made, the ^{87}Sr/^{86}Sr ratio was also calculated from the isotope dilution run. Correction for effects of isotope fractionation and mass discrimination was made by normalizing to ^{88}Sr/^{86}Sr = 8.3752.

The X-ray fluorescence data were obtained on a semi-automatic X-ray spectrometer equipped with a 2 kW Mo X-ray tube and a (200) LiF crystal for the analysis of Rb, Sr, Fe, Ti, Ca and K, a 2 kW Cr X-ray tube and a PE crystal for the analysis of Si and Al, or a 2 kW Cr X-ray tube and a KAP crystal for the analysis of Mg and Na. Mass-absorption corrections for Rb and Sr (both for external standard and sample) were made by measurement of the intensity of the Compton scattering of the Mo Kα primary beam. Matrix corrections for the major element analyses were calculated from tables of mass-absorption coefficients (Dewey et al., 1969). Assuming, initially, that there are no differences in mass-absorption between standard (G-2) and sample, an approximate composition of the sample is calculated by direct reference to the standard. From this approximate composition the total mass-absorption coefficient of the sample is determined for each operational wavelength. The original chemical composition is then corrected by the ratio of the calculated total mass-absorption coefficients of sample and standard, respectively. In turn, this corrected chemical composition is used to provide a yet more accurate value for the total mass-absorption coefficient at each wavelength. The process is repeated until a self-consistent analysis results.

Rb—Sr isochrons were computed as the best-fitted straight lines through the $^{87}Sr/^{86}Sr$ — $^{87}Rb/^{86}Sr$ data points, following the computation method of York (1966, 1967) and Williamson (1968). To each pair of coordinates the relative weight was assigned based upon estimated relative standard errors of 0.6 % and 2.0 % for the measured $^{87}Sr/^{86}Sr$ and $^{87}Rb/^{86}Sr$ ratios, respectively. The errors for the isochron ages and the initial $^{87}Sr/^{86}Sr$ ratios are quoted with 95% confidence limits as calculated from the analytical data.

Potassium determinations were made by flame-photometry with a lithium internal standard and a CsAl buffer. Argon was extracted in a bakeable glass vacuum apparatus and determined by standard isotope dilution techniques (using ^{38}Ar as tracer) in a Reynolds-type glass mass-spectrometer; the measurements were made by the static method.

Constants used.

All calculations were made using the following constants:

$$^{87}\text{Rb}: \lambda\beta = 1.47 \times 10^{-11} \text{ yr}^{-1};$$
$$^{40}\text{K}: \lambda e = 5.85 \times 10^{-11} \text{ yr}^{-1},$$
$$\lambda\beta = 4.72 \times 10^{-10} \text{ yr}^{-1}, \text{ and}$$
$$\text{abundance } ^{40}\text{K} = 0.0118 \text{ atom \% total K.}$$

Rb—Sr ages of the Dala porphyries and granites published by Welin et al. (1966) and Welin & Lundqvist (1970) are based upon a ^{87}Rb decay constant of 1.39×10^{-11} yr^{-1}. These ages have to be multiplied by 0.946 in order to make them comparable with the Rb—Sr ages of the present study.

Results and discussion.

The whole-rock Rb—Sr data are listed in Table 3, whilst Fig. 2 shows an isochron plot of the data points. In Table 4, the Rb—Sr and K—Ar data of the separated biotites are given.

The K—Ar ages of both biotites and the Rb—Sr age of biotite 68 Hed 27 are concordant at about 925 million years. Evidently, this date reflects the imprint of a tectonothermal event, causing resetting of the K—Ar and Rb—Sr clocks. This 925 million years old event can be placed within the Sveconorwegian (Dalslandian) orogenic period, the youngest Precambrian period of igneous and metamorphic activities wide-spread in southern and south-western Scandinavia (e.g., Broch, 1964; Welin, 1966). Biotite 68 Hed 29 has a somewhat lower Rb–Sr age; possibly, this phenomenon may be connected with its high degree of chloritization. (The concentrate still contains much chlorite, which is also reflected in its low potassium content).

When plotted in a diagram of $^{87}Sr/^{86}Sr$ versus $^{87}Rb/^{86}Sr$ (Fig. 2), it is evident that not all rocks follow the simple Rb–Sr systematics of a closed system. Twelve samples, i.e. all plutonic and volcanic rocks with lower Rb/Sr ratios and one granite sample (68 Hed 29) of higher Rb/Sr ratio, show an approximately linear arrangement. This confirms the allegedly co-magmatic origin of both rock types, as any time interval between eruption of the porphyries and intrusion of the granites and syenites must have been relatively short. However, two rhyolites of high Rb/Sr ratio (68 Hed 30 and, to a minor extent, 68 Hed 24) depart significantly to the right of the overall linear array of the other twelve samples.

All rocks in the area have experienced a tectonothermal event in Sveconorwegian time. Rocks of high Rb/Sr ratio (thus also of relatively high $^{87}Sr/^{86}Sr$ before metamorphism) are the most likely to have lost ^{87}Sr or exchanged strontium with less radiogenic strontium in surrounding rocks during the metamorphism. The two rocks falling significantly below the best-fitted straight line through the other twelve samples are both rhyolites of high Rb/Sr ratio. It is obvious that rocks belonging to a sequence of volcanic extrusions and eruptions, with many abrupt Rb/Sr discontinuities over relatively short distances, are the most vulnerable for modifications of strontium isotopic relationships between successive layers under conditions of metamorphism. An alternative explanation, that the lowering of the apparent Rb–Sr ages of the rhyolites 68 Hed 24 and 30 was entirely or mainly due to gain of rubidium, does not seem probable. For 68 Hed 30, such a process would have implied a doubling of the rubidium content in the rock system at the time of metamorphism, which is hardly feasible without assuming extensive metasomatism involving enrichment in potassium; no evidence whatsoever is available for such a process. Moreover, the two rhyolites of high Rb/Sr ratio are characteristically low in strontium and calcium, not high in rubidium and potassium.

If we omit the rhyolites 68 Hed 24 and 30, then the isochron age computed for the twelve samples of plutonic and volcanic rocks is 1541 ± 63 million years with an intitial $^{87}Sr/^{86}Sr$ ratio of 0.7051 ± 0.0039. It may be noted that granite sample 68 Hed 29, being a much higher point than the other eleven data-points, has a disproportionally strong influence on the slope of this isochron. If this sample is omitted from the computations, then the slope of the isochron would correspond to an age of 1428 ± 124 million years and the intercept to an initial $^{87}Sr/^{86}Sr$ ratio of 0.7089 ± 0.0052.

As all rocks could have behaved to some degree as open systems with regard to strontium during the metamorphism 925 million years ago, the isochron age of 1541 ± 63 million years might be a minimum estimate of the true age. In our opinion, however, this age should approximate the true age of the magmatism. This view is supported by the Rb—Sr measurements on the Dala porphyries and granites in Sweden, which form the continuation of the Trysil porphyries and granites. Measurements on these rocks have been made by Welin et al. (1966), Priem et al. (1968) and Welin & Lundqvist (1970); in total, eleven rocks (ten porphyries and one granite) from that area have been analysed. It was shown by Welin & Lundqvist that nine porphyry samples and the granite define a fairly good isochron. The slope of the isochron is computed by the present authors as corresponding to an age of 1590 ± 65 million years with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7042 ± 0.0062. Of the eleven samples measured from the Dala porphyries and granite, there is also one rhyolite sample, likewise of high Rb/Sr ratio, that falls significantly below the 1590 million years isochron (see the discussion by Welin & Lundqvist).

Within the limits of error, the isochron ages of the sub-Jotnian plutonic and volcanic complexes in the Trysil area (excluding samples 68 Hed 24 and 30) and Dalarna (excluding one rhyolite sample) are identical. If we take all samples from the Trysil area and Dalarna together, excluding the three data-points which do not conform to the linear array, then the slope of the computed isochron corresponds to an age of 1569 ± 42 million years and the intercept to an initial $^{87}Sr/^{86}Sr$ ratio of 0.7047 ± 0.0030.

Conclusions.

Rb—Sr isochron data of the sub-Jotnian plutonic and volcanic rocks in the Trysil area, eastern Hedmark, and in the adjoining Swedish province of Dalarna indicate an event of acidic magmatism 1570 ± 40 million years ago. However, contrary to Dalarna where intrusive and extrusive rocks are often closely associated, the plutonic and volcanic complexes in the Trysil area seem to be spatially separated; the volcanic rocks make up the eastern block which has been down-faulted with regard to the western block of plutonic rocks. It is obvious that the sub-Jotnian acidic basement complexes in the eastern Hedmark—Dalarna region represent magmas that were partly intruded at a high crustal level, and partly extruded at the Precambrian surface of the Earth.

This magmatism took place some 100 million years after the termination of the Svecofennian orogeny. The magmas range in composition from granitic to alkali granitic and alkali syenitic, in accordance with the alkaline affinities characteristic for most of the post-Svecofennian acidic magmatism in the Baltic Shield. It also reflects the anorogenic nature of the magmatic event, as syenitic and per-alkaline granitic magmatism are generally restricted to stable continental areas, characterized tectonically by simple fracturing.

The biotite Rb—Sr and K—Ar ages of around 925 million years reflect the imprint of a tectonothermal event in Sveconorwegian (Dalslandian) time. This younger event is probably also responsible for the migration of radiogenic strontium from some volcanic rocks.

Acknowledgements.

The authors are much indebted to Thor L. Sverdrup and Øyvind Gvein, geologists with the Norges Geologiske Undersøkelse (N.G.U.) in Trondheim, for their help and advice in collecting the specimens investigated and for helpful discussions. Field work in connection with this study was sponsored by N.G.U. The computer calculations for the isochron analyses were made at the Mathematical Centre, Amsterdam. The authors also acknowledge valuable discussions with Dr. Eric Welin, Stockholm.

This work forms part of the research programme of the «Stichting voor Isotopen-Geologisch Onderzoek», supported by the Netherlands Organization for the Advancement of Pure Research (Z.W.O.).

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Table I

Investigated plutonic and volcanic rocks from the Trysil area (see Fig. 1).

Sample Nr.	Rock type	Sampling site
67 Hed 1	Albite granite	Storeneset, E side Engeren Lake.
67 Hed 2	Porphyric rhyolite	On road Nr. 25, 6 km E from Nybergsund.
67 Hed 3	Porphyric rhyolite	Grønli on road Nr. 25.
68 Hed 20-A	Quartz-rich alkali syenite	On road Nr. 25, 14.5 km SW from Nybergsund.
68 Hed 20-B	Aplitic vein	idem.
68 Hed 21	Slightly mylonitized, stilpnomelane-bearing ferro-hastingsite granite	On road from Trysil to Gjevaldshaugen, 2.1 km SW from bridge at Trysil.
68 Hed 22	Mylonitized, stilpnomelane-bearing, quartz-rich alkali syenite	Flendalen, on road 7.8 km N from Trysil bridge.
68 Hed 23	Hornblende trachyte	Rundhöa, about 24 km N from Trysil.
68 Hed 24	Porphyric rhyolite	Rundhöa, about 1 km NNW from 68 Hed 23.
68 Hed 26	Slightly mylonitized, stilpnomelane-bearing biotite granite	Gammelsæterberget, about 24 km NNW from Trysil.
68 Hed 27	Biotite granite	On road Nr. 208, 20.5 km E from Midtskogberget.
68 Hed 28	Porphyric quartz-rich trachyte	Sagnfossen in Trysilelva.
68 Hed 29	Gneissose stilpnomelane-biotite granite	Vestsjöberget.
68 Hed 30	Rhyolite	Kolos, on road Nr. 26.

Table 2

Major element compositions of the investigated plutonic and volcanic rocks from the Trysil area (Wt. %).

Sample Nr.	G-2	Hed 1	Hed 2	Hed 3	Hed 20-A	Hed 20-B	Hed 21	Hed 22	Hed 23	Hed 24	Hed 26	Hed 27	Hed 28	Hed 29	Hed 30
SiO_2	69.19	71.15	69.86	73.37	63.50	73.58	71.57	64.01	62.01	74.17	72.67	70.92	67.19	75.66	75.88
TiO_2	0.53	0.46	0.33	0.27	0.58	0.16	0.40	0.62	0.82	0.24	0.27	0.46	0.36	0.11	0.22
Al_2O_3	15.35	13.65	14.73	13.57	17.30	13.71	13.61	16.89	16.53	12.80	13.63	13.82	16.14	12.44	11.33
Fe_2O_3*	2.77	2.83	3.03	2.20	3.52	1.39	2.79	4.05	5.83	1.89	2.01	2.80	3.34	1.41	3.21
MgO	0.78	0.43	0.54	0.51	0.79	0.25	0.24	0.79	1.61	0.10	0.28	0.42	0.89	0.05	0.08
CaO	1.99	1.28	1.36	1.30	2.80	1.25	0.81	2.09	3.61	0.72	1.09	1.12	2.83	0.48	0.11
Na_2O	4.16	3.73	4.02	2.75	4.56	4.03	4.09	4.09	4.51	3.11	3.87	4.16	4.87	4.61	3.68
K_2O	4.51	5.74	5.40	5.31	6.23	4.90	5.74	6.73	4.35	6.27	5.46	5.58	3.67	4.52	4.78

* Total Fe as Fe_2O_3.

Table 3

Rb-Sr whole-rock data of the investigated plutonic and volcanic rocks from the Trysil area.

Sample Nr.	Rb ppm Wt.	Sr ppm Wt.	$^{87}Sr/^{86}Sr$	$^{87}Rb/^{86}Sr$
67 Hed 1	209 **	166 **	0.7902°°	3.70
	209 *	163 *	0.7881°	
67 Hed 2	285 **	196 **	0.8013°°	4.30
	286 *	192 *	0.8000°	
67 Hed 3	199 **	196 **	0.7718°°	2.99
	201 *	194 *	0.7701°	
68 Hed 20-A	131 *	545 *	0.7235°	0.70
			0.7260°	
68 Hed 20-B	99.1*	355 *	0.7231°	0.81
			0.7262°	
			0.7227°	
68 Hed 21	139 *	67.7*	0.8380°	6.02
			0.8360°	
			0.8372°	
68 Hed 22	181 *	329 *	0.7425°	1.60
			0.7424°	
			0.7427°	
68 Hed 23	98.3*	615 *	0.7175°	0.46
			0.7192°	
68 Hed 24	236 *	61.8*	0.9399°	11.31
			0.9402°	
			0.9420°	
68 Hed 26	221 *	144 *	0.8095°	4.48
			0.8116°	
			0.8109°	
68 Hed 27	136 *	154 *	0.7701°	2.57
			0.7654°	
			0.7641°	
			0.7652°	
68 Hed 28	127 *	737 *	0.7198°	0.50
			0.7200°	
68 Hed 29	314 **	16.1**	2.222 °°	64.5
		16.2**	2.226 °°	
	313 *	16.2*	2.224 °	
68 Hed 30	247 **	13.5**	1.748 °°	58.5
	244 *	13.3*	1.748 °	

** Isotope dilution analysis.
* X-ray fluorescence spectrometry.
°° Calculated from isotope dilution run.
° Direct measurement on unspiked sample.

Table 4

*Rb-Sr and K-Ar data of separated biotites from the granites 68 Hed 27 and 29**

Rb-Sr data

	Rb ppm Wt.	Sr ppm Wt.	$^{87}Sr/^{86}Sr$**	Radiogenic ^{87}Sr*** ppm Wt.	Age**** million years
68 Hed 27	893	23.1	2.560	3.50	928
	906	22.7	2.576	3.47	
68 Hed 29	1962	17.1	8.896	6.95	842
	1996	16.5	9.529	7.02	

K-Ar data

	K % Wt.	Radiogenic ^{40}Ar ppm Wt.	Atmospheric ^{40}Ar (% total ^{40}Ar)	Age**** million years
68 Hed 27	6.88	0.594	11.4	942
	6.88	0.594	10.0	
68 Hed 29	3.98	0.326	4.0	908
	4.01			

* 68 Hed 27: Green biotite, $n_y = 1.633$. The concentrate contains very little chlorite.

68 Hed 29: Dark, greenish brown lepidomelane, $n_y = 1.672$. The concentrate contains many impurities, mainly chlorite and some stilpnomelane.

** Calculated from isotope dilution runs.

*** The initial $^{87}Sr/^{86}Sr$ ratios were calculated as 0.730 and 1.425, respectively, using a graphical analysis according to Compston, Jeffery & Riley (1960).

**** Maximum analytical error estimated at ± 4 %.

Manuscript received in April 1970.

GEOLOGY OF THE HØLONDA–HULSJØEN AREA, TRONDHEIM REGION

by
Josef Chaloupský[1])

Abstract.

The geology of an area south of Trondheim is described. A new stratigraphical division of the Early Palaeozoic is proposed and its structure outlined.

The entire Early Palaeozoic sedimentary and volcanic complex has been divided into four separate groups, differing in lithology and separated by conspicuous layers of synorogenic conglomerates:

1) The oldest Støren Group (presumably of Tremadocian age), 2) The Krokstad Group (largely of Arenigian, Llanvirnian and Llandeilian age), 3) The Lower Sandå Group (of Caradocian and perhaps partly of Ashgillian age), 4) The youngest Upper Sandå Group (presumably of Silurian age). The total thickness of the Ordovician-Silurian sequence overlying the Støren volcanic complex is about 800 m. The overall structure is interpreted as a series of closely packed tight to isoclinal folds accompanied by shearing and stretching phenomena, which are grouped into extensive anticlinorial and synclinorial belts.

Introduction.

In this paper the results are presented of geological investigations and mapping of an area 45 km SSW of Trondheim, carried out during the summer of 1969. The purpose of this work was to produce a geological map of the area Hølonda–Evjeseter–Hulsjøen–Lundseter.

Mapping data were plotted onto airphotos from which a geological map on a scale of 1 : 25 000 was subsequently compiled. The area investigated is of a low mountainous topography, the lower parts of the terrain being locally covered with Recent and Pleistocene deposits of considerable thickness.

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Fig. 1. Black square indicates location of the mapped area.

This extremely interesting sector of the Caledonides in the Trondheim region is distinguished by a very low grade of metamorphism and by the presence of thick Early Palaeozoic complexes of flysch-like character. Although the fold structure is relatively complicated, several fossiliferous horizons and distinctive marker beds, particularly conglomerates, make it possible to locate the individual members of the sequence within the Early Palaeozoic stratigraphical column with reasonable accuracy.

The interesting geology has attracted the attention of a number of geologists; among the modern ones Th. Vogt (1945) should be mentioned in particular. His division of the Early Palaeozoic complex into four series (Støren, Lower Hovin, Upper Hovin and Horg) and his dating of the early Caledonian epeirogenic phases has provided a basis for the stratigraphy of Early Palaeozoic successions in other parts of the Trondheim region Caledonides. Studies published after Vogt's detailed work, particularly those of C. W. Carstens (1951, 1952), H. Carstens (1960), Blake (1962), Chadwick et al. (1963), Chaloupský (1963) and Carter (1967) have provided a lot of new and valuable information but essentially no changes in the basic conception of the stratigraphy and structure of the area. The relevant results that modified the existing opinions were, for example, the recognition of the Hølonda porphyrite as being intrusive, and not effusive (C. W. Carstens 1951, Chadwick

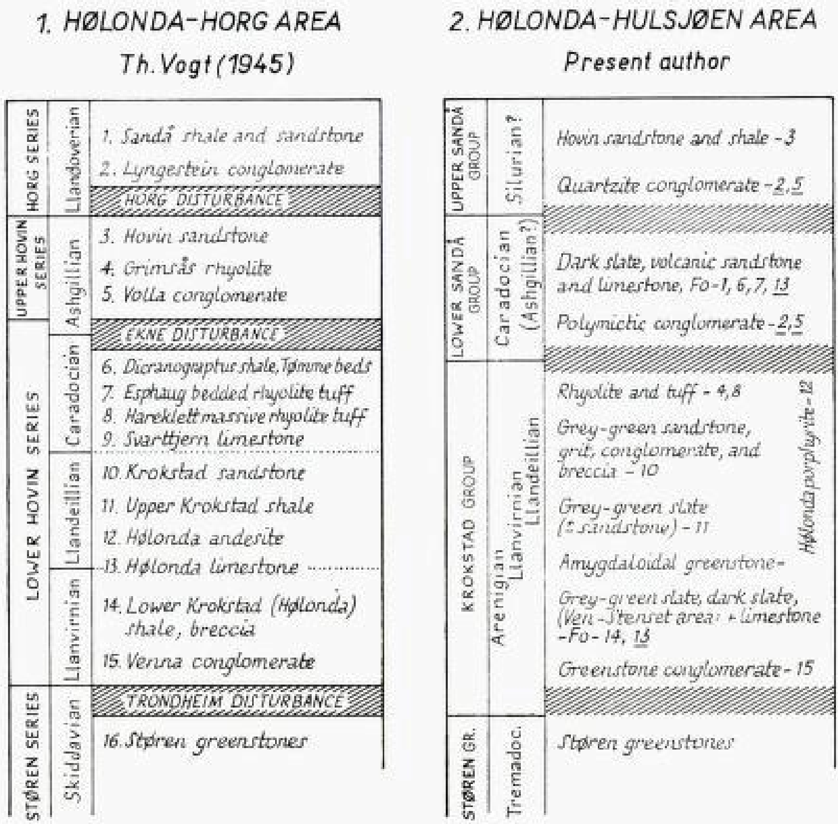

Fig. 2. Correlation of the stratigraphical schemes of Th. Vogt (1945) and the author. Numbers after the lithologies in column 2 correspond to numerals of equivalent rocks (or partly equivalent rocks when underlined) in column 1.
Fo = fossiliferous horizon.

et al. 1963); finds of graptolites in the Bogo shale, of Middle Arenigian age (Blake 1962), which allowed the boundary between the Støren and Hovin Series to be drawn more precisely; observation that the Hølonda limestone comprises more than one horizon (Chadwick et al. 1963); inclusion of the Jåren Beds in the Hovin Series (Carter 1967); incorporation of stratigraphically differentiated Sandå shale, Tømme black shale and Hareklett bedded rhyolite tuff in one complex showing tectonic repetition (Chaloupský 1963); a more precise stratigraphical positioning of some other members of the sequence (Oftedahl et al. 1969).

The stratigraphical correlation of beds from different sectors of the area studied is to be found in the various papers mentioned above. The basic stratigraphical division of the Early Palaeozoic complex, after Th. Vogt, is given in Fig. 2.

The author has mapped the area which is directly adjacent to the south-western border of that mapped by Th. Vogt. In describing the stratigraphy and structure of the area the author has made allowance for a different interpretation of some of the geological phenomena. A new stratigraphical division is put forward in the concluding chapters. Many questions, however, remain open. It would be desirable to continue the detailed and integrated investigation of the area under consideration, with particular emphasis on the composition and genesis of rocks, chiefly conglomerates, the sedimentary conditions and the newly found fossiliferous localities (their precise position is shown on aerial photos deposited in NGU, Trondheim). Only then it will be possible to make a more precise assessment of the stratigraphical position of individual members of the succession, and of the dating of volcanic activity and the extent and effects of the Caledonian deformation phases.

Acknowledgements.

I am greatly indebted to Norges geologiske undersøkelse for defraying the expenses of the investigation. My special thanks are due to Professor Chr. Oftedahl of the Technical University in Trondheim for all his assistance and interest in my work. I should like to express my special thanks to my colleague statsgeolog Fr. Chr. Wolff for his incessant help, collaboration and inspiring and friendly company during my stay in Norway. Finally, I cannot fail to mention all those Norwegian friends in NGU, Orkla Grube-Aktiebolag (Løkken Mine) and in the country who were always willing to help. The days spent among them will belong to the most cherished memories of my life.

Stratigraphy and rock characteristics.

The description of the rocks as given below is based on the stratal sequence presented in the stratigraphical column (see the map) and in Fig. 2. Although the author has endeavoured to use the generally accepted terminology,* in some cases the stratigraphical content and construction of formations and groups has had to be somewhat modified in order to fit the new geological interpretation.

* though with 'group' replacing 'series'.

Støren Group.

The volcanic complex of the Støren Group is the oldest member of the Early Palaeozoic succession occurring in the cores of the anticlinorial zones at the western and south-eastern margins of the mapped area. It is distinguished by the predominance of basic extrusive lavas accompanied by pyroclastics (agglomerates and keratophyric rocks), cherty beds and occasionally by sedimentary rocks (greyish-green or reddish slates). The lavas, which comprise massive and pillow varieties, were converted by low-grade metamorphism into typical greenstone of a characteristic mineral association: chlorite, epidote, actinolite, albite, calcite, quartz, etc.

The lack of precise palaeontological data makes the dating of the Støren Group difficult, but on correlation with other sectors of the Trondheim region it is assignable to the lowermost Ordovician, most probably to the Tremadocian.

Krokstad Group.

The term Krokstad Group is used to denote a relatively monotonous sequence of greenish slates and sandstones bearing intercalations of grit, conglomerate and breccia. Beds of grey to dark-grey slates occur in the lower part of the complex. This sedimentary sequence, roughly 400 m thick, overlies the older volcanic Støren Group, being locally separated from it by basal greenstone conglomerate. Sedimentation of the group was terminated by the extrusion of rhyolite lavas and by the intrusion of porphyrites.

In addition to the greyish-green colour, the flysch-like nature and abundant sedimentary structures (lamination, graded bedding, current and glide bedding) are characteristic features of the sedimentary rocks of the Krokstad Group. The greater part of the clastic material was undoubtedly derived from the Støren volcanic complex. The succession of rocks of the Krokstad Group is shown schematically in the stratigraphical column on the geological map and in Fig. 2.

Basal greenstone conglomerate.

The conglomerates immediately overlying the greenstone complex have been described from a number of localities in neighbouring areas, under different names: Venna (Stokvola) conglomerate (Th. Vogt 1945), Greenstone conglomerate (C. W. Carstens 1951, H. Carstens

1960) and Fjeldheim conglomerate (Chadwick et al. 1963). In the present area this lithology shows the character of a coarse-grained, unsorted, polymict conglomerate, composed mainly of greenstone, keratophyre, red jasper, sandstone, slate and limestone pebbles and other rock fragments. Pebbles are generally poorly rounded so that the rock has the appearance more of a volcanic breccia than a conglomerate.

Greyish-green slates.

These rocks constitute the greater part of the lower division of the Krokstad Group and are divided by a sequence of dark-grey slates. The development and appearance of the slates in both the lower and the upper parts is similar. Petrographically, they are very fine-grained sericite slates up to metasiltstones with well-developed cleavage planes, straight or strongly warped, often transverse to the original stratification. The latter is readily observable only where the slates bear intercalations of fine-grained sandstone alternating rapidly with thin slate beds. The greyish-green colour of slates is due to the increased proportion of chlorite and epidote.

Grey to dark-grey slates.

Grey slates up to several tens of metres in thickness occur in the lower part of the Krokstad Group; their transition into the underlying and overlying greyish-green slates is gradual. On account of the petrographical similarity and local inconspicuous change in shade, the two slate types are difficult to differentiate. The darker-grey or laminated varieties abounding in pyrite can easily be confused with the dark slates of the Lower Sandå Group described below.

Greyish-green sandstone and grit.

The upper part of the sedimentary complex of the Krokstad Group in represented by greyish-green, fine-to coarse-grained sandstone with interlayers of grit, conglomerate and breccia. This unit displays beautiful examples of graded bedding and also occasionally of current and glide bedding. The larger components of the sandstone and grit are quartz, quartzite, fine-grained slates, greenstone, keratophyre, and red jasper. Sericite, calcite, epidote, albite, chlorite, ore minerals and muscovite are common. The rocks can also be designated petrographically as lithic sandstone and greywacke (Fig. 3).

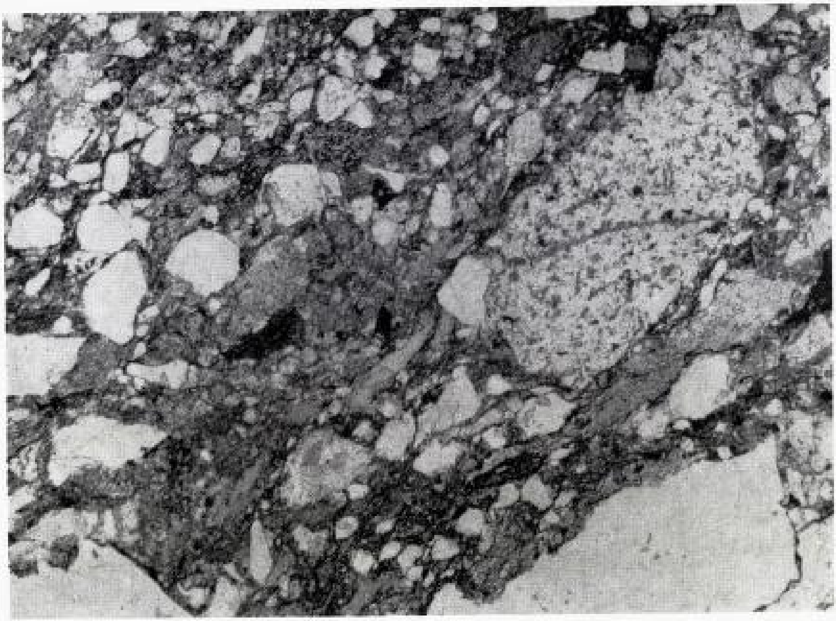

Fig. 3. Coarse-grained greyish-green sandstone of the Krokstad Group 3 km NNW of Kleivlykkja. x 22; parallel nicols.

Breccia.

Beds of fine-to coarse-grained breccia occur particularly in the lower part of the sandstone unit. Similarly as with the basal greenstone conglomerate, the breccia is composed of angular or sub-rounded fragments of greenstone, keratophyre, red jasper, quartzite, sandstone, slate and some limestone, embedded in a sandy matrix. The beds of breccia are not limited to one horizon; they usually form several layers separated by beds of greyish-green sandstone or greenish and reddish slate.

Amygdaloidal greenstone.

Sporadic amygdaloidal rocks occupy a separate position within the Krokstad Group. They constitute thin layers (up to several metres in thickness) within the greyish-green slates, near the upper boundary of the dark-grey slate unit.

The amygdaloidal structure of these greyish-green fine-grained rocks is striking (Fig. 4). Amygdales, 1–4 mm in mean size, are filled with calcite or an aggregate of calcite, albite and chlorite grains. The ground-

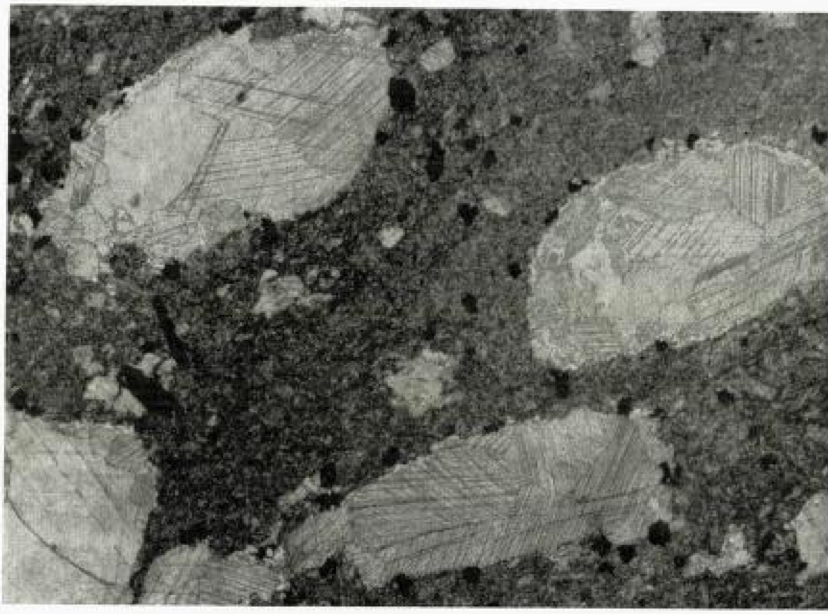

Fig. 4. Amygdaloid greenstone, Krokstad Group 1 km SE of Jonland. x 22; parallel nicols.

mass, with indications of ophitic texture, is formed of a mixture of thin laths of plagioclase-albite, chlorite and ore pigment. Quartz and apatite are usually present in accessory amounts. Regularly dispersed chlorite segregations seem to be pseudomorphs after primary mafic minerals, pyroxene or hornblende. Amygdaloidal greenstone presumably represents altered basic lavas which in their petrography are nearest to some types of the older Støren volcanic complex.

Hølonda porphyrites.

Weakly altered Hølonda porphyrites (porphyritic andesites) are found only in the northernmost part of the mapped area, and extend into the adjacent areas studied by Th. Vogt (1945), Chadwick et al. (1963) and Carter (1967). They are frequently represented by marked topographic ridges and summits of hills.

The porphyrites are composed of plagioclase and pyroxene phenocrysts enclosed in a groundmass of fine albite laths, chlorite, epidote, ore minerals etc. (Fig. 5). Plagioclases are partly or completely saussuritized

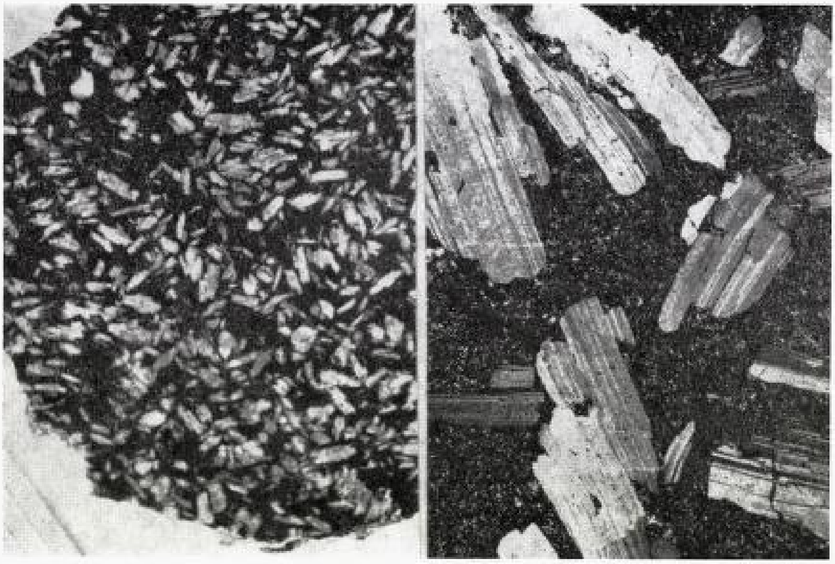

Fig. 5. Hølonda porphyrite. Road-cutting 2,5 km SSW of Hølonda church. Specimen of natural size on the left, thin-section of the same specimen on the right, x 15; crossed nicols.

and of albite composition. Pyroxene, which is colourless up to green is almost fully replaced by chlorite and uralite. Vogt (1945) differentiated two separate types of porphyrite; The Berg type, rich in plagioclase phenocrysts; and the Almås type, a little more basic and with more abundant phenocrysts of albite and altered pyroxene. Chadwick et al. (1963) observed the common occurrence of these two types within the same body.

It has not yet been decided indubitably whether the porphyrites are of intrusive or extrusive origin. Th. Vogt referred to them as lavas, and regarded the fragments of the Berg-type porphyrite in the neighbouring limestones as pyroclastic material from these lavas (Fig. 6). In contrast, C. W. Carstens (1951) considered the porphyrites to be intrusive discordant bodies. Chadwick et al. (1963) record the following criteria pointing to the intrusive character of porphyrites: lack of features distinctive of subaerial or submarine extrusive lavas, local strong baking of slates at the contact with the porphyrites, the formation of chilled and even glassy margins to the porphyrites, and occasionally the truncation of bedding in the sedimentary rocks.

Fig. 6. Fragments of porphyrite in limestone. Road-cutting, 2,3 km SSW of Hølonda church.

In the present area the porphyrites can be regarded as intrusive sill-like bodies, which were generally concordant with the bedding of the country rocks and subsequently deformed together with the sediments. Their occurrence is confined to a relatively narrow stratigraphical horizon, viz. the top part of the sandstone and slate unit of the Krokstad Group, at the contact with the overlying dark-slate and limestone beds of the Lower Sandå Group. In the vicinity of Hølonda, the limestones directly overlie the porphyrites and are not subjacent, as thought by Vogt (1935). From this it follows that the porphyrites occupy an analogous stratigraphical position to the rhyolites in the southern part of the area studied (described below). It is possible, however, that in neighbouring areas the porphyrites also occur at lower horizons in the Krokstad Group. It may be postulated that shallow intrusive porphyrite bodies were occasionally exposed during sedimentation and furnished detrital material for the sediments being deposited. Locally abundant fragments of porphyrites in the limestones in the vicinity of Hølonda are then interpreted as true pebbles and not as pyroclastic material (Fig. 6).

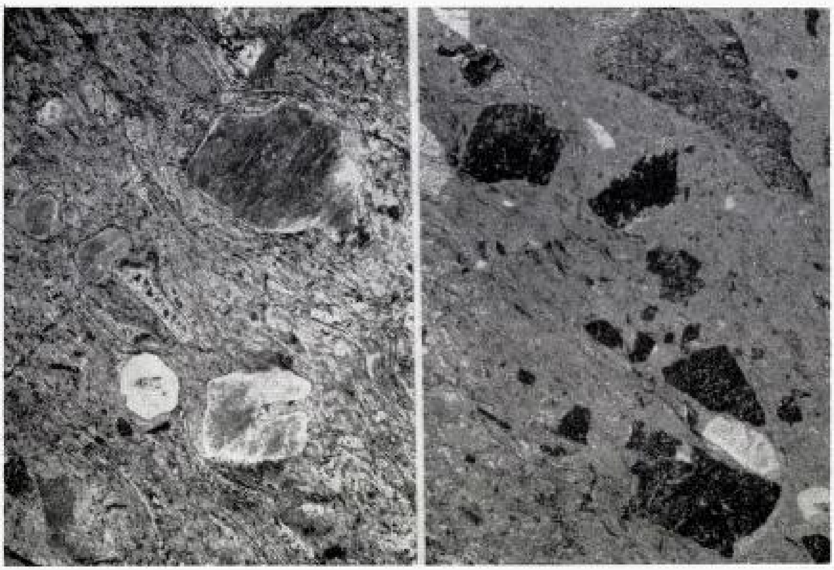

Fig. 7. Left: rhyolite showing fluidal (eutaxitic) texture. 1,5 km NW of Kleivlykkja. x 15; parallel nicols. Right: rhyolite with fragmentary texture. 3 km ENE of Evjeseter. x 22; parallel nicols.

Rhyolite and rhyolite tuff.

The rhyolites of the central part of the present area link up towards the west with two prominent bands of rock referred to by Vogt (1945) as Grimsås rhyolite and Hareklett massive rhyolite tuff. In his opinion these two rock-types are of different stratigraphical position and age. The present author, however, interprets the strips of rhyolitic rock as projections of one intensively folded layer which occupies a strictly definable stratigraphical position in the uppermost part of the Krokstad Group, directly below the basal conglomerate of the Lower Sandå Group. All rhyolites in the mapped area are of similar petrographical character.

The rhyolites are massive rocks, grey to greenish-grey in colour, without any definite traces of stratification. The groundmass is very fine-grained, composed of quartz, feldspar and sericite with an admixture of epidote, clinozoisite, sphene and magnetite, and contains quartz, albite and less frequently potassium-feldspar phenocrysts. Traces of fluidal texture and of younger recrystallization and silicification features are occasionally discernible in the groundmass (Fig. 7).

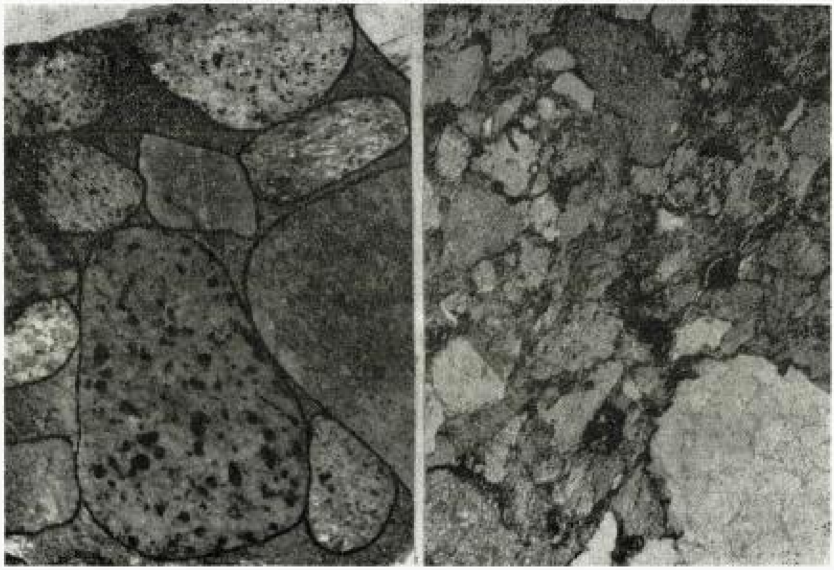

Fig. 8. Polymictic conglomerate, Lower Sandå Group. 1,5 km NW of Kleivlykkja. Specimen of natural size on the left, thin-section of the same specimen on the right. Abundant rhyolite fragments in groundmass. x 22; parallel nicols.

Types of fragmentary structure are common, this structure being particularly conspicuous on weathered rock surfaces. Rhyolite fragments are irregular, angular and several millimetres across. The abundance of fragmentary material and the structure of the rhyolites do not only provide evidence of their effusive character but suggest an origin as tuffs and agglomerates for at least a part of them. Because of folding and slight metamorphism the differentiation into lava and tuff layers is now almost impossible.

The age of the Krokstad Group can be estimated from the palaeontological finds recovered from equivalent horizons in adjacent areas. From the basal part of the Krokstad Group (the Fjeldheim Beds of Chadwick et al. 1963), Blake (1962) has described numerous graptolites of Middle Arenig age from the dark Bogo shale. Dark Bjørgen roofing slate bearing Upper Arenigian fauna (Størmer, 1932) also corresponds to the lower part of the Krokstad Group. Another significant fossiliferous formation is the limestone with intercalated dark-grey slate occurring in the more easterly Ven-Stenset area. These rocks were dated as either Upper Llanvirnian up to the Lower Llandeilian (Vogt 1945) or Llandeilian (Strand

Fig. 9. Polymictic conglomerate, Lower Sandå Group. 3 km ENE of Evjeseter.

1949): the limestones are analogous in stratigraphical position to the dark-grey slates in the lower part of the Krokstad Group in the present area. In the author's opinion, the entire Krokstad Group underlies the Lower Sandå Group, the Caradocian age of which has been confidently established. Consequently the sedimentary complex of the Krokstad Group, volcanics included, can be placed within the period embracing the Arenigian, Llanvirnian and Llandeilian.

Lower Sandå Group.

The Lower Sandå Group comprises about 250 m of dark slates with intercalations of grey fine-grained sandstone and locally of limestone. Polymictic conglomerates form the basal unit of the group.

Polymictic conglomerate.

Basal polymictic conglomerates are of only limited distribution. They overlie the greyish-green sandstones and slates, rhyolites and rhyolite tuffs, which are the youngest members of the Krokstad Group. The largest proportion of the pebble material is of light-coloured quartzite; fine- to medium-grained granitic to dioritic rocks, rhyolite, felsitic

Fig. 10. Dark banded slate, Lower Sandå Group. 750 km NW of Kleivlykkja. Flaggy parting on the left, splintery disintegration on the right.

porphyry, sandstone, slate and more rarely greenstone, limestone etc. are present in variable amounts. Pebbles are several centimetres across on the average, well worn, densely packed or scattered in a matrix of greywacke type. The matrix is occasionally composed almost exclusively of minute fragments of rhyolite rocks (Figs. 8, 9).

Dark slate.

Dark-grey to black slates constitute the main part of the Lower Sandå Group. They disintegrate into splinters mostly along false-cleavage surfaces (Fig. 10). Rusty weathered grains of pyrite and pyrrhotite occur in abundance. Locally a prominent banding or lamination is produced by the alternation of thin grey silty laminae or bands with black slate. The boundaries between the slates and the silty laminae are generally sharp on one side and diffuse on the other. As well at these indications of graded bedding, flame structures have also been observed. The banded slates are most common within the main occurrence of rhyolite and rhyolite tuffs in the central part of the area, where they contain in addition numerous layers of fine-grained *volcanic sandstone* (Figs. 11 and 12). These are up to several metres thick, dark-grey and partly

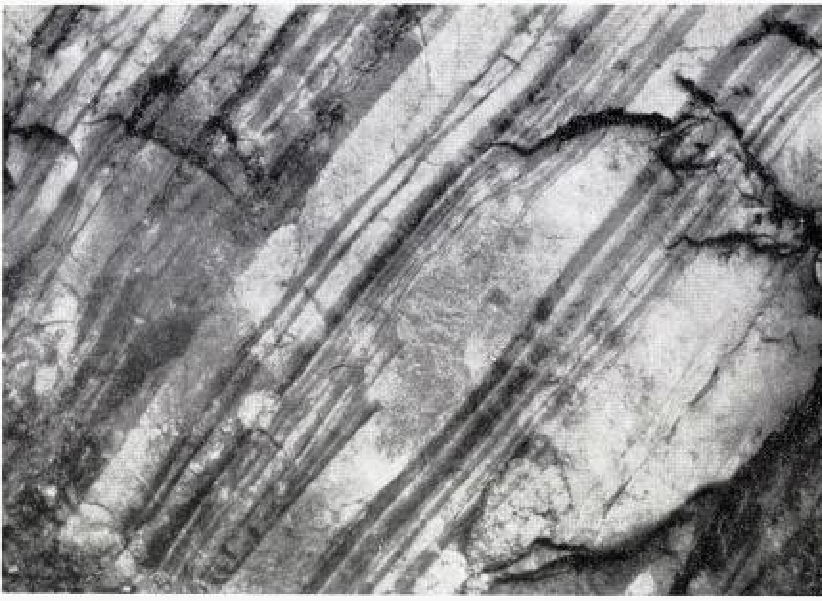

Fig. 11. Interbanded grey sandstone and dark slates, Lower Sandå Group; the sequence is here inverted as denoted by the prominent load cast near the centre of the picture. Graded bedding and current ripple lamination are also present. Road-cutting near Sjursmoen, west of Hovin. Photo, Dr. M. R. Wilson.

greenish in colour. The greenish-grey sandstones are difficult to distinguish from the strongly folded greyish-green sandstones of the older Krokstad Group, and it cannot be excluded that they were partly confused even during the mapping. The sandstones consist essentially of angular or more or less rounded quartz grains and felsite fragments which contain strongly decomposed plagioclase and resemble closely the underlying rhyolite rocks. Calcite, albite, epidote, fragmentary flakes of muscovite, chlorite, ore pigment, fragments of shale, sandstone, and rarely tourmaline and apatite occur in subordinate or accessory amounts. The admixture in dark banded slates shows a similar composition. Fragments of volcanics suggest that pyroclastic material is possibly present. It is, however, more plausible that it represents redeposited material of older volcanic rocks, viz. rhyolites and rhyolite tuffs.

Dark slates with sandstone intercalations make up a uniform unit, stratigraphically equivalent to the following rocks in the area mapped by Th. Vogt: the Dicranograptus shale, the Tømme black shale and

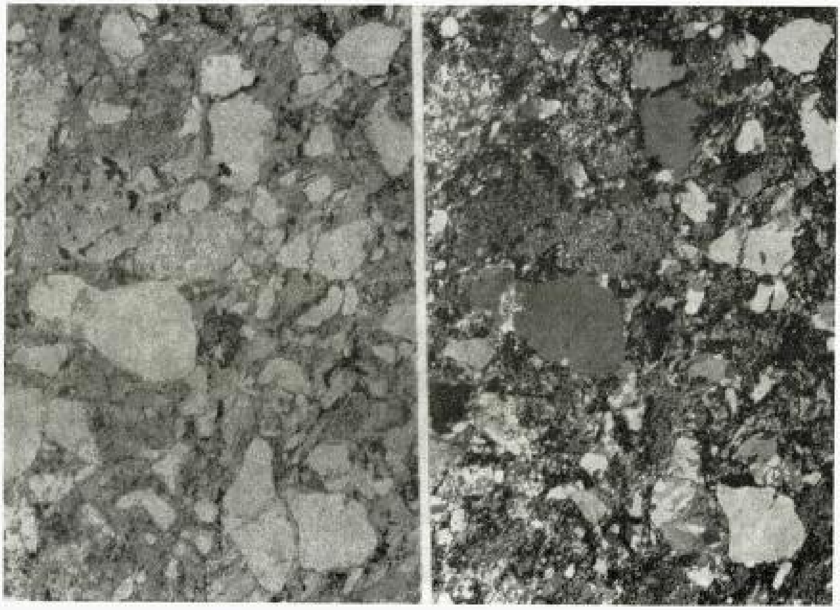

Fig. 12. Volcanic sandstone, Lower Sandå Group. 3 km W of Kleivlykkja. x 22; parallel nicols on the left, crossed nicols on the right.

mudstone, the Esphaug bedded rhyolite tuff, the Sandå shale with the associated Lundamo tuff (Oftedahl et al. 1969) and part of the Hølonda shale (Skjegstad shale). The first three of these formations have yielded fossils of Caradocian age. The author has also found poorly preserved graptolites in the dark slates, north of the Hulsjøen Lake. Dr. I Chlupáč of the Geological Survey, Prague, identified them as monoserial graptolites, but neither generic nor specific determination was possible: both a Silurian and an Ordovician age can be considered.

Limestone.

The Lower Sandå Group also contains limestones occurring in the lower part of the dark slates, west of Kleivlykkja and close to Hølonda. They are light- to dark-grey, generally fine-grained and partly of crystalline texture. In places they are nodular and contain intercalations of dark slates. Their mean true thickness is a few metres. West of Kleivlykkja fossils have been found at several localities in the limestones. The localities are plotted on the geological map and their precise positions

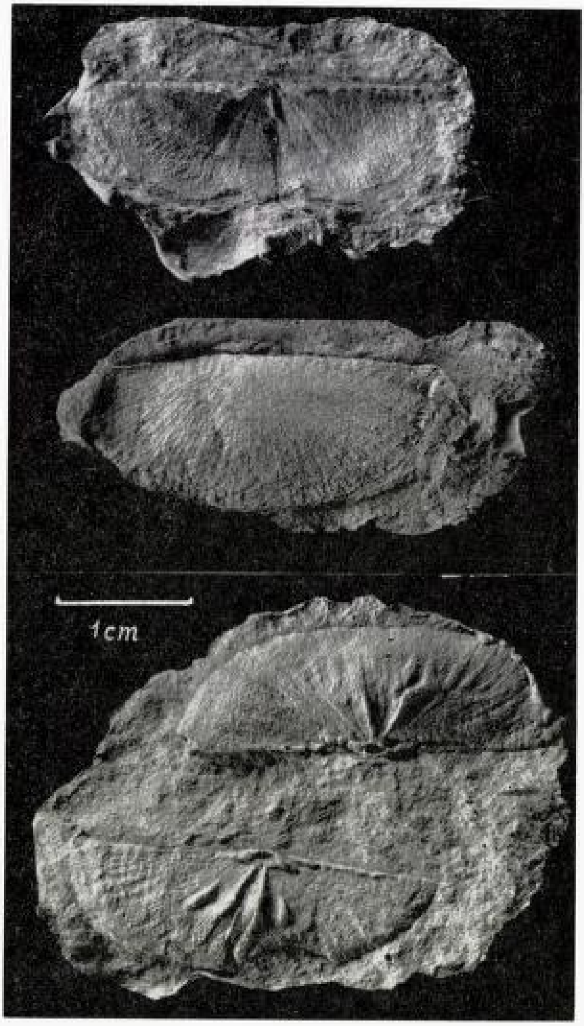

Fig. 13. Latex casts of brachiopods Sowerbyella (Sowerbyella) Jones, 1928, in limestone, Lower Sandå Group. 1,5 km ESE of Lundseter.

are shown on the air-photos deposited in the NGU, Trondheim. The best preserved fossils (brachiopods) were collected from the westernmost locality, ESE of Lundseter (Fig. 13).

According to Dr. V. Havlíček of the Geological Survey, Prague, the brachiopods belong to one species, classed undoubtedly in the Sowerbyellidae family. They can be assigned with a high degree of certainty to the genus and subgenus Sowerbyella (Sowerbyella) Jones, 1928, which is widely distributed throughout the world except for the Mediterranean province where it is lacking altogether. The stratigraphical range of this subgenus is considerable covering the Middle and Upper Ordovician (Llanvirnian to Ashgillian). From the Ordovician (Caradocian) of Norway it has been recorded, for example, by Spjældnes (1957) from the environs of Oslo.

As indicated by the reconnaissance mapping of the geologists of the Orkla Grube-Aktiebolag (the group working at the Løkken mine), the limestones described link up in the west with the Kalstad limestone, the fauna of which is probably of Caradocian up to Ashgillian age (Kjær, 1932). In an analogous stratigraphical position are the limestones from the vicinity of Hølonda which Vogt (1945) correlated with the limestones from Ven-Stenset yielding fossils of Upper Llanvirnian to Llandeilian age (Vogt, 1945, Strand, 1949). The fossiliferous limestones of the Ven-Stenset area, however, seem to be earlier, being the facies equivalent of the dark grey slates in the lower part of the Krokstad Group.

From the palaeontological finds it can be inferred that the entire Lower Sandå Group is most probably of Caradocian age, extending partly into the Ashgillian.

Upper Sandå Group.

The Hovin sandstone, beginning with basal conglomerates, is one of the youngest members of the Early Palaeozoic complex in the mapped area. It is preserved in the cores of the synclinal folds in the south.

Quartzite conglomerate.

The basal quartzite conglomerate, up to tens of metres in thickness, is a distinctive horizon separating the dark slates of the Lower Sandå Group from the overlying Hovin sandstone. It is composed mainly of well-worn, generally closely packed boulders of light-coloured quartzite, and subordinately of various volcanic and sedimentary rocks. (Fig. 14). The thickness of the conglomerate and its composition, size and proportions are inconstant. In lithology the rock often resembles the generally polymictic conglomerate from the base of the Lower Sandå Group. Where the conglomerates are not intimately associated with the typical Hovin sandstone, their precise stratigraphical position is difficult to

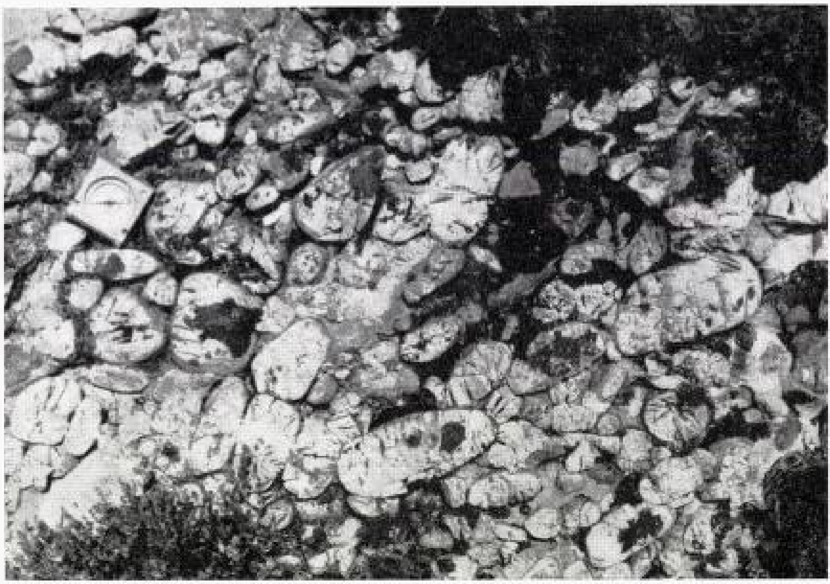

Fig. 14. Quartzite conglomerate, Upper Sandå Group. 2 km W of Kleivlykkja.

establish (e.g. north-west of the Hulsjøen Lake). Reliable criteria for the differentiation of the two conglomerates are obtainable only after a detailed petrographical examination.

Hovin sandstone.

The sandstones overlying the quartzite conglomerate are distinguished by the alternation of sandstone beds, several dm- to m-thick, with thin beds of dark-grey slate. Sandstones are grey and fine- to medium-grained. They contain an increased proportion of calcite and consist mostly of the angular quartz grains, fragments of slate, feldspar and scattered flakes of muscovite and/or decomposed biotite. In appearance the Hovin sandstone looks quite like the grey fine-grained sandstone intercalated in the dark slates of the Lower Sandå Group. The criteria enabling the Hovin sandstone to be distinguished from the older sandstone in the field are, in addition to its close connection with the quartzite conglomerate, its relatively strong weathering, low strength, typical slabby jointing and a fairly large admixture of macroscopic muscovite flakes.

From the Hovin sandstone no determinable fossils have so far been recovered. On the basis of the new stratigraphical table, the Hovin sandstone together with quartzite conglomerate represents the youngest Early Palaeozoic formation in the mapped area. It overlies the Lower Sandå Group of unquestionably Caradocian and possibly up to Ashgillian age, so that the Upper Sandå Group can quite likely be dated as Silurian.

History of the sedimentary complex and its stratigraphical correlation.

The major part of the Early Palaeozoic sedimentary complex overlying the Støren volcanic group is distinguished by a flysch type of sedimentation. As remarked by Chadwick et al. (1963) «evidence such as the great variety of rock types and rapid facial changes along the strike indicate that sediments were probably laid down in an unstable near-shore environment with a certain amount of pene-contemporaneous erosion in some places». Moreover, the products of magmatism – rhyolite and rhyolite tuff, amygdaloidal greenstone and intrusive porphyritic andesite – are not scarce.

The whole Early Palaeozoic complex has been divided into four separate groups differing in lithology and separated by conspicuous layers of syn-orogenic conglomerate (Fig. 15).

The oldest *Støren Group* (presumably of Tremadocian age) consists almost entirely of basic extrusive rocks and their pyroclasts; acid extrusives are present to a small extent.

The *Krokstad Group* (largely of Arenigian, Llanvirnian and Llandeilian age) is about 400 m thick. It is composed of basal greenstone conglomerate, slates and in the upper part of sandstones with intercalations of grit, conglomerate and breccia. The clastic material was derived mostly from the rocks of the Støren Group. The deposition of this group was terminated by intense volcanic activity (extrusions of rhyolite and rhyolite tuff) and by intrusions of porphyrite.

Fragments of these volcanic rocks appear in the basal polymict conglomerate of the *Lower Sandå Group* and form a substantial admixture in the volcanic sandstone interbedded in the younger dark slates of the same group. The Lower Sandå Group is decidedly of Caradocian and perhaps partly of Ashgillian age.

The youngest *Upper Sandå Group* (presumably of Silurian age) is of restricted distribution. The thick basal conglomerate contains well-sorted material composed dominantly of fragments of Precambrian rocks. The

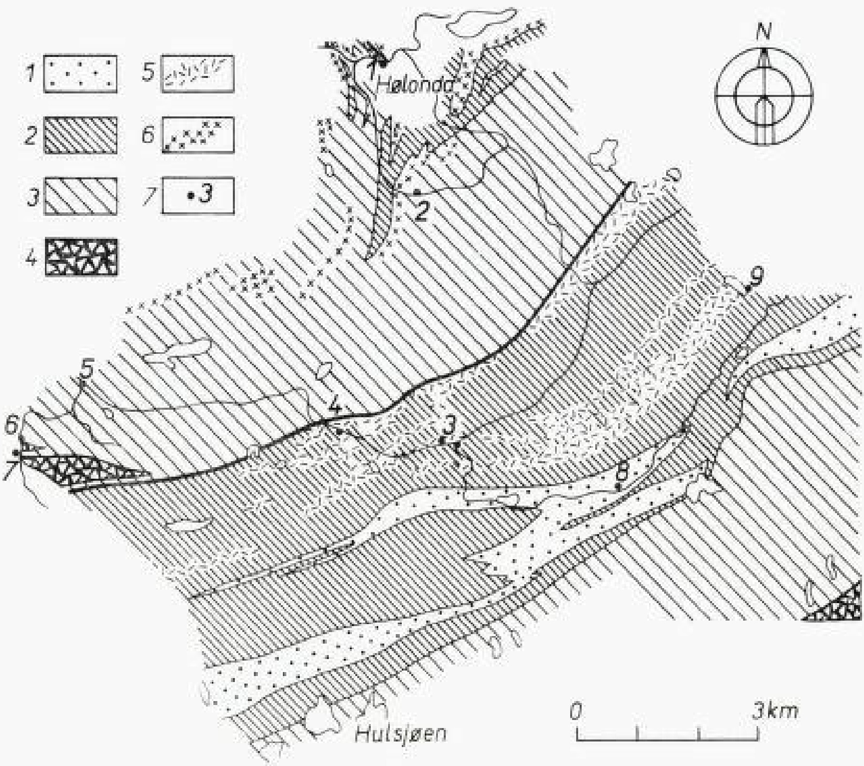

Fig. 15. Distribution of principal stratigraphical units in the Hølonda—Hulsjøen area. 1 – Upper Sandå Group; 2 – Lower Sandå Group, partly folded together with the upper part of the Krokstad Group; 3 – Krokstad Group; 4 – Støren Group; 5 – Rhyolite and rhyolite tuff; 6 – Hølonda porphyrite; 7 – localities on the excursion route.

distinctive nature of this series is made more pronounced by a moderate unconformity between its basal beds and the older formations.

The whole complex displays rapid *facies changes* and locally also striking changes in the thickness of individual beds and even of entire groups. The thickness of the Krokstad Group, for example, decreases markedly in the envelope of the Støren volcanic complex at the western margin of the mapped area. The abrupt change in thickness of the group, however, is not always primary, in some cases being of tectonic origin as evidenced by the dragging out or truncation of beds along thrust planes, increased intensity of folding pattern, or even by differences in the intensity of denudation preceeding the deposition of a new formation.

The unconformable nature of individual groups is generally well marked. The apparent unconformity at the contact of, for instance, slate and conglomerate beds, is invariably attributable to disharmonic folding of the more or less competent rocks. The moderate unconformity of regional scale (particularly of the Upper Sandå Group) is only inferable from an examination of the geological map.

The total thickness of the Ordovician-Silurian sequence overlying the Støren volcanic complex is about 800 m. Th. Vogt estimated the thickness of the same complex to be much greater, at about 2000—3000 m. This discrepancy in estimates is due primarily to the different concepts of the structure (see the chapter on structural geology).

As regards the facies development of the rocks, it should be noted that certain rock sequences show changes of lithology in the north-western and south-eastern parts of the area (Fig. 15). The most striking features are the following: characteristic of the central and south-eastern parts of the area is the predominance of acidic rhyolite volcanism in the upper part of the Krokstad Group, the abundance of volcanic sandstone in the Lower Sandå Group (with a considerable admixture of rhyolitic rocks) and also the limited distribution of the youngest Hovin sandstone formation, a possible result of a primary geographical restriction of the sedimentary area. In contrast, the northern part of the area is distinguished by the common occurrence of more basic intrusive porphyrite bodies, widespread carbonate sedimentation, and a lack of conglomerate beds of any great thickness.

Of interest is the *stratigraphical correlation* of beds with those from other parts of the Trondheim region, especially of the Meråker area, east of Trondheim, which has recently been studied by Fr. Chr. Wolff and his team of co-workers (1967). The correlation shows a good agreement in the general succession of strata, although there are some differences in the facies development and thickness of individual formations. Quite remarkable is the analogy in the lithological composition of the upper part of the Krokstad Group, i.e. of greyish-green slates and sandstones with grit, conglomerate and breccia intercalations, and of intrusive porphyrites with similar rocks (including gabbro-diorite sills), all constituting the Kjølhaugene Group. The lower part of the Krokstad Group, from the basal greenstone conglomerate up to and including amygdaloidal greenstone, can then be a stratigraphical equivalent of the Sulåmo Group of the Meråker area. The Caradocian and tentatively Silurian Lower and Upper Sandå Groups can possibly be correlated to a large extent with the youngest Slågån Group, in the dark slates of which Getz (1890) found graptolites of Llandoverian age (near Kjølhaugen). The differing sedimentary environment of the Meråker area is manifested particularly in the lack of prominent conglomerate horizons in the youngest parts of the Ordovic-Silurian sequence, the paucity of acid volcanics and the absence of marked unconformities.

In concluding, the problem of *diastrophism* will be discussed briefly. In Vogt's opinion (1945) the coarse conglomerates represent important erosional breaks. They have been distinguished as the Trondheim disturbance (Late Skiddavian), the Ekne disturbance (broadly between Caradocian and Ashgillian) and the Horg disturbance (Lower Llandoverian). The new stratigraphical positioning of the formations in the area studied necessitates some modifications in the dating of these orogenic disturbances.

The deposition of the coarse greenstone conglomerate above the Støren volcanic complex was undoubtedly a response to an orogenic movement of great intensity. This conglomerate is the first member of the sedimentary cycle of the Krokstad Group, the sediments of which contain abundant material eroded from this volcanic complex. On account of the palaentological finds, Blake (1962) placed this disturbance roughly at the Tremadocian/Arenigian boundary, although Skevington (1963) and Skevington and Sturt (1967) have challenged this and locate the disturbance actually within the Arenig.

The sudden movements of the land mass were also connected with strong magmatic activity, extrusions of rhyolite lava and tuffs, and probably also porphyrite intrusions; the deposition of coarse-grained sediments and polymict conglomerates in the uppermost parts of the Krokstad Group and at the base of the Lower Sandå Group, close to the Llandeillian/Caradocian boundary, can also be due to orogenic unrest.

The third conspicuous disturbance took place either in the latest Ordovician or earliest Silurian time. It may be postulated that a re-delimitation of the Early Palaeozoic sedimentary basin and the partial erosion of beds underlying the basal quartzite conglomerates of the Upper Sandå Group were associated with this disturbance.

A more precise dating of the extent and effects of the individual orogenic disturbances must await a detailed correlation of the area with other parts of the Norwegian Caledonides.

Remarks on structural geology and metamorphism.

The Early Palaeozoic complex of the present area occurs within the zone of minimum metamorphism in the Trondheim region, the grade of metamorphism not exceeding the lowest chlorite-muscovite subfacies of the greenschist facies. Pelitic sediments with well-developed cleavage surfaces which frequently transect the original bedding, have the character of slates or phyllite slates. Basic volcanics bear the typical mineral assemblages of greenstones. Sedimentary structures, clastic minerals and textures of magmatic rocks are still well preserved.

The various rock types and the slaty cleavage strike predominantly NE—SW to E—W, dipping moderately to steeply to the north-west or south-east. The boundary between the sectors showing opposing dips runs approximately across the middle of the area along the border between the Lower Sandå and the Krokstad Groups, at the northern margin of the northernmost rhyolite belt. This boundary seems to be an important tectonic line, a zone of steep dips, tightly packed folds and numerous thrust faults. It is possible that it may have been one of the paths of ascent for rhyolite extrusions and the site of the striking reduction in the thickness of the Krokstad Group, at least in the west.

The main geological structure of the area, which as yet has not been studied in detail, takes the form of large-scale tight folds overturned either to the SE or to the NW. Individual beds were at first mostly interpreted as separate stratigraphical horizons and the total thickness of the Early Palaeozoic formations above the Støren Group was then estimated at many hundreds of metres (in the area studied by Vogt, about 2000 to 3000 m). Later, on discovering that a number of lithologically identical units join up around tight anticlinal and synclinal closures, it became clear that the geological structure was more complex (Fig. 16). In the author's view the overall structure is interpreted as a series of closely packed tight to isoclinal folds accompanied by shearing and stretching phenomena, which are grouped into extensive anticlinorial and synclinorial belts. This style of folding has much in common with that described by Roberts (1967, 1969) for the north-western part of the Trondheim region. A more detailed analysis of tectonic elements and deformation phases is given in the papers of Carter (1967) and Rutter et al. (1967).

Aerial photographs show that the area is transected by numerous joints and faults, roughly perpendicular to the strike of the beds, but

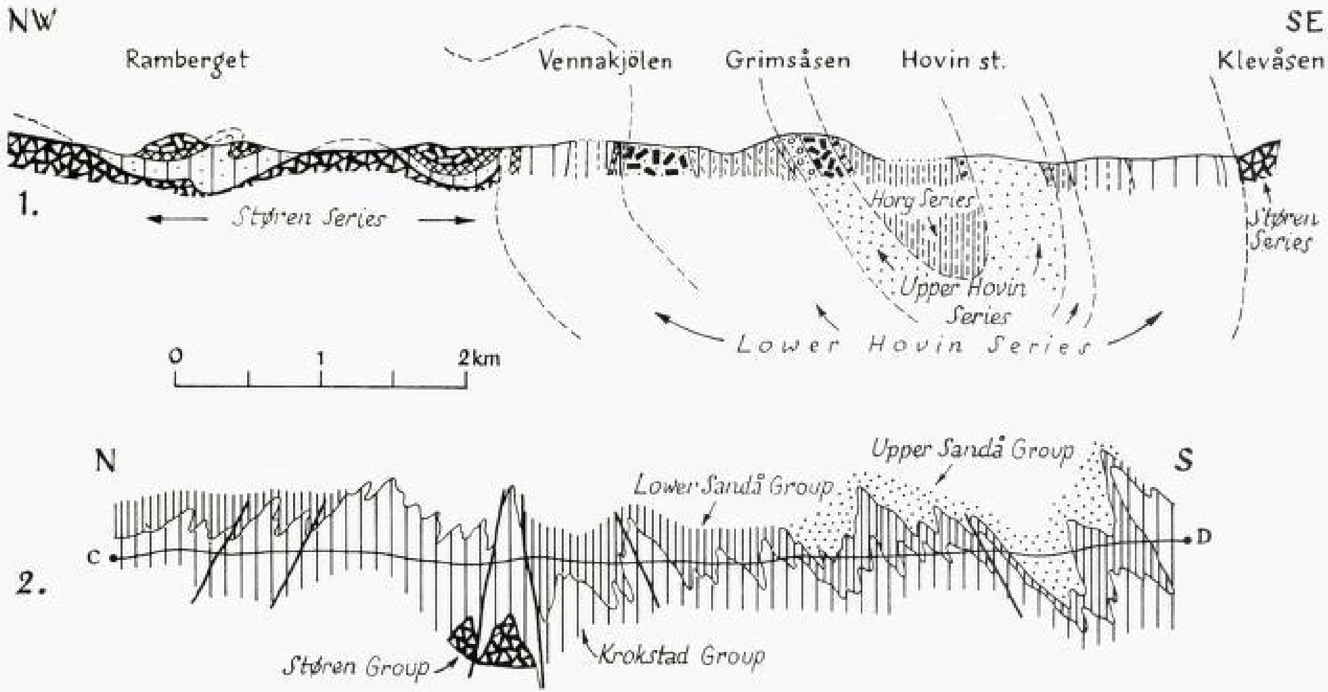

Fig. 16. 1: Geological section through the Hølonda–Horg area constructed from Th. Vogt's (1945) data. For explanation see the geological map. 2: Geological section across the adjacent Hølonda–Hulsjøen area, according to the author. For explanation see fig. 15.

which do not produce any marked lateral offset; these have not therefore been indicated on the map. The only clear-cut fault is located south of Hølonda. The abrupt termination of two strips of the Upper Sandå Group in the heavily drift-covered terrain 3,5 km east of Kleivlykkja can also be interpreted in terms of a fault or a transverse fold structure.

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Appendix.

One-day excursion to easily accessible localities in the principal formations of the mapped area.

Excursion route: Trondheim—Hølonda—Lundseter—Hovin—Trondheim.

1. A cutting at the sharp bend of the main road, 1 km WSW of Hølonda church. Partly exposed anticline with Hølonda porphyrite in its core. The envelope of the porphyrites is made up of limestone and younger dark slates of the Lower Sandå Group. Scattered porphyrite fragments are present in the limestones.

2. Road-cutting 2,3 km SSW of Hølonda church, in a slightly elevated terrain. Breccia from the lower part of the greyish-green sandstones (upper part of the Krokstad Group). The breccia, with characteristic fragments of red jasper, contains intercalations of greyish-green slate.

3. Road-cutting, 1,5 km NW of Kleivlykkja. Poorly exposed small outcrop, about 300 m WSW of the confluence of the stream flowing out from the Kleivlykkja Lake, with the Skolda river. Polymict conglomerate from the base of the Lower Sandå Group. About 20 m to the north, in a wood, the underlying rhyolite is exposed in outcrops and is also present as abundant boulders. About 100 m south of the confluence of the streams near a broken bridge, dark grey laminated slates of the Lower Sandå Group are exposed in the stream bank.

4. Road-cutting, 1,5 km north of Skara. In a ca. 100 m profile, strongly folded calcareous slates are exposed; these are interbedded with fossiliferous limestones high above the road.

5. Locality, 1,5 km north-west of Lundseter. In the river bottom close to the bridge, near the road junction, beds of greyish-green sandstone and grit interbedded with greyish-green slates are well exposed (upper part of the Krokstad Group).

6. Locality 1750 m west of Lundseter. In the bed of the stream, about 35 m north of the bridge along a side-road, the contact of the Støren greenstone with the basal beds of the Krokstad Group (greyish-green slates, and grey slates at the northern end of the profile) is exposed. At the contact there are several metres of greenish, locally reddish, chert and greenstone conglomerate.

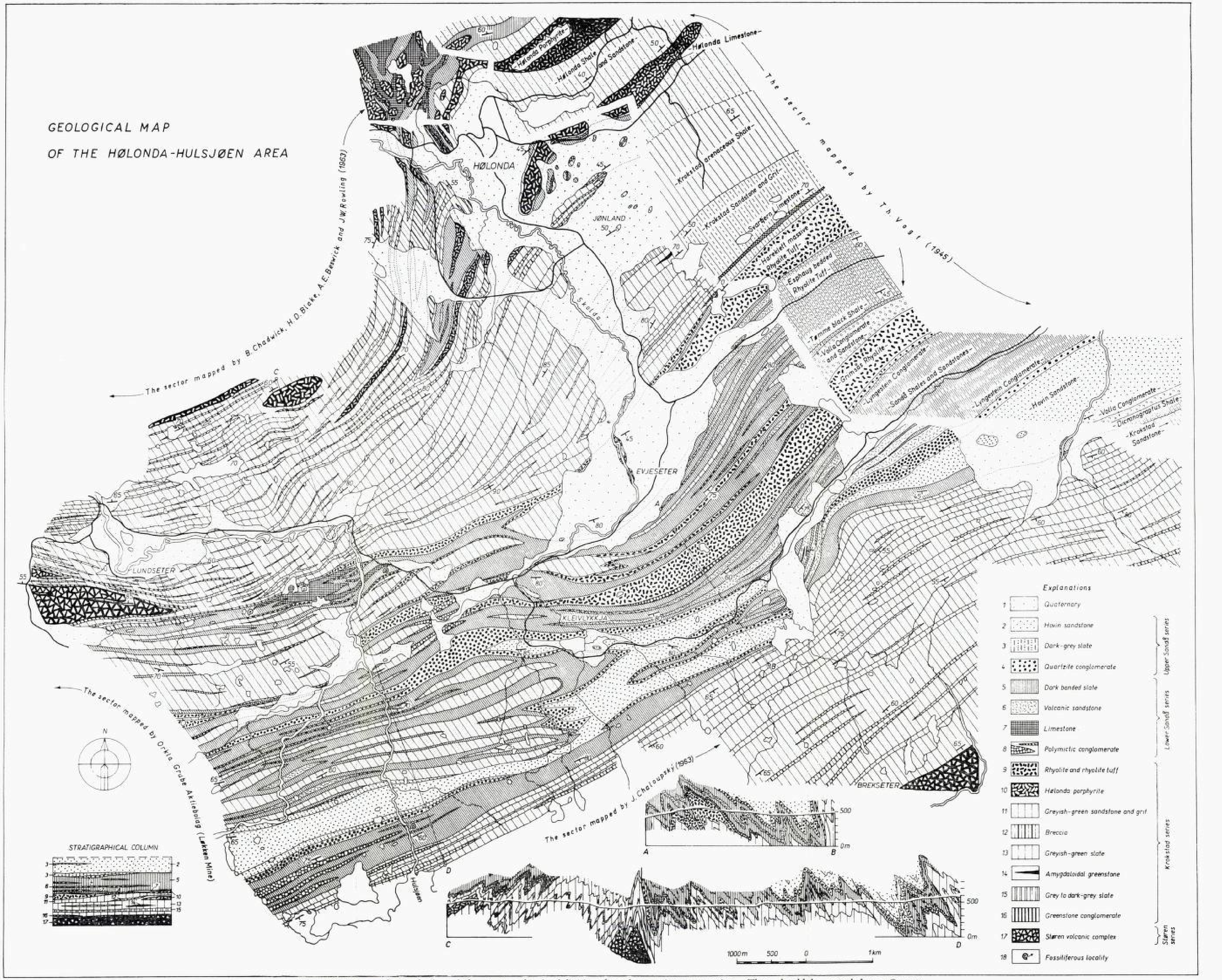
7. About 300 m to the north of locality 6, the first conspicuous outcrop above the road. Typical basal greenstone conglomerate (breccia) at the contact with the Støren greenstone; abundant fragments of greenstone and red jasper and also some limestone.

8. Roadside exposure approximately 1,7 km east of the Kleivlykkja Lake. The boundary of the quartzite conglomerate with the overlying Hovin sandstone, the latter

containing intercalations of dark slates (Upper Sandå Group). A number of exposures of quartzite conglomerate occur in the neighbourhood.

9. Supplementary locality. South-eastern shore of the Lake on the Grimsåsen ridge, 5 km north-east of Kleivlykkja. Basal polymict conglomerate of the Lower Sandå Group, directly overlying the rhyolites and rhyolite tuffs.

Manuscript received May 1970.



Note added in proof: In the legend the stratigraphical subdivisions have been written as «series». These should be amended to «Group».

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ÅRSBERETNING FOR 1969

NGU's administrasjon.

Ved kgl. res. 29. august 1969 ble direktør Egil Alnæs oppnevnt til styreformann ved Norges geologiske undersøkelse etter professor dr. Jens A. W. Bugge.

Styrets sammensetning i 1969 var følgende:

Professor Jens A. W. Bugge, Universitetet i Oslo, formann inntil 29/8 1969.
Direktør Egil Alnæs, Ila Jern A/S, formann fra 29/8 1969.
Direktør Olav Øverlie, Christiania Spigerverk, varaformann.
Professor Steinar Skjeseth, NLH, Ås.
Dosent Markvard Sellevoll, Universitetet i Bergen.
Administrerende direktør Karl Ingvaldsen, NGU, Trondheim.

Varamenn:

Direktør Leiv Løvold, Folldal Verk A/S, Folldal.
Professor Rolf Selmer-Olsen, NTH, Trondheim.

Det ble holdt i alt 4 styremøter.

Adm. direktør Karl Ingvaldsen har hatt den daglige ledelse av institusjonen. Forøvrig har ledelsen bestått av direktør dr. Harald Carstens ved Geologisk avdeling, direktør Inge Aalstad ved Geofysisk avdeling og direktør Aslak Kvalheim ved Kjemisk avdeling. Kontorsjef har vært Per Kr. Gundersen.

NGU gjentok i 1969 sitt forslag om å få opprettet et disposisjonsfond ved institusjonen, og for 1970 er det under kap. 943 Post 20 Spesielle undersøkelser bevilget kr. 125 000,—.

I forståelse med Industridepartementet fremmet NGU i brev av 10. februar 1969 forslag om et prosjekt for inventering av mineralske råstoffer i Nord-Norge. Overensstemmende med St.prp. nr. 161 for 1968—69 bevilget Stortinget 30. mai kr. 400 000,— for dette prosjekt under post 21 Undersøkelser i Nord-Norge. Forutsetningsvis skal denne plan gjennomføres i løpet av 5 år, og for 1970 er budsjettet kr. 500 000,—.

Planleggingen av det nye laboratoriebygg ved NGU er ført frem til skisseutkast ved siden av at det er utarbeidet en terrengmodell over Østmarkneset med påstående bebyggelse. Statens bygge- og eiendomsdirektorat har beregnet omkostningene for bygget til kr. 6 mill., og det regnes med en første byggebevilgning fra 1971.

Ved utgangen av 1969 hadde NGU ialt 148 stillinger, hvorav 127 fast organiserte og 21 helårsengasjementer. Økningen fra forrige år var 1 praktikant i helårsengasjement og fra 2. halvår 1969 helårsengasjementer for 1 konstruktør og 1 laborant til bearbeiding av materialet fra boringene på kontinentalsokkelen. Budsjettet for 1969 omfattet også omgjøring av 1 fast organisert stilling som vitenskapelig assistent til stilling som statsgeolog II. Se forøvrig egen personalliste.

Til særskilte prosjekter har NGU knyttet til seg engasjerte medarbeidere, og fra 1970 vil institusjonen ansette en egen gruppe medarbeidere for Nord-Norges prosjektet.

Budsjett og regnskap.

Statsbudsjettets kap. 3943

	Inntekter:		Budsjett		Regnskap
1.	Oppdragsinntekter	kr.	870 000,—	kr.	907 178,05
2.	Salg av kart og publikasjoner	»	40 000,—	»	108 881,17
3.	Salg av instrumenter	»	75 000,—	»	89 828,40
4.	Andre inntekter	»	15 000,—	»	5 543,80
		kr.	1 000 000,—	kr.	1 111 431,42

Statsbudsjettets kap. 943

Utgifter:		Budsjett		Regnskap
01. Lønninger	kr.	6 046 700,—	kr.	5 902 802,60
10. Kjøp av kontorutstyr	»	30 000,—	»	33 416,31
11. Kjøp av feltutstyr	»	105 000,—	»	104 752,92
12. Kjøp av instrumenter	»	252 000,—	»	251 662,25
13. Kjøp av maskiner	»	65 000,—	»	62 186,63
14. Gjenansk. av instrumenter ..	»	450 000,—	»	296 841,80
15. Vedlikehold	»	170 000,—	»	161 380,07
21. Unders. i Nord-Norge	»	400 000,—	»	167 140,64
29. Andre driftsutgifter				
291. Kontorutgifter	»	181 300,—	»	206 577,20
292. Trykningsutgifter ..	»	120 000,—	»	101 939,88
293. Bygningers drift	»	180 000,—	»	193 839,70
294. Reiser og forpleining ..	»	759 000,—	»	767 070,82
295. Forbruksvarer	»	423 000,—	»	422 855,12
296. Ymse driftsutgifter ..	»	607 500,—	»	595 105,83
	kr.	9 789 500,—	kr.	9 267 571,77
Hydrologisk dekade	kr.	350 984,83	kr.	202 795,26

Etter avvikling av A/S Vaddas Gruber er kr. 72 907,70 stilt til NGU's rådighet for bearbeidelse av Bidjovaggemateriale. Av disse midler gjensto pr. 31/12—69 kr. 16 581,66.

Berggrunnskart.

I 1969 har det vært arbeidet innen følgende kartblad i målestokk 1 : 250 000 (AMS 515).

Vadsø (NR 35, 36—5). Av kartbladets 8000 km² landareal er nå 4700 km² ferdig kartlagt.

Karasjokk (NR 35, 36—7). Kartbladet er ferdig kartlagt.

Narvik (NR 33, 34—1 og 2). Kartleggingen er kommet så langt at en regner med at kartet blir ferdig i løpet av ytterligere 2 sesonger.

Mo i Rana—Saltdal (NQ 33, 34—1 og 2). Dette kartbladet er foreløpig bare på planleggingsstadiet.

Mosjøen—Vega (NQ 31, 32—12 og NQ 33, 34—9). Av kartbladets 11 600 km² er nå ca. 9000 km² kartlagt.

Kristiansund (NP 31, 32—3). Av kartbladets 2530 km² er ca. 1300 km² kartlagt.

Trondheim—Østersund (NP 31, 32—4 og NP 33, 34—1). Av kartbladets 14 453 km² er nå ca. 9950 km² kartlagt.

Årdal (NP 31, 32—11). Av kartbladets ca. 17 000 km² er ca. 8300 km² kartlagt.

Bergen (NP 31, 32—14). Av kartbladets 5400 km² er ca. 5000 km² kartlagt.

Hamar (NP 31, 32—16). Av kartbladets ca. 8100 km² er 5310 km² kartlagt.

Sauda (NO 32—1). Av kartbladets 12 652 km² er ca. 7500 km² kartlagt.

Måløy (NP 31, 32—10) som omfatter 4900 km² er kartlagt, men vil først foreligge ferdig trykket i 1970.

Kvartærgeologiske kart.

Følgende kartblad i målestokk 1 : 250 000 er under arbeid:

Trondheim (AMS NP 31, 32—4) 3000 km² er kartlagt.
Stavanger (AMS NO 31—6). Kartleggingen er nesten fullført.
Bergen (AMS NP 31, 32—14). Kartleggingen er påbegynt.
Odda (AMS NP 31, 32—15) 1000 km² er kartlagt.
Landgeneralkart Jotunheimen. 15 000 km² er kartlagt.

Flymålinger og aerogeofysiske kart.

De magnetiske målinger fra fly ble utført med et 2 motors Piper Navajo fly leiet fra Nor-Fly A/S, Hønefoss og et nyanskaffet Varian proton magnetometer. Det ble fløyet i alt 430 timer og målt en samlet profillengde på 62 000 km.

Over land ble den systematiske dekning med profilavstand 500 meter fortsatt over et ca. 3000 km² stort område av Vest-Agder og Rogaland og et mindre område i Nord-Trøndelag (Fig. 1).

Over Smøla ble det utført en ekstra detaljert måling med profilavstand 200 meter.

Varangerhalvøya ble dekket med målinger med profilavstand 1 kilometer for den sydlige del og 2 kilometer for den nordlige del.

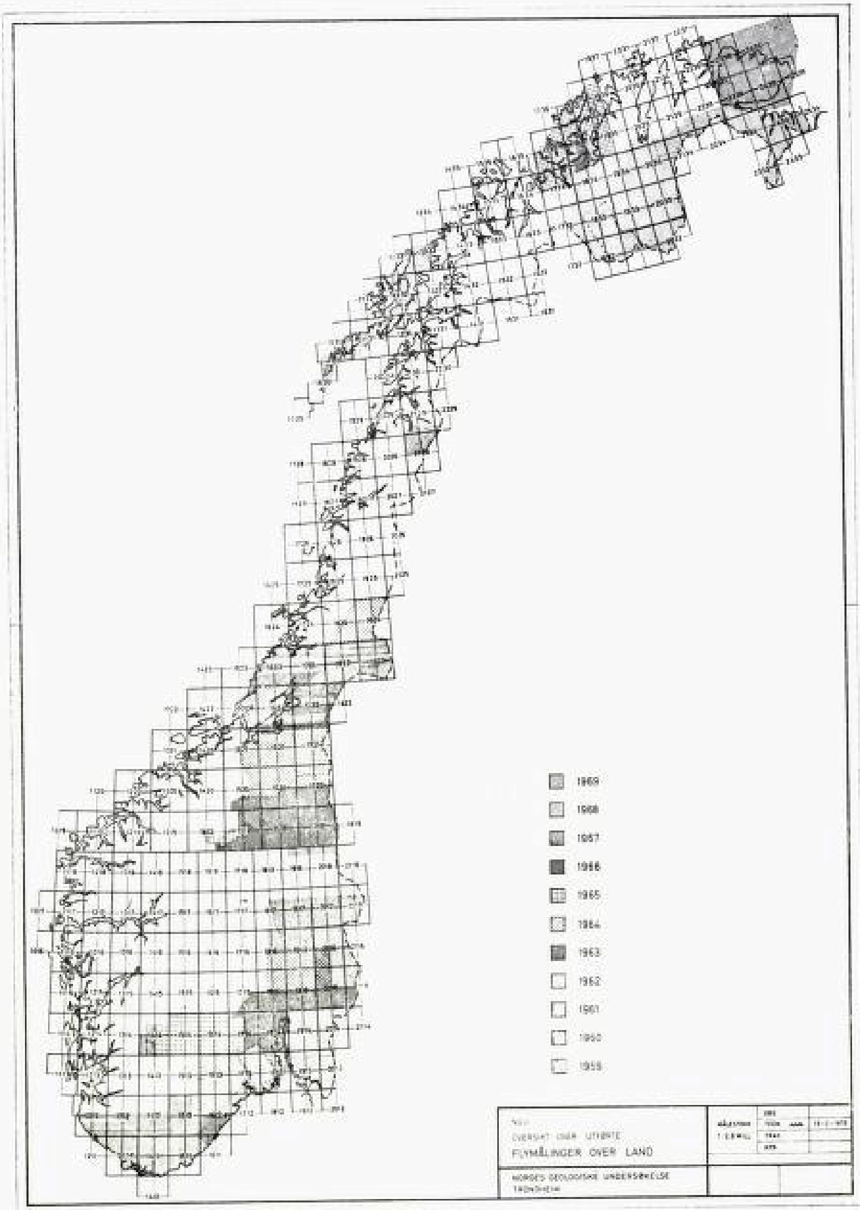
1969
1968
1967
1966
1965
1964
1963
1962
1961
1960
1959
OVERSIKT OVER UTFØRTE
FLYMÅLINGER OVER LAND
NORGES GEOLOGISKE UNDERSØKELSE
TRONDHEIM

Fig. 1.

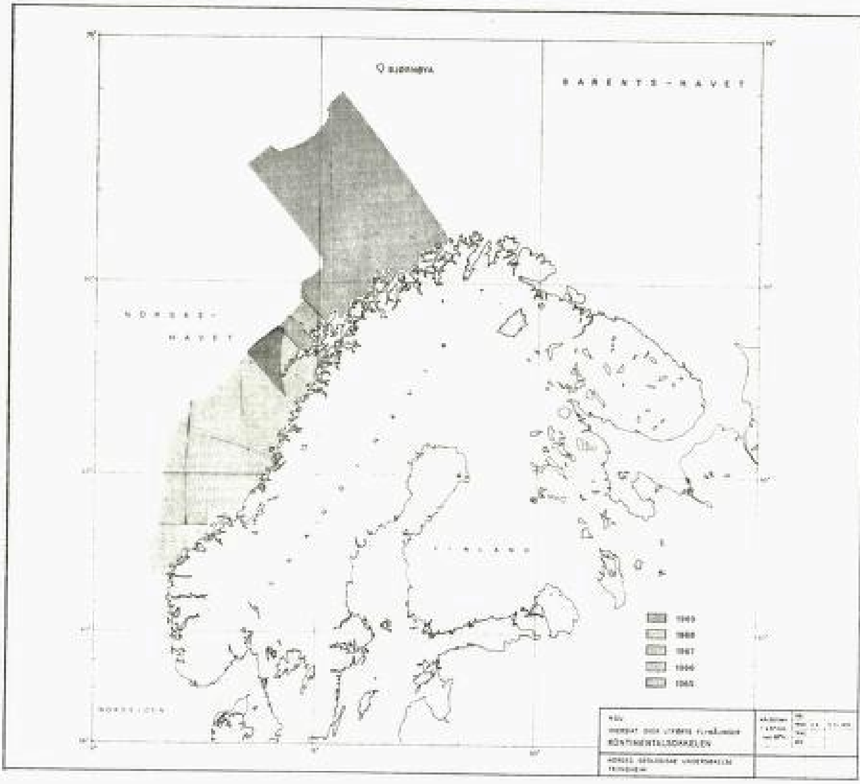

Fig. 2.

Som oppdrag for Sulitjelma Gruber A/S ble målinger utført i området Sulitjelma—Junkerdalen.

De magnetiske målinger av kontinentalsokkelen med økonomisk støtte fra NTNF ble fortsatt nordover fra Vesterålen og det ble målt ialt 33 000 kilometer profil (Fig. 2).

I løpet av året ble 36 aeromagnetiske kartblad i målestokk 1 : 50 000 offentliggjort og antallet slike kartblad er dermed kommet opp i 225.

Hydrogeologi.

Seksjonen har i 1969 vært beskjeftiget med oppdrag i forbindelse med grunnvannsforsyning ved boring i fjell og løsavleiringer, og det er utført en rekke befaringer i forbindelse med planleggingen av enkelt- og fellesvannforsyningsanlegg over hele landet. En har dessuten vært engasjert i langtidsplanlegging for grunnvannsforsyning innen større regioner.

De to rørbrønnene for Kongsvinger Vannverk ble i slutten av året satt i produksjon med et uttak på 2300 l/min., mens kapasiteten er vesentlig større. Vannverket forsynes nå bare med grunnvann.

Til fordel for annen hydrogeologisk virksomhet ble det i 1969 forsøkt å få en viss reduksjon i antall befaringer til småanlegg i områder der de hydroleologiske forhold er gunstige.

Seksjonens arbeid utover oppdragene har i første rekke konsentrert seg om oppgaver innen den Internasjonale Hydrologiske Dekade og vesentlig i Øvre Romerike. Kvartærgeologisk kartlegging innen den Internasjonale Hydrologiske Dekades felt på Filefjell er også startet.

Til brønnboringsarkivet kom det i 1969 inn opplysninger om 424 nye borebrønner, slik at arkivet nå omfatter i alt 7427 borebrønner.

Seismiske målinger.

Seismiske refraksjonsmålinger ble utført på en rekke mindre oppdrag. Problemstillingen har som regel vært å bestemme beliggenheten av sjiktgrenser i undergrunnen. Ved et oppdrag for Statens vegvesen i Nordland var forholdene noe spesielle. Det gjaldt her å finne fast fjell under en meget grovblokket steinur, og det lot til at metoden var meget vel anvendbar. Noen kontroll av resultatene foreligger imidlertid ikke. To oppdrag for Trondheim Elektrisitetsverk hadde noe større volum enn de øvrige. Begge gjaldt kraftverksplaner, det ene ved Jonsvannet og det andre i Rennebu.

Geofysiske bakkemålinger og borhullsmålinger.

For Folldal Verk A/S ble det utført elektromagnetiske målinger av en rekke felter i herredene Dovre, Folldal og Alvdal med et samlet areal på 36 km^2. Elektromagnetiske målinger i borhull ble utført for samme oppdragsgiver i Grimsdalen.

I Grongfeltet ble det foruten orienterende slingram og egenspenningsmålinger på en rekke steder, foretatt detaljerte elektromagnetiske og egenspenningsmålinger over et område ved Skåleseter—Jule i Sørli.

For Elkem A/S, Skorovas Gruber ble utført oppdrag med måling av indusert polarisasjon (IP) både i Meråker og i Skorovas.

Oppdrag med IP-målinger ble utført for Orkla Grube A/B over et par lokaliteter i Løkkenområdet. For samme oppdragsgiver ble det også utført elektromagnetiske borhullsmålinger i gruben på Løkken.

Som et ledd i NGU's kromprosjekt ble det gjort forsøk med IP-måling av et mindre område i Feragen.

Tyngdemålinger ble utført på en rekke punkter på Smøla og øyene utenfor som en del av et felles geologisk-geofysisk prosjekt med detaljert undersøkelse av dette område.

Geokjemisk prospektering.

Det er lagt ned et stort arbeid i planlegging av elektronisk databehandling av geokjemiske observasjonsdata. Det er igang samarbeid med Norsk Regnesentral i denne sak, og dette har resultert i utvikling av programmer som synes lovende. EDB-behandling vil være nødvendig p.g.a. den store mengde analysedata (i 1969 ca. 110 000 analytiske enkeltbestemmelser).

Fastfjellsgeokjemi er kommet igang. I forbindelse med NGU's Grongfelt-prosjekt har en studert anvendelsen av geofysikk på geokjemiske anomalier og en har fått resultater som synes å vise nye muligheter for å kunne skille mellom interessante og uinteressante anomalier. Kombinasjonen geokjemi/geofysikk har dermed fått betydelig øket aktualitet i malmletingen.

I NGU's blyprosjekt i det sydlige Norge har geokjemisk prospektering, både regionalt fra bil langs veier og mer detaljert, ført til funn av blymineraliseringer som er verd å undersøke nærmere.

Med en industribedrift har en et større geokjemisk samarbeidsprosjekt gående, og fra flere bergverksselskaper har en hatt geokjemiske prospekteringsoppdrag. I noen oppdrag har oppdragsgiveren selv utført feltarbeidet.

Ingeniørgeologi.

I oppdrag for Statskraftverkene ble det utført et omfattende ingeniørgeologisk kartleggingsarbeide i forbindelse med kraftutbyggingsplaner i Øst-Jotunheimen. En del tunneler ved Øvre- og Midtre Tesse kraftanlegg ble i denne forbindelse undersøkt.

Et annet oppdrag for Statskraftverkene ble utført i Ryfylkeheiene i forbindelse med utbyggingsplanene for Ulla—Førre vassdragene. Her ble det utført ingeniørgeologiske undersøkelser ved 7 forskjellige damsteder.

Som et ledd i den systematiske ingeniørgeologiske kartlegging av tunneler er det i oppdrag for Statskraftverkene kartlagt ca. 5 km av Breiskar-tunnelen ved Trollheim kraftanlegg.

Malmgeologiske undersøkelser.

NGU har intensivert de geologiske og geokjemiske undersøkelsene av blyforekomstene langs fjellranden i Syd-Norge. I denne forbindelse er det anmeldt 24 punkter på kartbladet Søndre Osen, 10 punkter på kartbladet Jordet og 15 punkter på kartbladet Gjøvik 1816 I.

En detaljundersøkelse er foretatt av Feragens kromittforekomster, og en rekke befaringer av serpentin-kromittlokaliteter i Syd-Norge har funnet sted.

Radiometriske anomalier registrert ved flymålinger er undersøkt i fylkene Hedmark og Sør-Trøndelag.

Rapport over undersøkelsene i Bidjovagge er utarbeidet og snart ferdig til trykking.

Mineralske råstoffer og bygningssten.

Det har her vært foretatt befaringer og undersøkelser av kvartsittforekomster på Våland i Holt, skifer i Oppdal og i eokambriske sparagmitter i Syd-Norge, undersøkelser for Leangen Fabrikker A/S (Elkem A/S) for å finne bergarter i Sør-Trøndelag som egner seg for steinullfremstilling og befaringer i områder i Hedmark og Sør-Trøndelag for å finne materiale egnet for faste veidekker.

Med støtte fra NTNF og Ferrosilisiumprodusentenes centralkontor har NGU arbeidet med problemet oppsprekking av kvarts. Videre har det vært utført et betydelig arbeid i forbindelse med Nord-Norge-prosjektet, se eget avsnitt.

Diamantboringer.

I løpet av året ble det diamantboret ca. 7000 meter fordelt på 9 forskjellige oppdrag. De to største oppdrag var for Folldal Verk A/S og har foregått i Tverrfjellet gruve og i Grimsdalen.

Laboratorier.

Spektrografisk laboratorium.

Med optiske spektrografer, kvantometer og røntgenfluorescens-apparatur har en fortsatt metodestudier og analyser som vesentlig gjelder bestemmelse av hovedbestanddeler og bibestanddeler i malmer, mineraler og bergarter. Det er anskaffet et nytt instrument Perkin-Elmer 303 for atomabsorpsjon som brukes av spektrograf-laboratoriet og de kjemiske analyselaboratorier i fellesskap. Det er lagt ned et stort arbeid for å tilpasse dette instrumentet til bergartsanalyser og liknende, ved metodeforskning som bl.a. omfatter oppslutningsmetoder og undersøkelse av reproduserbarhet og nøyaktighet.

Året har vært preget av utviklingsarbeid og mindre av serieanalyser i oppdrag.

Kjemiske analyselaboratorier.

Det har i året vært liten etterspørsel etter serieanalyser som kan utføres i rutine. Dette henger sammen med pause eller stopp i kunders diamantboringsprogram. Oppdragene har vært mer av uvanlig og tidkrevende art. Det er blitt tid til en del nødvendig utviklingsarbeid, dels vedrørende uvanlige bestemmelser, dels for å studere de nye muligheter som anskaffelsen av et instrument nr. 2 for atomabsorpsjon har gitt, og dels for å ta i bruk en ny analyseteknikk som er basert på jonespesifikke elektroder. Videre er det lagt ned en del arbeid på innføring av også andre metoder som er mer hensiktsmessige enn de som hittil er brukt. Det er også utført analyser i internasjonalt laboratorie-samarbeid vedrørende bergartsstandards og malmer.

Geokjemisk analyselaboratorium.

Analyseringen av geokjemiske prøver har foregått praktisk talt utelukkende med atomabsorpsjon. Metoder er etablert for mange elementer. Økningen i antall enkeltbestemmelser viser at rutinene går bra. Den store mengde analysedata (ca. 110 000) gjør direkte «tape-punching» av analyseresultater meget ønskelig for den videre tallbehandling.

Oversikt over analyseoppdrag:

1. Enkeltbestemmelser utført av geokjemisk laboratorium i forbindelse med geokjemisk seksjons prospekteringsoppgaver ca. 110 000.
2. Enkeltbestemmelser forøvrig 4477, som fordeler seg slik:

Utført av			Utført for			
Kjem.lab.	Spektr.lab.	Geokj.lab.	Kjem.avd.	Geof.avd.	Geol.avd.	Kunder
387	2880	1310	55	143	2699	1680

En meget stor del av analysene for Geologisk avdeling gjelder Grong-prosjektet.

Laboratorium for ildfaste og keramiske materialer.

Bestemmelse av varmeledningsevne, bråkjølingstall o.l. for varianter av forsteritt-stein har fortsatt, og det er funnet interessante sammenhenger mellom de målte verdier og brenningstemperaturen.

For Sjøfartsdirektoratet har en gjort de årlige undersøkelsene av norske malmkonsentrater med hensyn til stabilitet ved transport til sjøs som funksjon av konsentratets vanninnhold.

Forøvrig er det gjort fysikalske undersøkelser av forskjellige slag i forbindelse med institusjonens øvrige arbeid.

Metallurgisk laboratorium.

Det er gjort forsøk med galvaniske elementer av systemet kis/jern for å studere disses elektromotoriske kraft og evne til å gi strøm. I denne forbindelse forsøker en å studere årsaken til at magnetkis forbrenner i fuktig luft ved å bygge konsentrasjonsseller av kis og studere oksygenets og fuktighetens rolle i reaksjonen. Samtidig har en hatt forsøk gående for å studere lysets innvirkning på oksydasjonen av Fe(II) til Fe(III). Resultatene hittil tyder på at lys er nødvendig for reaksjonen.

Av interesse for geokjemisk prospektering er det utført diverse adsorpsjonsforsøk. Arbeidet over adsorpsjonen av kobber og sink på feltspat er trykt i årboken og en har fortsatt med en undersøkelse av adsorpsjon av Fe^{3+} ved forskjellig pH. Videre er det gjort forsøk med å skille den organiske fraksjon og mineralfraksjonen i humusprøver for å muliggjøre bestemmelse av tungmetallenes fordeling mellom disse to bestanddeler av normale humusprøver.

Geologisk avdelings kjemiske laboratorium.

I løpet av året er det ved dette laboratorium blitt utført 219 silikat-analyser og 381 andre analyser. Alle analysene er utført for institusjonens eget behov.

Annen kjemisk virksomhet som rensing og konsentrering av stoff, etsing av mineraler osv. blir det stadig større behov for, slik at det er brukt mye tid på den slags problemer.

Arbeidet med å samle og systematisere alle bergartanalyser fortsetter i den utstrekning det er tid til det.

Laboratoriet har drevet en del undersøkelser over svovelbestemmelser i kalkstein. Dette arbeidet er snart ferdigskrevet.

Røntgenlaboratoriet.

I løpet av 1969 er det tatt 708 opptak av mineraler.

Jordartslaboratoriet.

Ved laboratoriet er det utført 146 sprøhets- og flisighetsanalyser. Av disse er 33 utført for Geofysisk avdeling. Det er videre utført 441 kornfordelingsanalyser av sand og moreneavsetninger.

Mineralseparasjonslaboratoriet.

Ved laboratoriet har en i inneværende år separert en rekke prøver. Metodene har vært tunge væsker, magnetseparasjon, vaskebord og tørrristebord for glimmer.

Uranlaboratoriet.

Ved uranlaboratoriet er det i inneværende år blitt utført ca. 1200 uranbestemmelser.

Preparantverkstedet.

Verkstedet har i løpet av året laget 2300 tynnslip, 422 polerslip og 174 kombinasjonsslip. Videre har verkstedet utført en rekke poleringer av bergarter.

Verksteder og elektronisk laboratorium.

Verkstedet har som vanlig utført vedlikehold av kjøretøyer, maskiner, instrumenter og utstyr for hele institusjonen.

Måleutstyret for IP-målinger ble videre utviklet og et nytt sett bygget og anvendt ved årets målinger.

For salg ble blant annet fremstilt et sett slingramutstyr og en susceptibilitetsmåler. Av magnetometre ble solgt 53 stk., hvorav 52 utenlands.

Bergarkivet.

Bergarkivet inneholder 5581 rapporter, hvorav 3447 vedrører malmarkivet, og resten, 1133, mineralske råstoffer og bygningssten.

Tilveksten i 1969 var relativt stor, 387 rapporter, hvorav 316 rapporter vedrører malm og 71 mineralske råstoffer. I tillegg er 103 kart innkommet i arkivet i 1969.

Biblioteket.

Tilveksten av periodisk litteratur var 4382 bind, og samlet antall pr. 31/12 1969 er 39 326 bind.

Boktilveksten var på 1394 bind, og samlet bestand ved utgangen av året er 6225 bind.

I løpet av året er katalogisert 1064 boktitler med 1399 bind, og av periodika 352 titler med 4174 bind.

Pr. 31/12 1969 er det ialt katalogisert 6035 titler med 45 551 bind.

Det er sluttet 3 nye bytteforbindelser i 1969, og NGU har nå ialt 304 slike avtaler. Det var i 1969 abonnert på 124 tidsskrifter.

Undervisning.

Statsgeolog Magne Gustavson har undervist i petrografi (mikroskopering) ved NTH.

Statsgeolog Thor L. Sverdrup har i høstsemestret undervist i industrielle mineraler og bygningssten ved NTH.

Ved Jordskjelvstasjonen, Universitetet i Bergen har fysiker A. Breen forelest om elektromagnetiske målinger og direktør I. Aalstad om magnetometri.

Direktør I. Aalstad og geofysiker G. Hillestad har forelest i anvendt geofysikk ved NTH.

Geolog H. Barkey har undervist i fotogeologi ved NTH.

Geofysiker G. Hillestad har forelest i anvendt seismikk ved Danmarks Tekniske Høiskole i København.

Geokjemiker B. Bølviken har forelest i geokjemisk malmleting ved NTH som del av kurset Malmgeologi II.

Direktør A. Kvalheim har forelest spektroskopi ved NTH og NLHT med øvelser henlagt til NGU's laboratorium.

Utenlandske møter, kongresser og ekskursjoner.

Geolog S. Svinndal foretok en studiereise til Warren Spring Laboratory i England, 7.–9. januar, arrangert av Forskningsgruppe for Sjeldne Jordarter.

Adm. direktør K. Ingvaldsen og direktør I. Aalstad deltok i «International Oceanology Conference» i Brighton, England, 6.–21. februar. Dessuten besøkte de NAM's gassfelter i Groningen og Chevron Oil Co. i Haag. For Aalstad ble turen bekostet av NTNF's kontinentalsokkelkontor.

Vit.ass. I. Hultin deltok som NGU's representant ved Kongressen i malmmikroskopering i København i tiden 1.—4. mai.

Fysiker A. Breen og geofysikerne G. Sakshaug, A. Sindre og K. Åm deltok i det 31. EAEG-møte i Venezia 21.–23. mai.

Direktør A. Kvalheim og laboratorieingeniørene G. Faye og M. Ødegård deltok i Internasjonalt spektroskopisk kollokvium i Madrid 26.–30. mai.

Direktør A. Kvalheim deltok i IUPAC-konferanse i Cortina d'Ampezzo (betalt av IUPAC) 28. juni–3. juli, og besøkte på tilbakereisen Macaulay Institute i Aberdeen.

Statsgeolog dr. D. Roberts deltok i tiden 27. juli–6. august i en ekskursjon til de svenske Kaledonider (Jämtland, Västerbotten og Norrbotten) arrangert av Sveriges geologiska undersökning.

Direktør I. Aalstad og geofysikerne H. Håbrekke og K. Åm deltok i et tokt med forskningsfortøyet «Vema» tilhørende Lamont-Doherty Observatory (Columbia University) i Norskehavet i tiden 12.–19. september.

I tiden 6.–10. oktober deltok geolog S. Svinndal, geokjemiker B. Bølviken, vit.ass. A. Bjørlykke og statsgeolog T. Sverdrup i en ekskursjon til Bolidens gruber i Laisvall og Vassbo, Sverige.

Direktør I. Aalstad og geofysikerne H. Håbrekke og K. Åm deltok i «Colloquium on Basement Mapping by Magnetics» i Århus, Danmark, 27.–30. oktober, bekostet av NTNF's kontinentalsokkelkontor.

Geofysiker A. Sindre deltok i «Colloquium on Seismic Deep Sounding in Northern Europe» i Uppsala 1.–2. desember.

Innenlandske møter.

I et Nordisk geokjemikermøte som ble arrangert ved NGU 10.–11. november deltok direktør A. Kvalheim, geokjemikerne B. Bølviken og R. Sinding-Larsen og laboratorieingeniør J. R. Krog.

Spesielle prosjekter.

Nord-Norge-prosjektet.

Basert på arkivmateriale, publikasjoner, konferanser med bergmestre og befaringer i fylkene er det utarbeidet samlerapporter over hva som i dag er kjent innen Finnmark og Troms fylker. (Nordland fylke er under bearbeidelse).

Videre er opplysninger vedrørende bly i fjellranden samlet i en rapport.

NGU har i forbindelse med prosjektet mottatt ønsker om undersøkelser fra departement, utbyggingsavdelinger og fra en rekke privatpersoner. Det ble foretatt befaring av en rekke av disse lokaliteter.

I Troms fylke er det foretatt uranundersøkelser i Njallaavzze og i indre Kvænangen. Videre er et anomaliområde i Orrefjellet i Salangen undersøkt.

I Finnmark fylke ble store deler av det indre området prøvetatt med henblikk på materiale for faste veidekker. En del områder i ytre kyststrøk står fortsatt tilbake å undersøke.

Det er også foretatt en rekke befaringer av skiferfeltene i Troms og Finnmark for å vurdere de arbeidsoppgaver institusjonen vil bli stillet overfor i disse fylker.

Grongfeltet.

Gjennom en årrekke har NGU for statlige midler utført betydelige undersøkelser i Grongfeltet omfattende geologisk kartlegging, flyfotografering, aerogeofysiske målinger, geofysiske bakkemålinger og geokjemisk prospektering m. v., slik at det samlet foreligger et meget omfattende materiale. NGU har i 1969 tatt opp som et spesielt prosjekt bearbeiding av dette materialet kombinert med supplerende undersøkelser, slik at det kan utarbeides en så vidt mulig fullstendig oversikt over mulighetene for funn av ytterligere malm og dermed styrke grunnlaget for ytterligere malmleting i Grongfeltet. For gjennomføring av dette prosjekt er det engasjert 2 spesielle medarbeidere i tillegg til det som settes inn av NGU's ordinære kapasitet.

Samarbeidsutvalget.

Samarbeidsutvalget ved NGU hadde i 1969 denne sammensetning:
For administrasjonen:

Adm. direktør K. Ingvaldsen	Varamann	kt.sjef P. K. Gundersen
Direktør A. Kvalheim	»	lab.ing. G. Faye
Direktør I. Aalstad	»	geofysiker G. Hillestad
Direktør H. Carstens	»	statsgeol. T. L. Sverdrup

For de ansatte:

Statsgeolog Ø. Gvein Embetsmennenes Landsforbund	»	statsgeol. M. Gustavson
Geofysiker H. Håbrekke NIF's etatsgruppe v/ NGU	»	konstruktør K. Solem (NITO)
Sekretær C. Singsaas Statstjenestemannsforbundet	»	laborant R. Rossing
Tekniker J. Staw Norsk Tjenestemannslag	»	tegner T. Solvang

Som formann i utvalget fungerte direktør A. Kvalheim. 4 møter ble holdt, og det ble behandlet ialt 18 saker. De viktigste saker var budsjett, arbeidsprogram, byggesak og forskyvning av arbeidstiden.

NGU's informasjonsblad NGU-nytt utkom i 1969 med 1 nummer.

Publikasjoner.

Følgende artikler og avhandlinger er publisert av institusjonens medarbeidere:

Birkeland, Tore og Carstens, Harald: Decoration of Dislocations in Quartz. *NGU 258*, 372—374.

Bryhni, I., Bollingberg, H. J. og Graff, P.-R.: Eclogites in quartzofeldspathic gneisses of Nordfjord, West Norway. *NGT 49*, 193—225.

Carstens, Harald: Arrays of Dislocations Associated with Healed Fractures in Natural Quartz. *NGU 258*, 368—369.

— Dislocation Structures in Pyropes from Norwegian and Czech Garnet Peridotites. *Contr. Mineral and Petrol. 24*, 348—353.

— Dislocation free Quartz Crystals. *NGU 258*, 370—371.

Gustavson, Magne: The Caledonian Mountain Chain of the Southern Troms and Ofoten Areas. Part II, Caledonian Rocks of Igneous Origin. *NGU 261*, 1—110.

Hysingjord, Jens: Edel granat fra Otterøy ved Molde. *NGU 255*, 5—9.

Nilsen, R.: Adsorpsjon av kopper og sink på feltspat. *NGU 258*, 360–367.

Nissen, A. L.: A new Norwegian occurrence of scheelite. *NGU 258*, 116–123.

Osland, Lidvin M.: Olje og oljegeologi, oljeleting i Nordsjøen. *Naturen 1969, 7*.

Roberts, David: Trace Fossils from the Hovin Groups, Nord-Trøndelag, and their bathymetric significance. *NGU 258*, 228–236.

— Preferential veining in a conglomerate from the Trondheim region, Central Norway. *Geologische Rundschau, 58*.

— Deformation structures in the Hovin Group schists in the Hommelvik–Hell region; a discussion. *Tectonophysics 8 (2)*, 157–162.

— Lenticular and lenticular-like bedding in the Precambrian Telemark suite, Southern Norway; a comment. *N.G.T. 49*, 433–435.

Sverdrup, Thor L.: Mineralindustrien i Norge. *Bergverksnytt nr. 10*, 1968.

NGU's publikasjonsserie.

I 1969 er det ferdigtrykt og levert følgende publikasjonsnumre i NGU's serie: Nr. 258, 259, 260 og 261. Se forøvrig NGU's *publikasjonsliste* i NGU Nr. 258 som omfatter numrene 1 til 261.

Foredrag.

Under årsmøtet 1969 i Studieselskapet for Nord-Norsk Næringsliv holdt administrerende direktør Karl Ingvaldsen et foredrag i Nordreisa 9. juni med titelen: «Mineralske råstoffer i Nord-Norge – Kjente og ukjente aktiva i landsdelen.» Publisert i «Nord-Norge – Næringsliv og økonomi» Skrift nr. 38, oktober 1969.

Rapporter utarbeidet ved NGU i 1969.

Geologisk avdeling:

Mineraler, bergarter og grus.

Forfatter	Arbeidets art/arbeidssted	Oppdragsgiver	Oppdragsnr.
Gvein, Ø.	Kjerneboring i kvartsittskifer i Oppdal og Snåsa, Trøndelag	Stenkontoret	850
	Skiferundersøkelser i området vest for Trondheimsfjorden, S.-Trøndelag fylke	NGU	887
	Storelvdal nyttbare bergartsforekomster, Hedmark fylke	A. Mathiesen & Co. A/S	
	Kjerneboring i Sæterfjellet, Oppdal, Sør-Trøndelag	Oppdal Almenning	907
	Skifer i Rindal, Øyer, Oppland fylke		
Gustavson, M.	Herjangen distenforekomst, Ankenes, Nordland fylke		
Hultin, I.	Prosjektering av skråsynk Klefstadlykkja Talkgruver, Fron, Oppland fylke	Smedstad & Safa A/S, Kvam	863
	Diamantboringer i Hald kalkfelt Inderøy, Nord-Trøndelag	N. Buch, Trondheim	813 A
Hysingjord, J.	Fykanfjell beryllforekomst Glomfjord, Nordland fylke		
Nissen, A.	Sandnessjøen og Bardal distenforekomster, Nordland fylke		

Forfatter	Arbeidets art/arbeidssted	Oppdragsgiver	Oppdragsnr.
Rye, N.	Geologiske undersøkelser av grusforekomster ved Geilo, Buskerud fylke	Lars Rønsgård, Geilo	933
Sverdrup, T. L.	Geologiske undersøkelser av bergartsforekomster for steinullfremstilling, Sør- og Nord-Trøndelag	Leangen Fabr. A/S, Trondheim	
Sørensen, E.	Geologisk undersøkelse av anorthositt til vegformål, Rogaland fylke	Grubernes Sprengstoffabrikker A/S	873
	Geologisk undersøkelse av anorthosittfelter til vegformål, Sogn og Fjordane fylke	NGU	880
	Malmer		
Gust, J.	Egenpotensialmålinger i Tingstad Grube, Levanger, Nord-Trøndelag	NGU	
Hovland, R.	Lesja jernmalmfelt, Oppland fylke	NGU	
	Strømme	NGU	
	Notat om Eikeland Jernverks gruber, Aust-Agder fylke	NGU	
	Notat om Gaustadfeltet, Bø, Nordland fylke	NGU	
	Notat om Finberger ilmenitforekomst Øksnes, Nordland fylke	NGU	
Reite, A.	Undersøkelse av is og blokktransport i Joma-området, Røyrvik, N.-Trøndelag	NGU	

Geofysisk avdeling:

Forfatter	Arbeidets art/arbeidssted	Oppdragsgiver	Oppdragsnr.
Barkey, H.	Geologiske og ingeniørgeologiske undersøkelser i Vest-Jotunheimen	NVE	817 A
	Div. feltbefaringer i forb. med seismiske grunnundersøkelser i Vest-Jotunheimen	NVE	817 B
	Div. befaringer og ing.geologiske undersøkelser i forbindelse med fyllingsmasser for damstedene ved Raudalsvann, Styggvann og Høydalsvann i V.-Jotunheimen	NVE	817 C
	Geologiske undersøkelser Småengtjønna, Stoksund i Åfjord	Åfjord kommune	904
	Geologisk befaring av påtenkt magasin for Vanviken Vassverk	Leksvik kommune	909
Breen, A.	Magnetiske borhullsmålinger, Rana	A/S Norsk Jernverk, Rana Gruber	858
	Magn. borhullsmålinger Tverrfjellgruva, Dovre	Folldal Verk A/S	892
Eidsvig, P.	IP-målinger Storhusmannsberget, Meråker	Elektrokemisk A/S, Skorovas Gruber	839
	IP-målinger Løkken	Orkla Grube-AB	854/862
	Måling av spesifikk motstand på prøver fra Feragen kromfelt	NGU	883
	Beregning av dyp til loddrettstående plate på grunnlag av IP-målinger med eksponderende elektrodesystem	NGU	889

Forfatter	Arbeidets art/arbeidssted	Oppdragsgiver	Oppdragsnr.
Hillestad, G.	Seismiske unders. over grusforekomster Berger–Asak, Frogner i Sørum	NGU	801
	Seismiske målinger Stokkan–Dragvoll, Trondheim	Siv.ingeniør Kummeneje, Trondheim	827
	Seismiske unders. Sandnessundet, Tromsø	Sandnessundforbindelsen A/S	832
	Seismiske undersøkelser Kløfta	NGI	846
	Seismiske undersøkelser Oppdal (Driva kraftanlegg)	Sør-Trøndelag Elektrisitetsverk	852
	Seismiske undersøkelser Torsetsund, Aure	Torsetsund bruselskap A/L	857
	Seismiske unders. Ytre Bangsjø, Snåsa	Nord-Trøndelag Elektrisitetsverk	860
	Seismiske undersøkelser Eidsøra, Nesset	Siv.ingeniør Kummeneje, Trondheim	864
	Seismiske unders. Nidelv bro, Brattøra, Trondheim	NSB	865
	Seismiske unders. Dalseiddalen, Vaksdal	NSB	868
	Seismiske undersøkelser Kambo, Moss	NSB	903
	Seismiske unders. Nupen i Skjerstad	Statens vegvesen	932
Melleby, P.	Seismiske undersøkelser Vikanholmen–Dalviknesset, Kristiansund N.	Werring & Søn A/S	891
Sakshaug, G. F.	Geofys. unders. Villumelv–Corisvann–Stålhaugen Gr.–Storforsdalen, Saltdalen	A/S Sulitjelma Gruber	830
Schröder, M.	An aerial photogeological interpretation and field investigations in an area N of Lillehammer	NGU	893 A
Sindre, A.	IP-målinger Luostegaissa, A/S Sydvaranger, Porsanger		840

Forfatter	Arbeidets art/arbeidssted	Oppdragsgiver	Oppdragsnr.
Sindre, A.	IP-målinger St. Knuts Grube, Alvdal og Stormyra i Grisungdalen, Dovre	Folldal Verk A/S	847
	Seismiske målinger Løkken	Orkla Grube-AB	937
Singsaas, P.	E. m. målinger Njallavarre kobberskjerp, Ballangen	A/S Bjørkaasen Gruber	848
	E. m. borhullsmålinger Nordgrube- og Storwartsområdet, Røros	A/S Røros Kobberverk	851
	E. m. borh.målinger Tverrfjellet, Dovre	Folldal Verk A/S	855
	E. m. borhullsmålinger Gjertrudfjell, Sulitjelma	A/S Sulitjelma Gruber	870
Svinndal, S.	Geol. kartlegging Ulla—Førre-anleggene	NVE	818 A
	Undersøkelser etter sjeldne jordartselementer i Fensfeltet, Ulefoss, sept.—des.	FSJ	820 B
	Kjerneboring i Frankipel Ila pir, Tr.heim	A/S Trondheim Kornsilo	894
Tan, T. H.	Undersøkelser i forb. med kobbermineraliseringen i Ucca Vuovdas, Kautokeino	NGU	510 A
	—»—	NGU	548 D
	Oppfølging av elektromagn. flyindikasj. Caskias grønnsteinsområde, Kautokeino	NGU	756 A
	Moreneundersøkelser (mikroblokkleting) rundt kobbermineraliseringen ved Ucca Vuovdas, Kautokeino	NGU	756 B
	Kjemisk avdeling:		
Bølviken, B.	Geokjemisk metodestudium Laksådal/Oterstrand	A/S Sulitjelma Gruber	837
Krog, R.	Geokjemiske unders. Haustasjøen V, Kvikne VI	A/S Folldal Verk	829

Personale.

Ansettelser i 1969.

Administrasjonskontoret:
Aune, Dagny, kontorfullmektig II, 25. august

Geologisk avdeling:
x) Kildal, Ellen S., statsgeolog II, 1. januar
Petersen, Sigrid, sekretær I, 6. juni
Furuhaug, Leif, laborant I, 1. april
Haugum, Ellen, tegner II, 1. april
Askvik, Helge, statsgeolog II, 1. juli
Torske, Tore, statsgeolog II, 1. juli
x) Kirkhusmo, Lars Anders, statsgeolog II, 1. august
Rustung, Bodil, kontorfullmektig II, 15. september

Geofysisk avdeling:
Åm, Knut, geofysiker II, 4. januar
Åsheim, Arnt, laborant II, 6. januar
Brevik, Bjørn, mekanikerformann, 1. mai

Kjemisk avdeling:
Kuldvere, Arnold, laboratorieingeniør II, 1. februar
Berge, Frank Gabriel, laboratorieassistent I, 1. mars
Thoresen, Hans Henry, laboratorieassistent I, 1. juli
Sinding-Larsen, Richard, geokjemiker II, 1. oktober
Søberg, Baard, laboratorieassistent I, 1. desember

Avskjed i 1969.

Administrasjonskontoret:
Karlsen, Inger, kontorassistent I, 16. august

Geologisk avdeling:
Hovland, Roar, vitenskapelig assistent I, 16. april
Broch, Olaf Anton, statsgeolog I, 30. april
Kulvik, Kari, kontorassistent I, 13. juli

x) Disse personer er gått over fra annen stilling eller engasjement ved institusjonen.

Kjemisk avdeling:

Skarholt, Siri, laboratorieassistent I, døde 9. februar
Hvatum, Ole Øivind, geokjemiker II
Berner, Beate, konstruktør I, 20. oktober
Tan, Brith Helen, laboratorieassistent I, 30. november

Permisjoner i 1969.

Hagemann, Fredrik, statsgeolog I, perm. u/lønn fra 1/1—31/12—69

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Ansatte pr. 31/12—1969.

Den oppførte ansettelsesdato angir tidspunktet da vedkommende ble knyttet til NGU.

Administrasjonskontoret:

Adm. direktør:	Ingvaldsen, Karl, siv.ing., a. 1. januar 1958
Bergingeniør:	Welde, Harald, siv.ing., a. 1. januar 1965
Kontorsjef:	Gundersen, Per Kristian, c.j., a. 1. oktober 1960
Forvalter:	Thorvaldsen, Arvid, a. 1. juli 1956
Bibliotekar:	Ryssdal, Marit, a. 1. oktober 1963
Fotograf:	Aamo, Ingemar, a. 1. august 1962
Regnskapsfører:	Sandvold, Morten, a. 1. juni 1968
Kasserer:	Aursand, Marit, a. 1. august 1965
Kontor-fullmektig I:	Ristan, Anne Margrethe, a. 1. mai 1961
Kontorassistent/fullmektig II:	Aune, Dagny, a. 25. august 1969
	Bakken, Solveig, a. 12. mai 1966
	Odde, Lise, a. 19. juni 1967
	Haugan, Anne Margrethe, a. 1. sept. 1968 midl.
Vakt- og varmemester:	Wold, Jostein, a. 15. august 1961

Geologisk avdeling:

Direktør:	Carstens, Harald, dr. philos., a. 1. juli 1968 (2. mars 1955)
Statsgeolog I:	Holmsen, Per, c.r., a. 1. juli 1939
	Hagemann, Fredrik, c.r., a. 1. mars 1957

	Sverdrup, Thor Lorck, c.r., a. 16. november 1958
	Bryn, Knut Ørn, c.r., a. 1. januar 1959
	Wolff, Fredrik Christian, c.r., a. 16. februar 1960
	Gustavson, Magne, c.r., a. 1. januar 1961
	Roberts, David, Ph. D., a. 1. januar 1965
Statsgeolog II:	Skålvoll, Harald, c.r., a. 1. juli 1957
	Thorkildsen, Christian Dick, c.r., a. 1. febr. 1960
	Hysingjord, Jens, c.r., a. 15. august 1961
	Gvein, Øivind, c.r., a. 11. desember 1963
	Reite, Arne, c.r., a. 5. april 1965
	Rye, Noralf, c.r., a. 1. juli 1965
	Siedlecka, Anna, dr.ing., a. 1. mai 1966
	Kildal, Ellen Sigmond, c.r., a. 1. oktober 1965
	Askvik, Helge, c.r., a. 1. juli 1969
	Torske, Tore, c.r., a. 1. juli 1969
	Kirkhusmo, Lars Anders, c.r., a. 16. mai 1965
	Nissen, August L., c.r., a. 1. januar 1964
Midlertidig statsgeolog:	Poulsen, Arthur O., cand min.,
Vitenskapelig assistent:	Hultin, Ivar, c.r., a. 1. august 1966
	Bjørlykke, Arne, siv.ing., a. 15. juni 1968
	Osland, Lidvin M., c.m., a. 1. februar 1968, midl.
Laboratorie-ingeniør I:	Graff, Per-Reidar, c.r., a. 1. april 1964
Konstruktør II:	Klemetsrud, Harald Tiedemann, a. 1. juli 1957
	Sørensen, Erling, a. 1. mai 1963
Teknisk assistent I:	Hatling, Harald, a. 1. februar 1961
	Gust, Johan, a. 1. oktober 1962
	Røste, Johannes Rye, a. 9. desember 1963
Preparant I:	Jacobsen, Tom, a. 1. mai 1962
	Iversen, Egil, a. 1. august 1965
Laborant I:	Bøhle, Anne-Elise, a. 4. mars 1968
	Furuhaug, Oddvar, a. 16. april 1968
	Furuhaug, Leif, a. 1. april 1969
Tegner I:	Vikholt, Hallfrid, a. 1. mars 1955
	Torjussen, Eli, a. 1. september 1968

Tegner II:	Heming, Astri, a. 1. januar 1962
	Pedersen, Agnar, a. 1. februar 1967
	Haugum, Ellen, a. 1. april 1969
Laboratorie-assistent I:	Berg, Tom, a. 16. januar 1967
Sekretær I:	Petersen, Sigrid, a. 6. januar 1969
Kontorassistent/fullmektig II:	Wettavik, Vigdis, a. 1. mars 1964
	Iversen, Bjørn Sverre, a. 1. september 1966, midl.
	Rustung, Bodil, a. 15. september 1969, midl.

En del geologer ved andre institusjoner og viderekomne studenter har vært knyttet til avdelingen som vitenskapelige medarbeidere under sommerens markarbeid. Videre har diverse personell vært ansatt i korttidsengasjementer.

Geofysisk avdeling:

Direktør:	Aalstad, Inge, c.r., a. 1. okt. 1962 (15. juli 1952)
Geofysiker I:	Sakshaug, Gunnar, siv.ing., a. 1. juli 1936
	Singsaas, Per, a. 1. september 1937
	Hillestad, Gustav, siv.ing., a. 20. januar 1953
	Håbrekke, Henrik, siv.ing., a. 17. august 1959
Fysiker I:	Breen, Arne, siv.ing., a. 1. desember 1940
Geolog I:	Svinndal, Sverre, c.r., a. 1. juli 1961
Geofysiker II:	Moxnes, Hans Petter, c.r., a. 6. juli 1959
	Sindre, Atle, c.r., a. 24. mai 1961
	Eidsvig, Per, siv.ing., a. 1. januar 1967
	Åm, Knut, siv.ing., a. 4. januar 1969
Geolog II:	Barkey, Henri, c.r., (nederl. eks.) a. 1. desember 1963, midl.
	Tan, Tek Hong, c.r. (nederl.eks.) a. 23. april 1959
Avdelings-ingeniør II:	Uddu, Odd, a. 1. oktober 1952
	Haugan, Arne, a. 1. juni 1961
Konstruktør I:	Brandhaug, Kolbjørn, a. 1. september 1958
Konstruktør II:	Opsahl, Henrik, a. 21. april 1958
	Gausdal, Odd, a. 20. september 1957

Borformann:	Vassbotn, Sven, a. 1. september 1963
Teknisk assistent I:	Dalsegg, Einar, a. 1. mai 1966
	Melleby, Peter, a. 14. november 1955, midl.
Tekniker I:	Blokkum, Oddvar, a. 17. januar 1961
	Staw, Jomar, a. 18. juni 1956
Tekniker II:	Holiløkk, Lars, a. 1. juni 1959
Laborant II:	Opdahl, Ragnar, a. 23. oktober 1957
	Åsheim, Arnt, a. 6. januar 1969
Laboratorie-assistent:	Johansen, Hermann, a. 1. april 1963, midl.
Tegner i særklasse:	Grønli, Gunnar, a. 12. januar 1956
Tegner I:	Haugen, Torbjørn, a. 3. juni 1959
	Solvang, Terje, a. 1. januar 1961
	Godø, Rolf, a. 1. januar 1966
Tegner II:	Østby, Solveig, a. 14. august 1961
	Sagflaat, Hans, a. 1. desember 1962
Tegner-assistent:	Andreassen, Birgith, a. 1. februar 1964, midl.
	Hove, Erling, a. 1. desember 1965, midl.
Mekaniker-formann:	Brevik, Bjørn, a. 1. mai 1939
Mekaniker i særklasse:	Pettersen, Reidar, a. 25. mars 1952
	Gravseth, Odd, a. 10. november 1953
Instrumentmaker i særklasse:	Kirkeby, Kåre, a. 15. september 1951
Snekker:	Pettersen, Norman, a. 18. februar 1946
	Tetli, Alf, a. 1. oktober 1958
Sekretær I:	Singsaas, Cathrine, a. 1. oktober 1953
Kontorassistent/ fullmektig II:	Pettersen, Bodil, a. 1. juni 1968

Kjemisk avdeling:

Direktør:	Kvalheim, Aslak, siv.ing., a. 1. oktober 1947 (1. oktober 1937)

Laboratorie-ingeniør I:	Faye, Gjert Chr., siv.ing., a. 10. desember 1958
	Andreassen, Birger Th., siv.ing., a. 15. febr. 1961
	Nilsen, Rolf, siv.ing., a. 1. april 1963
Geokjemiker I:	Bølviken, Bjørn, siv.ing., a. 1. mars 1954
Midl. laboratorie-ingeniør:	Grennes, Johannes, siv.ing.
Laboratorie-ingeniør II:	Ødegård, Magne, siv.ing., a. 1. mai 1961
	Krog, Jan Reidar, siv.ing., a. 1. mai 1964
	Kuldvere, Arnold, c.r. (svensk eks.) a. 1. febr. 1969
Geokjemiker II:	Sinding-Larsen, Richard, siv.ing., a. 1. okt. 1969
Konstruktør I:	Næss, Gunnar, a. 16. januar 1960
	Solem, Knut, a. 1. januar 1961
	Flårønning, Asbjørn, a. 1. juni 1964
Konstruktør II:	Bremseth, Asbjørn, a. 9. november 1959
Konstruktør III:	Sivertsen, Tove, a. 9. januar 1958
Teknisk assistent I:	Wik, Jon M., a. 23. november 1953
Tegner I:	Holmberget, Edna, a. 1. september 1960
Første-laborant:	Storvik, Arne, a. 1. mars 1964
Laborant i særklasse:	Horgmo, Birger, a. 1. mars 1953
	Ekremsæter, Jørgen, a. 1. september 1960
Laborant I:	Wolden, Odd, a. 11. mars 1963
	Kalvøy, Henry, a. 24. mai 1965
	Rossing, Rolf, a. 16. januar 1967
Laboratorie-assistent I:	Taftøy, Inger Johanne, a. 1. februar 1966
	Berge, Frank Gabriel, a. 1. mars 1969
	Thoresen, Hans Henry, a. 1. juli 1969
	Søberg, Baard, a. 1. desember 1969
Sekretær I:	Bersvendsen, Jørgen H., a. 1. juni 1957
Kontorassistent/fullmektig II:	Vongraven, Eva, a. 28. oktober 1968

Engasjert personell i spesielle prosjekter:

Geofysiker Ørnulf Logn	Grongprosjektet fra 1. april 1969
Bergingeniør Reid K. Kvien	Grongprosjektet fra 1. oktober 1969
Vit.ass. Svein R. Østmo	Hydrologisk dekade fra 15. desember 1966
Vit.ass. Geir Goffeng	Hydrologisk dekade fra 1. januar 1967
Konstruktør Ole Nashoug	Hydrologisk dekade fra 4. juni 1968
Vit.ass. Odd Øvereng	Nord-Norge-prosjektet fra 1. juli 1969
Vit.ass. Arild Palmstrøm	Oppdrag for Statskraftverkene fra 1. januar 1968
Geolog Tore Birkeland	Kvartsundersøkelsene fra 1. novemb. 1968
Geolog Carl O. Mathiesen	Fra 1. februar 1969 engasjert for å fullføre den endelige rapport fra undersøkelsene i Bidjovagge med gjenstående midler fra A/S Vaddas Gruber Tidligere engasjert v/ Kautokeino Kobberfelter — Statens Undersøkelser 15. juni 1956, underlagt NGU 1. juli 1964 (Bidjovagge-undersøkelsene)

Dessuten har dr. Stanislaw Siedlecki fra 10. juli 1966 vært engasjert som geolog I i forbindelse med kartlegging av kartblad Vadsø. Arbeidet er gjort mulig ved tilskudd fra A/S Sydvaranger.