Structural Geology, Metamorphism, and Rb/Sr Geochronology of East Hinnøy, North Norway

by

John Michael Bartley

B.S., University of Washington
(1976)

SUBMITTED IN PARTIAL FULFILLMENT
OF THE REQUIREMENTS OF THE
DEGREE OF

DOCTOR OF PHILOSOPHY IN
GEOLOGY

at the

MASSACHUSETTS INSTITUTE OF TECHNOLOGY

February 1981

Signature of Author ______________________________
Department of Earth and Planetary Sciences
August 8, 1981

Certified by ______________________________
B. C. Burchfiel
Thesis Supervisor

Accepted by ______________________________
Theodore R. Madden
Chairman, Departmental Graduate Committee

Structural Geology, Metamorphism, and Rb/Sr Geochronology

of East Hinnøy, North Norway

by

John Michael Bartley

Submitted to the Department of Earth and Planetary Sciences on August 8, 1980 in partial fulfillment of the requirements for the degree of Doctor of Philosophy in Geology

ABSTRACT

Rocks of east Hinnøy represent three Caledonian tectonic settings: (1) pre-Caledonian crystalline basement (Lofoten terrain), (2) its autochthonous Eocambrian/Cambrian(?) sedimentary cover (Storvann Group), and (3) Caledonian allochthons (Narvik, Stangnes, and Salangen Groups). Affinities of the Storvann Group for the Eocambrian and Cambrian sedimentary rocks of the Baltic foreland support the hypothesis of Griffin and others (1978) that the Lofoten terrain was part of the Baltic craton in pre-Caledonian time. The allochthons are considered to have been derived from west of the present Norwegian coastline, and at least in part to have originally formed in an oceanic setting.

Five Caledonian deformations are identified on east Hinnøy, termed D_1 to D_5. D_1 and D_2 were synchronous with amphibolite facies metamorphism, while D_3 through D_5 post-dated the metamorphic peak. D_1 is the thrust emplacement of the Caledonian allochthons upon the Lofoten terrain. D_2 produced thrusting of the Lofoten basement rocks and recumbent folding and interleaving of the basement with its structural cover on a scale of at least several kilometers. Analysis of minor structures indicates ESE -directed transport during D_2. Penetrative deformation associated with D_1 and D_2 dies out structurally downward in the rocks of the Lofoten terrain, away from their contacts with metasedimentary rocks which underwent prograde metamorphism during Caledonian orogenesis. This suggests involvement of the basement in deformation was controlled by introduction of volatiles liberated by prograde metamorphic reactions in the metasedimentary rocks.

D_3 through D_5 formed multiple sets of upright to overturned folds, superposed on earlier structures.

Several NE-trending high angle dip-slip faults of probable early Cenozoic age cut east Hinnøy into several blocks. The present study indicates that this faulting was more intense, and of a more uniform trend and sense of movement than previously recognized.

The Caledonian metamorphic peak reached amphibolite facies (kyanite grade) in all rocks of east Hinnøy. Sillimanite was recognized at one locality. Estimated peak metamorphic conditions are 550-600°C and 5-9 kb. These conditions were reached during and/or immediately following D_2. D_3 through D_5 occurred under greenschist facies conditions (biotite grade) during cooling from the metamorphic peak.

Rb/Sr whole rock ages of 1726 ± 31 Ma and 1559 ± 155 Ma were determined for the Middagstind Quartz Syenite and Melaa Granite, respectively, supporting the structural interpretation that these bodies belong to the pre-Caledonian basement terrain. The former age indicates that the Middagstind syenite formed as part of the regionally important intrusive episode which formed roughly half of the Lofoten terrain 1800 to 1700 Ma ago (Griffin and others, 1978). The large error for the age of the Melaa Granite is due to disturbance of whole rock Rb/Sr systems by metamorphism. Consideration of structural relationships and possible sources of disturbance suggests the actual age may lie at the younger end of this range, and the Melaa Granite may be correlative with the 1400 Ma Lødingen Granite. Rb/Sr whole rock study of the Ruggevik Tonalite Gneiss, which intruded the Stangnes Group prior to its D_1 emplacement upon the Lofoten terrain, did not yield an isochron, but suggests a possible late Precambrian age. Four Rb/Sr biotite/whole rock systems from the Middagstind syenite and Ruggevik tonalite give ages of 362 to 347 Ma, indicating cooling following the metamorphic peak occurred during Devonian time.

Consideration of the regional geology of the Caledonide orogen suggests that the part of the mountain belt now preserved in Scandinavia developed mainly in the underthrusted plate of a continental collision. Large scale geometry and timing of structures in the region including east Hinnøy (Lofoten-Rombaken transect) have affinities for accretionary prism or foreland thrust belt tectonics, despite a position within the metamorphic core region of the orogen. This has two fundamental implications of potentially general application. First, orogenic belts may be broad zones of lithospheric deformation (the "megasuture" of Bally, 1975) only within the upper 15-20 km of the crust. Deeper lithospheric levels may remain rigid, so that at depth, the plate tectonic hypothesis of narrow boundaries between rigid plates may be appropriate for convergent as well as divergent and transform plate margins. Second, the commonly observed phenomenon of detachment of crystalline thrust sheets at mid-crustal levels may result from decollement above an anhydrous, rigid lower crust which does not participate in deformation due to a lack of volatiles to promote recrystallization. This implies control of this detachment by distribution of chemical species rather than thermal structure, and suggests the brittle/ductile transition may not play a fundamental role in this system.

Thesis supervisor: B. C. Burchfiel
Title: Professor of Geology

TABLE OF CONTENTS

ILLUSTRATIONS

Figures

Tables

Plates (in pocket at back)

ACKNOWLEDGEMENTS

Many people have contributed material and/or spiritual assistance during this study; I cannot hope to fully enumerate them, and apologize in advance to those who have been omitted. It is not through a lack of gratitude, but rather a lack of space.

I would like to thank my advisor, B. Clark Burchfiel, for involving me in a project that has proven rewarding, for productive and enjoyable visits in the field, and for illuminating discussions of structural geology and tectonics of north Norway in particular and of orogenic belts in general. The Corps of Clones, including Kip Hodges, Gary Axen, Scott Cameron, Peter "the General" Guth, John Sharry, Jon Spencer, Brian Wernicke, and James "Igor" Willemin, have provided numerous valuable discussions and criticisms of the ideas presented here, as well as endless distractions and moral support. Kip Hodges in particular deserves my appreciation, for assistance in the field and for allowing me to freely use the results of his own field studies in Norway. Peter Vrolijk helped keep me on an even keel during the writing of this work, through discussions short and long during early morning runs. Frank Spear and Stan Hart provided invaluable assistance in understanding the petrology and geochronology of the study area, and in making available to me their laboratory facilities. Mary Roden and Julie Morris supplied aid, instruction, and sympathy in the laboratory during the isotopic analyses performed at MIT.

Bill Griffin of the Mineralogisk Museum in Oslo was a great help in getting started in Norway. Bill and Kerstin, his wife, provided a reliable and invaluable way station for itinerant graduate students for which I can scarcely thank them adequately. Brenda Jensen was very generous with her time in helping with the many day to day problems of living in a foreign country, as well as extraordinarily efficient in supervising and instructing me with the XRF equipment at the Museum. Arne Raaheim, Bjørn Sundvoll, Steinar Sollheim, Toril Enger, and Steve Ehrenberg were helpful with problems in the isotopic analyses I did at the Museum. Terje Foss and Borghild Nilssen provided abundant good natured assistance in rock preparation, mineral separation, and miscellaneous technical problems. John Green and Guro Vikør, our neighbors at Bjerke Studentheim in Oslo, did much to keep our spirits up at difficult times during our stay in Oslo.

Lars and Mette Fruergaard-Jensen provided a haven of rest and friendship during the fieldwork on Hinnøy; their support and friendship are deeply appreciated. Karl-Annar and Ivar Mathissen at Harstad campground were generous and friendly hosts during our weekly visits for rest and relaxation. Harstad Oppland Rutebil A/S provided reliable, inexpensive, and generally good natured bus transportation to and from the field.

I want to thank my family as a whole, and especially my sister and her husband, Kathy and Bob French, and my parents, Jack and Anita Bartley, for support, encouragement, and understanding during my graduate studies.

Kristy Bartley, my wife and closest friend, has assisted me in every phase of this work: in planning, in fieldwork, in the laboratory, at the drafting table, at the typewriter, and especially in those inevitable grim moments when it seemed like completion of the work was impossible. Without her aid, this study would very likely have been impossible.

Financial support for the study came from a variety of sources. Field expenses during the first field season and part of the second were supplied by a grant to Clark Burchfiel by M.I.T. Two G.S.A. Penrose grants (Nos. 2296-78 and 2420-79) and a grant from the Crown Princess Märtha Friendship Fund of the American-Scandinavian Foundation supported field studies during the second and third seasons of fieldwork.

CHAPTER I: INTRODUCTION

Purpose

This study has two main focuses. It examines the structural behavior of pre-existing continental crust in the internal portion of a collisional orogenic belt: specifically, (1) the extent of basement involvement in metamorphism and deformation, (2) the factors controlling basement involvement and the structural styles developed, and (3) the structural relationship between basement and cover rocks. The geometry and structural history of the associated structural cover, and the relationships between structural and thermal events, have also been investigated. With these goals in mind, I have studied the structural geology, petrology, and geochronology of east Hinnøy, an area which has proven to be important to understanding the tectonics of the Scandinavian Caledonides. Since there are a finite number of orogenic belts on earth and their variability is large, a contribution to the understanding of processes in any one belt can be a significant step in understanding tectonics in general.

The Scandinavian Caledonide orogen is a major geotectonic belt, extending 1500 km from southwest Norway to northern Norway, and eastward into western Sweden (Figure 1). It constitutes the eastern part of a two-sided collisional orogen of early Paleozoic age, the western portion of which now lies in east Greenland (Figure 2). Structures in Scandinavia verge mainly east; those in Greenland verge mainly west. The Cenozoic opening of the Norwegian Sea split the orogen longitudinally through its central part, so that in Scandinavia, the internal zones lie in the coastal regions and the external zones are inland. The depth of erosion leaves cover and basement rocks exposed in subequal amounts, making it ideal for studying their structural relationships.

The Scandinavian Caledonides can be divided into four fundamental domains passing westward from the foreland (Figure 1): (1) the parauthochthon, (2) the eastern or lower nappes, (3) the western or higher nappes, (4) the western gneiss terrain. Domains (1) to (3) are a succession of progressively structurally higher and more allochthonous thrust complexes which pinch out westward so that higher allochthons rest directly on the basement in the west (Gee, 1975; Binns, 1978). In general, higher

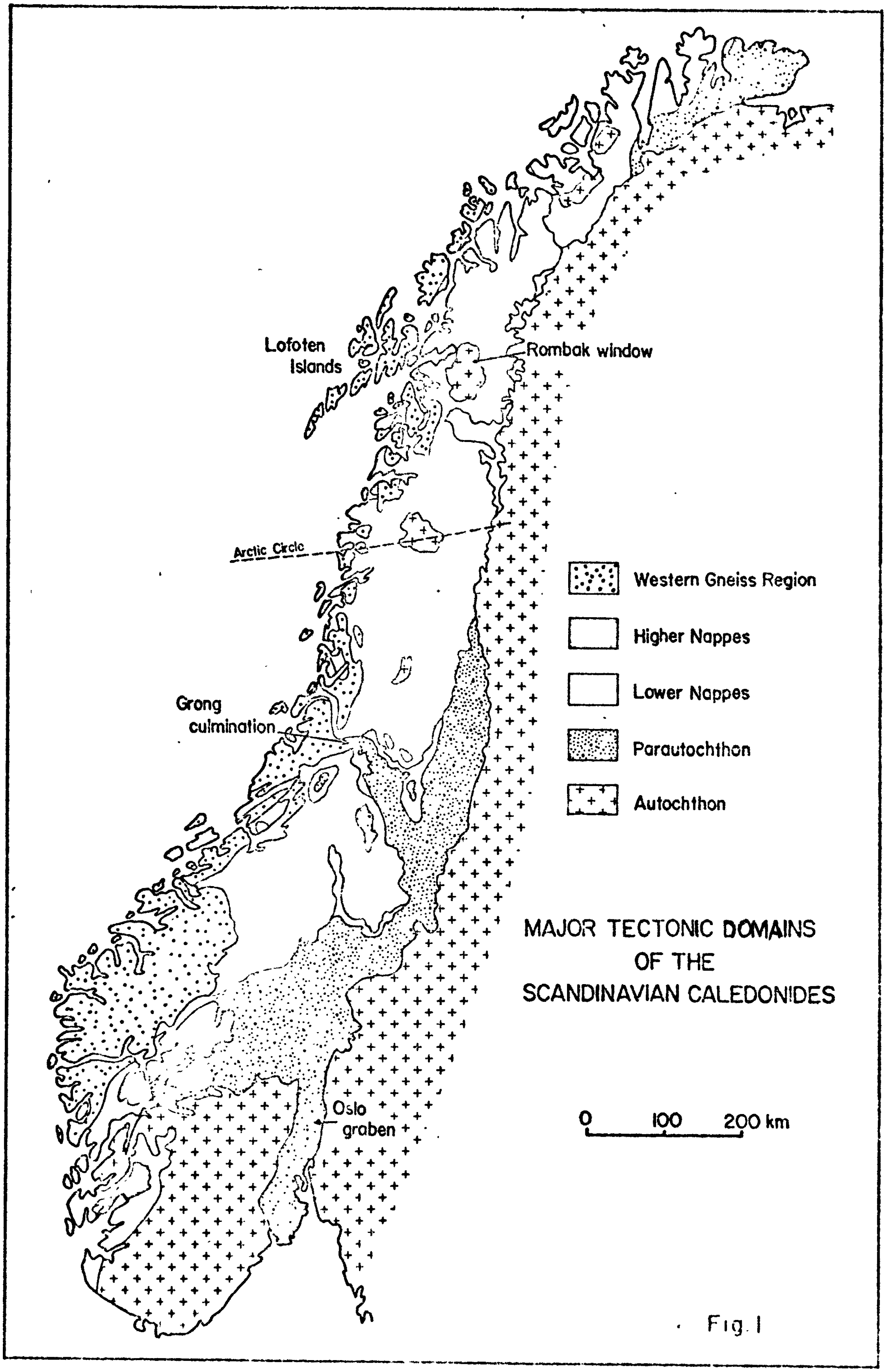
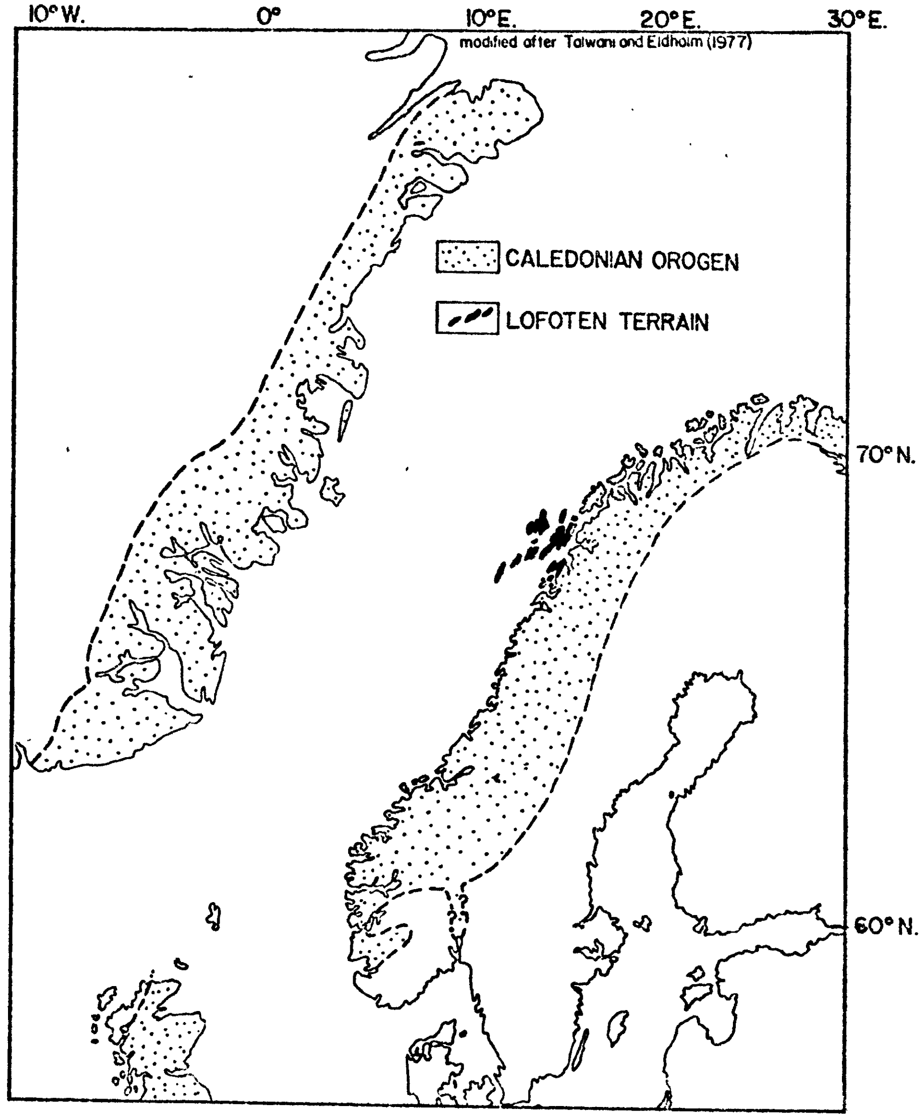

Lofoten
Islands
Rombak window
Arctic Circle
Western Gneiss Region
Higher Nappes
Lower Nappes
Parautochthon
Autochthon
Grong
culmination
MAJOR TECTONIC DOMAINS
OF THE
SCANDINAVIAN CALEDONIDES
0 100 200 km
Oslo
graben
Fig. 1

Figure 2.



POSITION OF THE LOFOTEN TERRAIN WITHIN THE NORTHERN CALEDONIDES PRIOR TO OPENING OF THE NORWEGIAN SEA sheets are at higher metamorphic grades, although the timing of metamorphism relative to emplacement is not often well known. Basement rocks are involved in thrusting, mainly as thin sheets, but the extent of this is still uncertain. Domain (4) is a complex of Precambrian basement rocks which in general lie structurally beneath nappes of domain (3) (Wilson and Nicholson, 1973). The presence of an older continental crystalline terrain to the west of a belt of nappes including rocks of apparent oceanic origin (eg. Gale and Roberts, 1972, 1974) led Dewey (1969) to propose that the western gneiss terrain was sutured to Scandinavia in the Caledonian orogeny. However, subsequent study has shown that the nappes are derived from the west of the modern Norwegian coastline, and passed over the western gneiss terrain which apparently belonged to the Scandinavian craton in Pre-Caledonian time (Gee, 1975).

The structural position of the western gneiss terrain under the metamorphic nappes has led to the assumption that much of its intense metamorphism and deformation are of Caledonian age, generated at the deepest exposed levels within the Scandinavian Caledonide orogen. However, Griffin and others (1978) showed that in exposures of the western gneiss terrain in the Lofoten Islands (Figure 1), little or no Caledonian metamorphism and deformation could be recognized; the structures and mineral assemblages are Precambrian in age. This led them to hypothesize a shallow (3 to 9 km) structural level for the Lofoten block during Caledonian orogenesis, in direct conflict with the notion that it was buried by nappes which themselves were at medium-pressure amphibolite facies conditions during emplacement.

The boundary between the Lofoten terrain and the Caledonian nappes lies on eastern Hinnøy (Figure 3). This study examines this boundary and the structure of the rocks on either side of it, in order to:

(1) resolve the dilemma regarding the Caledonian structural position of the Lofoten block posed by the work of Griffin and others (1978);

(2) provide constraints on the pre-Caledonian position of the Lofoten block. If it is not structurally equivalent to the rest of the western gneiss terrain, it could be exotic to Scandinavia;

(3) establish the extent and nature of basement involvement in Caledonian deformation on east Hinnøy;

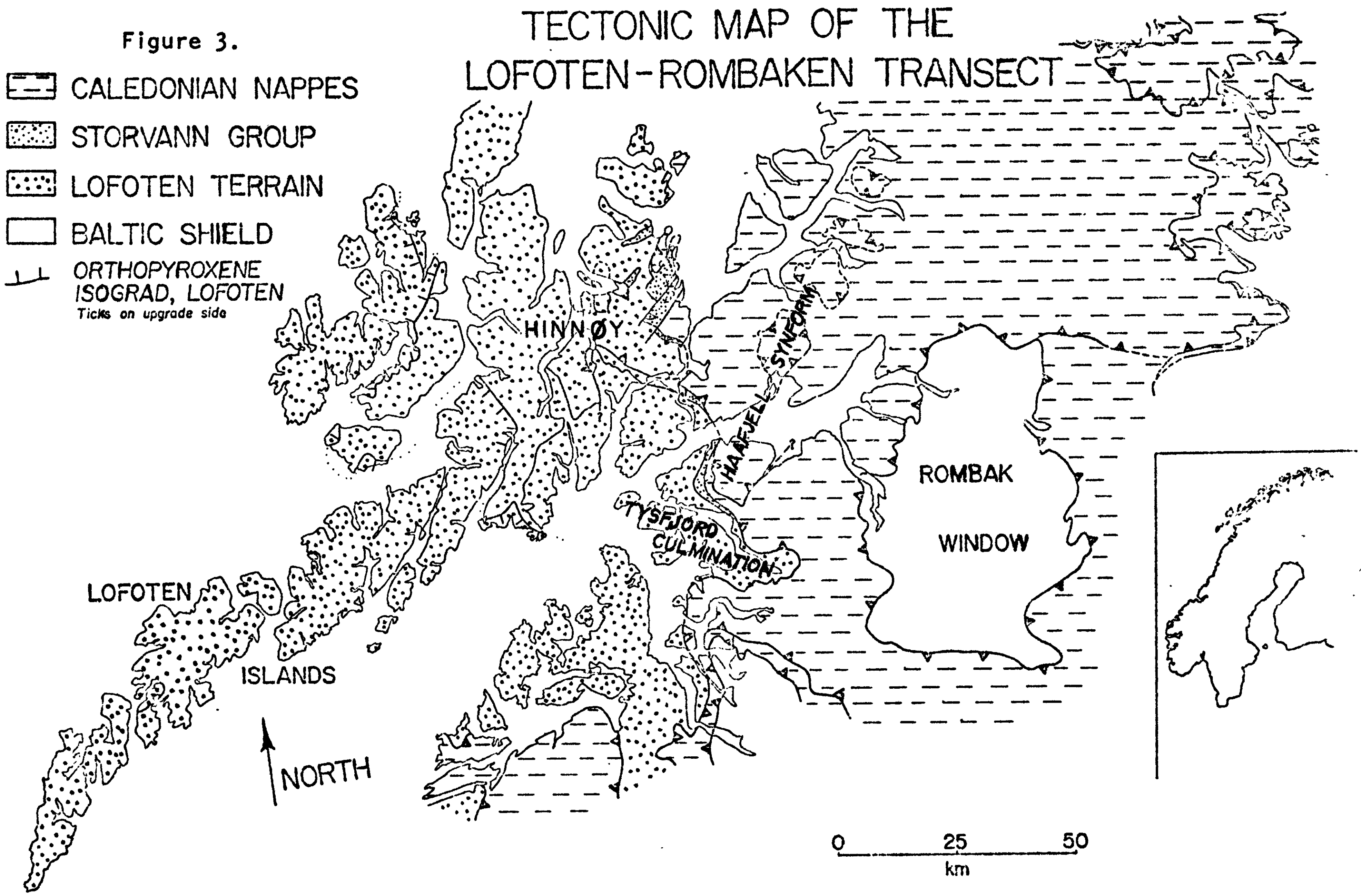

TECTONIC MAP OF THE
LOFOTEN-ROMBAKEN TRANSECT
CALEDONIAN NAPPES
STORVANN GROUP
LOFOTEN TERRAIN
BALTIC SHIELD
ORTHOPYROXENE
ISOGRAD, LOFOTEN
Ticks on upgrade side
HINNØY
HAAFJELL SYNFORM
TYSFJORD
CULMINATION
ROMBAK
WINDOW
LOFOTEN
ISLANDS
NORTH
0
25
50
km

Figure 3.

(4) establish the structural and metamorphic history of the Caledonian nappes and their geometrical relationships on east Hinnøy to improve understanding of the kinematics of the Caledonian deformations; and

(5) constrain the conditions and timing of deformation and metamorphism through metamorphic petrology and Rb/Sr geochronology.

Location and Access

Hinnøy (ca. 69°N, 16°E) is the island nearest the mainland of the archipelago called Lofoten/Vesteraalen. On its northeast corner is the town of Harstad, population ca. 20,000, where shipbuilding is the main economic activity. Harstad was the main base of operations for fieldwork. The eastern quarter of the study area is occupied principally by small farms. Summer and permanent homes are concentrated along the shoreline, much of which is thus privately owned. The remainder of east Hinnøy is sparsely populated, being mainly boggy valleys, forested hillsides, and alpine uplands.

The study area appears on portions of the following 1:50,000 topographic map sheets, available from Norges Geografiske Oppmaaling: Harstad (1332 IV), Tjeldsund (1332 III), Gullesfjord (1232 II), and Kvaefjord (1232 I). A major glacial valley running southwest from Harstad to the head of Kvaefjord (Plate I) forms the northern boundary of the map area. The western limit was taken as the shore of Gullesfjord and Austerfjord, and thence south to Kongsvik on Tjeldsund. Along this side, I connected my mapping as much as possible with that of Hakkinen (1977) on west Hinnøy. The southern and eastern boundaries were defined by Tjeldsund, the channel which separates Hinnøy from the mainland to the east and Tjeldøy to the south. One small area across Tjeldsund on the mainland was also mapped (Plate I), and proved to be of vital importance for structural interpretations.

Because of the populated nature of the area, roads are fairly extensive, though often unpaved and concentrated along shorelines where the majority of the people live. There are few places in the study area more than five kilometers from a road. Public bus service is excellent, and for the summers of 1978 and 1979 this was nearly the sole means of transportation for fieldwork.

Physiography, Vegetation, and Exposures

North Norway around 69°N is characterized mainly by extreme glacial topography, with deep, steep-sided fjords separating alpine peaks. Most of the study area has more moderate relief, with valleys 1 to 2 km wide and rather rounded hills 500 to 600 m high. The southwestern portion underlain by Precambrian granite (Plate I) approaches the more alpine topography with peaks as high as 1100 m. The moderation of the topography is a result of bedrock lithology to some extent, the schists and marbles of the Caledonian cover rocks being less resistent to geomorphic processes. However, considerable areas of Precambrian basement also underlie subdued lowland areas in this vicinity.

Quality and nature of outcrop is mainly a function of elevation. Shoreline exposures are the cleanest exposures available for measuring structures and examining contacts, although the rocks are generally somewhat weathered. These are especially abundant along the eastern side of the study area from Harstad south to Sandtorg, the east shore of Storvann (S), at the head of Kvaefjord, and across Tjeldsund around Fjeldal. In most other shoreline areas, Quaternary sediments cover the bedrock.

Inland from the shoreline are a number of lowlying areas (lower than 100 m) of fluvial sediments or fluvially-reworked glacial sediments. Where well drained, these areas are largely cleared farmland, providing pasturage for livestock and poor exposures for bedrock geologists. Where poorly drained, extensive bogs and marshy lakes are developed. This is probably the least productive elevation zone for bedrock geology, and is only redeemed occasionally by roadcut exposures.

Most hills which rise above the low valleys reach 300 m or more. The slopes of these hills are generally either heavily vegetated or are vegetated talus piles. In either case, outcrops are a little more abundant than in the low valleys, but are generally awkward to find and reach. Vegetation is mainly deciduous trees a few meters in height, especially birch and willow; the occasional thickets of evergreen are mainly the result of artificial reforestation efforts.

Treeline varies from about 300 to 400 m depending upon direction of exposure, bedrock lithology, drainage, etc. Above treeline are three principal zones. Where little soil has developed over cliffy topography resulting from a resistant lithology, outcrop is abundant but generally rather lichen-encrusted and weathered. Marbles occurring within such areas are generally (but not always) less resistant and consequently are poorly exposed. Where soil is better developed, rolling heaths are present with variable amounts of scrubby trees near treeline, and it is not uncommon for marble to be better exposed here because the carbonate-rich soil is less fertile and supports little vegetation. This process is encouraged by domestic sheep which are pastured in higher meadows during the summer. The sheep dig out shelter wherever the soil is loose, which often occurs at marble layers; for many marble outcrops I am entirely indebted to my woolly friends!

At elevations above about 550 m there is little or no vegetation, and outcrop becomes nearly 100% except for slope deposits. For example, the western slope of Middagstind (914 m) is almost devoid of outcrop because it is mantled by a talus cone which extends all the way to the lake at its foot.

Previous Investigations

The earliest studies in Lofoten-Vesteraalen were by Helland (1897), Kolderup (1898), and Vogt (1909), who considered the terrain to be one of basic "eruptive" rocks. Heier (1960) showed that the rocks of Langøy, and Lofoten-Vesteraalen as a whole, are mainly intermediate to acid in composition and owe their dark color to granulite facies conditions of crystallization. This led to an extensive effort in the late 1960's and early 1970's by the Norges Geologiske Undersøkelse (NGU) and the Geologisk Museum in Oslo, which was headed by Knut Heier and W. L. Griffin, and included many others. The objectives of their effort were to map the geology and analyze the granulite facies rocks petrologically, geochemically, and isotopically. The major conclusion from the point of view of Caledonian tectonics was that the history recorded in Lofoten is nearly entirely Precambrian; Caledonian effects appear to be few and minor. The results of this project are summarized in Griffin and others (1978), which includes references to the many other papers, published and unpublished, which were produced by the project.

Also included in this effort were structural studies by Rice University students of B. Clark Burchfiel. J. F. Tull (1972, 1977) mapped the geology of Vestvaagøy in western Lofoten, and J. W. Hakkinen (1977) worked on west Hinnøy directly west of the present study area. This author relied heavily on Hakkinen's work in his own study, although interpretations may not always be identical.

The Caledonian nappes carrying metasedimentary rocks, exposed to the east of Hinnøy, were first studied by Vogt (1942, 1950) to the north of Ofotfjord, and by Foslie (1941, 1949) south of Ofotfjord. The metasedimentary rocks were considered to be Cambro-Silurian in age, and to overlie the Tysfjord Granite, a Precambrian granite gneiss whose precise relationships to Lofoten rocks further north is still a subject for some debate. The contact between the Lofoten terrain and the nappes, which is exposed on east Hinnøy, was little studied until recently. An unpublished reconnaissance map of the Harstad 1:100,000 sheet by Th. Vogt (1955) on file at NGU in Trondheim shows only four units: granite, marble, mica schist, and Quaternary deposits. Gustavson, in preparation of the Narvik 1:250,000 map sheet (1974a), produced two maps of east Hinnøy (1966, 1974 b and c) at different stages of his work, as well as synthesizing in English the earlier work of Vogt and Foslie on the mainland (Gustavson 1966, 1969, 1972). Gustavson's mapping was clearly of a reconnaissance nature; location of lithologies as well as structural interpretations were found unreliable by the present author. Consequently, major differences will be recognized between Gustavson's work and my own, both in data and interpretations.

Methods and Scope

The primary effort in this work was the preparation of a geologic map of the study area with attention to major lithotectonic assemblages and structural relationships. The mapping was done an topographic maps at a scale of 1:50,000 available from Norges Geografiske Oppmaaling, prepared in the 1930's and revised in 1952 by the US Army Corps of Engineers. Contour interval is 30 meters; the 1:50,000 sheets are photographic enlargements of 1:100,000 originals. The maps are consequently not up to the standard set by USGS maps in the U. S., and the precision of location of geologic features suffers accordingly. Partial air photo coverage by Widerøe Flyveselskap was available from Norges Geologiske Undersøkelse. However, availability and use of these photos was restricted by the Norwegian Government for military security reasons (two sizable military installations and several small ones are located in the study area). Hence, air photos were used only as an aid in reconnaisance and location, and not for an actual map base.

East Hinnøy is geologically a very complex area, and this study can claim no more than to have reconnoitered some of its intricacies. The major successes here have been identification of several fundamental terrains and their contact relationships, and a clarification of the nature and sequence of structural and metamorphic events affecting these terrains. There remain many potentially fruitful studies of the details of this area which simply were beyond the time limits imposed here.

Note On Geographic Names

The Norwegian people are fond of names; every minor geographic feature, be it physical or cultural, is named on their maps. The names are often descriptive, such as Rundfjell ("Round Mountain") or Breivik ("Broad Cove"). The descriptive nature of these names together with a history of physical isolation of communities has led to duplication of names even on a very local scale. For example, in the study area there are two lakes named Storvann ("Big Lake") within several kilometers of each other (Plate I). For the sake of clarity, I have in this study distinguished them as Storvann(N) and Storvann(S). Other similar duplications occur, but have here been avoided for geographic references.

CHAPTER 2: LITHOLOGIES AND STRATIGRAPHY

Introduction

Rock units of East Hinnøy can be divided into three fundamental assemblages: (1) pre-Caledonian crystalline basement rocks, which are considered to be an eastward extension of the Lofoten terrain (Griffin and others, 1978); (2) the Storvann Group, a sequence of metasedimentary rocks in depositional or modified depositional contact with the pre-Caledonian basement, and which is believed to be its Cambrian or Eocambrian sedimentary cover; and (3) Caledonian allochthons, which in the study area comprise at least three distinct assemblages of metasedimentary and lesser metaigneous rocks.

The lithostratigraphic complexity of East Hinnøy is markedly greater than previously recognized. All metasedimentary rocks and most amphibolite units had been assigned a Cambro-Silurian age and assumed to be practically all allochthonous along Caledonian thrusts (e.g., Gustavson 1972, 1974 a, b, c). Many of these rocks have been found to be intruded by Precambrian granitoid bodies and hence are themselves part of the pre-Caledonian basement. In this study, six different assemblages of rocks have been considered elements of the basement complex: (1) Archaean gneisses, (2) the Hesjevann assemblage, (3) the Kvaefjord Group, (4) the Middagstind Quartz Syenite, (5) the Austerfjord Group, and (6) the Lødingen and Melaa Granite(s). Three of the above rock associations include metasedimentary rocks previously considered part of the Caledonian nappes of Paleozoic metasedimentary rocks.

Cover rocks and their relationships are also more complex than previously described. Gustavson (1972) reported a single stratigraphic sequence for metasedimentary rocks of East Hinnøy, which he apparently considered adequate to describe all allochthonous and autochthonous units. In fact, the quartzite he considered to be autochthonous Eocambrian cover of the pre-Caledonian basement has proven to be intruded by Precambrian Melaa Granite, and is here included in the Hesjevann assemblage. An extensive autochthonous sequence of quartz-rich terrigenous metasedimentary rocks and subordinate marbles does occur in this area (the Storvann Group), which is petrographically quite distinct from the allochthonous units and the pre-Caledonian metasedimentary rocks. Allochthonous terrains include

(1) slivers of pelitic schists and gneisses correlated with the Narvik Group, (2) banded amphibolite intruded by a semiconcordant tonalite gneiss body, here named the Stangnes Group, and (3) marbles and mica schists of the Salangen Group. At the boundary between the clearly allochthonous and clearly autochthonous rocks are local occurrences of calcareous schists with some marble and amphibolite; the affinities and significance of these latter rocks are uncertain.

Pre-Caledonian Basement

Introduction

The basement terrain of East Hinnøy includes six different rock associations, from oldest to youngest: (1) Archaean gneisses, which include migmatitic gneiss, a hornblende diorite body possibly related to the migmatites, and the Gullesfjord Gneiss; (2) the Hesjevann assemblage, an association of quartzite, schists, marbles, and amphibolite, which probably do not constitute a petrogenetic suite but are commonly associated rocks, present in pendants and blocks in the basement granitoid gneisses; (3) the Kvaefjord Group, a sequence of quartzofeldspathic, micaceous, and mildly calcareous terrigeneous metasedimentary rocks of at least middle Precambrian and perhaps older age; (4) the Middagstind Quartz Syenite, a pluton similar in age and composition to the mangerites which form the bulk of the Lofoten terrain to the west; (5) the Austerfjord Group, a diverse group of metasedimentary rocks dominated by biotite-rich schist and calcareous para-amphibolite, with subordinate quartzite, quartzofeldspathic schist, and marble; and (6) the Melaa and Lødingen Granites, which are considered correlative; because this correlation cannot yet be proven, the local terminology is retained for the present. Age relationships between these units are not all well established; the order above is in some cases more a matter of best guess than hard data. However, two fundamental observations appear to be valid and of foremost importance: (1) All these units form a single continuous pre-Caledonian terrain, and (2) this terrain is continuous with the Lofoten-Vesteraalen province of Griffin and others (1978). Hence, conclusions reached about the structural position and tectonic significance of the basement terrain of East Hinnøy are valid for the entire Lofoten terrain.

Archaean Gneisses

Introduction

Three units are included in the Archaean Gneisses: migmatitic gneisses, hornblende diorite, and the Gullesfjord Gneiss. The possibility exists that the migmatites and hornblende diorite are genetically related; however, the massive character of the diorite argues against this. The Gullesfjord Gneiss intrudes both of these units; since the Gullesfjord Gneiss in its type area on West Hinnøy has yielded a late Archaean Rb/Sr whole rock age (Griffin and others, 1978), all these units are assigned to the Archaean.

Migmatitic Gneiss

Migmatitic gneiss is very extensive on West Hinnøy and was described in detail by Hakkinen (1977). On East Hinnøy it is limited to the area around Revnes and the Melaa, and along a few streams southeast of Storvann (S). It was not studied in detail and is considered only briefly here.

The migmatitic gneiss is typically a hornblende-biotite-two feldspar medium to coarse-grained gneiss, commonly of granodioritic composition but spanning a large range of silica contents. It is heterogeneous on all scales from hand specimen to mountainside (Figure 4A). Characteristically it weathers to light or medium gray with layers, stringers, or irregular masses of darker material within. Plagioclase porphyroblasts approximately 1 cm in length are common. The Gullesfjord Gneiss and Melaa Granite also have migmatitic portions locally which makes identification of these units in the field sometimes problematic. Two criteria were generally used: (1) the Melaa and Gullesfjord units both contain important perthitic microcline, generally as the dominant feldspar, whereas the migmatitic rocks more commonly are plagioclase-rich; and (2) the granite gneisses are dominantly very massive, homogeneous rocks with only local migmatitic areas which seem to result from near-complete digestion of xenoliths in a melt rather than the <u>in situ</u> anatexis postulated as the origin of the migmatites (Hakkinen, 1977; Griffin and others, 1978).

On east Hinnøy, the only unit with which the migmatites are in contact is the Gullesfjord Gneiss. The relations are obscure and appear generally gradational, but in one locality near the Melaa, an outcrop of migmatites

Figure 4A: Archaean migmatite.

Outcrop is in stream bed, about 1 km SE of Storvann(S).

Figure 4B: Possible graded bedding, Hesjevann assemblage quartzite.

Darker bands are more micaceous than lighter, more pure quartzite bands, suggesting fining of the protolith. Bedding would be upright here. Outcrop lies on the crest of a broad, low ridge about 1 km west of Gausvik.

Figure 4A.

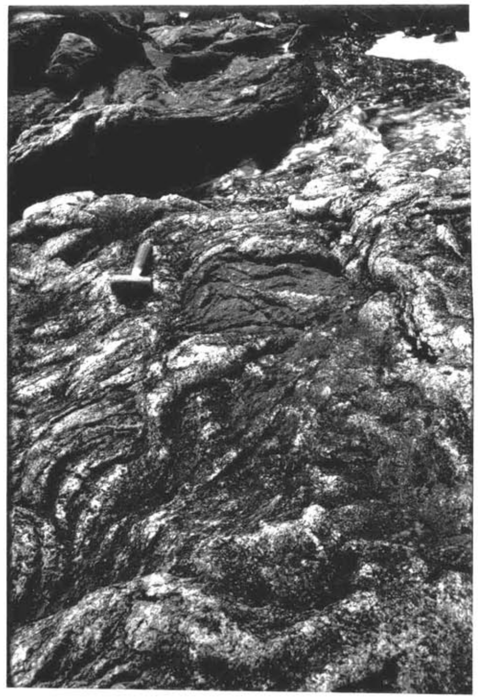
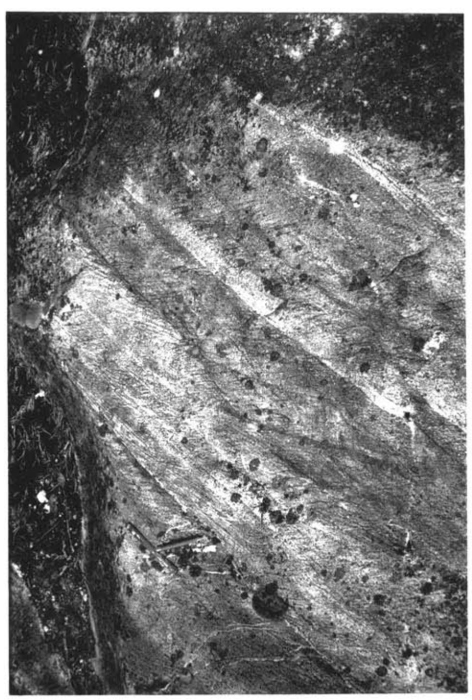


Figure 4B.



with transposed dikes of granite suggests that the granite gneiss intrudes the migmatites, a relationship consistent with the isotopic data of Griffin and others (1978), discussed below.

The Archaean age of the migmatites is based on lithologic correlation and mapped continuity with rocks on West Hinnøy for which Griffin and others (1978) determined an age of 2685 ± 65 Ma from a Pb/Pb whole rock secondary isochron. This age is believed to indicate isotopic homogenization in an anatectic event so that the original rock-forming age would be somewhat earlier.

Jacobsen and Wasserburg (1978) published Sm/Nd whole rock analyses of four migmatite samples and one Gullesfjord Gneiss sample from West Hinnøy and Langøy, which gave model ages from 2600 to 2670 Ma. Two other migmatite samples, one from Langøy and one from Moskenesøy, gave younger ages (2390 and 2040 Ma), which the authors attributed to later migmatization; disturbed isotopic systematics by granulite facies metamorphism at 1800 Ma cannot be ruled out. Linear regression of the five "good" points yields an age of 2600 ± 360 Ma. Using only the Sm/Nd systematics would not lead to a confident age assignment, but their consistency with Pb and Rb/Sr (see Gullesfjord Gneiss below) systematics is reassuring.

Hornblende Diorite

The hornblende diorite is present in the study area only to the west of Middagstind, mainly on and around the peak called Hornet. It may also be an important protolith of the mafic to intermediate hornfels extensively developed to the east and south of the Middagstind syenite pluton; however, the fact that it shows only moderate contact metamorphic overprint and maintains its own recognizable identity west of Middagstind up to rather near the syenite pluton argues against this hypothesis.

The hornblende diorite is a dark gray to black, massive, medium -grained rock, with weak to moderate foliation defined by hornblende and biotite. It is not uncommon for foliation to be poorly developed. The age of this fabric is pre-1700 Ma because the adjacent Middagstind syenite (1726 Ma old; see Chapter 5) cuts this foliation, and is itself unfoliated.

Major minerals in the hornblende diorite include plagioclase (35-65%), hornblende (15-20%), biotite (15-20%), magnetite (ca. 1%), ± quartz (0-5%), ± epidote minerals (0-20%). Plagioclase is typically oligoclase, variably replaced by fine granular aggregates or dispersed grains of clinozoisite. Polysynthetic twinning is common. Prior to static metamorphism, textures were granoblastic to hypidiomorphic granular. Hornblende is blue-green, and equant to stubby prismatic. Its color suggests a metamorphic rather than igneous origin, so that it is not clear that the original igneous assemblage included amphibole. Hornblende is now generally ragged due to partial replacement by granular aggregates of clinozoisite. Biotite is brown and lepidoblastic, and may be mildly chloritized. Small amounts of granular quartz are sometimes present. Magnetite is euhedral to subhedral and shows locally complex laminar or skeletal textures in association with another oxide (ilmenite?).

The age of the hornblende diorite is presumably late Archaean. It is not obviously migmatitic, though somewhat heterogeneous; however, it appears to be intruded by the Gullesfjord gneiss on the basis of granite dikes in the diorite and hornblende-bearing xenoliths in the gneiss near their contact. The simplest interpretation is perhaps that the hornblende diorite is a more mafic portion of the mignatite complex which experienced less anatexis and thus shows less leucosome development.

Gullesfjord Gneiss

The Gullesfjord Granodiorite Gneiss appears to be the dominant lithology of the basement terrain comprising the southwestern half of the study area. During much of the field study, the granite gneiss here was believed younger (Melaa Granite; see Chapter 5), and related to the Lødingen Granite. That some younger granite is present here is still believed by the author, but the details of contact relationships between it and the Gullesfjord Gneiss are not well understood. The Gullesfjord Gneiss was originally described and named by Hakkinen (1977) from his study of West Hinnøy. He considered it to extend in the upper plate of the Austerfjord Thrust onto East Hinnøy where it was mapped in this study. The Gullesfjord Gneiss is a massive granitic to granodioritic gneiss, white to pink weathering, medium-grained, ranging texturally from mildly foliated to blastomylonitic. Xenoliths of biotite rich material, from a few centimeters to a few meters in long dimension, are common but not ubiquitous. Augen gneiss is common. The extreme pencil lineation locally found by Hakkinen, and attributed by him to intersection of two foliations, was not observed, although an intersection lineation between Precambrian and probably Caledonian fabrics is present.

The Gullesfjord Gneiss is in contact with the Austerfjord Group along the Austerfjord thrust all along its southwest edge in the study area. This thrust has been traced across Tjeldsund to the mainland (Fjelldal area) where the thrust becomes involved with other Caledonian structures. The gneiss at Fjelldal is in contact with Storvann Group metasedimentary rocks. The basal quartzite unit of the Storvann Group is all but missing here, so that the contact is probably not strictly depositional but rather a tectonic slide of sorts (Fleuty, 1964); however, the Storvann Group more generally unconformably overlies the pre-Caledonian basement in the area. Along the southeast part of the study area, the Gullesfjord Gneiss is in contact with rocks of the Hesjevann assemblage. The contact with the Hesjevann assemblage quartzite is not generally well exposed, but is more or less concordant, not intensely tectonized, and has been previously interpreted as stratigraphic and to be the basal unconformity of the Caledonian geosyncline (Gustavson, 1972). However, in an outcrop near Gausvik the weak compositional banding of the quartzite appears truncated along the contact, and the quartzite unit is clearly cut out westward along this contact, leaving the structurally overlying amphibolite in direct contact with the Gullesfjord Gneiss. Westward across Storvann(S), the Hesjevann assemblage is clearly intruded by Melaa Granite, a Precambrian granite possibly correlative with the 1400 Ma Lødingen granite, establishing a Precambrian age for the Hesjevann assemblage rocks (see also below). Hence, the contact in the Gausvik area is not the basal Caledonian unconformity; however, whether the contact in the Gausvik area is intrusive, depositional, or a metamorphically-obscured tectonic juxtaposition is unclear.

The only other unit the Gullesfjord Gneiss clearly contacts is the Archaean migmatites. As described above, it is believed that the Gullesfjord Gneiss intrudes the migmatites.

The granite gneiss to the west, south, and east of Storvann(N) is of questionable affinity. It is shown as Melaa Granite on Plate I, but the possibility cannot be excluded that it belongs to the Gullesfjord Gneiss; problems of petrographic distinction of these units will be dealt with under "Melaa Granite" below. If this gneiss is the Gullesfjord Gneiss, the contact relations described here would further support the unity of the pre-Caledonian basement of East Hinnøy as a single terrain, continuous with the Lofoten terrain. Similarly, the correlation of the belt of granite gneiss stretching from Storvann(S) to Tjeldsund, north of the belt of Hesjevann assemblage rocks and south of the Storvann Group outcrops, is ambiguous. Probably the only solution to this dilemma will be detailed geochronologic study.

Hesjevann Assemblage

General Relationships

The Hesjevann assemblage is a group of rocks that occurs near Hesjevann as pendants in the Melaa Granite, and with more ambiguous contact relationships with Precambrian granite(s) west and south of Gausvik. For convenience, other minor blocks of metasedimentary rocks in the basement granites, e.g., those in the Gullesfjord Gneiss north of Revsnes, have been included here. The Hesjevann assemblage consists of amphibolite, quartzite, calc-silicate marble and calc-silicate rocks, and a variety of mica schists. Most of these rocks where recognized by Gustavson (1974 a, b) were considered to be allochthonous Caledonian cover rocks, although Gustavson placed some of the amphibolites within the Precambrian basement. Detailed mapping has shown all these rocks to be Precambrian, and in general intrusively enclosed by the basement granites. There is no evidence that these rocks represent a genetic association, and it is likely they are not all the same age. The designation Hesjevann assemblage is informal and used here for convenience to refer to the association of rocks which appear to crop up repeatedly within the basement terrain of Hinnøy. These rocks may correlate with the quartzite/amphibolite/marble assemblage at Flesnes on west Hinnøy described by Hakkinen (1977).

Quartzite

The quartzite is a generally massive to weakly banded rock, pure white in color. Locally it has rhythmic interbeds a few tens of centimeters thick of quartz-biotite schist suggestive of graded beds (Figure 4B), and in a few outcrops west and south of Gausvik contains quartz-pebble conglomerates with pebbles up to 8 cm in diameter. Medium to coarse grain size (usually ca. 1 mm) gives the rock a sugary appearance in weathered outcrop. Near Gausvik, some outcrops appear cataclasized but more generally have a penetrative schistosity which may transect compositional banding defined by varying amounts of mica in the rock.

Compositionally the quartzite has variable but significant amounts (up to 10%) of either or both microcline and muscovite. Penetrative schistosity is moderately developed to absent. Quartz and feldspar grains are generally equant and granular. Quartz often has sutured grain boundaries and protomylonitic crush zones, indicating late deformation post-dating the thermal metamorphism. Detritial zircon is a prominent accessory mineral.

Quartzite of the Hesjevann assemblage is distinguished from the basal quartzite unit of the Storvann Group by its pure white color, coarser grain size, more homogeneous composition, and associated rock types. The protolith was probably a moderately well-sorted quartz sandstone. The basement on which it was deposited is undetermined, but Hakkinen (1977) suggested on the basis of stratigraphic facing that a similar quartzite at Langvassbugt may have been deposited on the Archaean rocks of west Hinnøy. It is worth noting that apparent graded beds mentioned above also indicate an upright facing along the contact of this unit with the Gullesfjord Gneiss near Gausvik (see p. 34 above), suggesting this contact may be in fact unconformable. The quartzite is structurally overlain by amphibolite near Gausvik, by marble at Hesjevann, and completely enclosed by granite north and west of Hesjevann. Hence structural dismemberment is frequent, and most contacts are probably tectonic to a lesser or greater degree.

Marbles

Hesjevann marbles are generally siliceous, with typical mineral assemblages that include calcite $\pm$ dolomite, tremolite, and white mica. Clinozoisite is common. Textures are usually coarse-grained granoblastic; compositional banding is weak to absent except near Gausvik where pink color bands are present. Tectonite fabrics are generally weak, though a lineation defined by 1 cm long tremolite needles is present in a road-cut in Gausvik. The marbles at Gausvik on lithology alone could belong to the Salangen Group (see below). However, it is difficult to envision how

Salangen Group rocks could be present in such a structural position. The presence of similar marbles in the Hesjevann area which clearly are part of the basement complex supports a Precambrian age for the marbles at Gausvik as well.

The protolith of this unit was a calcareous sedimentary rock. Although the siliceous component in the rocks may be partly original sedimentary material, the presence of aluminous phases like epidote and the occurrence of these rocks enclosed in granite suggests metasomatic modification of bulk composition probably occurred.

White Calc-Silicate Rock

The calc-silicate rocks are pure white, porcellainous rocks which were initially mistaken in the field for quartzite. Thin section study revealed these rocks to be composed of tremolite, clinozoisite, microcline, plagioclase, and quartz, with carbonate and sphene as important minor phases. The texture is hornfels-like, with randomly-oriented euhedra of tremolite and clinozoisite in a fine, granoblastic matrix of quartz and feldspar (Figure 5). The texture and composition suggest these rocks are contact metamorphic rocks related to intrusion of the Melaa Granite. If so, these rocks have preserved nearly unchanged Precambrian contact metamorphic textures through the Caledonian orogeny.

Schists

Various types of mica schists occur with the Hesjevann assemblage, all of them as more or less isolated bodies. West of Hesjevann, a pendant includes a brown quartz-biotite schistose hornfels with chlorite pseudomorphs after small (3 mm) garnet porphyroblasts. The chlorite and some biotite develop as part of a spaced cleavage which overprints a static, decussate texture thought to reflect contact metamorphism by the Melaa Granite (Figure 6). The absence of any evidence of penetrative schistosity development is striking in so micaceous a rock. The limits of Caledonian fabric development in basement rocks and its significance will be further considered in Chapters 3 and 6.

West of Gausvik, outcrops of light gray quartz-muscovite schist are included in this association. These are very micaceous rocks, showing strong microfolding of probable late Caledonian age. The rocks are

Figure 5: Photomicrograph of static metamorphic texture, calc-silicate rock, Hesjevann assemblage.

Needles and elongate prisms are tremolite; stubby prisms are clinozoisite. Matrix is quartz, microcline, plagioclase, and opaques. Plane polarized light, 75X.

Figure 6: Photomicrograph of schistose hornfels, Hesjevann assemblage.

Granoblastic and decussate textures of quartz and biotite in lower left are typical of bulk of rock. Trace of the late spaced solution cleavage is at upper left. 33X, Plane polarized light.

Figure 5.

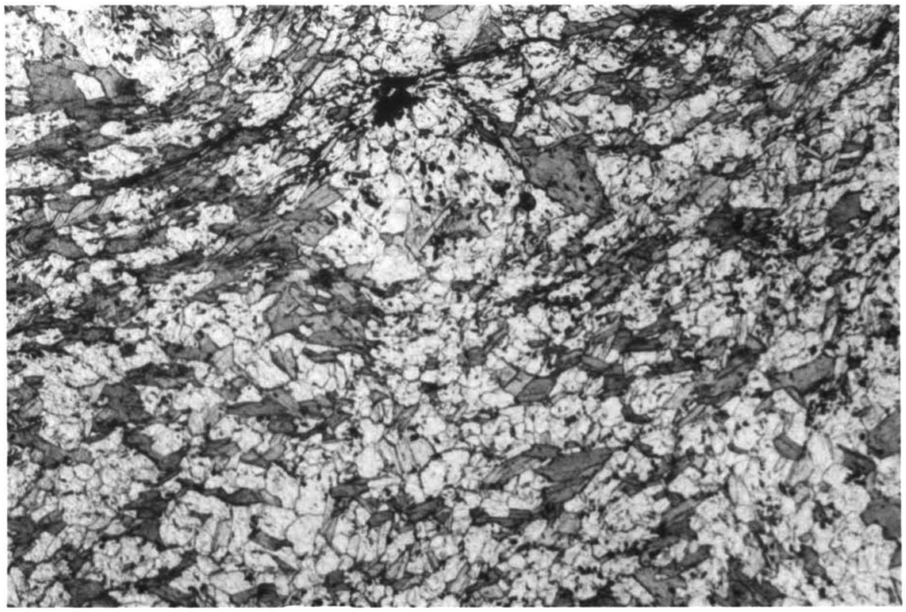
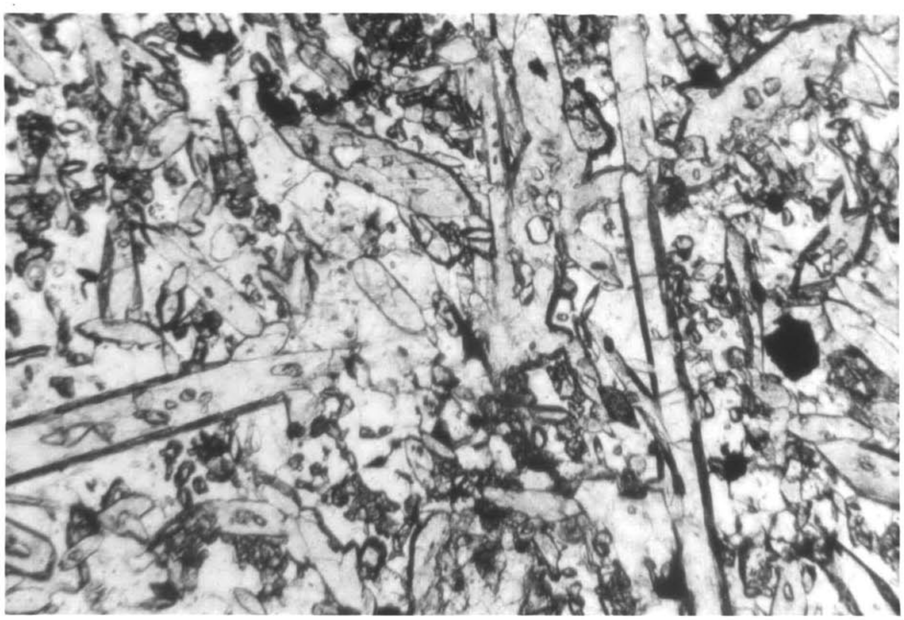


Figure 6.



strongly schistose, with dimensional preferred orientation of both quartz and mica. Micas are polygonized around microfolds, and locally flakes grow parallel to axial planes of the folds. Textures are granoblastic to lepidoblastic. Quartz grains have straight boundaries indicating an absence of post-crystalline strain.

Amphibolite

A major component of the Hesjevann assemblage is dark gray to black, massive and homogeneous amphibolite. Compositional banding is rare, and tectonic fabrics are generally weak or absent. Where present, a lineation defined by hornblende elongation is generally better developed than any planar fabric. An exception to this is in some of the amphibolite outcrops southeast of Storvann(S) where a spaced cleavage is locally prominent.

Petrographically, this lithology is dominated by blue-green hornblende and plagioclase, with clinozoisite, biotite, and garnet variably present as minor phases. Accessories include sphene, apatite, opaques, and rutile. Cleavage, where developed, is defined by spaced (solution?) surfaces with development of retrogressive minerals that include epidote after plagioclase and hornblende, and chlorite after biotite and hornblende. Later fractures in some specimens contain calcite, quartz, and tourmaline. Textures are granoblastic to decussate, and grain size ranges from 0.5 to 2mm. Randomly-oriented subhedral porphyroblasts of hornblende (up to 8mm) are sometimes present. These textures indicate recrystallization under static conditions; whether this is preserved from contact metamorphism by the Melaa Granite or is a result of Caledonian heating without penetrative deformation is unclear.

The massive, homogeneous appearance of this rock, together with its characteristic associated lithologies (white quartzite, calc-silicated marble) allows easy distinction of the rock from other amphibolites in the area, and suggests an intrusive origin.

Kvaefjord Group

General Characteristics

This group of mainly metasedimentary rocks crops out primarily at the head of Kvaefjord. Paragneisses in the core of the synformal anticline east and south of Harstad are also believed to belong to this group. The Kvaefjord Group comprises hornblende-epidote paragneisses, semipelitic paragneisses and schists, and quartzofeldspathic gneisses, with rare calcareous schists and marbles. A granite-pebble conglomerate (or pseudoconglomerate?) is present on the west side of Torskvatsfjell. It is unclear to what extent the quartzofeldspathic gneisses may include tectonized granitoid rocks. In part this unit has unavoidably become a "garbage-basket" designation.

No rocks from western Hinnøy or Lofoten are easily correlated with rocks assigned to the Kvaefjord Group. They constitute an apparently genetically distinct group in the Lofoten terrain, and consequently the name Kvaefjord Group is proposed. Possible correlatives may be the early Proterozoic supracrustal sequence of Lofoten (Griffin and others, 1978), but the absence of iron formation and metavolcanic rocks makes this correlation questionable.

The Kvaefjord Group is clearly intruded by the Middagstind Quartz Syenite, and hence is at least early Proterozoic in age. Contacts between the Kvaefjord Group with other basement rocks are seldom well-exposed and generally ambiguous due to intense deformation, mostly prior to 1700 Ma. On Torskvatsfjell, xenoliths of rocks similar to the Kvaefjord Group occur in the Melaa(?) Granite, and the contact of the granite on the north side of Torskvatsfjell appears to be gradational through a migmatitic border zone into Torskvatsfjell paragneisses.

The Kvaefjord Group has experienced a long, complex history including polyphase folding and probably faulting, polymetamorphism, and intrusion by granite and syenite. No primary structures have been preserved. Consequently, no lithologic sequence has been established, and no meaningful thickness can be determined for the group. Order of discussion below is based on the areal importance of rock types and implies no relative age interpretation.

Semipelitic Paragneiss and Schist

Two main lithologies are included in this unit which were not separated on the map: a quartz-muscovite-biotite schist and gneiss, and a quartz-garnet-mica gneiss. Neither show significant compositional banding, but minerals are segregated into lensoid clusters on the order of a centimeter in length. Compositionally these rocks, especially the garnet-bearing variety, are similar to the quartz-garnet schist of the Storvann Group (see below). The distinction is made on two characteristics: these rocks are generally coarser-grained and more micaceous, and the associated rock types are distinctly different.

The quartz-two mica schist is dark gray or brown and often slightly rusty in outcrop. It is very micaceous so that it is usually intensely crenulated. Schistosity is defined by anastomosing bands of mica separated by lensoid aggregates of granular quartz. Quartz veins one to several centimeters thick are ubiquitous, but feldspar-bearing pegmatites are generally absent. Thin layers (5-50 cm thick) of quartzite with micaceous partings are occasionally present.

In thin-section, these rocks are very quartz-rich, containing up to 70% quartz with micas making up essentially the rest of the rock. Muscovite generally dominates over biotite. Quartz is granoblastic, with sutured grain boundaries indicating some late to post-metamorphic strain. Micas are lepidoblastic, forming bands of both micas intimately intergrown with occasional cross-micas. Zircon and opaque minerals are the main accessory phases.

The garnet-bearing rocks generally are lighter colored and appear more quartzose and less micaceous in the field. This apparent difference is not confirmed by thin section study, however. Garnet makes up about 5% of the rock in the form of 1 cm subhedral porphyroblasts. Micas are intimately intermingled in intensely microfolded lepidoblastic films. Accessory phases include zircon, blue-green tourmaline, and opaque minerals.

The protoliths of these rocks were most likely semipelitic and pelitic terrigenous sedimentary rocks. With the absence of relict sedimentary structures nothing can be said about original depositional environment.

Hornblende-clinozoisite Paragneiss

This rock type forms the bulk of the outcrops in the core of the Kanebogen synformal anticline east of Harstad. Especially good exposures are present in recent road cuts along the main highway leading south from Harstad. The rock is medium gray to gray-green in color, and banded on the scale of one centimeter. A strong penetrative gneissosity is typical, although less micaceous varieties do not generally break along this foliation.

Leucocratic bands are composed of quartz (40-60%), plagioclase (0-40%), microcline (0-10%), and clinozoisite (0-20%). Mafic bands are dominantly biotite (20-30%), blue-green hornblende (0-10%), clinozoisite (10-30%), and plagioclase (0-40%). Accessory minerals include calcite, sphene, white mica, pyrite, and opaque oxides. The rocks are generally even-grained and granoblastic in the quartzose bands, decussate in mafic bands with strong dimensional and lattice preferred orientation of mica and hornblende defining the schistosity. Occasionally small subhedral hornblende porphyroblasts (up to several millimeters long) are present, parallel to the primary schistosity but not generally aligned. Some fine-grained specimens show thin quartz ribbons suggesting deformation was in part mylonitic, but in thin section all grains are now uniformly polygonal, indicating post-kinematic annealing.

The protolith of this unit is believed to be a somewhat calcareous terrigenous sedimentary rock on the basis of its high silica content, high percentage of Ca-silicates, and the presence of calcite. Alternatively, the protolith could have been a volcanogenic sediment. The highly banded nature of the rock is also supportive of a sedimentary origin, although metamorphic differentiation cannot be ruled out.

Quartzofeldspathic Gneiss

Fine-grained, light-colored quartz-feldspar-biotite schists and gneisses are common on east Hinnøy, especially in association with the rock types described immediately above. These gneisses almost certainly include both feldspathic metasedimentary rocks and some fine-grained granitoid rocks, but to distinguish between these two rock types in the field can be difficult in the extreme. This led Gustavson (1972, 1974a, b, c) to include extensive terrains in a blanket designation of "meta-arkose and fine-grained granites". Much of these areas have been more effectively reinterpreted in this more detailed study, but the problem is real and has not been fully resolved during this study.

Reconnaissance petrographic study has shown that most of the fine-grained, well-banded feldspathic rocks are probably from sedimentary protoliths. Quartz contents range up to 75%, which seems unlikely for a granitoid rock, and alkali feldspar, prominent in basement granitoids, is generally present only as an accessory. Intimate association of the quartzofeldspathic gneiss with definite metasedimentary rocks is interpreted to indicate a primary stratigraphic relationship.

Similar lithologies have often been considered "meta-sparagmites" in Scandinavia, since in non-metamorphic areas, the Eocambrian rocks locally include thick sequences of feldspathic sandstones, which would form quartzofeldspathic gneisses when dynamothermally metamorphosed. The Kvaefjord Group quartzofeldspathic gneisses are traceable into the contact aureole of the Middagstind Quartz Syenite where they are truncated and contact metamorphosed. Clearly these gneisses are compositionally similar to "sparagmite" but of early Proterozoic age. Hence, great care must be taken in correlating feldspathic metasedimentary rocks with "sparagmite" on lithologic grounds alone.

Middagstind Quartz Syenite

Introduction

In the northwest corner of the study area, and extending northward from it onto northern Hinnøy, is a massive, coarse-grained syenite pluton which appears to be post-kinematic. Other than rare, local shear zones of protomylonitic to mylonitic texture, the syenite is completely without tectonic fabric. It crosscuts structure in surrounding country rocks, and related dikes and apophyses are unfolded and generally unfoliated. It developed an extensive contact aureole of hornfels on its south and east sides (mysteriously, it is thinner on the west side). The hornfels is also generally massive and unfoliated, although some areas are distinctly sheared, as are the contacts between the syenite and its wallrocks in many localities on the east side of the pluton. It was thus surprising for this author to find that the Middagstind Quartz Syenite gives an Rb/Sr whole rock age of 1726 my (see Chapter 5), thus predating Caledonian events.

The syenite is generally pink to tan with rounded dark dots of mafic minerals 5 mm across (Figure 7). It is not extremely resistant to weathering, perhaps due to its low quartz content, and forms extensive talus slopes and boulder fields. In the study area, it seems to be scarcely

Figure 7: Middagstind Quartz Syenite, hand specimen.

Note massive, unfoliated texture and irregular mafic clots. Actual size.

Figure 8: Photomicrograph of exsolution textures in alkali feldspar, Middagstind Quartz Syenite.

Three types (episodes?) of plagioclase exsolution are recognized:

(1) Rounded blebs (here visibly saussuritized); (2) irregular laminae, oriented vertically in this view; and (3) very fine planar lamellae, oriented horizontally in this view. Crossed polars, 75X.

Figure 7.

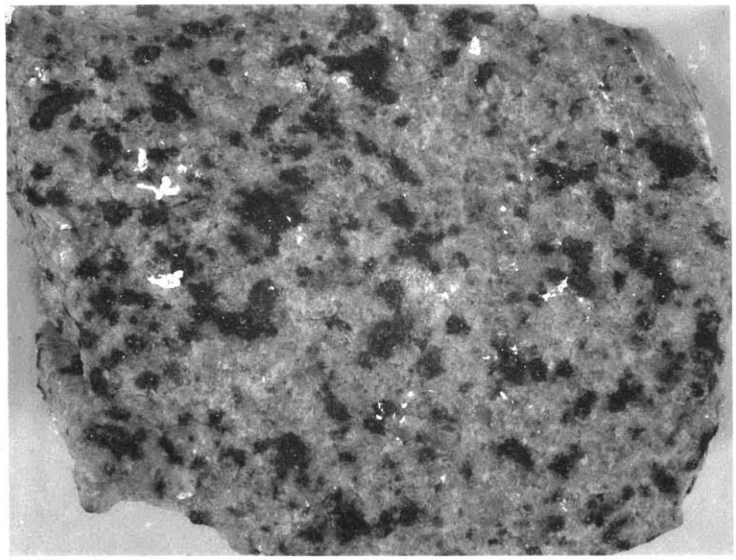


Figure 8.

unroofed. On Middagstind, a small erosional remnant of the roof is perched directly on top of the mountain, with a subhorizontal contact with the underlying syenite (see Plate IIC, section H-H').

The syenite pluton is quite homogeneous. Chilled border zones are slightly more fine-grained and slightly more mafic, but are only a few meters to a few tens of meters thick. The southwestern portion is slightly more leucocratic; amphibole is less prominent, and one specimen (291) may have contained primary quartz.

The only area where the syenite is deformed significantly is the apophysis 2 km southeast of the summit of Middagstind, where steep north-dipping shear zones are present (see Plate I). Here a mylonitic fabric is developed, imparting an augen gneissic texture to the rock. Porphyroclasts of feldspar are separated by mylonite zones (1-5 mm thick) which nucleated on, and then obliterated mafic clots.

The intermediate, alkali feldspar-rich composition, age, and Sr isotopic initial ratio (0.7061; see Chapter 5) are all consistent with the Middagstind Quartz Syenite having been part of the major plutonic episode which emplaced mangeritic rocks in Lofoten-Vesteraalen from 1800 to 1700 Ma (Griffin and others, 1978). Some interesting contrasts exist, however. The mangerite suite of Lofoten was intruded under regional granulite facies conditions, yet the Middagstind pluton developed a contact aureole. Both clino- and orthopyroxenes are essential constituents in the Lofoten mangerites, but no pyroxene has been observed in the Middagstind body.

Petrography

The syenite is composed of complexly exsolved alkali and plagioclase feldspars, surrounding mafic clots with complex mineralogy and striking textural relationships. Microcline is the dominant feldspar, comprising 50 to 70% of the rock. It is invariably perthitic, in every stage of plagioclase exsolution. Stringers and blebs of plagioclase in perthite grains can be traced into areas of subequal patchy intergrowth of two feldspars, and further into separate grains of plagioclase. Some microcline grains suggest as many as three phases of plagioclase exsolution on the basis of different morphology and orientation of exsolution lamellae, and cross-cutting lamellae (Figure 8). The complexity of these relations argue for a complex thermal history, which is consistent with relations of mafic minerals.

Plagioclase is relatively abundant for a syenite (15-35%); some individual specimens might be more correctly termed monzonite. The present composition is estimated to be calcic oligoclase (An_{20}-An_{30}) determined by optical extinction angle. The complex exsolution textures make it difficult to determine how much plagioclase was present in the primary igneous mineral assemblage. Alteration of plagioclase to form fine-grained, disseminated clinozoisite is ubiquitous; polysynthetic twinning is only moderately developed. Consequently, the easiest way to distinguish plagioclase from microcline in orientations not favorable for twinning is by the extensive retrogression of igneous plagioclase to clinozoisite.

The mafic clots appear to be corona-type structures, but are no longer due to near complete replacement of the original mineralogy (Figure 9). The Middagstind syenite was almost certainly pyroxene-bearing at the time of its crystallization, (see "Hornfels" below, and Chapter 4), but no pyroxene relicts have been found in any of the thin sections examined. The central elements in these annular structures are either an opaque oxide (magnetite $\pm$ ilmenite) or a mesh of biotite + quartz. The opaque oxide is invariably rimmed with sphene or leucoxene, as granular aggregates or radiating bladed collars. The sphene is then commonly overgrown by biotite in radiating sprays. Intergrown meshes of quartz and biotite may be peripheral to the opaque/sphene/biotite aggregates, or may grow independently. Small, equant grains of clinozoisite are commonly dispersed in minor amounts in both types of biotite aggregates.

Small amounts of ferrohastingsite (0-5%) are common, and typically occur as rims of the quartz-biotite mesh-like aggregates. Grains now physically separated by quartz and biotite are in optical continuity, suggesting partial replacement. Textures are not conclusive, but ferrohastingsite is a common amphibole in alkalic plutonic rocks and the preferred interpretation is that this is primary amphibole which formed as late igneous overgrowths on orthopyroxene. The orthopyroxene has been replaced entirely in retrogressive metamorphism to biotite and quartz, but the amphibole was still a stable phase and persists, rimming the pseudomorphs.

Muscovite and garnet occur in accessory amounts. Muscovite generally appears as coarse euhedra within the mafic clots, growing across the other grains; it appears to represent a late metamorphic phase. Garnet also

Figure 9: Photomicrograph of corona-like mafic clots in Middagstind Quartz Syenite.

Note biotite + quartz intergrowths which probably pseudomorph primary orthopyroxene, and ferrohastingsite rim. High relief mineral rimming opaque grains is leucoxene. Plane polarized light, 21X.

Figure 10: Photomicrograph of granoblastic texture of hypersthene hornfels.

High relief, colorless grains are hypersthene; colored, transparent grains are brown hornblende; low relief, colorless mineral is plagioclase. Plane polarized light, 75X.

Figure 9.

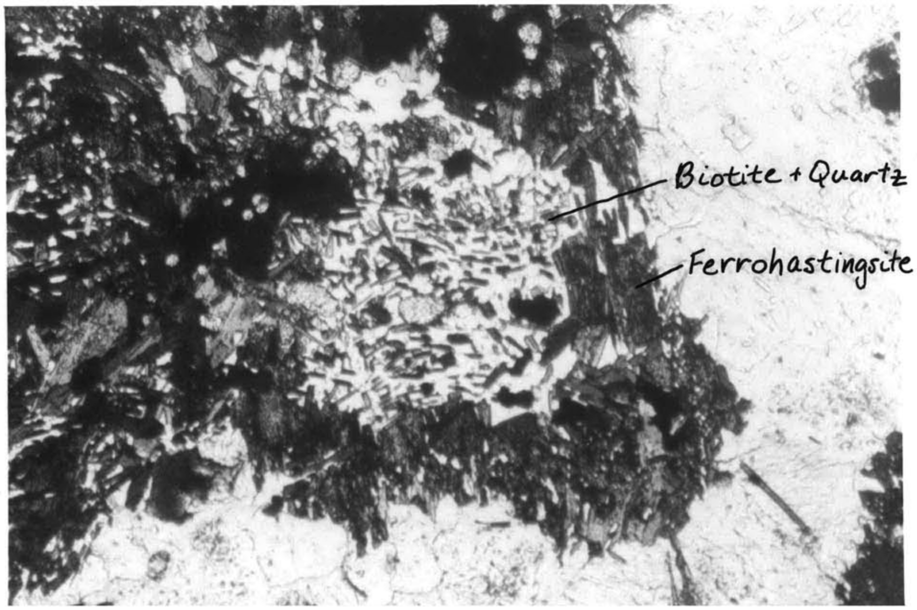
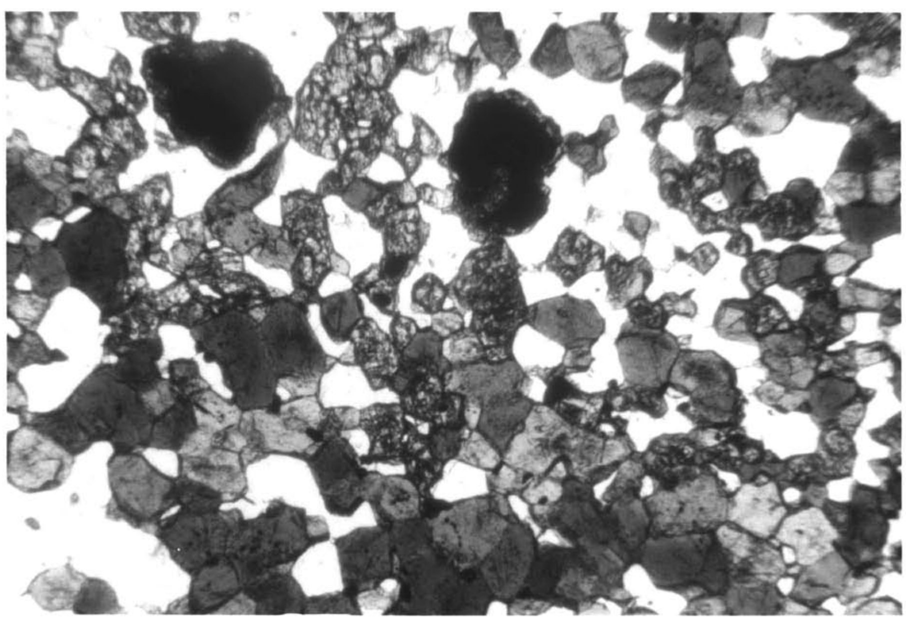


Figure 10.



generally occurs in association with other mafic phases, especially as a rim on sphene. Garnet may also occur as isolated grains in the feldspar matrix; its paragenetic significance is not entirely clear.

The most prominent other accessory phase is apatite which locally makes up nearly 1% of the rock. It commonly occurs as euhedra, up to 1 mm in diameter, which are poikiloblastically enclosed in all other phases described above.

Hornfels

The eastern and southern sides of the Middagstind pluton, and to a lesser extent the western side, preserve a hornfelsic contact zone in wall rocks. The hornfels consists mostly of mafic to intermediate rocks, possibly developed from the Archaean hornblende diorite, although more quartzofeldspathic rocks of the Kvaefjord Group are also affected. The hornfels is a massive charcoal gray or black, fine-grained, generally structureless rock in the field. Locally, pre-contact metamorphic compositional banding can still be faintly recognized. In some areas, especially southeast of Middagstind, the hornfels has been sheared to form a fine-grained, finely-foliated and lineated amphibolite. This effect is very zonal, and appears to be the result of ductile shearing.

The areal extent of the hornfelsic rocks is surprising on the east and southeast sides of the pluton: the exposed width of the contact zone is one to two kilometers. This is most likely due to low dip of the contact. The thickness of the contact zone is much less than its outcrop width, perhaps only a few hundred meters (see Plate IIA, A-A'; IIC, H-H').

The hornfels was not extensively sampled, so that only general comments about the character of the rocks can be made. No isograds could readily be mapped in the field, but different grades of contact metamorphism were recognized by petrographic study. The highest grade assemblage observed came from the immediate contact zone of the pluton on the east side of Middagstind itself (sample 16D; see location map, Figure 67). Iron-rich hypersthene ($2V \approx 40°$), green-brown hornblende, plagioclase (An_{60}, by Michel -Levy optical method), and opaque oxide coexist in a granoblastic aggregate (Figure 10). Retrogressive reaction has occurred to a limited extent, developing thin garnet rims on the opaque phase, surrounded by aggregates of plagioclase. Hypersthene is often sieved with plagioclase.

Other thin sections examined were not orthopyroxene-bearing. A typical assemblage is blue-green hornblende + plagioclase + sphene + opaque ± biotite. One thin section included two coexisting amphiboles, actinolite and grunerite. Actinolite may rim the grunerite, or the two may be irregularly or tabularly intergrown. Textures are generally granoblastic to decussate; rocks with mesoscopically visible relict fabric often preserve some amphibole preferred orientation, both dimensional and lattice, which suggests static metamorphic overprint occurred at conditions not very different from those of the synkinematic metamorphism which predated the syenite intrusion.

Samples of the hornfels which have been subsequently foliated also display amphibolite facies mineralogy, but the textures clearly indicate synkinematic metamorphism, with strong dimensional and lattice preferred orientation of hornblende. The lack of any static overprint on these samples implies that deformation post-dates the pluton and may well be Caledonian in age. This is supported by the fact that sheared contact rocks from the immediate contact of the pluton have precisely the same assemblage, blue-green hornblende + plagioclase + sphene, all trace of pyroxene having disappeared. Structural arguments for the timing of these shear zones are presented in Chapter 3.

Austerfjord Group

Introduction

The Austerfjord Group consists of a dismembered sequence of schists, amphibolite, quartzite, and marble, first described by Hakkinen (1977). In the present study, the Austerfjord Group was traced southeastward in the lower plate of the Austerfjord Thrust, from the type area to Kongsvik, and thence across Tjeldsund to the Fjelldal area. The rocks that crop out in the extreme northwest of the map area are also tentatively correlated with the Austerfjord Group, but have not been investigated in detail.

In the map area, the main lithologies represented are sericite quartzite and calcareous amphibolite (the iron-stained amphibolite of Hakkinen, 1977). These have been traced from Hakkinen's map area and constitute the two structurally highest map units of his Østerdalen sequence (of the Austerfjord Group). The only locality where lower units are exposed to any extent is the stream bed between Tverrfjell and Kongsviktind (Kongsvikdalen). The near absence of the biotite schists which dominated the type section is probably the result of displacement by the intrusion of the Lødingen Granite. In outcrops two kilometers west of the summit of Kongsviktind, Austerfjord Group amphibolite and intensely tectonized Lødingen Granite are interlayered, apparently by intrusion of the granite. The contact between massive granite and amphibolite is not exposed, but numerous deformed small (<1 m thick) granite dikes inject the amphibolite.

A minimum age for the Austerfjord Group is established by an unpublished Rb/Sr whole rock age of 1415 ± 80 Ma on the Lødingen Granite by P. N. Taylor (pers. comm. to Hakkinen, 1977) on samples from the granite where it intrudes the Austerfjord Group in its type locality at the head of Austerfjord. The conclusion of Hakkinen that the Austerfjord Group underwent prograde metamorphism during Caledonian time suggests its deposition post-dates the 1700 Ma high-grade metamorphism. Thus, the age of the Austerfjord Group can be bracketed with reasonable confidence within a period of 300 Ma.

The base of the Austerfjord Group is not preserved anywhere yet studied, nor have any primary sedimentary structures been recognized to indicate stratigraphic facing. From place to place, the precise sequence of lithologies is variable, although the overall order is consistent (Hakkinen, 1977); inversions in the order of units are rare. It is likely that the sequence is not strictly stratigraphic. Hakkinen (1977) suggested that most contacts are in fact low-angle faults (tectonic slides) which have "shuffled" the sequence. On what kind of basement, and in what type of depositional environment, the sequence originally formed, is uncertain.

Lower Units (Kongsvikdalen sequence)

In the stream cut section between Tverrfjell and Kongsviktind, the following sequence of lithologies was observed passing upward (Table 1): (1) massive quartzofeldspathic schist with minor fissile biotite schist, ca. 30 m, grading upward into hornblende-bearing, darker colored schists, 20 m; (2) a thin (3 m) calcareous quartz-muscovite schist that grades upward into massive to banded buff-colored tremolite-phlogopite marble,

TABLE 1: Comparison of lithologic successions of Austerfjord Group rocks.

Østerdalen sequence (Hakkinen, 1977)

Unit 7	
Quartzite	(10m)
Quartzofeldspathic schist	(35m)
Fine grained biotite schist	(20m)
Unit 6	
Iron-stained amphibolite	(120m)
Unit 5	
Tremolite-bearing marble	(40m)
Calcareous schist	(40m)
Biotite rich amphibolite	(10m)
Unit 4	
Garnet muscovite schist	(55m)
Biotite schist, minor amphibolite	(10m)
Sericite quartzite	(10m)
Unit 3	
Garnet amphibolite, intercalated with marble	(45m)
Unit 2	
Chlorite biotite schist	(90m)
Unit 1	
Kyanite garnet biotite schist	(20m)
Biotite schist, finely laminated	(150m)

Kongsvikdalen sequence (This study)

Schistose amphibolite	(20m)
Quartzofeldspathic schist with marble layers	(25m)
Quartz muscovite schist	(10m)
Tremolite phlogopite marble	(20m)
Calcareous schist	(3m)
Hornblende bearing biotite schist	(20m)
Quartzofeldspathic schist	(30m)

20 m; tremolite prophyroblasts grow unaligned but within the primary schistosity; (3) quartz-muscovite schist with minor randomly-oriented porphyroblasts, 10 m; (4) 1 m of white quartzite, overlain by 1 m of calcite marble, in turn overlain by 25 m of quartzofeldspathic schist with 5 to 25 centimeter thick layers of pink marble; (5) schistose amphibolite, apparently part of the calcareous amphibolite described in more detail below, ca. 20 m. This amphibolite is succeeded by fine-grained, strongly foliated granite gneiss of the upper plate of the Austerfjord Thrust. The actual thrust contact is not well exposed here, but is very precisely locatable on the map scale.

It is difficult to correlate precisely these units to those described by Hakkinen, although the overall character of the section is similar. The second unit correlates well with Hakkinen's unit 5 (calcareous schist and tremolite-bearing marble), but other units do not match well (Table 1). On such limited data, a unique interpretation is impossible.

Calcareous Amphibolite

This is the dominant map unit in the area studied, reaching a thickness of about 250 meters in the stream cuts on the west side of Kongsviktind. It is mainly moderately to very schistose, dark gray to black amphibolite, with bands of buff-colored marble from several centimeters to about two meters thick. The marble is generally dolomitic, and may be tremolite-bearing. The amphibolite is generally carbonate-bearing. The presence of the marble bands is especially diagnostic, allowing ready distinction of this unit from other amphibolite units of east Hinnøy, especially in the critical Fjelldal area. The abundance of carbonate is also believed to indicate a sedimentary origin for this unit, consistent with Hakkinen's (1977) interpretation that this unit (his iron-stained amphibolite) was a para-amphibolite.

In thin section, the amphibolite is composed of blue-green hornblende, biotite, epidote and clinozoisite, calcite, quartz, and plagioclase. Retrogression of mafic phases to chlorite is extensive in rocks from the Fjelldal area, possibly an effect of the Fjelldal high-angle fault, since similar retrogression of Storvann Group rocks has also occurred here (see Storvann Group, quartz-garnet schist below).

Sericite Quartzite

This milky white quartzite is exposed discontinuously under the Austerfjord thrust on the west side of Tverrfjell, and in the Fjelldal area. Its mesoscopic appearance is porcellainous to finely granular; generally it is finer-grained than the quartzite of the Hesjevann assemblage, lacking a sugary appearance in weathered outcrop. Fine-grained white mica defines the variable-developed foliation. Hakkinen described this unit as flaggy, presumably due to mylonitic deformation along the Austerfjord Thrust. Flaggy structure is variably developed in the present study area; the quartzite is locally quite massive (e.g., the lens at the west end of sections C-C' and G-G', Plates IIA and IIC respectively). If as concluded in Chapter 3, the Austerfjord Thrust does not have large displacement, it may be that the contact between the quartzite and the overlying Gullesfjord Gneiss is not always the most important level of transport. This would explain the fact that commonly on Tverrfjell, the tectonite fabric is strongest <u>away</u> from the granite/metasediment contact; the upper contact of the lens just described shows no evident tectonization where well-exposed near its southern end. Although this does not demonstrate that the contact is primary, it suggests the possibility must be considered. It should be noted that in the Fjelldal area, the zone of contact between Austerfjord Group rocks and the small klippe of Gullesfjord Gneiss is a zone of intense mylonitization, so that significant movement would seem to be localized at this level in that particular area.

The idea that the contact between the uppermost Austerfjord Group quartzite and the Gullesfjord Gneiss may be locally sedimentary suggests a further speculative possibility. If the Østerdalen sequence is considered as a stratigraphic succession, albeit somewhat dismembered, the units of sandy, quartz-rich and of calcareous protoliths are concentrated toward the top while more pelitic and impure psammitic units, with more poorly sorted or fine-grained terrigenous protoliths, are at the bottom. Such a sequence would appear to constitute a regressive historical development. Such sequences are preserved in the stratigraphic record, but less commonly than transgressive sequences because sediment just deposited is more likely to be re-exposed and eroded during a regression. If, however, the upper contact of quartzite with granite was originally primary and

and stratigraphic, then the sequence can be interpreted as inverted. An inverted Østerdalen sequence makes a reasonable transgressive sequence. Thus, the possibility exists that the Austerfjord rocks record a post -Svecofennian (1700 Ma) transgression, terminated by recurrence of magmatic activity at 1400 Ma with the emplacement of the Lødingen (and Melaa?) granite(s).

Melaa Granite

The Melaa Granite is the name proposed for the massive granite exposed on the east side of the Melaa Vannene (Melaa Lakes) area, near Hesjevann. The precise areal extent of this unit is difficult to establish without further geochronological study. During mapping, I assumed all the granite gneiss in the western part of the study area was essentially the same unit, because compositional variation was limited, and textural variation was clearly not a reliable criterion for subdivision since texture and fabric are locally very variable. However, existing geochronological data (Chapter 5) suggest the situation is more complex, and that much of the granite is of Archaean age (Gullesfjord Gneiss). The present author considers that a younger granite, the Melaa Granite, is present, but that its extent is more limited than previously hypothesized.

The Melaa Granite is clearly intrusive into rocks of the Hesjevann assemblage in the Hesjevann area. Dikes of granite cut bodies of amphibolite and quartzite, marble bodies are converted to calc-silicate mineral assemblages along their margins, and all metasedimentary units appear to be "floating" in a "sea" of granite on the map scale (see Plate I, and Plate IIC, section H-H'). Contact relations with other units, in particular the Gullesfjord Gneiss and the Storvann Group, are uncertain. Presumably, the Melaa Granite intrudes the former and lies unconformably beneath the latter, but this cannot be documented at this time.

The Melaa Granite is compositionally a true granite, with 10 to 30% quartz, 40 to 60% perthitic microcline, 15 to 30% plagioclase (calcic oligoclase determined by optical methods), 2 to 10% biotite, and accessory clinozoisite, sphene, muscovite, zircon, apatite, green tourmaline, and garnet in aplitic phases. Opaque oxides are noticeably scarce.

The alkali feldspar is entirely microcline, generally fresh, and not uncommonly rims plagioclase to form rapakivi textures (Figure 11). Perthi-

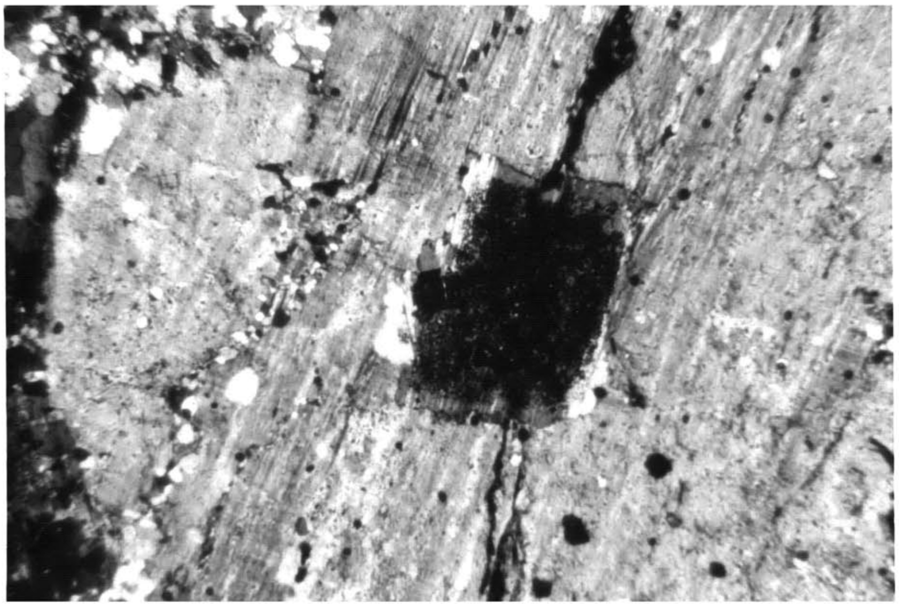


Figure 11: Photomicrograph of Rapakivi feldspar in Melaa Granite.

Dark center is plagioclase strongly retrograded to clinozoisite. Rim is microcline. Note incipient cataclastic band in microcline at left. Crossed polars, 30X.

tic plagioclase exsolution, usually in rounded blobs but also in stringers as well, is ubiquitous. Large phenocrysts are absent, but the alkali feldspar is generally coarser-grained than other phases (up to 1 cm) and in some specimens textures are suggestive of accumulation of early-formed alkali feldspar (R. Wiebe, pers. comm., 1978). This is consistent with the notion that this is a true granite, of probably crustal anatectic origin (see also Chapter 5), with alkali feldspar near the liquidus in the crystallization of the magma.

Plagioclase, calcic oligoclase by optical determinations, is polysynthetically twinned and mildly sieved with retrograde clinozoisite (probably a product of Caledonian thermal overprint); the clinozoisite grains allow ready distinction of microcline from plagioclase in orientations where twinning is ambiguous. Rarely, plagioclase is antiperthitic.

Biotite is brown, tabular, and randomly oriented in small aggregates in undeformed examples. As a whole textures are hypidiomorphic granular; over an area of a square kilometer or more directly north of Hesjevann, no tectonite fabric whatsoever can be distinguished in the Melaa Granite. Discussion of the development of the tectonite fabrics in this granite and the Gullesfjord Gneiss is deferred to Chapter 3, because it is an integral part of the structural picture of east Hinnøy.

Petrographic distinction between the Gullesfjord Gneiss and Melaa Granite is tenuous at best; one gneissic granite often looks much like another, explaining the lack of any striking contact in the field. Three possible criteria exist: (1) alkali feldspar is clearly dominant in the Melaa Granite, while plagioclase is slightly dominant in the Gullesfjord Gneiss (Hakkinen, 1977; this is consistent with the author's own observations); (2) rapakivi textures are present, but not ubiquitous, in the Melaa Granite, but have not been reported from the Gullesfjord Gneiss; and (3) the Gullesfjord Gneiss appears to have been penetratively deformed in Precambrian time, while the Melaa Granite locally is undeformed, preserving primary granitic texture. None of these criteria are particularly useful for field distinctions between gneissic granites. A solution may only be obtained by detailed multi-system geochronological study.

The Melaa Granite is considered tentatively correlative with the Lødingen Granite on the basis of gross lithologic similarity, and similar age (see Chapter 5). If this correlation is valid, it further supports the argument developed in Chapter 3 that the Austerfjord Thrust is not a zone of major translation. Further geochronological and chemical study would better evaluate this possibility.

Caledonian Cover Rocks

Introduction

The Caledonian cover rocks of east Hinnøy were described by Gustavson (1972) as a single stratigraphic sequence, which he considered to comprise the Harstad and Straumsbotn Nappes. The present study concludes that these cover rocks include elements of at least four, and perhaps more, distinct stratigraphic sequences, including both autochthonous and allochthonous units. Hence, Gustavson's nomenclature must be abandoned; new structural nomenclature will be introduced in Chapter 3. Stratigraphic subdivisions established in this study are: (1) the Storvann Group, a newly-distinguished sequence of quartzite, schist, and marble which is considered to be the autochthonous Eocambrian or Cambrian sedimentary cover of the Lofoten terrain; (2) the Narvik Group, represented in the study area by pelitic schists and gneisses with minor intrusions of granitoids and amphibolites, and present only in two tectonic slivers at the base of the nappe pile; (3) the Stangnes Group, a newly-distinguished, thrust-bounded slab of layered amphibolite intruded by a semi-concordant tonalite gneiss pluton, the Ruggevik tonalite; and (4) the Salangen Group, a sequence of marbles and mica schists which comprise the highest unit of the nappe pile exposed on east Hinnøy. All of the above have been metamorphosed at amphibolite facies conditions during the Caledonian orogeny. Juxtaposition of the allochthons and the autochthon occurred prior to or synchronous with metamorphism, so that no abrupt breaks in metamorphic grade are observed.

The use of the term cover here is consistent with its customary use in the Scandinavian Caledonides, but different from its use in some other places. "Cover" refers to both sedimentary cover, that is, the sedimentary rocks deposited during an episode of continental margin development ("geosynclinal cycle"), and to structural cover, that is, the allochthon(s) emplaced upon the autochthonous continental crust during an orogenic event. The structural cover can, and often does, include rocks from the pre-existing continental crust, that is, stratigraphic basement. Due to the intense metamorphism and deformation of most of the rocks in the Scandinavian Caledonides, recognition of this basement involvement in the allochthons is often very difficult. As a consequence, the use of "cover" as primarily a structural term is less ambiguous, if confusing to those accustomed to classical Alpine usage. Thus, the Storvann Group is actually part of the structural basement in this usage.

Storvann Group

Introduction

The Storvann Group consists of a sequence of metasedimentary rocks derived from mainly quartz-rich terrigenous protoliths. It is exposed extensively in the eastern half of the study area from the southern outskirts of Harstad to Storvann(S), and in some critical outcrops in the extreme southeast of the study area near Fjelldal and on Ramboheia. Outcrops along the eastern shore of Storvann(S) are designated as a proposed type section, since a relatively complete and coherent (though still strongly deformed) sequence of rocks from its contact with pre-Caledonian basement to thrust-truncated top is exposed (Figure 12A). Another excellent set of exposures, tectonically thinned but essentially complete, is present along the shoreline and in roadcuts east of Kanebogen to the Harstad NAF campground (Figure 12B).

Contact Relationships

The basal contact of the Storvann Group is exposed at several places: (1) on the eastern shoulder of Vikelandsfjell, (2) the north side of Middagsfjell, (3) on the shoreline at Kanebogen, (4) on the ridge west of Kanebogen (the other limb of the Kanebogen synformal anticline from the shoreline section), (5) in a road cut and on the shoreline 1.5 km south of Kanebogen, (6) northwest of Sørvikfjell where the basal quartzite unit is so extensively exposed, (7) the summit area of Finnslettheia, (8) the east shore of Storvann(S), (9) in roadcut and shoreline exposures at Tjeldsund on strike from the Storvann(S) section (Figure 13), (10) on the shoreline near Fjelldal, and (11) on the southwest side of Ramboheia. In all cases, the compositional layering of the basal quartzite is parallel

Figure 12.

LITHOLOGIC SUCCESSIONS, EAST HINNØY

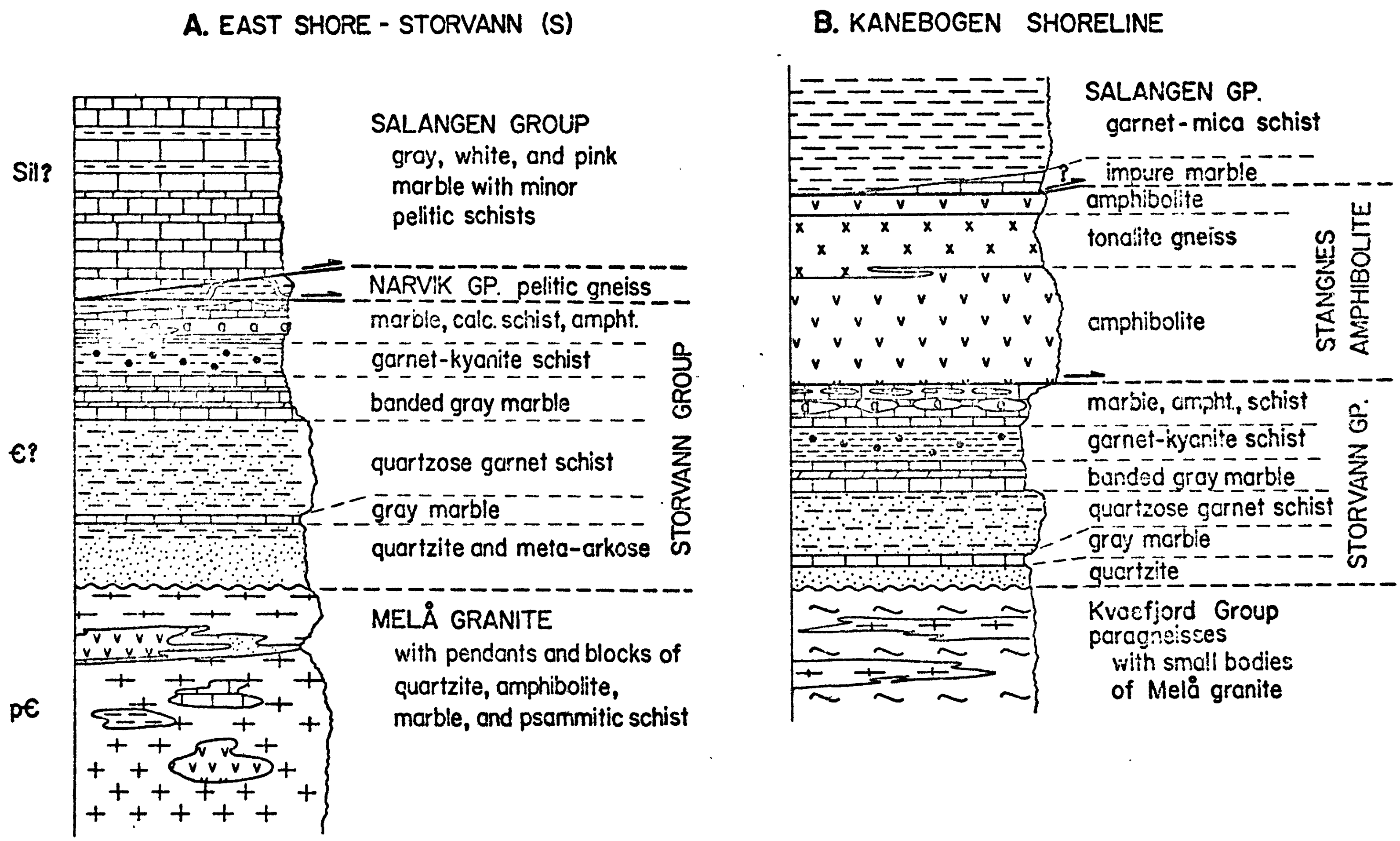


Figure 13: Basal contact of the Storvann Group at Tjeldsund.

Contact is at center of photograph, with strongly deformed basement granite at left, white quartzite and quartzofeldspathic schist at right. Folds are F_3 (see Chapter 3).

Roadcut is along highway on west side of Tjeldsund, about about 1 km south of Tjeldsund bridge.

to the contact, and although the rocks are tectonized, there is no suggestion of concentrated strain at the contact. On the map scale, lithologic units of the Storvann Group in general carry through parallel to the basal contact. Three of the above exposures (2, 10, and 11) are anomalous in that little or none of the basal quartzite unit is present, and the quartz-garnet schist (see below) is in direct contact with the basement or separated from it by only a few centimeters to a meter of quartzite. At these places, it is inferred that the contact was sheared early during the penetrative deformation and thrusting so that no mylonitic fabric was preserved, or perhaps mylonites were never formed. Rocks at the base of the Storvann Group were removed in the process; these contacts are thus considered tectonic slides in the sense of Fleuty (1964). Strictly speaking, it might be preferable to consider the Storvann Group rocks at these places, Ramboheia in particular, as parautochthonous rather than autochthonous.

The Storvann Group comprises a transgressive sequence. At the base is an impure quartzite which is overlain by progressively more aluminous, though still quartz-rich, rocks, interrupted by two marble horizons. This compositional change is inferred to reflect fining upward of the terrigenous clastic input to the sedimentary basin, consistent with a transgressive setting. How much of the sequence has actually been preserved is unclear, because the stratigraphically upper contact is everywhere a thrust fault.

The combination of the above relations, that is, (1) lack of discordance of lithologic units at the map and outcrop scales, (2) consistency of rock sequence above the pre-Caledonian basement, (3) lack of evidence of concentrated strain at the basal contact, and (4) the transgressive nature of the sequence, are taken to indicate that the basal contact of the Storvann Group is a nonconformity.

The upper contact of the Storvann Group is always marked by a thrust fault, but is not necessarily the same thrust everywhere. On the Stangnes peninsula, the Storvann Group is overlain by the Stangnes Amphibolite with an intervening zone of complex mixing of lithologies (see "Calcareous schists" below for further discussion of this zone). On the east shore of Storvann(S), an entirely different assemblage of rocks overlies the Storvann Group. Above the highest schist ("Pelitic schist" below) which is considered part of the Storvann Group, is an assemblage of mixed lithologies similar to that on the Stangnes peninsula, which is in turn overlain by a slice of Narvik Group pelitic gneiss and schist, which is then overlain by marble of the Salangen Group. The allochthony of the latter two groups is well established (Gustavson, 1966, 1972; J. Tull, pers. comm., 1979; K. Hodges, work in progress), as well as being clearly demonstrable an east Hinnøy (see descriptions of these units below). An interesting point is that such completely different rocks occur at the upper contact of the Storvann Group (see Figure 12). This will be discussed further with regard to the geometry of the Caledonian nappe pile (Chapter 3).

Correlation and Thickness

The impure, often feldspar-rich, nature of the basal quartzite, and the mainly terrigenous character of the section, combine to suggest the correlation of this sequence with the autochthonous Eocambrian/Cambrian sections of the Scandinavian foreland. The basal "sparagmite" unit typical of the autochthonous sections preserved at the eastern thrust front is feldspathic quartz sandstone and arkose, which would form a feldspathic quartzite upon metamorphism. At this latitude (68-70°N), autochthonous cover sequences along the eastern thrust front have been described by Moberg (1908) and Vogt (1918, and unpublished data summarized by Gustavson, 1966). Gustavson (1966) has described autochthonous sequences from tectonic windows through the nappe pile. The sections are all largely sandstone and shale, fining upwards, with minor limestone horizons. The sections described by Vogt are fossiliferous, bearing Cambrian faunas (*Platysolenites antiquissimus*, *Torellella sp.*, *Hyolithes sp.*, and *Strenuella*). Gustavson's section from the Dividalen window is similarly dominated by sandstone and shale, while the other sections from windows are too incomplete for reasonable comparison. The overall similarities are encouraging. Correlation unit by unit is not attempted, due to (1) geographical separation (more than 100 km), and (2) the difference in metamorphic grade (unmetamorphosed versus amphibolite facies). Nevertheless, I would assert strongly the correlation of the Storvann Group with the autochthonous sequences of the Scandinavian foreland, supporting the hypothesis of Griffin and others (1978) that the Lofoten terrain was indeed part of the Scandinavian craton in pre-Caledonian time. This is also consistent with structural arguments, to be presented in Chapter 3, that the Lofoten block is not a far-travelled nappe.

The overall thickness of the Storvann Group is highly variable, due to large variations in strain history from one locality to another. Original thicknesses can only be conjectured. At Kanebogen, the entire section from the basal contact to somewhere within the upper marble unit (i.e., nearly the entire known section) is about 300 meters thick (Plate IIA, section A-A'). On Middagsfjell, the quartz-garnet schist unit alone is about 700 meters thick (also see section A-A').

Lithologies

Quartzite

The basal unit of the Storvann Group is a heterogeneous mixture of micaceous quartzite, vitreous quartzite, and quartz-feldspar-biotite schist, with local occurrences of quartz-biotite schist and garnet-mica schist. A separate upper member consisting of the latter lithologies was mapped separately in the area about 2 km south of Storvann(N).

The vitreous quartzite is usually present as compositional bands a few centimeters thick, separated by more micaceous layers one mm to several cm thick. The vitreous layers are fine-grained and range in color from white to bluish gray and rarely dark gray. The micaceous quartzite is similar but has mica in discontinuous films and as disseminated grains defining the schistosity. The compositional banding on a mesoscopic scale has been transposed by isoclinal folding (F_1 and F_2) and has little primary significance except that its strict concordance at the immediate contact with the basement supports the interpretation that the contact is stratigraphic.

Rocks with higher feldspar contents are present in this unit, especially in its lower parts. In particular, on the north side of Finnslettheia it is locally difficult to locate the basement/cover contact due to the compositional similarity of the feldspathic cover (here a "meta-arkose") with the granitic basement. In such cases, the contact was drawn on the basis of: (1) the appearance of vitreous or micaceous quartzite interlayers, and (2) the better layering and finer grain size of the metasedimentary rocks (in most cases). On and near Torskvatsfjell where the basement is dominated by metasedimentary rocks of the Kvaefjord Group, the identification of the basal contact can become difficult, and locally can only be made by assuming the structural geometry is reasonably simple. However, in general the basal contact can be traced into areas where the lithologies are more distinctive and the contact more easily located, so that its geometry can be determined with confidence.

In thin section, the feldspathic quartzites contain both microcline and saussuritized plagioclase in subequal amounts, forming from a few percent up to perhaps 25% of the rock. In general, microcline is dominant in less feldspathic rocks while plagioclase is more important in the "arkosic" rocks. The thoroughly recrystallized nature of the rocks indicates that the saussuritization is a result of late retrogression rather than reflecting the sediment source. A change from muscovite to biotite as the dominant mica also accompanies the transition from less to more feldspathic compositions. The micas, though concentrated along compositional bands, are commonly individually rotated into an orientation parallel to axial planes of late folds. The rotation of micas and the resulting intersection lineation commonly are present even in outcrops which show no mesoscopic late folding.

The basal several meters of the quartzite unit usually consist of vitreous quartzite or meta-arkose, but on the east shore of Storvann(S) and at the north end of the Stangnes peninsula, a quartz-biotite schist with small feldspar porphyroblasts is present at the base. It is considered a local facies of the basal Storvann Group, but it is also possible that these are lenses of Precambrian metasedimentary rock occurring locally below the unconformity. Volumetrically, these rocks represent a rather minor element of the sequence, so the resolution of this question is of modest importance.

A layer of quartz-biotite-garnet schist may be locally present at the top of the quartzite unit. On the east shore of Storvann(S), about 5 m of a schist very similar to the quartz-garnet schist unit (see below) is present at the top of the quartzite, while two km south of Storvann(N), 50 m or more of this unit occurs as part of a separate, upper member of the quartzite unit. This lithology is in a sense transitional to the quartz -garnet schist present higher in the sequence, although generally separated from it by the lower marble unit.

The other lithology present in the upper member of the quartzite unit is a quartz-biotite schist. This rather dark-colored, fissile schist is an atypical rock type in the Storvann Group, but clearly occurs stratigraphically between vitreous and feldspathic quartzite more typical of the quartzite unit and the lower calcite marble. It is thus considered a facies of the basal quartzite unit.

Thickness of the quartzite unit varies, probably mainly due to finite strain, from zero (due to tectonic sliding) to about 250 m northwest of Sørvikfjell. A "typical" thickness for the unit would be about 50 m. The possibility that some of the thickness variations in the quartzite, including its absence in some areas, are due to facies variations (for instance, due to infilling of topography) cannot be ruled out.

The protolith of the quartzite unit was of variable composition but generally a rather mature sandstone with subordinate shaly facies. The absence of a basal conglomerate is notable, and unfortunate because the presence of locally-derived pebbles would reinforce the interpretation that the basal contact is an unconformity. It is important to note that the Storvann Group basal quartzite is readily distinguished from the quartzites of the basement complex, on the basis of lithology and associated rock types, as well as structural position.

Lower Calcite Marble

Often present at the top of the basal quartzite unit is a thin (5-10 m) gray calcite marble. It is commonly impure, containing several percent white mica, minor quartz, and occasional accessory pyrite and/or graphite. Compositional banding is defined by varying concentrations of mica and by subtle shading of color from light to medium gray. Early isoclinal fold hinges are moderately common, indicating transposition of bedding by folding to produce the present compositional banding. Texturally, the calcite is granoblastic, with lepidoblastic mica films. Quartz occurs as disseminated equant grains. Calcite is generally intensely twinned, indicating late to post-metamorphic strain.

A somewhat different facies of this unit is present on the west flank of Finnslettheia. Outcrop is moderate to poor, but normal marble appears to pass laterally into a calc-silicate rock composed of tremolite, calcite, plagioclase, quartz, epidote, and sphene. Phlogopite occurs in trace amounts in one sample. The fabric development of this unit is highly variable (Figure 14), apparently in spatial relation to its position within an antiform on which it mainly occurs in the west limb (Plate IIA, section C-C'). Away from the hinge little penetrative fabric is present and the texture in thin section is granoblastic to decussate. Apparently, the static metamorphism (see Chapter 4) obliterated much of the Caledonian schistosity. Nearer the fold hinge, a strong foliation is developed which is defined by dimensional preferred orientation of grains in rocks of essentially identical mineralogy to those which are unfoliated. Late fold hinges will be discussed further in Chapter 3.

The lower marble unit is not present everywhere. This is at least in part due to tectonic slides between the quartzite and quartz-garnet schist units. In the inverted section on top of Finnslettheia (Plate IIA, C-C'), only three lenses of this marble occur, while the intervening areas where the marble is absent commonly show more intense foliation. This could result from either large-scale boudinage of the marble or tectonic sliding. From what is known of relative ductility of calcite- and quartz-rich rocks at amphibolite facies conditions, it seems unlikely that marble would be the boudin-forming lithology. Hence, a tectonic sliding interpretation is preferred.

The protolith of the calcite marble was presumably a shallow water biogenic limestone with some terrigenous component. The reducing conditions which produced the pyrite and graphite may have resulted from incomplete oxidation of organic carbon. The calc-silicate facies was in part dolomitic (as evidenced by Mg-bearing phases such as tremolite), and bears no phase indicating low f_{O_2}. It is possible that local secondary dolomitization occurred under oxidizing conditions to form the protolith of the calc-silicate facies, at the same time oxidizing all remaining carbon in the rocks.

Quartz-garnet Schist

The quartz-garnet schist is the areally most extensive unit of the Storvann Group, and one of the most distinctive lithologies. In outcrop it is a massive, resistant light gray rock with foliation defined by lensoid or sigmoidal clots of mica one centimeter or less in length. Quartz veins a few mm to a few cm thick, and several cm to several tens of cm long, are

Figure 14: Photomicrographs of calc-silicate facies of lower marble of the Storvann Group. Plane polarized light, 30X.

A: Static texture. Radiating needles are tremolite, in a matrix of calcite and quartz.

B: Foliated specimen. Tremolite needles have been aligned in discrete bands (one at center of photo), with intervening bands of granoblastic calcite and quartz.

Figure 14A.

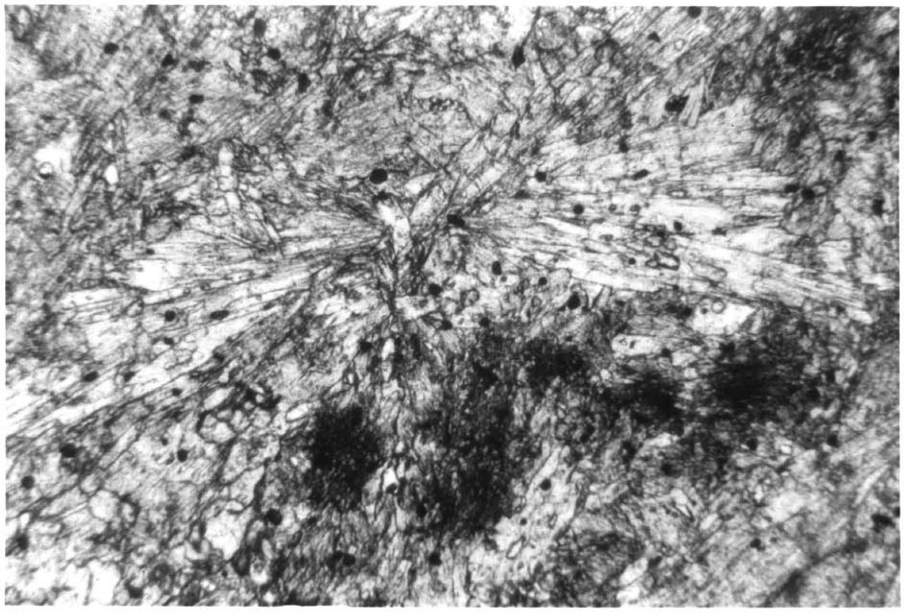
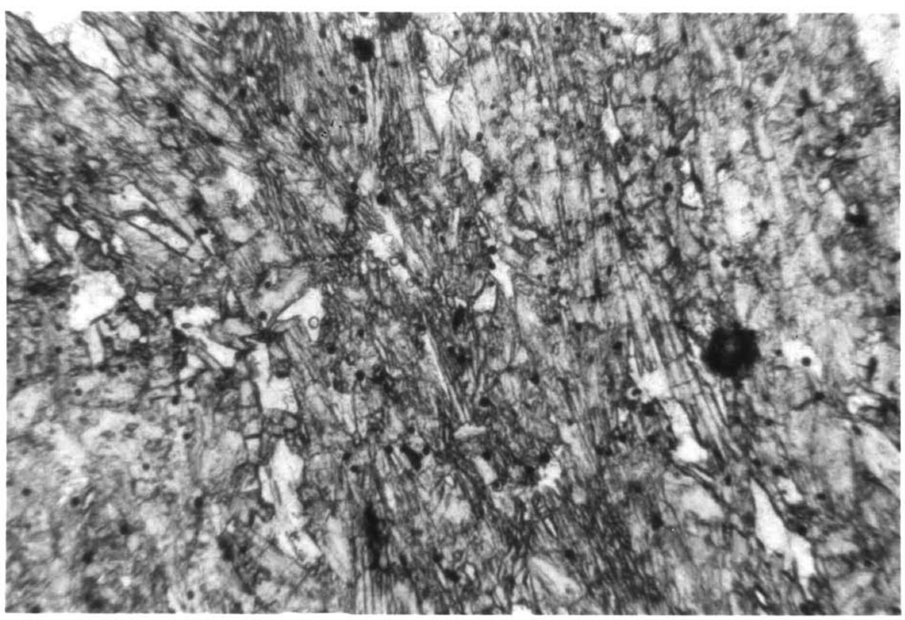


Figure 14B.



common. Easily confused with, but distinct from, these veins are bands of quartzite on the same scale. The quartzite bands are more laterally continuous and contain mm-scale color bands parallel to the schistosity.

Quartz generally constitutes 60 to 80% of the rock, occasionally ranging up to more than 90% (forming essentially a garnetiferous quartzite). Quartz generally shows only weak dimensional preferred orientation, mainly appearing as granoblastic aggregates. Sutured grain boundaries are not uncommon. Undulatory extinction is locally present. The textures suggest an annealing recrystallization after schistosity formation, followed by late to post-metamorphic strain.

Garnet is ubiquitous as small, subhedral porphyroblasts (1-3 mm, occasionally up to 5 mm), composing up to 10% of the rock. The garnets are randomly dispersed throughout the rock. Where late spaced cleavage planes intersect garnet, rims of chlorite $\pm$ epidote are developed. Although commonly sieved with equant blebs of quartz, trails of aligned inclusions have not been observed in the garnet within this unit.

Both biotite and muscovite are present, but in relatively minor amounts, together comprising no more than 10% of the rock as a rule. Proportions of the two micas vary, biotite usually dominating. Continuous, through-going mica films are seldom present, making the schistosity often obscure in outcrop. A late spaced cleavage defined by mica concentrations is often more prominent than the earlier foliation. In one specimen, randomly-oriented, interkinematic biotite porphyroblasts are present.

Green tourmaline is a ubiquitous accessory phase, occurring as small (<1 mm) euhedral or subhedral grains. Other accessory minerals include zircon, sphene, and apatite.

Three rock types with somewhat different compositions occur locally in this unit. On Finnslettheia, a plagioclase-porphyroblastic, graphitic schist is interlayered with more typical quartz-garnet schist near the base of the unit. The prominent foliation is marked by lepidoblastic intergrowths of muscovite and graphite, with subordinate biotite. Sparse, albite-twinned plagioclase porphyroblasts are a few mm across, anhedral, and randomly-oriented. The textural relationships in this rock are complex and significant for the microstructural development of the study area; they will be described in detail in Chapter 3.

In the valley midway between Sørvikfjell and Storvann(N), a hornblende -porphyroblastic facies of this unit is locally present near its base. Hornblende occurs as stubby, randomly-oriented porphyroblasts up to 1 cm across. In the shoreline exposures at Fjelldal, the stratigraphically higher part of the quartz-garnet schist unit becomes more pelitic, with the minerals kyanite and staurolite appearing. The rocks here are more retrograded in late events, developing extensive retrograde white mica, biotite, and chlorite. Equilibrium coexistence of kyanite, staurolite, and garnet can be documented (Figure 15). Due to the intense retrogression, it cannot be demonstrated that biotite coexisted with these other phases, though this seems likely.

The thickness of this unit ranges from about 150 to more than 700 m. It may be further thinned in the complex folding at the southeast corner of Storvann(N). Relevance of these thicknesses to original stratigraphic thickness is not certain, but it is clear that this was and is the thickest unit within the Storvann Group.

The protolith of this unit was probably a chemically mature siltstone with some intermixed sand. The presence of abundant garnet in a relatively mica-poor rock suggests high alumina relative to alkalis, and hence a clay -rich, feldspar-poor composition. The feldspar and amphibole bearing lithologies may reflect a variation in provenance, or a moderate amount of carbonate which reacted with the silicate phases to form Ca-bearing silicates during metamorphism.

Upper Calcite Marble

The second gray calcite marble is generally similar to the first, but may be usually distinguished by its more pure composition and more prominent color banding. A few percent white mica are present and minor quartz, but graphite and sulfides are absent and the rock is generally more than 95% calcite. While the lower unit is generally homogeneous in color, the upper marble commonly is banded gray and buff on a scale of one to five cm. Textures are granoblastic; the early foliation can be recognized by the color bands and by sparse micaceous layers.

The thickness of the upper marble is difficult to estimate because the upper contact is generally poorly exposed or tectonic. Only a few meters' thickness is present in the Kanebogen shoreline section, while its thick-

Figure 15: Photomicrograph of kyanite, staurolite and garnet in quartz -garnet schist at Fjelldal.

Ragged appearance of staurolite and kyanite is due to retrograde metamorphism to muscovite, quartz, and chlorite. Plane polarized light, 21X.

Figure 16: Photomicrograph of fibrolitic sillimanite in pelitic schist at Storvann(S).

Fibrolite is intergrown with lepidoblastic muscovite. Plane polarized light, 300X.

Figure 15.

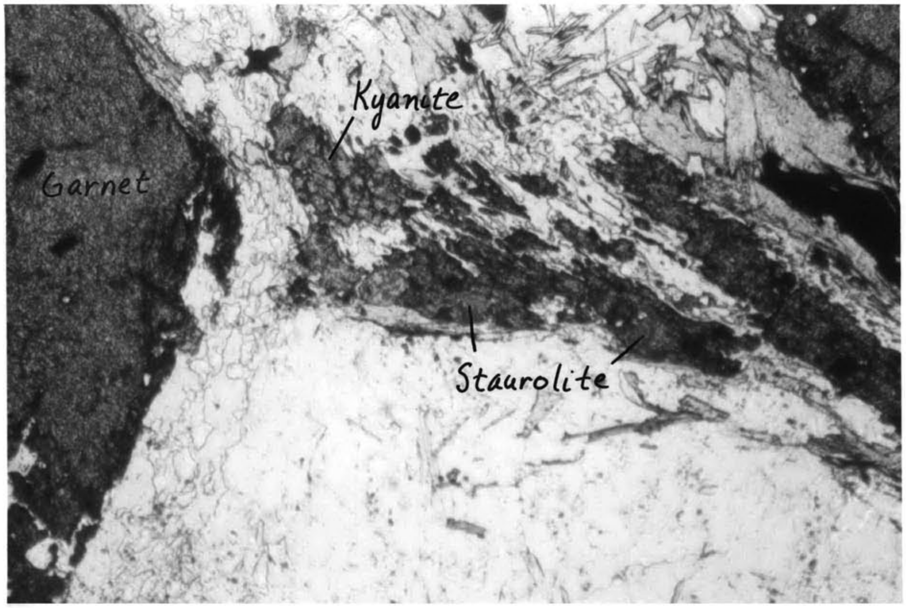


Figure 16.

ness is considerably increased by folding in most other areas of favorable exposure (e.g., Plate IIA, B-B'). Perhaps a representative thickness is southwestward along the strike from Ruggevik, where the marble is about 100 meters thick (Plate IID, L-L').

Pelitic Schist

This highly garnetiferous schist is the structurally highest unit I feel confident in assigning to the Storvann Group. It was probably a mechanically favorable level for detachment, since it commonly occurs as a thin (10-20 m) layer immediately under the lowest allochthonous unit of the area. In some areas (e.g., Kanebogen) it appears to be involved in the zone of tectonic mixing described in the next section. The consistent presence of a pelitic schist, often rusty-weathering and bearing coarse garnet porphyroblasts, at the top of the Storvann Group section, regardless of the overlying units, leads me to believe this schist belongs to the Storvann group.

The rock has a well-developed schistosity with the foliation defined by mica films which anastomose and wrap around the subhedral to euhedral garnets. The garnets range in size from 0.5 to 2 cm in diameter. In thin section, major minerals include quartz, muscovite, biotite, garnet, and often kyanite. One specimen contains fibrolitic sillimanite intergrown with muscovite. Accessory minerals include green tourmaline, zircon, and sphene.

The garnet porphyroblasts include trails of elongate quartz grains, recording a two-stage growth history. The inner portion includes abundant linear trails, suggesting growth post-dating an early schistosity. These inclusion trails continue into the outer zone where inclusions are more sparse, and curve into a sigmoid pattern indicating synkinematic growth. Where late spaced cleavage which truncates the internal fabrics of the garnet intersects the garnets, retrograde chlorite is developed.

Biotite occurs as two growths. It is present as small, randomly -oriented porphyroblasts which are bent and truncated by the late cleavage, and also in secondary quantities to muscovite in the mica films which define that cleavage.

Muscovite forms both the through-going mica films, and also locally remains in the matrix preserving an earlier foliation in short segments.

It is not clear how this older mica fabric relates to the internal fabric of the garnet due to strong disruption by the late cleavage, but these older fabrics are assumed to be related.

Kyanite occurs as anhedral relicts, replaced in part by white mica, quartz, and chlorite. The retrogression is interpreted to be related to late cleavage forming events, although it is not clearly related to cleavage traces as the garnet retrogression clearly is.

Sillimanite occurs as fine-grained intergrowths with muscovite in one kyanite-bearing thin section (Figure 16). The kyanite is replaced in part by muscovite, suggesting that the kyanite to sillimanite reaction occurred by two coupled reactions involving muscovite, similar to the relationships described by Carmichael (1969). This is the first occurrence of sillimanite grade Caledonian metamorphism in this part of the Scandinavian Caledonides.

On the east shore of Storvann(S), the pelitic schist unit is locally calcareous in its upper part. Calcite occurs as discontinuous lensoid interlayers a few mm thick, composed of fine-grained granoblastic aggregates. No calc-silicate minerals are developed.

The structural thickness of this unit is highly variable and in most areas very uncertain. It is very non-resistant to weathering so that exposures tend to be rare except along some shorelines. Areas of the map and cross-sections showing apparently large thicknesses of the pelitic schist unit (ss_2) are largely the result of generalization from a few outcrops which were impossible to map individually on the scale of the study (e.g., Plate IIA, B-B'). It is probable that in these wooded areas of poor exposure, the structure is far more complex than that shown on the map. However, the data do not permit a more sophisticated interpretation.

The protolith of the pelitic schist unit was a shale, deposited in part under reducing conditions, and in part in environments where limestone was also deposited. It appears to be an expression of continued transgression of the Scandinavian continental margin in early Paleozoic time. Its position and lithology would be consistent with at least a possible correlation with the Cambrian "alum shale" of the autochthonous foreland sections (Vogt, 1918; Gustavson, 1966), particularly since the alum shale also is commonly the basal decollement horizon for the frontal thrust belt.

Calcareous Schist and Sliver Zone

A complex mixture of lithologies forms a zone a few meters to perhaps 100 m thick above the top of the Storvann Group at four places, all but one of these clearly related to the lowest thrust of the Caledonian nappe pile. These are: (1) the east shore of Storvann(S) (Figure 12A); (2) the roadcut and shoreline exposures along Tjeldsund along strike from the Storvann(S) section; (3) under the Stangnes Thrust at Kanebogen (Figure 12B); and (4) under the Vikeland Thrust at the head of Kvaefjord. Where calcareous schist is the dominant lithology (although still containing a mixture of other rocks such as marble and amphibolite), the unit has been mapped as calcareous schist (cs). Where mixing has been more intense, it has been mapped as a sliver zone (sz). The lower contact of these rocks is more or less gradational with the pelitic schist unit of the Storvann Group, but is probably a thrust fault, but subsequent penetrative deformation and metamorphism have obscured relationships on the outcrop scale.

Typically, this unit is an intimate mixture in varying proportions of four lithologies: (1) quartz-plagioclase-calcite-epidote schists, sometimes containing one or more of the following: actinolite, muscovite, biotite, and in one case diopside, with accessory sphene and apatite; (2) garnet -two mica pelitic schists similar to the pelitic schist unit of the Storvann Group; (3) amphibolite, commonly garnetiferous; and (4) marbles, both calcitic and dolomitic, often rather impure and grading into the other three lithologies. Boudinage is often intense and ubiquitous. It is not certain to what extent the extreme mixing of lithologies may reflect sedimentary versus tectonic mixing, but the position of this unit below thrust faults suggests tectonic mixing is important. The interpretation of this zone as one of tectonic mixing along a major thrust boundary will be further discussed in Chapter 3.

Narvik Group

Introduction

The rocks of the Narvik Group were first described by Vogt (1922, 1942) from a type locale around the town of Narvik, about 80 km east of the study area on east Hinnøy; Vogt termed these rocks the "Narvik schists" or "Narvik Series". Strand (1960) renamed this assemblage the Narvik Group in accordance with present stratigraphic nomenclature. Gustavson summarized these earlier works and gave a generalized stratigraphic sequence for the Narvik Group, based on the assumption that no major inversions or repetitions occur within this group, from its type area westward to the Tysfjord culmination (Figure 3) or northward on the east side of the Haafjell synform. However, work in progress by Hodges in the Tysfjord area indicates that large scale recumbent folding is probably present in the Narvik Group, and that its extent and contact relations as interpreted by Gustavson (1966, 1972, 1974a, b, c) are incorrect. Hence, the stratigraphic and structural nature of the Narvik Group is currently in a state of flux.

The principal lithologies of the Narvik Group are quite distinctive, which allowed its recognition on east Hinnøy despite very restricted outcrop. The predominant rock type is a dark-colored, medium grained, often rather massive pelitic schist or gneiss, with abundant small garnet porphyroblasts, and sparse to abundant augen-shaped feldspar porphyroblasts. Kyanite is very common, occurring as disseminated bladed porphyroblasts and as coarse blades in quartz veins. Dikes and dikelets of pegmatitic and aplitic material, both pre- to synkinematic with respect to the schistosity, are nearly ubiquitous (Figure 17). In some areas, the dikes are abundant enough to give the rock a migmatitic appearance; in the Tysfjord area, unambiguously migmatitic rocks occur (Hodges, work in progress). Sheets and lenses of amphibolite are common, but texturally variable ranging from semi-aphanitic to blastoporphyritic to even-grained phaneritic. Other lithologies occur in subordinate amounts in the Tysfjord area, but do not occur on east Hinnøy and will not be considered here.

Occurrences on east Hinnøy

On east Hinnøy, Narvik Group rocks occur in restricted outcrops in two places: (1) shoreline exposures at Storvann(S), and (2) in a road cut at the power station one km west of Kilbotn, together with a string of natural outcrops leading south from the roadcut for perhaps 0.5 km. At the first locality, the rocks have a concordant contact with the calcareous schist unit below. The upper contact is unexposed, but the Narvik Group rocks are clearly structurally overlain by Salangen Group marbles. This lens is of very restricted extent, no more than 50 m thick and not trace-

Figure 17: Narvik Group pelitic gneiss.

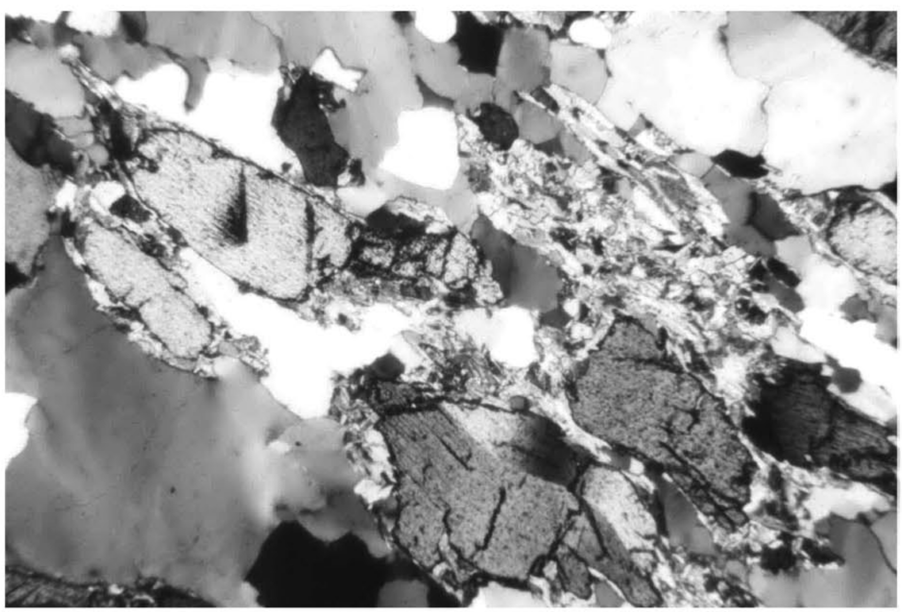
Exposure is in roadcut by power station, 1 km west of Kilbotn. Isoclinally folded white stringers are aplitic dikes.

Figure 18: Photomicrograph of twinned and bent kyanite, Narvik Group pelitic gneiss.

Kyanite grains are elongate parallel to primary schistosity in the rock. Texture is suggestive of boudinage: grains at upper left may have once been continuous with those at lower right. Plane polarized light, 30X.

Figure 17.

Figure 18.



able along strike from the shoreline outcrops. The only lithology present in the outcrops on the east shore of Storvann(S) is the pelitic gneiss unit; kyanite was not observed at this locality.

The power station locality is more varied and interesting. The contact with the Storvann Group (?) to the west is not exposed, but at the east end of the roadcut, a contact with the Stangnes amphibolite is exposed. The contact is sharp, and strongly foliated so that all structures are flattened into parallelism. The contact is considered tectonic due to the limited extent of typical Narvik Group lithologies and the lack of similarities between Stangnes Group rocks and typical Narvik Group amphibolites. However, the possibility cannot be ruled out that the Stangnes Group is an atypical part of the Narvik Group, at least on the basis of present knowledge.

The Narvik Group lithologies at the power station locality include (1) pelitic gneiss, in part kyanite-bearing both in veins and as disseminated porphyroblasts, (2) blastoporphyritic foliated amphibolite, and (3) a one meter thick, fine-grained, unfoliated metadolerite dike injected parallel to foliation.

The pelitic gneiss is composed of quartz, garnet, white mica, biotite, plagioclase, and kyanite, with chlorite and white mica occurring as retrograde phases formed by late cleavage development. Accessory minerals include green tourmaline, zircon, apatite, opaques, rutile, and sphene.

Garnet occurs as euhedral to subhedral porphyroblasts (3-5 mm), showing two stages of growth. The early phase forming the cores is sieved with equant, non-aligned quartz blebs while the later growth is generally inclusion-free. No trails of aligned inclusions are observed here. Locally, rims overgrow biotite flakes which lie in the schistosity, suggesting the outer zone at least is post-kinematic with respect to the early penetrative deformation. In addition to quartz, core zones occasionally include plagioclase, white mica, epidote, and rutile. Where late cleavage traces intersect garnets, retrograde chlorite is often formed. No grain breakage was observed in thin section, but at some localities crushed garnets are commonly observed in hand specimen examination.

Kyanite occurs as coarse blades in quartz veins, and locally as disseminated porphyroblasts (ca. 5 mm long) parallel to the predominant schistosity. It is commonly twinned and bent (Figure 18), and is generally rimmed by variable amounts of retrograde white mica. Garnet/kyanite grain boundaries which are straight and without intervening phases are rarely preserved, documenting equilibrium coexistence.

Muscovite and biotite both occur in two habits indicating two phases of growth. The early phase consists of coarse books or porphyroblasts, apparently aligned in the schistosity, which are now truncated and bent along the late cleavage. The second phase of growth is fine-grained, apparently retrogressive, and related to formation of the late cleavage. The second growth forms through-going mica films, and rims kyanite porphyroblasts.

Plagioclase occurs as saussuritized porphyroblasts in granoblastic aggregates of quartz. The time of saussuritization is not clear. It may be related to retrogression during the latest phase of deformation, because the plagioclase grains are truncated by late spaced cleavage surfaces. Alternatively, evidence from the associated amphibolite at this exposure (see below) and from Hodges' work near the Tysfjord culmination suggests the Narvik Group may have experienced a higher grade metamorphism prior to emplacement in its present location in the Caledonian nappe pile, resulting in superposition of the kyanite grade mineralogy on previously higher grade rocks. Hence, the present Caledonian mineral assemblage may result from a retrogressive metamorphism in the Narvik Group, and thus have been responsible for the saussuritization of the plagioclase.

The foliated, blastoporphyritic amphibolite is a massive, dark gray medium-grained rock with white irregular spots several mm across (pseudomorphs after original plagioclase phenocrysts?). It is intercalated with the pelitic gneiss with the contacts entirely transposed. Foliation is less intense largely due to the absence of mica. The rock contains subequal proportions of blue-green hornblende, sodic andesine, and clinozoisite. Accessory minerals include quartz, sphene, and zircon.

Hornblende shows weak preferred orientation, both crystallographically and dimensionally. No retrogression has affected the hornblende, nor are relicts preserved of primary igneous phases it must have replaced.

Textural relationships of the plagioclase and clinozoisite are more complex. The plagioclase has been determined by the Michel-Levy method to be about An_{35}. It is partly replaced by granular aggregates of clinozoisite, often leaving isolated remnants of plagioclase in optical conti-

Figure 19: Photomicrographs of plagioclase partially replaced by clinozoisite, Narvik Group amphibolite. 75X.

A: Plane polarized light. High relief grains are clinozoisite. Remainder of view is plagioclase.

B: Same view, crossed polars. Note optical continuity of plagioclase as indicated by simultaneous extinction.

Figure 19A.

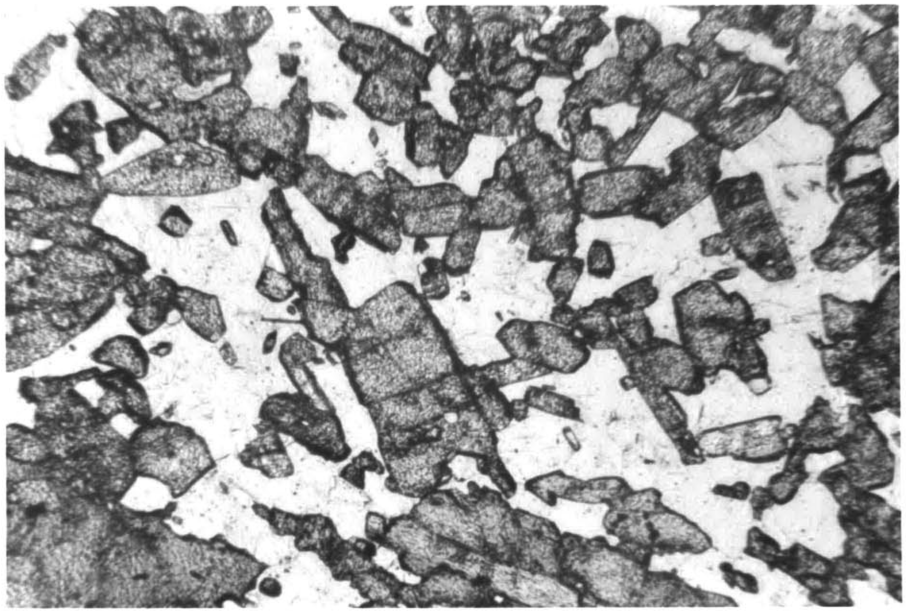
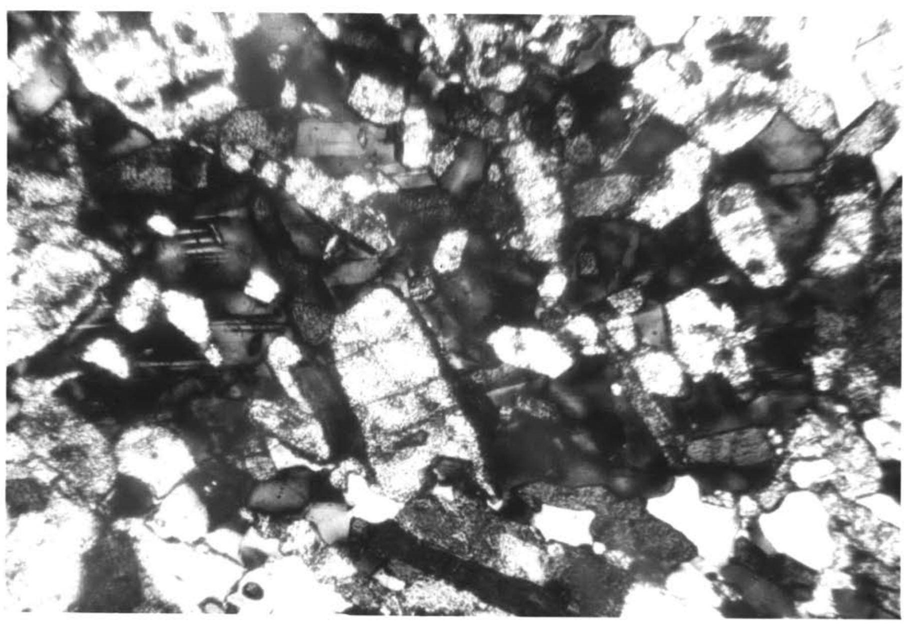


Figure 19B.



nuity (Figure 19). The clinozoisite is evenly disseminated throughout the specimen and not related to any discernable late cleavage development. It is interpreted to be a product of the kyanite grade metamorphism, consistent with the observation that metaigneous rocks throughout the study area bear the assemblage blue-green hornblende + plagioclase + clinozoisite as a product of the Caledonian thermal peak. The question remains at what time and under what circumstances the now partly replaced plagioclase grains originally formed. These could be relicts of primary igneous plagioclase, but the composition is rather sodic for a mafic igneous rock and the absence of zoning would be surprising for igneous plagioclase in such a body. The absence of any relicts of primary mafic phases also would argue against relict igneous mineralogy being preserved. The preferred, though tentative, interpretation is that these are relicts of the pre-kyanite grade metamorphism hypothesized to account for relations in the Tysfjord area (Hodges, pers. comm., 1979).

Perhaps the most curious feature of the outcrops at the power station is the meter-thick, blastoporphyritic metadolerite dike exposed only in a small outcrop (less than a meter high). As noted above, it is injected parallel to foliation into the host Narvik Group rocks but is itself unfoliated. It preserves chilled margins against its walls: at the edges, the matrix is essentially aphanitic, and relict phenocrysts are small (ca. 3 mm) and sparse (1% of the rock). Toward the middle of the dike, the matrix becomes fine-grained phaneritic and appears to mimic primary microporphyritic texture, while phenocrysts are larger (5 mm) and more abundant (ca. 5%). These relations suggest that the host rock had already cooled considerably from the metamorphic peak at the time of the dike injection.

The present mineralogy of the dolerite dike is actinolite, albitic plagioclase, epidote and clinozoisite, with biotite, rutile, and sphene as accessory minerals. Actinolite composes up to 70% of the rock, occurring as aggregates of randomly-oriented, stubby prismatic to equant anhedral crystals, often intimately intergrown with epidote. In actinolite-rich samples, plagioclase is interstitial to the actinolite, while in plagioclase-rich samples, the actinolite encloses plagioclase laths apparently mimetically preserving primary microporphyritic texture (Figure 20).

Plagioclase laths have been determined by the Michel-Levy method to be roughly An_{10} in composition. However, many grains show zoning; oscillatory

Figure 20: Photomicrograph showing relict microporphyritic texture in metadolerite dike at Kilbotn power station.

Laths of plagioclase, now albite, are enclosed in matrix of fine-grained granoblastic actinolite and epidote. Crossed polars, 17X.

Figure 21: Photomicrograph of blastomylonite gneiss, Stangnes amphibolite.

Porphyroclast is hornblende and plagioclase. Ribbons are hornblende, plagioclase, and clinozoisite. Crossed polars, 30X.

Figure 20.

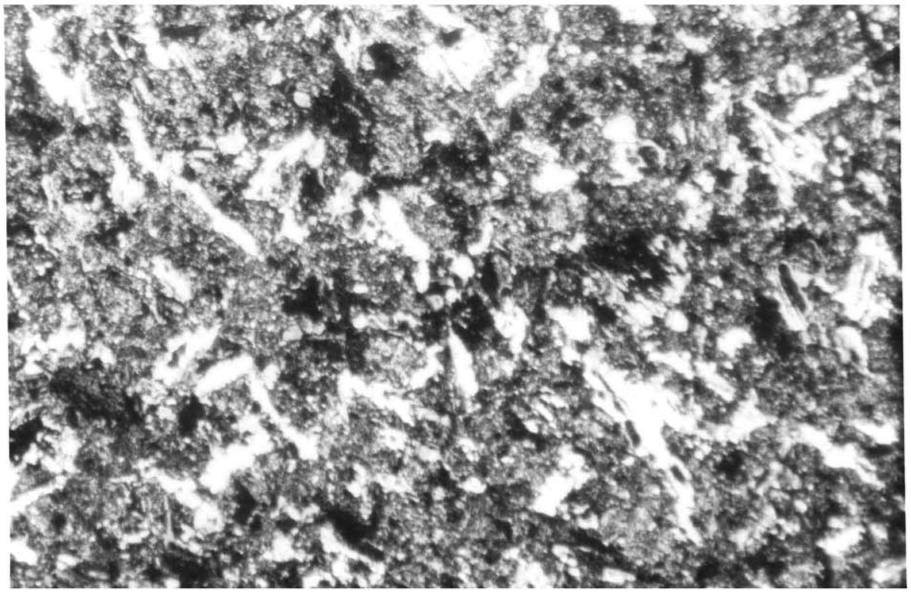
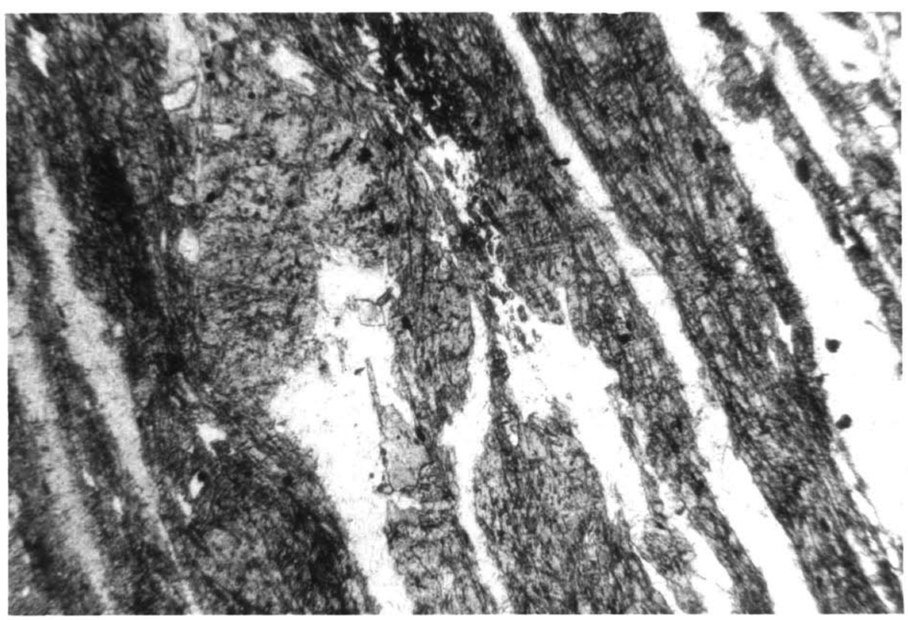


Figure 21.



zoning has been locally recognized. Plagioclase phenocrysts have been replaced in amounts varying from 0 to 100%, commonly retaining subhedral shape at all stages of replacement.

The mineral assemblage is more suggestive of the greenschist facies rather than amphibolite facies, and the absence of a tectonite fabric strongly favors the interpretation that this dike was intruded after the main penetrative deformation was completed on east Hinnøy. Rb/Sr whole rock/biotite ages from Ruggevik Tonalite samples from a quarry a few hundred meters from the outcrop of the dike gave ages of 358 $\pm$ 4 and 362 $\pm$ 5 Ma (Chapter 5), which are taken to record the time of cooling of this area below approximately 260° C (Hodges and others, 1980). Thus, the metamorphism of the dike predates this age. The dike is thus interpreted to be a late Caledonian intrusion.

Stangnes Group

Introduction

The Stangnes Group contains rocks that are anomalous among the nappes of east Hinnøy. Rock units in all the other tectonic units are dominated by silica- and alumina-rich terrigenous metasedimentary rocks and marbles, suggesting close association with continental margin environments. The Stangnes Group is composed of banded amphibolite, probably mafic metavolcanic rocks, intruded by the Ruggevik Tonalite Gneiss, a semiconcordant gneiss sheet of tonalitic to trondhjemitic composition. The tonalitic composition alone suggests a rather primitive magma source; present day $^{87}Sr/^{86}Sr$ ratios as low as .7047 (initial ratio is undetermined because age studies are to date unsuccessful; see Chapter 5), support this interpretation. It is unlikely that this suite of basic metavolcanic rocks, intruded by primitive granitoids, were associated with continental crust at the time of their formation. Hence, on the basis of simple petrochemical considerations, it seems likely that the Stangnes Group rocks belong to an exotic tectonic unit.

The Stangnes Group occurs as a thrust-bounded slab, about 0.5 km thick, between the autochthonous Storvann Group and the overlying allochthon composed of the Salangen Group. Structural arguments for the interpretation of the upper and lower contacts of the Stangnes Group as thrust faults are presented in Chapter 3 (D_1-Thrust Faults).

Along the basal thrust, lenses of coarse-grained (up to 3 cm), massive hornblendite are present. The origin of these bodies is unknown, but their lack of fabric suggests they crystallized after the main movement on the thrust. It is possible that enhanced permeability along the thrust zone increased fluid mobility at this level, leading to growth of very coarse -grained rocks.

The Stangnes amphibolite is lithologically distinct from those amphibolites the author has observed in the Narvik Group, so that the separation of these rock sequences is preferred here. Hence, the name the Stangnes Group is proposed, with the Stangnes peninsula directly east of Harstad designated as the type locality. The Stangnes Group is also exposed in the core of the overturned antiform opening northeastward from Middagsfjell toward Harstad, and in a belt trending southwestward from shoreline exposures near Ruggevik then curving around Kilbotn. On the southwest side of Sørvikfjell, the Stangnes Group rocks are truncated by the Storvann Fault and do not reappear on the other side. This abrupt disappearance of the Stangnes Group rocks at what is clearly a late high-angle fault can only be termed fortuitous. The Storvann Fault must nearly coincide with the lateral pinch-out of the Storvann Group by the joining into a single thrust of the two thrust faults that bound it.

The age of the Stangnes Group is unknown. Rb/Sr whole rock studies on the Ruggevik Tonalite Gneiss have not yielded an isochron (Chapter 5). The entire complex underwent thorough Caledonian metamorphic reconstitution and penetrative deformation, for which Rb/Sr whole rock/biotite ages give a minimum age of about 360 Ma. The tonalite is not likely to be older than about 1000 Ma. Since tonalitic rocks with primitive $^{87}Sr/^{86}Sr$ ratios are generally intruded into crust which was relatively young at the time of intrusion, this approximate lower age limit is considered to apply to the protoliths of the amphibolite as well.

Stangnes Amphibolite

The Stangnes Amphibolite is a fine-grained phaneritic amphibolite which is thin-banded black and white or pale green on the scale of a few mm to a few cm. The mechanical anisotropy introduced by this layering has led to the formation of abundant late upright to overturned chevron folds, giving the rock a very distinctive appearance in outcrop. This unit is easily distinguished from the other amphibolites of east Hinnøy on this basis.

The Stangnes Amphibolite is composed of blue-green hornblende (40-60%), plagioclase (35-55%), clinozoisite (0-20%), garnet (0-10%), and quartz (0-5%). Calcite and chlorite occur as minor alteration products. Accessory minerals include opaques, apatite, and sphene or rutile. One anomalous specimen has 40% clinozoisite and only 5% plagioclase. This rock may have resulted from local Ca enrichment during low-grade metamorphism prior to amphibolite facies conditions (Smith, 1968). Another possible explanation for the epidote enrichment is suggested by Holland and Norris (1979): the epidote-enriched bands may be the remains of pillow margins in strongly deformed and metamorphosed pillow lavas. The latter possibility is especially interesting in the light of the recent discovery of pillow lavas in a similar position in the nappe pile at Sulitjelma, 180 km to the south (Boyle and others, 1979).

The texture of the Stangnes Amphibolite ranges from mylonite gneissic to protomylonitic to nearly granoblastic. Near the basal thrust contact (ca. 20 m structurally above), an intense schistosity defined by grain elongation involves all major minerals. Sparse augen-shaped porphyroclasts of hornblende and plagioclase are set in a fine grained, intensely tectonized matrix of synkinematic hornblende, plagioclase and clinozoisite (Figure 21). The rock is mylonitic in the sense that considerable tectonic grain size recuction has occurred; however, it appears that annealing through grain boundary migration, plastic flow, and sub-grain growth kept pace with deformation. Reduction in grain size is much less in other samples (Figure 22). Textures of samples collected structurally higher (40-50 m above the thrust plane) are protomylonitic. Mechanical twins and mortar texture are developed in coarser grained hornblende bands, while plagioclase has been reduced to fine-grained, mylonitic bands in between.

At distances of more that 50 m from the thrust contact, the mylonitic texture largely disappears (Figure 23). Mineral alignment is still present, defining a strong schistosity, but the textures are granoblastic to subidioblastic.

The amphibolite is compositionally heterogeneous, which is expressed in the estimated modal ranges of minerals. The scale of the variations (typically a centimeter to a meter) suggests the differences are related

Figure 22: Photomicrograph of protomylonitic Stangnes amphibolite.

Note thin discontinuous and anastomosing bands of mylonitic material. Plane polarized light, 13X.

Figure 23: Photomicrograph of non-mylonitic Stangnes amphibolite.

Uniform grain size is smaller than the coarse relict grains in figures 21 and 22, suggesting that tectonically-induced grain size reduction may have occurred here also, but more uniformly and more completely annealed during deformation. Plane polarized light, 30X.

Figure 22.

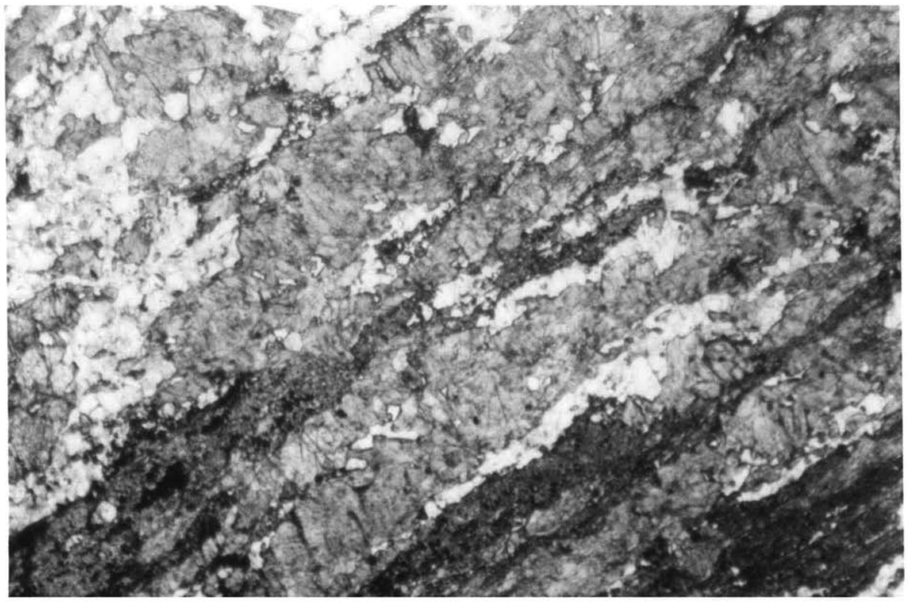
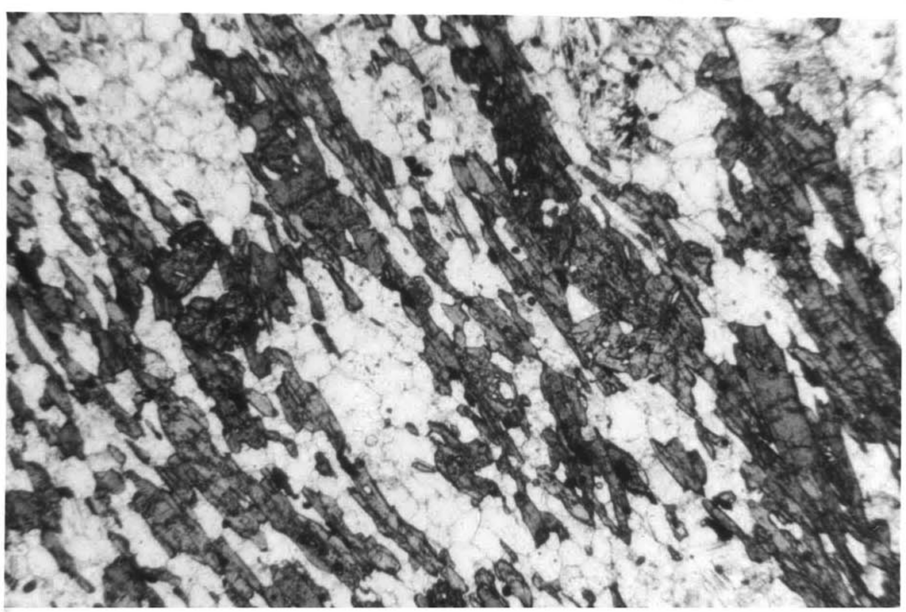


Figure 23.



to primary layering (bedding?), and hence that the protolith was a surface -deposited rock, probably mafic volcanic and volcanogenic sedimentary rocks. The tectonic setting where this sequence formed is uncertain, but it seems probable that it was an oceanic or transitional rather than a continental environment.

Ruggevik Tonalite Gneiss

The Ruggevik Tonalite Gneiss is a pale gray to white-weathering, medium grained leucocratic orthogneiss which occurs widely as a semiconcordant sheet in the middle to upper part of the Stangnes Amphibolite. The contacts are strongly transposed, but local truncation of layering in the amphibolite at the contact indicates an intrusive rather than metavolcanic origin for the tonalite gneiss (Figure 24). Irregularity of the outcrop pattern in some areas (e.g., one km north of Middagsfjell) also supports an intrusive origin. The tonalite gneiss is invariably well foliated, with moderately developed concentrations of hornblende and biotite into a gneissosity. On the outcrop scale it is even-grained and homogeneous; rarely, layers or lenses of amphibolite, often now boudins, occur in the tonalite gneiss.

The Ruggevik Tonalite Gneiss is well exposed in a quarry one km southwest of Kilbotn, in shoreline exposures on the headland north of Ruggevik, and in roadcuts and shoreline exposures at Stangnes. Ruggevik is chosen as the type locality because the other two areas were already designated as type localities for other units (Stangnes Amphibolite and Kilbotn Schist).

The tonalite is composed of plagioclase (ca. 40-50%), quartz (10-20%), biotite (1-15%), blue-green hornblende (0-20%), and clinozoisite (2-30%). One specimen contains significant muscovite (5%). Chlorite occurs in variable but minor amounts after hornblende and biotite. Accessory minerals include sphene, allanite, zircon, calcite, apatite, and opaques.

Plagioclase is entirely metamorphically recrystallized. Twinning is rare, and grains are subequant and granoblastic. Since no stained sections were prepared and twinning is rare in the plagioclase, the precise proportions of quartz and plagioclase are uncertain.

Hornblende is blue-green and generally has both dimensional and crystallographic preferred orientation parallel to the schistosity. Boudinage of individual crystals is common. Hornblende poikiloblastically encloses

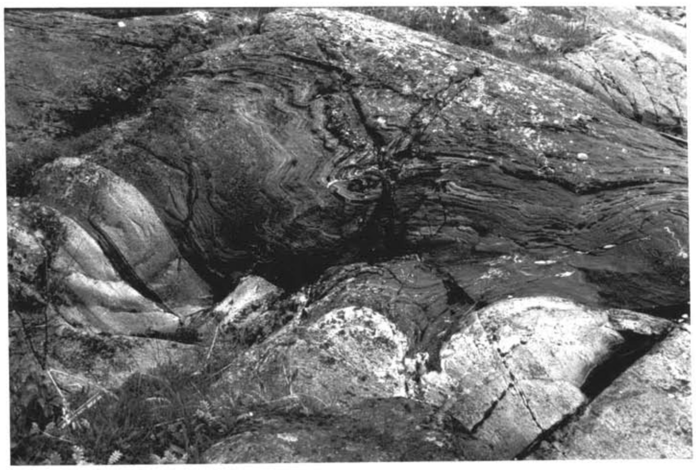


Figure 24: Intrusive contact of Ruggevik Tonalite Gneiss into Stangnes Amphibolite.

Shoreline exposure is on southeast side of Stangnes peninsula. Note disharmony of folding (F_4; see Chapter 3) due to contrast of banded amphibolite and massive tonalite gneiss.

quartz and plagioclase, and to lesser extent biotite. Biotite and chlorite appear as retrograde phases after hornblende in rocks with a well developed late cleavage.

Clinozoisite occurs as subidioblastic to granoblastic crystals, and clearly is a member of the amphibolite facies mineral assemblage. New growth related to late cleavage formation is also recognized in the form of finer grained, granoblastic aggregates along cleavage surfaces.

Biotite and muscovite (where present) generally show two phases of growth, one in the primary schistosity, the other in the late cleavage. There is little difference in appearance between the two ages of biotite: both are brown, and occur in unaltered lepidoblastic films and as isolated plates parallel to one of the two foliations. The younger fabric can be distinguished by consistent cross-cutting grain relationships. Where late microfolds are present, the micas are polygonized around them.

Phanerozoic tonalites are typically formed in convergent margin arcs, such as the tonalite-dominated Southern California Batholith. Tonalitic intrusions are uncommon in well developed continental crust. The relatively non-radiogenic Sr of the Ruggevik tonalite (initial $^{87}Sr/^{86}Sr$ was probably less than .704; see Chapter 5) also supports formation of this rock in island arc or very young continental crust. Consequently, it is suggested as a working hypothesis that the Ruggevik tonalite and the Stangnes Amphibolite represent a tectonic sliver from an ensimatic arc terrain, analogous to the Trondheim Nappe 500 km to the south.

Salangen Group

Introduction

The Salangen Group comprises the highest unit of the nappe sequence in the study area. As a whole the rocks are poorly exposed on East Hinnøy and structurally complex. Until relationships in this sequence are better established in better exposed, structurally less complex areas, it was considered less valuable to study these rocks in detail on East Hinnøy, so that less time was spent on these units. No stratigraphy or structural sequence was identified here, although the lithologies observed can be related to the sequence developed by J. Tull and M. Steltenpohl (pers. comm., 1979) in the Haafjell area (see below).

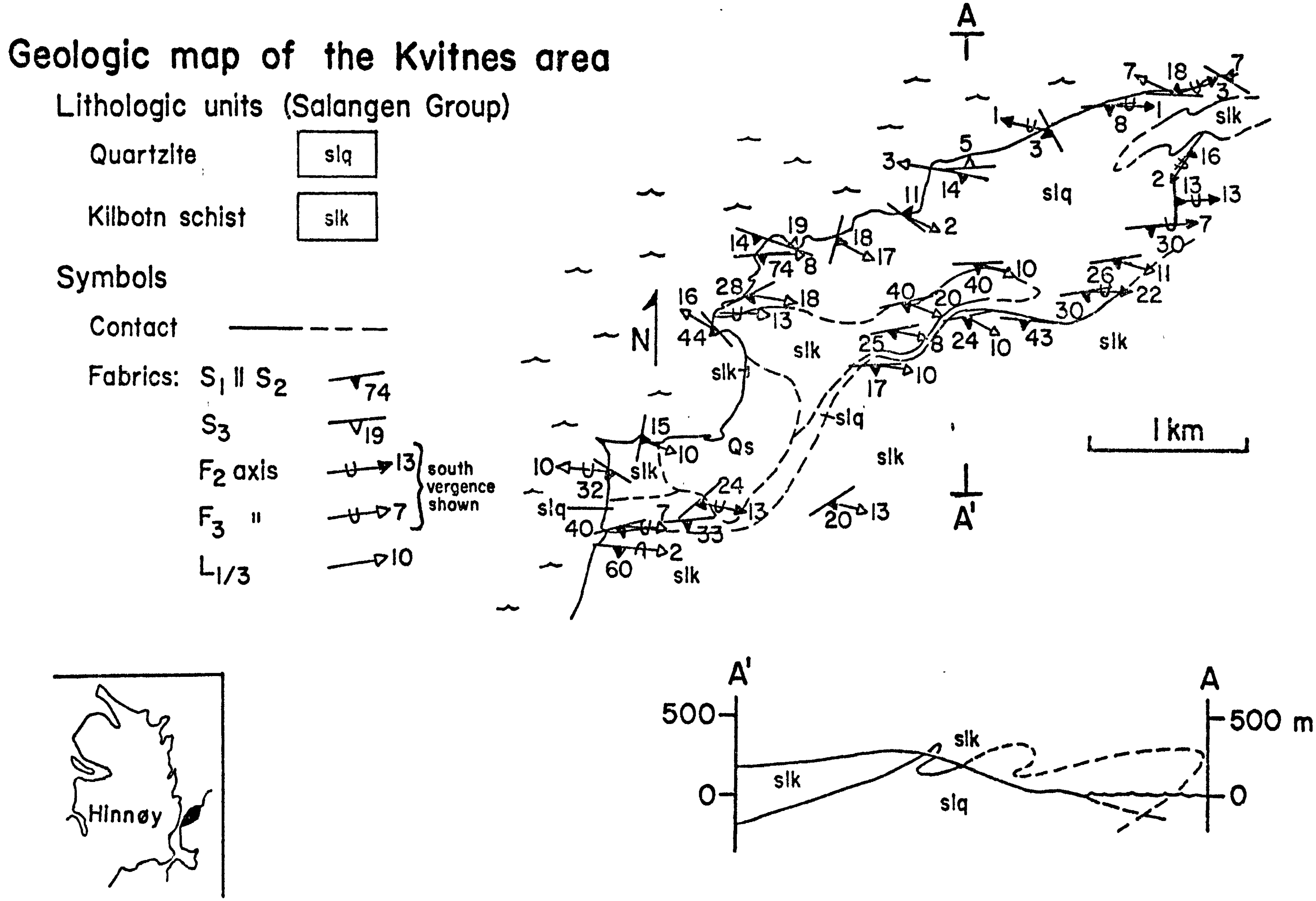
The term Salangen Group was introduced by Gustavson (1966) to include the Evenes Group and overlying Bogen Group of previous authors (e.g., Strand, 1960), on the grounds that these rock sequences were not distinguishable from one another. In fact, given the evidence for large scale inversions in the nappe pile (Chapter 3; also Hodges, unpublished data), it is likely that the similarities between rocks of the Evenes and Bogen Groups could be at least partly due to structural repetition. In this case, the more general term would be the only correct usage. The Salangen Group is consequently used here, at least pending completion of more detailed work on the structure and stratigraphy of these units.

The Salangen Group regionally is composed of marbles, subordinate schists, with one or more conglomerates locally present at the base and a quartzite unit near the middle. On East Hinnøy, three lithologic types are represented: (1) the Harstad Conglomerate, which appears to be the locally preserved "basal" conglomerate; (2) a series of variegated marbles which appear to be the Evenes marbles of Strand and Henningsmoen (1960), or the lower marble sequence of the Haafjell area as defined by Tull (pers. comm., 1979); and (3) the Kilbotn schist, a unit composed of garnet-mica schist with minor calc-schist and garnet amphibolite, which is correlated on lithologic similarity with the mica schist overlying the Evenes marbles in the Haafjell area (the basal unit of the Bogen Group in the terminology of Strand, 1960). The anomalously large apparent thickness of this unit may be due to a location in the core of a recumbent synform, or may be due to thrust slicing of the sequence. A further observation is consistent with this correlation. Overlying the apparently correlative schist in the Haafjell area is a pure white and tan quartzite with sericitic partings. Across Tjeldsund to the east of the study area, reconnaissance by the author indicates that Kilbotn schist is complexly infolded with a lithologically identical quartzite (Figure 25).

The basal contact of the Salangen Group on east Hinnøy is a thrust contact which brings Harstad conglomerate, calcite marble, and Kilbotn Schist into contact with the Stangnes Amphibolite (Chapter 3). If the above correlation scheme is correct, this contact is highly transgressive across Salangen Group units (Figure 31).

The age of the Salangen Group is unknown. As a major metasedimentary assemblage of the Caledonian nappe terrain, it has been assumed to be lower

Figure 25.


Geologic map of the Kvitnes area
Lithologic units (Salangen Group)
Quartzite
slq
Kilbotn schist
slk
Symbols
Contact
Fabrics: S1 ll S2
74
S3
19
F2 axis
13
F3 "
7
south vergence shown
L1/3
10
A
A'
N
1 km
Qs
slq
slk
Hinnøy
500
0
500 m
0

Paleozoic ("Cambro-Silurian") in age. The possible correlation of the Salangen Group with the Ullsfjord Nappe (Binns, 1978) of Finnmark bears on this. The marbles of the Ullsfjord Nappe are less intensely metamorphosed (greenschist facies) and have recently yielded tabulate corals (Favosites) of Late Ordovician to Early Silurian age (Binns, 1978) and an early Paleozoic shelly and coral fauna (Olaussen, 1977). Recent studies of the Salangen Group by Tull, Steltenpohl, and Andresen (pers. comm., 1979) in the Haafjell area tentatively support this correlation. Hence, on Plate I, a Silurian (?) age has been assigned to these rocks.

Harstad Conglomerate

The Harstad Conglomerate is a polymictic, matrix-supported metaconglomerate. It is best exposed in outcrops in and around the community park in downtown Harstad, and is also known by this author to occur on the west side of Svartdalsaasen (Plate I; Plate IID, section L-L'), and on the Trondenes Peninsula north of Harstad (not formally mapped in this study, as noted on Plate I). The unit was first described by Vogt (1922); its age, in particular relative to the Evenes marbles (lower marbles of the Salangen Group), has shifted back and forth since then (Vogt, 1941, p. 205 -208; Gustavson, 1966, p. 44; Gustavson 1972, p. 11). Regardless of its age, the conglomerate unit's structural position is consistently at the bottom of this nappe in all localities, except at Harstad where the entire nappe sequence is inverted (see Chapter 3) so that at the time of nappe emplacement, the conglomerate sat at the base of the nappe. Gustavson's (1972) primary evidence for the Harstad conglomerate being younger than the Evenes marbles is the occurrence of pink "color-banded" marble pebbles in the conglomerate at Harstad and Evenskjaer (outside the map area). These pebbles have not been observed by the author at either locality, but assuming their presence, their age significance remains non-diagnostic, since pink marbles occur in the basement rocks on east Hinnøy as well as in the Evenes marbles (see Hesjevann assemblage-Marbles; Austerfjord Group-Lower Units).

In the Haafjell area, Tull (pers. comm., 1979) has observed two distinct types of conglomerates: a lower unit with a fine-grained amphibolite matrix, enclosing mainly clasts of mafic and acid igneous rocks, and an upper unit with a quartzitic matrix with mainly white to buff-colored marble clasts. Both types are present in the Harstad area; on Svartdalsaasen, a third type has been observed with a calcareous schist matrix; this seems to grade laterally into the amphibolite-matrix type and is thought to be a facies of it. The mafic igneous clasts observed by Tull near Ballangen in the Haafjell area have not been observed by the author in the Harstad area; fine-grained acidic igneous rocks, and possible quartzite clasts, dominate.

The origin of the protolith of this unit is of considerable interest. It has been suggested that this could be an Eocambrian tillite (E. Tveten, W. C. Griffin, pers. comm.); however, since the unit is in a sequence dominated by carbonate rocks of possible Silurian age, this seems unlikely. B. Sturt (pers. comm., 1979) has observed on the Lyngen Peninsula in Finnmark what he interprets to be an unconformable relationship between rocks of the Ullsfjord Nappe (possible correlative of Salangen Group marbles, see above) and the underlying Lyngen gabbro. The basal rocks above the unconformity consist of fluviatile sandstone and conglomerate, with locally derived clasts from the underlying gabbro. The presence of mafic clasts in the conglomerate at Haafjell thus suggests a possible environment of formation of the Harstad conglomerate, and supports the proposed correlation.

Marbles

A variety of carbonate rocks are included here, some of which have been mapped separately (Plate I); no coherent structural succession has been recognized. The dominant lithology is a rather massive, pure, gray calcite marble, mainly homogeneous with sparse white mica impurities, but occasionally color banded. Locally, tremolite porphyroblasts occur. This lithology underlies much of Harstad, and essentially the entirety of Voldstadheia and Kvitfjell (Plate I; Plate IIB, section D-D' and F- F'). Subordinate marble lithologies include a white to buff-colored dolomitic unit, and the pink color-banded unit mentioned above. Tull and Steltenpohl (pers. comm., 1979) report that this pink marble only occurs in their lower marble group, that is, the Evenes marbles. This is one of the main reasons for correlating these marbles with the Evenes marbles.

A distinctive unit which reconnaissance suggests may be useable as a marker horizon is a spectacular zone of schist, amphibolite and skarn-like rocks best exposed in a roadcut ca. 2 km southeast of Sørvik. The most distinctive lithology is composed of euhedral garnets 1 to 5 cm in diameter, randomly-oriented hornblende needles up to 3 cm in length (including veins of coarse, acicular hornblendite), in a matrix of quartz and white mica. Other lithologies include coarse granoblastic garnet amphibolite, garnetite veins, and garnet-muscovite schist. Outcrops of similar rocks were observed along the shoreline to the east during reconnaissance, suggesting this may be a mappable horizon.

In thin section, the marbles are typically granoblastic, with generally intensely twinned grains of calcite $\pm$ dolomite. White mica and quartz are present in all specimens, but in most cases constitute less than 5% of the rock. In more siliceous specimens, accessory amounts of plagioclase, tremolite, and/or epidote or clinozoisite are present, and myrmekitic intergrowths of quartz and feldspar have been observed. Biotite, chlorite, and rutile occur as accessory minerals.

Kilbotn Schist

General Characteristics

The Kilbotn Schist is the proposed name for the schist that surrounds the bay at Kilbotn, extending southward nearly to Sørvik and cropping out locally in the vicinity of Harstad. As noted above, it is considered correlative with the basal schist unit of the Bogen Group of Strand and Henningsmoen (1960), and hence to overlie the Evenes marbles. The Kilbotn Schist is mainly a highly micaceous, fine- to medium-grained schist with ubiquitous small garnets. It is commonly calcareous, at the same time becoming hornblende-bearing. Boudinaged layers 10 to 100 cm thick of garnet amphibolite, almost certainly of metasedimentary origin, occur in the calcareous parts. This calcareous nature is consistent with the carbonate-rich nature of the entire Salangen Group.

The thickness of this unit is difficult to determine since the compositional banding is so strongly overprinted by the late cleavage that it is difficult to decide what surface(s) to refer to in estimating a structural thickness. The lateral extent, considering overall dip of the area in which it occurs, suggests a thickness as much as 1 km. This is far greater than the thickness observed for the correlative unit at Haafjell, suggesting that structural thickening of the unit, for instance in a major recombent fold hinge, probably occurred on east Hinnøy.

Petrography

Mica Schist

The bulk of the Kilbotn schist is composed of quartz, garnet, muscovite, biotite, and minor plagioclase. Clinozoisite is a common accessory mineral, and trace amounts of detrital zircon, allanite, green tourmaline, and rutile are also present.

Garnet is subhedral to anhedral, appearing as small porphyroblasts 0.5 to 2.0 mm in diameter. It is typically sieved with tabular quartz grains which preserve the trace of a foliation discordant to the dominant fabric in the matrix. Rarely inclusion trails show slight rotation, but they are generally planar.

Muscovite and biotite occur generally in subequal amounts, mainly concentrated in anastomosing, through-going mica films which define the main mesoscopic foliation. Biotite is not chloritized, except where crenulation of this foliation is intense, when chlorite may grow axial planar to these refolds. Micas are typically polygonized around such folds and may occasionally be recrystallized into axial planar orientation.

The protolith of this rock was a siltstone or shale. No evidence of graded beds or other sedimentary structures was noted in this study. Its close association with quartzite and abundant carbonates suggests a shallow marine depositional environment.

Calcareous Schist

The calcareous portion of the Kilbotn Schist is typically composed of quartz, hornblende, biotite, and plagioclase; calcite is generally present, and muscovite and garnet are common. Chlorite after biotite and hornblende, and sericite and clinozoisite after plagioclase are common late alterations. Accessory minerals include abundant detrital zircon, green tourmaline, rutile, allanite, and apatite.

Hornblende is blue-green, and similar to garnet, encloses planar inclusion trails of tabular quartz, and occasionally biotite, preserving a foliation discordant to the external mica fabric. Hornblende porphyroblasts are now elongate in the second schistosity but show little evidence of crystallographic preferred orientation. Late microfolds and crenulations are associated with growth of chlorite after hornblende, and locally bend or break hornblende grains.

Biotite, and muscovite when present, show two phases of growth, both post-dating the internal fabric of hornblende porphyroblasts. Early biotites are coarse, and may be kink-banded; they often poikiloblastically include quartz and plagioclase, as well as rimming hornblende. Late mica growths are fine-grained, lepidoblastic films concentrated at hinges of late microfolds and crenulations. Replacement of biotite by mats and lepidoblastic aggregates of colorless to pale green chlorite is commonly associated with the late folds as well.

Plagioclase is present as anhedral, strongly polysynthetically twinned isolated porphyroblasts, fresh in weakly crenulated specimens but strongly sericitized in intensely crenulated rocks. Shared grain boundaries with hornblende and early biotites are straight and unarmored, indicating the feldspar was produced by the early, higher-grade metamorphism.

Calcite may occur as either medium-grained lensoid granoblastic aggregates, or as fine-grained granular aggregates replacing hornblende or plagioclase. While its abundance does not exceed 10% in the specimens examined, it constitutes an essential phase in most rocks.

Garnet Amphibolite

It is of particular interest to determine if the lenses and boudins of garnet amphibolite in the Kilbotn Schist are igneous or sedimentary in origin. Their modal mineralogy strongly suggests the latter, as does the observation that their contacts with calcareous schists are entirely gradational in thin section. Major phases include hornblende, garnet, quartz, and calcite; minor constituents include plagioclase, clinozoisite, actinolite, and opaques with biotite and rutile as accessories. Chlorite occurs in small amounts as an alteration of hornblende.

Hornblende is blue-green, and occurs as medium-grained (ca. 1 mm) decussate aggregates, moderately elongate parallel to foliation. Locally, it shows lamellar intergrowth with a pale green amphibole, probably actinolite (Figure 26). Garnet occurs as sparse, subhedral porphyroblasts (ca. 3 mm), and unlike in the schists, it overgrew and enclosed the dominant external

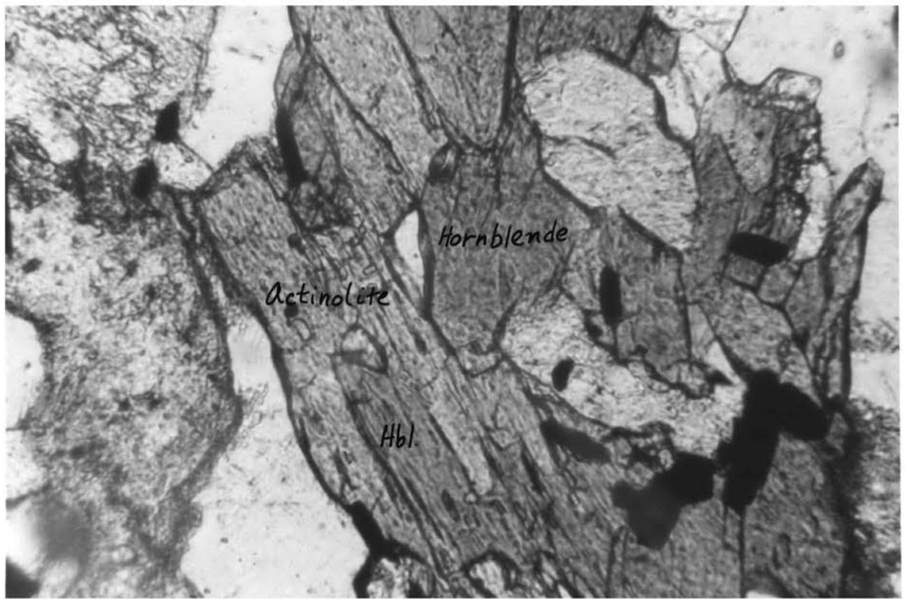


Figure 26: Photomicrograph of two amphiboles in garnet amphibolite within the Kilbotn Schist.

Plane polarized light, 300X.

schistosity. The second cleavage, so strongly developed in the schists, is practically absent here. Calcite and clinozoisite occur generally together as fine-grained granoblastic lensoid aggregates, in textural equilibrium with the other major phases of the rock. Quartz and plagioclase occur disseminated evenly through the rock; plagioclase surprisingly makes up less than 5% of the mode.

The garnet amphibolite appears to be of sedimentary derivation, presumably formed from a limey mud. Because mica is absent, whereas mica is very abundant in the associated schists, there is a large gradient in the effects of the late refolding. This will be further examined in the next two chapters, on structural geology and metamorphism.

CHAPTER 3: STRUCTURAL GEOLOGY

Introduction

Structures of Precambrian, Caledonian, and perhaps Cenozoic age are present on east Hinnøy. Precambrian structures are not understood in detail, partly because these were not the primary focus of this study, and partly because the relationships are not easily deciphered. Hakkinen (1977) was able to identify three phases of Precambrian folding on west Hinnøy in the early Proterozoic supracrustal rocks ("veined and layered paragneisses"), in areas where Caledonian effects were not present. Presumably, the Archaean gneisses of east Hinnøy experienced the same deformations, but these are all relatively massive units with little layering: macroscopic structures cannot be mapped, and the age of the foliation(s) present is difficult if not impossible to determine. As Hakkinen noted (1977, p. 68-70), in these units it is impossible to separate the effects of the different Precambrian deformations.

Where the Precambrian basement lithologies are more likely to record multiple deformations (e.g., the Kvaefjord Group metasedimentary rocks), the exposures are poor and the rocks are strongly overprinted by Caledonian events. As a consequence, only rather general remarks can be made about Precambrian structures.

Caledonian structures on east Hinnøy comprise a complex combination of thrust and fold nappe tectonics, overprinted by multiple refolding events. Five deformations, at least two involving thrust faulting, are recognized. Each deformation produced its own set of minor structures and fabrics; major structures from all but the youngest deformation (D_5) have been recognized. Table 2 summarizes these relationships.

The structural picture of the Caledonian orogeny developed here deviates from previous interpretations in four significant ways. First, the number, identities, and geometry of the thrust nappes is quite different from that proposed by Gustavson (1972) who published the only previous structural study of east Hinnøy. Second, the emplacement of the major allochthonous sheets is clearly pre- to synmetamorphic, unlike relationships further east in Troms (Gustavson, 1972; Olesen, 1971), where thrusting clearly post-dates the main stage of metamorphism. Third, the stack of early thrust nappes is recumbently folded on a large scale (> 5 km in amplitude), so that over extensive areas (indeed, much of the eastern half of the study area) the nappe pile is inverted. The basement is extensively involved, occurring in the cores of the recumbent anticlines. Fold nappe tectonics have already been described in north Norway, in the Sørfinnset area roughly 200 km to the south, by Wells and Bradshaw (1970). However, recognition of major inversions of the Caledonian nappe pile is something new, and may have great importance for the tectonics of the northern Scandinavian Caledonides (see Chapter 6). Fourth, contrary to previous statements (e.g., Gustavson, 1972) that only F_1 produced penetrative schistosity on east Hinnøy, as many as <u>three</u> different fold phases (F_1, F_2, and F_3) produced penetrative or semi -penetrative fabrics in some rocks (e.g., see Kilbotn Schist, Chapter 2). This obviously requires a more sophisticated approach to understand the fabric relationships.

An equally significant observation, first made by Hakkinen (1977) on west Hinnøy but greatly extended here, is that Caledonian deformation is zonal and limited in extent in the basement rocks (Figure 27). Hakkinen considered this to be a function of structural level: he argued that the Lofoten block was a high-level nappe emplaced along the Austerfjord thrust, over-riding the metamorphic nappe pile, so that metamorphic fabric died out upwards as a result of lower P/T conditions. Results of the present study appear to make this interpretation untenable (see "Austerfjord Thrust" below); a different explanation is required. The evidence for the limits of Caledonian deformation in the basement of east Hinnøy will be presented below (see also Chapter 2, Middagstind Quartz Syenite; Melaa Granite). Possible explanations will be considered in the latter part of this chapter, and the regional tectonic implications discussed in Chapter 6.

An additional consequence of the variable and zonal deformation of the basement is that one deformation may produce different types of structures in basement and cover rocks, depending on previous structural histories. This requires temporal correlation of, for instance, thrust faults, ductile shear zones, and folds which appear to all have developed during the second deformation phase (D_2), but represent quite different failure mechanisms.

To further understand the geometry and interrelationships of structures and fabrics, the study area was divided into 15 subareas (Figure 28), plus the Kvitnes area on the mainland which was only briefly examined and not included on Plate I (see Figure 25). The boundaries were chosen to achieve structurally homogeneous domains, often following late fault contacts which

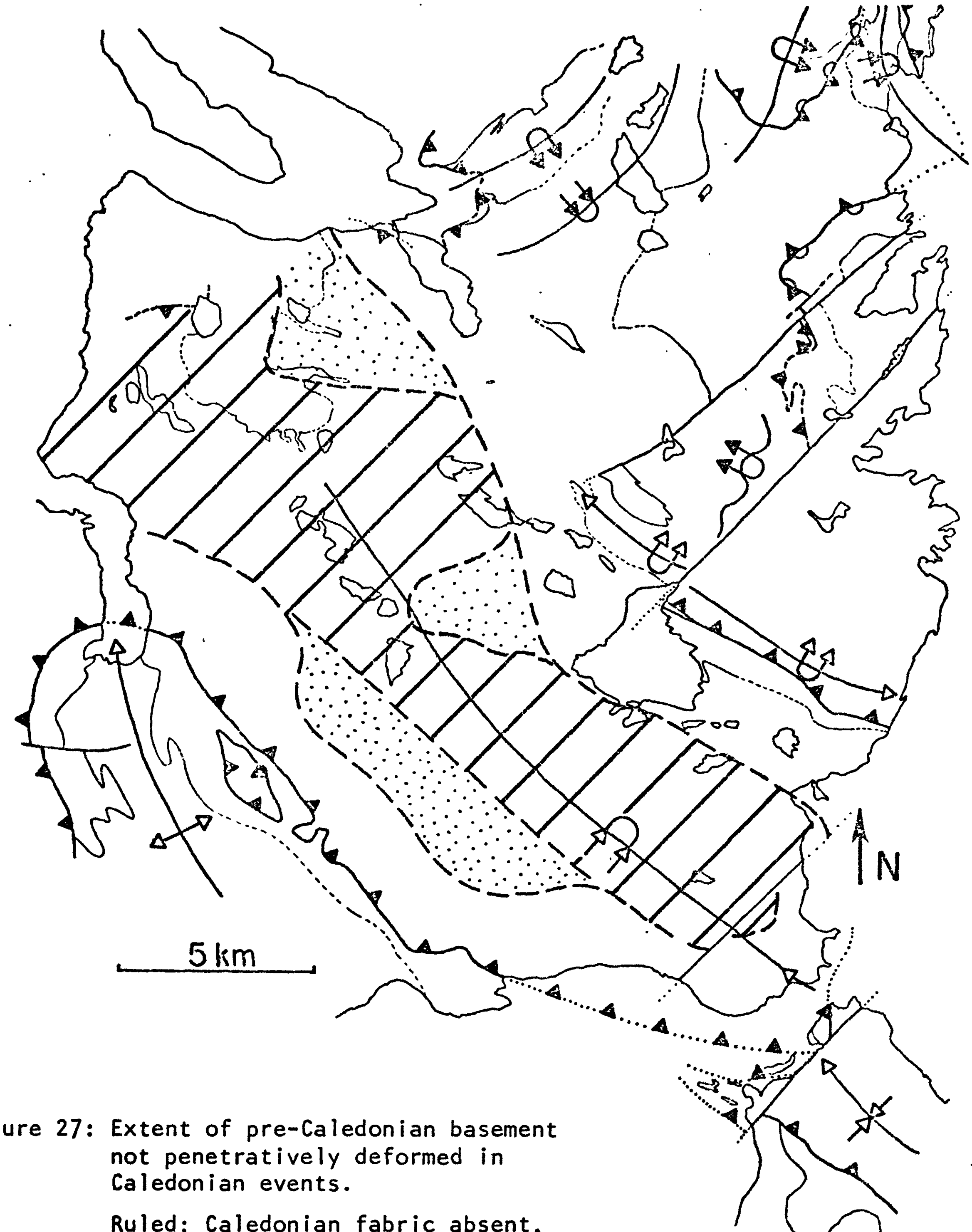
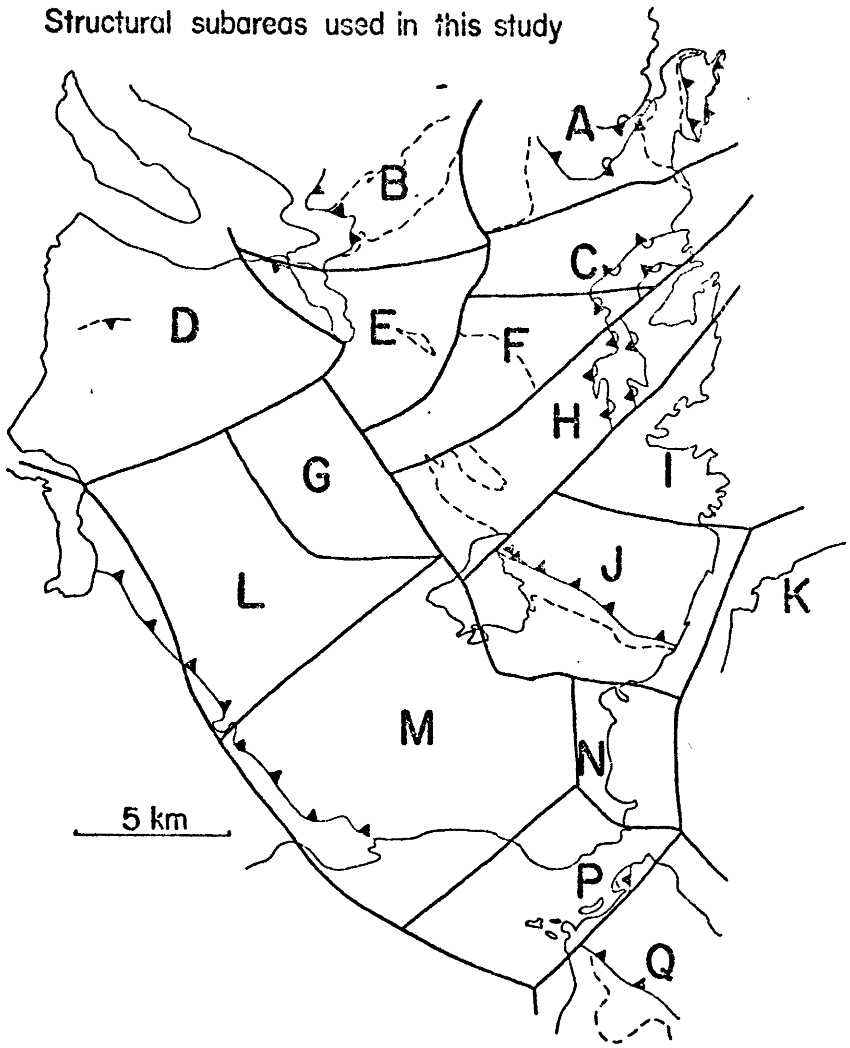


Figure 27: Extent of pre-Caledonian basement not penetratively deformed in Caledonian events.

Ruled: Caledonian fabric absent, or extremely weak

Stippled: Caledonian fabric zonal with foliated zones separated by volumes of undeformed rock

Figure 28.


Structural subareas used in this study
A
B
C
D
E
F
G
H
I
J
K
L
M
N
P
Q
5 km

juxtapose structurally different blocks. There remains considerable scatter of the data in some of the domains, in part due to the numerous successive deformations. Truly homogeneous domains would require subdivision beyond the level of detail achieved in most of this study. The results of the analysis are plotted partly on Plate IV, with the remainder distributed in figures throughout the chapter. The plots are discussed where appropriate throughout the text.

High angle faults of probable early Cenozoic age were discussed briefly, by Gustavson (1972). However, the faults are more abundant than indicated by Gustavson, and the regional pattern, in the present author's interpretation, is considerably more consistent (Figure 55). The interpretation presented here integrates all faults into a single kinematic picture which is compatible with plate tectonic considerations.

Precambrian Structures

The preservation of Precambrian folds and fabrics is clearly documented in the Middagstind area. As noted in Chapter 2, the Middagstind Quartz Syenite is a discordant pluton with no internal fabric which cross-cuts foliations and contacts in the surrounding rock units; since the Middagstind Quartz Syenite yielded a Rb/Sr whole rock age of 1726 ± 31 Ma (Chapter 5), the structures it truncates must be at least early Proterozoic in age. The absence of Caledonian fabric in or near the Middagstind pluton is indisputable.

The most widely developed Precambrian structure is a penetrative schistosity or gneissosity. It is defined by alignment of micas and feldspar elongation in the Gullesfjord Gneiss and the migmatite gneiss, and by hornblende alignment in the hornblende diorite. Within the contact hornfels on the east side of the syenite pluton, diffuse compositional banding is rarely preserved, and in thin section, weak hornblende alignment within this banding locally preserves the schistosity. Eastward the hornfels grades outward into rocks unaffected by the contact metamorphism; it is unknown how the Caledonian fabrics overprint the Precambrian schistosity here, but the two appear to be subparallel, perhaps explaining the subtlety of the transition.

All of the rocks in the western half of the field area are allochthonous along the Austerfjord Thrust (Plate I; Plate IIB; Plate II C, G-G', H-H'), a demonstrably Caledonian structure (see below). Above the thrust a mylonitic zone is developed that is up to 1 km in structural thickness. The mylonitic foliation dies out upward into a weaker subparallel penetrative gneissosity which is indistinguishable from the foliation cut by the Middagstind syenite to the north, and apparently in continuity with it. It is my tentative conclusion that the main fabric extending in the core of the broad synform from the Middagstind area to Saetertind (Plate III) is probably a Precambrian fabric, rather than Caledonian. More detailed work will be required to fully evaluate this possibility.

A few Precambrian folds have been recognized. Occasionally, open, diffuse mesoscopic folds of schlieren in the Gullesfjord Gneiss are present. The style of these folds is unlike Caledonian open folds observed in the cover rocks (F_3, F_4, and F_5): they are sinusoidal, and only mildly asymmetrical, whereas the Caledonian late folds in comparable rocks tend to be chevron-like or concentric and strongly asymmetrical. Axial trends on these folds are rarely measurable, because they are generally exposed on glacially polished surfaces so that the third dimension can not be observed. Hence, the possibility that these folds are Precambrian in age cannot be evaluated on a geometrical basis.

On a larger scale, it appears that the structural basin in which the Middagstind syenite intrudes is at least partly Precambrian. Foliations dip inward around the pluton, appearing to define a sharp-hinged east-west trending synform and a more open northwest-southeast synform (Plate III). The latter structure is Caledonian in age (F_3), since it folds the Austerfjord Thrust in the Tjeldsand area. The former fold appears to be intruded by the Middagstind syenite, which suggests a Precambrian age, post-dating the formation of the Precambrian penetrative fabric discussed above.

Poles to the Precambrian foliation around Middagstind are plotted stereographically on Plate IV (subarea D). There is a definite suggestion of a great circle girdle centered around an ENE-trending, gently plunging axis, although later deformations have complicated the picture. This fold phase appears also to be present further south in subarea M. Comparison of poles to the Precambrian fabric to poles to the Caledonian S_2 fabric indicates folding around a similar ENE-trending fold axis predated formation of the Caledonian S_2 fabric. In subarea M, it cannot be ruled out that the folding is an early Caledonian event.

Similarly, the diffuse, complex N-S girdle of poles to S-surfaces in subarea G may in part reflect the presence of an E-W Caledonian fold phase. However, the presence of Caledonian lineations (Caledonian L_{2a}, see below) and probable Caledonian D_3 shear zones indicates a unique interpretation of the fabrics in subarea G is not possible. Similarly, the fabrics in subarea N are complex and poorly understood; it is considered likely that these include both Precambrian and Caledonian fabrics, as yet unresolved due to poor exposure and inadequate study.

The two structural events described above, the penetrative foliation and the folds that fold this foliation, appear to correlate with events (i) and (iv) of Hakkinen's pre-Caledonian history for west Hinnøy (1977, p. 69). No structures considered to correspond to the other events in Hakkinen's structural sequence have been recognized.

Caledonian Structures

Introduction

Five deformational phases of Caledonian age have been recognized on east Hinnøy, which are summarized in Table 2. These can be broadly split into two groups: synmetamorphic events (D_1 and D_2), and late metamorphic events (D_3 through D_5). The early events comprise the construction of the nappe pile and its infolding (and thrusting) with the subjacent basement. The late events redeformed these structures, complicating the picture but producing no major tectonic transport.

A significant break in time occurred between the synmetamorphic and late metamorphic events. Most porphyroblasts grew under static conditions after the synmetamorphic deformations, followed by cooling as much as 200°C (see Chapter 4) before the beginning of late deformations. However, D_1 and D_2 are together veiwed as two parts of a single continuum of early deformation, and D_3 through D_5 are similarly considered different portions of a continuous period of late deformation.

Structures assigned to a particular phase of deformation in any specific location may not have formed strictly synchronously with structures at another location assigned to the same phase. However, over an area the size of the present study, in the absence of a source of strong thermal gradients (e.g., plutonic emplacement), the temperature history can be assumed

TABLE 2: Caledonian deformations on east Hinnøy.

Phase	Faulting	Folding	Fabrics
D_1	Emplacement of allochthons along Stangnes and Tjeldsund thrusts.	Folding of Storvann Group, possibly allochthonous rocks as well: A*: up to several hundred meters O*: unknown V*: unknown	Penetrative schistosity in cover, extends less than 1 km downward into basement Mylonitic fabric near thrust contacts
D_2	Austerfjord and Vikeland thrusts Steep ductile shear zones near Middagstind	Recumbent folding, refolding D_1 thrusts, involving basement and cover together A: up to 10 km O: variable V: ESE	Transposition and overprinting of S_1 to variable degree to produce new schistosity Sparsely developed amphibole lineation Boudinage
D_3	Low T mylonitic shear zones in granite gneisses of basement	Upright to overturned asymmetric folds A: up to 1 km O: WNW-ESE V: Mainly SSW	Spaced cleavage, intense in micaceous rocks, weak to absent in mica-less rocks
D_4		Upright to overturned asymmetric folds A: up to several hundred meters O: NE-SW V: variable	Spaced cleavage, similar to S_3
D_5		Local mesoscopic upright to overturned folds A: up to few meters O: NW to NNW V: variable	Crenulation cleavage in Kilbotn Schist

* A - amplitude of folds
O - axial orientation of folds
V - vergence of folds

to have been relatively uniform. Consequently, similar timing of structures relative to metamorphism implies similar time of formation if not struct synchroneity.

During the earliest event recognized, D_1, the three major allochthons in the map area (Narvik Group, Stangnes Group, Salangen Group) were emplaced along two major thrust planes, here named the Stangnes Thrust and the Tjeldsund Thrust (Plates II and III). Fold interference structures indicate that some folding developed prior to D_2 structures; these folds have been tentatively attributed to the D_1 event.

During the second deformational event, the basement, its autochthonous sedimentary cover (the Storvann Group), and the allochthons were recumbently infolded. Schistosity development was variable in extent during this event, but a widely developed hornblende lineation was produced. The Austerfjord Thrust is considered to have been emplaced during the second event, and is considered to be a response within the basement equivalent to the recumbent folding to the northeast. Discrete subvertical shear zones in the basement rocks of the upper plate of this thrust are also assigned to this deformational event.

The thrust locally mapped at Kvaefjord, named here the Vikeland Thrust, developed a thick mylonite zone similar to the Austerfjord Thrust. The thrust appears to truncate D_2 folds at a low angle, but is folded by open D_3 folds. It is tentatively assigned a late-D_2 age, and thought to be related to the Austerfjord Thrust.

During the third deformational event, D_3, WNW-ESE-trending cross folds were formed. These are the most prominent late folds in most of the study area. The folds are mainly of chevron type, and primarily verge to the southwest. This folding event is tentatively correlated with the formation of the Tysfjord Culmination (Figure 3) to the southeast of the study area. Some D_3 deformation in the basement was accommodated by development of mylonitic shear zones rather than folding.

The fourth deformational event, D_4, is a second late fold event which produced folds trending nearly perpendicular to those formed in D_3. Folds of D_4 age are only prominent in the northeastern part of the field area. These are upright folds that trend NE-SW with variable vergence. D_3 and D_4 structures rarely overlap in outcrop, so that chronological distinction is difficult. Structural analysis supports this relative timing assignment;

however, the two events are thought to be closely related in time. D_3 and D_4 produced widespread semi-penetrative axial planar spaced cleavage; in micaceous rocks this cleavage can be the dominant planar fabric (e.g., Kilbotn Schist).

The fifth deformational event, D_5, locally developed small scale folds and crenulation. In the Kilbotn Schist, this crenulation cleavage can strongly disrupt the spaced cleavage(s) formed in D_3 and D_4. No macroscopic structures of this age have been recognized.

In the discussion below, the following standard abbreviations will be used to refer to fabric elements:

S: foliation, either compositional banding, schistosity, or cleavage.

L: lineation, either mineral, intersection, or stretching

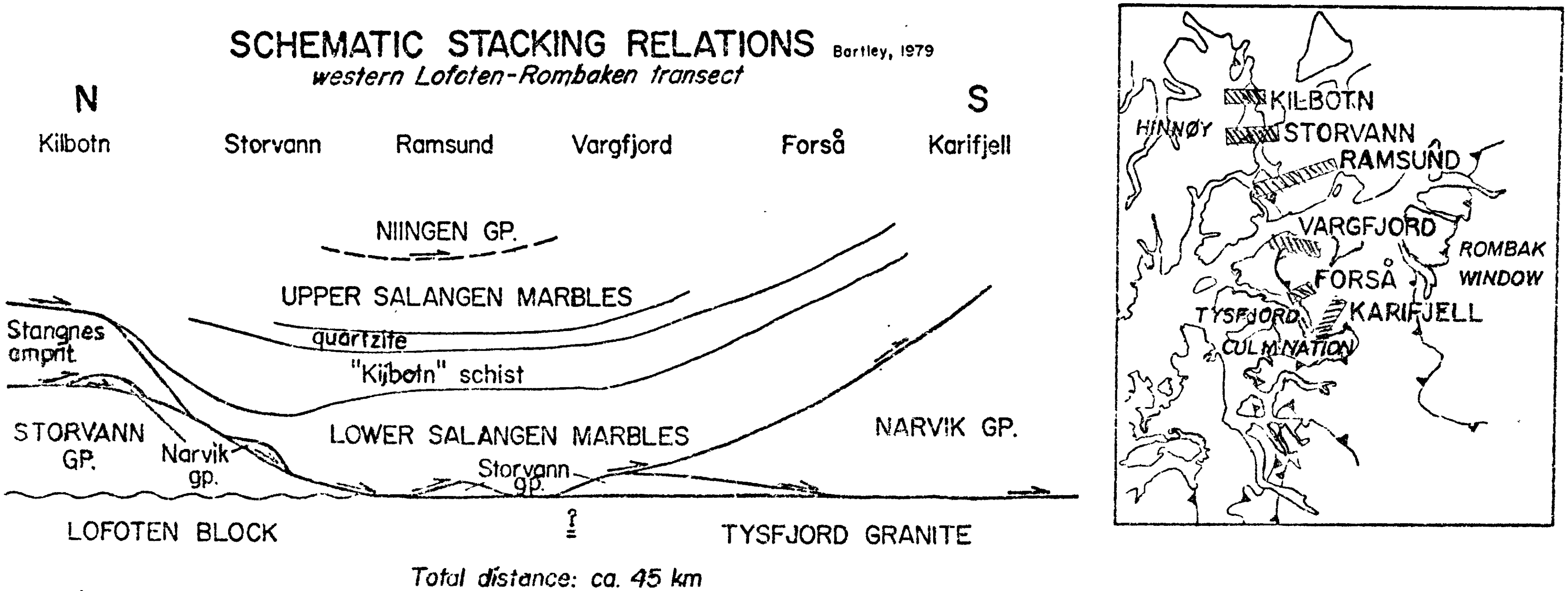
F: folds

Subscripts refer to the deformation phase in which the fabric element was produced. Thus, S_3 is the foliation produced in the D_3 deformation; it is not necessarily the third S-surface developed. This approach is deemed less confusing than simple serial numbering of successive fabrics of each type, since D_3 fabrics could include (for example) S_2, F_3, and L_1. In the case of intersection lineations, the ages of both S-surfaces involved are appended in subscript; hence, $L_{1/3}$ refers to the intersection of S_1 and S_3.

Nappe Sequence and Nomenclature

Figure 29 shows the preferred interpretation of the vertical sequence of rocks on east Hinnøy at the close of D_1, and its correlation with the nappes of the Haafjell area to the south. This interpretation differs radically from that presented by Gustavson (1972). Gustavson recognized two nappes on east Hinnøy, the Harstad Nappe and the Straumsbotn Nappe. The Harstad Nappe was considered to include all of the rocks included here in the Storvann Group, the Narvik Group, the Stangnes Group, and the Salangen Group, plus some of the subjacent basement. On Gustavson's map (1972), the basal thrust was drawn from Gausvik to Storvann(S), dipping northward, and thence northward toward Kvaefjord, interrupted by the Straumsbotn Nappe (see below). The Hesjevann assemblage and Kvaefjord Group rocks between Rundfjell and the head of Kvaefjord were interpreted as thrust-bounded and part of a klippe of the Harstad Nappe. The west boundary of this klippe was drawn under the Melaa Vannene, and between

Figure 29.


SCHEMATIC STACKING RELATIONS
Bartley, 1979
western Lofoten-Rombaken transect
N
S
Kilbotn
Storvann
Ramsund
Vargfjord
Forså
Karifjell
NIINGEN GP.
UPPER SALANGEN MARBLES
quartzite
"Kilbotn" schist
Stangnes
STORVANN GP.
Narvik gp.
LOWER SALANGEN MARBLES
Storvann gp.
NARVIK GP.
LOFOTEN BLOCK
?
TYSFJORD GRANITE
Total distance: ca. 45 km
HINNØY
KILBOTN
STORVANN
RAMSUND
VARGFJORD
ROMBAK WINDOW
FORSÅ
KARIFJELL
TYSFJORD CULMINATION

the Middagstind Quartz Syenite (not distinguished by Gustavson) and its contact aureole; the east boundary ran through the east-facing glacial scarps north of Hesjevann, and disappeared in an unclear manner near Straumsbotn. The thrust thence was drawn in the valley at the north side of the present study area and in the hills further north.

The difficulties of this interpretation can be summarized as follows: (1) There is very little evidence for a thrust fault anywhere along this proposed trace. Gustavson (1972) cited repetition of the basement/cover sequence across the valley between Storvann(S) and Gausvik, but as noted in Chapter 2, there is no repetition; the southern belt of metasedimentary rocks and amphibolite comprises a different suite of lithologies (Hesjevann assemblage) from the northern belt of metasedimentary rocks (Storvann Group), and is demonstrably part of the Lofoten terrain. The "klippe" is composed of entirely basement rocks, uncorrelative with any of the metasedimentary rocks of the cover. Most of the contacts bounding the "klippe" either are intrusive, as in the case of the Middagstind syenite and its contact hornfels, or non-existent, as in the case of the continuous marble band Gustavson (1974 a,b) shows bracketed by thrust faults running along the east side of Keipfjell. No marble is present except for isolated blocks; the cirque headwall south of Steinvann (labeled 185 m in Plate I), where Gustavson shows this thrust-bounded marble band present, is a cliff of uninterrupted granite gneiss. (2) The Harstad Nappe includes completely distinct sequences of rocks in mutual thrust contact, including both newly recognized units, the Storvann and Stangnes Groups, and units traced from other areas, the Narvik and Salangen Groups. Thus, the Harstad Nappe, as defined, includes both allochthonous and autochthonous units.

Included in Gustavson's Straumsbotn Nappe are the rocks bounded to the southeast by the Langvann Fault, and on the northeast by an irregular line drawn from directly north of Sørvikfjell to the south end of Storvann(N) and then west to the head of Kvaefjord.

The rocks of the Straumsbotn Nappe are not distinct from those of the autochthon, including largely basement granite, Kvaefjord Group, and Storvann Group rocks. The southwest boundary is clearly a high angle fault (see "High Angle Faults" below), cross-cutting all Caledonian structures. Where Gustavson shows the basal thrust of the Straumsbotn Nappe to turn to the northwest, it is drawn through a cliff of well-exposed, uninterrupted quartz-garnet schist of the Storvann Group. The "imbricate zone" on Finnslettheia (near Storjord, see Gustavson, 1972, p. 13) is incorrectly mapped; the map pattern Gustavson (1972, 1974 a,b) shows is entirely dissimilar from that on Plate I, inexplicably since exposures on the upper part of Finnslettheia are from good to excellent. The structures here are essentially all folds, with no evidence of imbricate thrusting (Plate II A, C-C').

Consequently, it is held that there is no evidence for a Straumsbotn Nappe. Since the Harstad Nappe as defined by Gustavson (1972) is also unviable, it is felt that the best approach is to start over with entirely new nomenclature. The structural nomenclature adopted here names the structures (i.e., folds and faults) rather than the rock bodies (i.e., nappes). The reason for this is simple: on east Hinnøy, the same rock bodies have been involved twice in nappe-forming events, first thrusting (D_1) and then large-scale recumbent folding and thrusting (D_2). Thus, the same rock bodies form parts of two different nappe sequences, and names become confusing. To avoid this, the folds and faults are identified, and the rock bodies by stratigraphic names only.

D_1: Nappe Emplacement

The significant structural features developed during D_1 are the emplacement of the three major allochthons, (isoclinal?) folding of the Storvann Group, and development of the most prominent regional schistosity. Folds (F_1) within the allochthons are also likely, but no folds unambiguously attributable to D_1 have been observed in these rocks.

Thrust Faults

Two major D_1 thrust faults are recognized on east Hinnøy. The lower Stangnes Thrust brings the Stangnes Group amphibolite and tonalite gneiss into contact with the autochthonous Storvann Group. The higher Tjeldsund Thrust emplaces the Salangen Group upon the Stangnes Group in the northern part of the study area, and upon the Storvann Group in the southern part of the area, where it thus becomes the basal thrust of the nappe pile (Figure 29). Lenses of Narvik Group rocks occur along both of these thrusts, supporting the interpretation that the Stangnes and Salangen Group rocks comprised higher structural units than the Narvik Group in the D_1 nappe stack.

Stangnes Thrust

The lowest thrust of D_1 age on the northeast part of east Hinnøy occurs at the base of the Stangnes Group, and is termed the Stangnes Thrust. It occurs upright and well-exposed on shorelines at the north and south ends of the Stangnes peninsula. The thrust resurfaces to the west and southwest in the core of a major F_2 recumbent fold (Plate IIA, A-A'), so that it is inverted in the intervening limb of the recumbent fold couple. The thrust merges southward with the Tjeldsund Thrust.

Three lines of structural evidence support the notion that the basal contact of the Stangnes Group is a thrust fault. The contact is concordant on the outcrop scale, but discordant on a regional scale. At the southeast corner of the Stangnes peninsula (Figure 12B), the Stangnes amphibolite occurs above a fairly complete sequence of Storvann Group rocks, plus a zone of mixed lithologies (interpreted as a zone tectonic slivers, see below). This contact cuts progressively down in the lower plate as it is traced to the north, until at the north end of the peninsula, only several meters of basal quartzite, a sliver of marble, and a few meters of the quartz-garnet schist unit are all that is left between the amphibolite and the basement. This discordance suggests either an unconformity, a fault, or an intrusive contact.

The Stangnes Amphibolite has well-developed lithologic layering, interpreted to reflect an original bedded nature and thus to indicate that this is a surface-depostied sequence of rocks; hence, an intrusive contact is rejected. There are no mafic feeder dikes for these metavolcanic rocks anywhere that cut the Storvann Group or in the Lofoten terrain on east Hinnøy, nor tonalitic bodies even vaguely resembling the Ruggevik Tonalite. This argues against an unconformable or intrusive relationship. Furthermore, the amphibolite near the contact is blastomylonitic, with the mylonitic character of the fabric dying out upward away from the contact (Chapter 2, p. 91). Hence, the lower contact of the Stangnes Group is considered a thrust fault, probably with large displacement.

The concordance of the Stangnes Thrust with lithologic banding and S_1 schistosity could result either from emplacement during formation of the S_1 schistosity, or from transposition of layering and superposition of the schistosity in already-juxtaposed rocks. The thrust emplacement is thought to be synchronous with S_1 formation. As described in Chapter 2, the texture of the amphibolite moving progressively upward from the thrust contact goes from blastomylonitic to protomylonitic to granoblastic (Figures 21 to 23). In the blastomylonitic samples, the same minerals have grown synkinematically in the penetrative schistosity as grew in the non-mylonitic schistosity of the rocks away from the thrust. There is no petrographic evidence for more than one schistosity in any of the amphibolite samples; the mylonitic fabric neither overprints nor is overprinted by the non-mylonitic schistosity. There is no angular discordance between the fabrics in the field. Consequently, a non-mylonitic penetrative schistosity, S_1, is considered to have formed concurrently with emplacement of the allochthons. This schistosity has been overprinted to varying extents by S_2, but the actual degree of this overprinting has been difficult to determine (see discussion of S_2 below).

The Stangnes Thrust is thought to have moved during the early part of the metamorphic peak. No contrast in metamorphic grade is observed across the contact. The dynamic recrystallization at amphibolite facies evidenced by the high-temperature mylonites of the thrust zone is regionally succeeded by refolding, presumably still at high temperature, in turn followed by static porphyroblast growth. Consequently, movement on the thrust was synchronous with the higher grade metamorphism in the area, but the thermal peak outlasted movement on the thrust.

No direct information on the direction of thrust emplacement was observed, due in part to the complex overprinting of this early deformation by subsequent events. Boudinage in the subjacent sliver zone is possibly related to movement on the thrust, but it is equally likely that the boudins are a result of stretching in the limbs of F_2 folds. Stretching lineations preserve, in general, a composite of superposed deformations; hence, these give no simple information as to the finite strain field of any one event. Regional relationships require that the Stangnes Group be more or less westerly derived. Since none of the Lofoten terrain comprises suitable basement for the Stangnes Group rocks (see "Stangnes Group," Chapter 2), a minimum distance of transport can be calculated assuming thrusting perpendicular to the modern limit of Lofoten basement some 60 km to the northwest. However, regional considerations (Chapter 6) indicate the Stangnes Group has almost certainly traveled much further than this.

Tjeldsund Thrust

The Tjeldsund Thrust is the name used here for the thrust at the base of the Salangen Group. Where exposed in roadcuts along the west side of Tjeldsund, north of Gausvik, it is a complex zone of mixed rocks, juxtaposing Salangen Group marbles with the lowermost part of the Storvann Group. In the mixed zone between are lithologies including thin-banded amphibolite similar to the Stangnes amphibolite. To the west, on the east shore of Storvann(S), the most complete Storvann Group section known occurs in the lower plate. Here, the sliver zone includes abundant calc-schists rarely present at Tjeldsund, and a lens of Narvik Group gneiss at the top, just underneath the Salangen Group marbles (Figure 12A). This regional discordance accompanied by tectonic mixing along the contact is, as with the Stangnes Thrust, interpreted to indicate tectonic juxtaposition.

North of the Storvann Fault, the Salangen Group is separated from the autochthon by the Stangnes Group. Two obvious possibilities exist for the structural position of the Stangnes Group: (1) the Stangnes Group sits depositionally at the base of the Salangen Group, perhaps unconformably, so that the Tjeldsund Thrust and Stangnes Thrust are equivalent; or (2) the Tjeldsund Thrust continues above the Stangnes Group, no longer the basal thrust in this area. The contact between the Salangen Group and the Stangnes Group is exposed in two locations: the southeastern corner of the Stangnes peninsula, and on a hill across the road from the water at Ruggevik. At the first location, the amphibolite is in contact with intensely tectonized marble of the Salangen Group. At the Ruggevik locality, exposures are poorer, and the contact appears almost gradational between Stangnes amphibolite and Salangen Group calcareous schist and marble. However, the contact is clearly discordant at map scale. At different locations, Kilbotn Schist, Harstad conglomerate, and calcite marble of the Salangen Group appear to be in contact with the amphibolite (Plate I). An unconformity would predict the contact to be discordant with units of the Stangnes Group, not the Salangen Group. Hence, it is concluded that this contact is also a thrust fault. The possibility cannot be excluded that the contact between the Stangnes and Salangen Groups is a relatively minor thrust and that the main thrust transport represented by the Tjeldsund Thrust in the southern part of the area is carried by the Stangnes Thrust in the northern part of the area. However, there is little lithologic affinity between the two groups. One possible connection is the occurrence of the mafic igneous clasts found by Tull in the Salangen Group meta-conglomerate near Ballangen (see Chapter 2, "Harstad Conglomerate"), which one might argue could have been derived from Stangnes amphibolite. However, the massive, unfoliated to weakly foliated clasts bear little resemblance to the well-banded and foliated Stangnes amphibolite, so that this interpretation seems unlikely.

Emplacement of the Tjeldsund Thrust appears to have occurred roughly synchronously with the Stangnes Thrust. The contact is concordant on outcrop scale with no observed contrast in metamorphic grade across it, and the rocks have been overprinted following juxtaposition with the same structural/metamorphic events.

The direction of transport and magnitude of displacement of the Tjeldsund Thrust are unknown. Relative to the autochthon, it is very far -travelled, presumably further than the next subjacent thrust sheet, composed of the Stangnes Group. However, the possibility of an earlier phase of thrusting pre-dating D_1 on east Hinnøy cannot be ruled out, so that stacking relatinships may be more complex than they appear in Figure 29.

Relationship Between D_1 Thrusts

As noted above, north of the Storvann Fault, two D_1 thrusts are present, while south of it, only the Tjeldsund Thrust is present (see also Plate IIA and B). Either the two thrusts merge as one, or one thrust is younger and truncates the older. The critical relationships which would resolve the nature of this transition have been apparently cut out by the Storvann Fault. However, the regional geometry of thrust faults is an anastomosing pattern with the nappes in between pinching out, and sometimes reappearing (Figure 29; Chapter 6). The preferred interpretation is that the thrusts are the same age and merge to form a single surface at Leikvik. This surface splays again into two thrusts south of Ofotfjord.

Significance of the Mixed Zone

A mixed zone of lenticular bands of calcareous and pelitic schist, marble, and amphibolite is typically associated with the lowest thrust contact of the nappe stack on east Hinnøy (Stangnes Thrust in the north, Tjeldsund Thrust in the south of study area). The lithologies and petrography of these rocks was described briefly in Chapter 2. Occasionally some of the rocks in the mixed zone have affinities with the adjacent autochthonous rocks. The extreme mixing of lithologies suggests tectonic interleaving. However, this is hard to prove on the outcrop; all contacts have been strongly metamorphosed, and probably metasomatized, since many contacts are now gradational.

Regional relationships strongly support the interpretation of this unit as a zone of tectonic slivers. As the Tjeldsund Thrust is traced southward from the southeast corner of the study area (Plate I; Figure 3), the thrust bifurcates, with a large wedge of Narvik Group rocks appearing in between. Directly south of Ofotfjord (Vargsfjord, Figure 29), another lens of parautochthonous Storvann Group rocks occurs underneath the basal thrust. Traced southward to Forsaa, the thrust progressively cuts out the Storvann Group rocks. Immediately north of the place where the Storvann Group rocks are cut out, Storvann Group lithologies (quartzite, quartz -garnet schist, calcite marble) are exposed in a zone of intense mixing identical in style to those exposed on Hinnøy. It is noteworthy that this locality is also at the base of the nappe pile.

Folding

Evidence for F_1 folding of the Storvann Group comes from fold interference patterns on both outcrop and map scales. In the basal quartzite unit of the Storvann Group, mesoscopic Ramsay Type 1 interference patterns (Ramsay, 1967, p. 521) are abundant in the Storvann(S) section, and common in the Torskvatsfjell area east of Kvaefjord (Figure 30). The domes and basins are recumbent, suggesting either two phases of recumbent isoclinal folds were superposed, or that the second folding flattened the earlier folds into a recumbent isoclinal geometry. The interference patterns are the product of F_2 (see below) superposed on earlier folds. The F_1 folds are tentatively correlated with the time of nappe emplacement. The S_3 axial plane cleavage (not readily visible in the photograph, Figure 30) clearly cuts obliquely across the interference structure and hence post -dates its formation. Vergence has not been determined for the F_1 folds, since the smooth-surfaced exposures of the quartzite generally make it impossible to determine whether an interference pattern closes into or away from the outcrop.

Figure 30: Interference between F_1 and F_2 folds in quartzite of the Storvann Group.

Ramsay type 1 (dome and basin) interference pattern. S_1 and S_2 are subhorizontal. S_3, difficult to distinguish in the photograph, dips gently to the left (north), and is superimposed on the interference pattern.

The outcrop pattern of Storvann Group rocks on Finnslettheia and Sørvikfjell (Plate I; Plate II A, C-C') is also suggestive of early fold superposition. On the east side of Finnslettheia, an eastward closure of the upper calcite marble is present, enclosed by quartz-garnet schist (ss_1). This is thus a syncline, with younger rocks in the core. To the east on Sørvikfjell, a similar eastward closure is present, but the fold is an anticline: the basal quartzite is enclosed in turn by the lower calcite marble and quartz-garnet schist. The fact that an anticline and a syncline close in the same direction implies that these folds are a dome and a basin. The western closures of these structures are truncated by the Storvann Fault. Note again that the steeply to moderately dipping axial planes of west vergen F_3 chevron folds are oblique to these early recumbent structures and hence not related to their formation. This implies the presence of larger scale F_1 fold structures as well, with amplitudes on the scale of a kilometer or more.

This early fold phase may be correlative with F_0 phase tentatively identified by Hakkinen (1977) on the basis of fold interference patterns in the Austerfjord area. However, the fact that, on east Hinnøy, D_3 structures can be demonstrated to be superimposed on the earlier interference patterns, and that the rocks folded are Caledonian metasedimentary rocks, allows greater confidence in asserting the presence of two early fold phases.

Fabrics

The most prominent foliation of the Caledonian nappe terrain and immediately subjacent basement formed at least in part during D_1. S_1 is a penetrative schistosity defined by dimensional and crystallographic preferred orientation of micas and amphibole, and to a lesser extent quartz and feldspar. Carbonate minerals have generally recrystallized to form equant granoblastic grains that do not define a schistosity. Compositional banding is parallel to S_1 except in F_1 fold hinges. However, most isoclinal folds observed are F_2 in age and fold the S_1 schistosity isoclinally.

The flattening of S_0 compositional banding into S_1 involved both simple rotation and transposition by folding. At the top of Finnslettheia, infolding of the quartzite/basement contact is minor, so that the contact observed in outcrop can be considered actually parallel to S_0, the original bedding. By contrast, the shoreline exposures at Tjeldsund show this contact to have suffered intense isoclinal folding, with the result that several alternating bands of quartzite and granite gneiss are present in a cross-strike traverse. S_0, properly defined here by the envelope of F_1 folds, is not parallel to S_1; its orientation has been lost by intense transposition.

In micaceous rocks, S_1 schistosity may be disrupted or even obliterated by later foliations, although compositional banding may be coherent enough to make S_1 still measurable in outcrop. However, in non-micaceous lithologies, especially amphibolite, S_1 is generally only folded by successive deformation phases with very little new fabric development.

No lineation clearly related to D_1 was recognized. This may be partly due to the presence of later well-developed lineations (L_2 amphibole and stretching lineations, $L_{1/3}$, $L_{2/3}$, and $L_{1/4}$ intersection lineations), which obscure earlier structures. If L_1 lineations were developed at a small to moderate angle to later lineations, which are nearly all subparallel, distinction of L_1 would be especially unlikely.

A systematic study of pebble elongations in the Harstad Conglomerate was not made, but brief field examination indicates a mildly constrictive strain field. Axial ratios are approximately 1:4:10, with the long axis trending N70E. This constitutes a composite of all five deformations recognized in the area; there is no obvious way to resolve the D_1 component present in this finite strain. Consequently, the characteristics of the D_1 strain field remain unknown.

D_2: Recumbent Folding and Thrusting

The second Caledonian deformation on east Hinnøy is a complex composite of deformation styles, including recumbent isoclinal folding on scales up to several kilometers, development of mylonitic thrust zones (including the Austerfjord and Kvaefjord Thrusts), and steep ductile shear zones within the basement transported by these thrusts. A variable but locally strong amphibole lineation developed, which together with analysis of F_2 folds indicates an overall transport direction of ESE for this deformation. A large-scale kinematic model for D_2 incorporating all these features is presented at the end of this section.

Recumbent Folding

Major Folds

Several large scale (wave lengths > 1 km) F_2 axial surfaces have been recognized in the study area (Plate III). The average direction of the traces of these is NE-SW, except the folded thrust near Fjeldal and the Kanebogen synformal anticline near Kanebogen, where large late upright folds have re-oriented the F_2 axial surfaces to NW-SE traces. The vergence of the F_2 folds is southeast, consistent with the transport direction determined below using small scale structures. The folds across the northern side of the area have been rotated to a downward facing orientation so that anticlines are synformal and synclines antiformal. The F_2 folds were nearly strictly recumbent at the time of their formation. Present steep dips and/or plunges are a result of refolding.

Hinges of large F_2 folds are exposed only on Middagsfjell and near Sorvikfjell. The nose of the antiformal syncline in the core of which Harstad is located can be walked around on the north side of Middagsfjell. This fold clearly folds the nappe-emplacing thrusts and associated S_1 schistosity (Plate II A, A-A'). The limbs are essentially isoclinal. The style of the fold is assumed to be more or less similar, but because it folds two thrust faults, thickness changes from limbs to core are not reliable indicators. The appearance of the Harstad Conglomerate in the western limb is not understood. This may be a second-order antiform on the west side of the main axial surface.

Hinges of other F_2 folds in the northern part of the study area are not exposed. The Kanebogen synformal anticline is recognized on the basis of mirror image sequences of Storvann Group and overlying allochthons across its axial surface. Its synformal geometry is required by the antiform to the west, and is consistent with the slight synformal convergence of dips on its limbs (Plate II A, A-A'). This fold is important because it unequivocally demonstrates important basement imvolvement in F_2 recumbent folding. Both Kvaefjord Group rocks and granite gneiss (Gullesfjord Gneiss?) occur in the core of this fold.

The synformal anticline at Storvann(N) is not readily recognized independently since it is mainly developed in massive Precambrian granite gneiss. An overturned synformal closure of the S_1 schistosity is apparent in the

Kvaefjord Group rocks immediately west of Storvann(N), but it is not obvious that this is an F_2 fold. However, this area lies between two F_2 antiforms so that it must be synformal in a general way; the simplest interpretation is that the closure in the Kvaefjord Group rocks on Torskvatsfjell is this synform. This would thus appear to be a second, structurally lower, basement-cored recumbent anticline underneath the Kanebogen anticline.

The hinge of the F_2 antiformal syncline at Vikelandsfjell is also unexposed, but it is believed to be antiformal for two reasons: (1) the structural plunge in this area is eastward (Plate IV, subarea B), and the limbs converge in that direction (Plate I); and (2) if it were synformal, it would constitute a major F_2 fold with opposite vergence from all other F_2 structures. The fold is now rootless, riding on the Kvaefjord Thrust (Plate II D, I-I'). This is the structurally lowest axial surface recognized in the F_2 fold stack.

Major early fold hinges are also exposed on Sørvikfjell and Finnslettheia. These are the hypothetical dome and basin structures described in the previous section as resulting from interference of F_1 and F_2. Consequently these closures are not purely F_2 in age. The style of the folds is isoclinal and similar, but now strongly distorted by F_3 refolds. The few outcrops of Stangnes(?) amphibolite on the SE side of Finnslettheia are tentatively considered to sit in the core of the basin, the upper calcite marble perhaps having been cut out along the Stangnes Thrust locally here.

Additional F_2 folds are recognized in this area on the east side of Sørvikfjell, where the Stangnes amphibolite is recumbently infolded with Storvann Group rocks. These folds sit structurally below the recumbent dome on Sørvikfjell; the synform cored by Stangnes Group rocks may be the western hinge of another northerly-closing basin, explaining the presence of Stangnes Group rocks to the west of their main outcrop belt near Kilbotn.

The relationship of the folds on Sørvikfjell to those near Harstad is not obvious but can be inferred by tracing rocks of the inverted limb between the Harstad antiform and Kanebogen synform southward. The Storvann Group rocks from the SE side of Storvann(N) south to the Langvann Fault (Plate III) are folded in a large overturned F_3 antiform but all belong to the overturned limb of this F_2 fold couple (Plate IID, J-J', K-K'). Since the Langvann Fault dies out not far westward and thus cannot have large displacement, the inverted basement/Storvann Group contact north of the fault reasonably matches the same contact south of the fault on the top of Finnslettheia. This places the Precambrian granite at the top of Finnslettheia in the core of the Kanebogen anticline (not synformal here!). The dome and basin structures underneath appear to be second-order structures in the inverted limb, because in passing structurally downward to the east, no major reversal of the stacking sequence occurs.

The upper, upright limb of the Kanebogen anticline is only recognized in one locality on east Hinnøy other than the Stangnes peninsula. Two kilometers east of the southern end of Straumsbotn, an F_3 synformal keel is cored by Storvann Group quartzite and quartz-garnet schist in upright sequence (Plate IID, I-I'). Perched on top of basement rocks which are in structural continuity with the basement which cores the Kanebogen anticline near Finnslettheia, these rocks can only belong to the upper limb of the Kanebogen anticline. Everywhere else on east Hinnøy this upper limb has been removed by erosion, exposing lower elements of the F_2 structural pile.

The Kilbotn Schist may include an unrecognized S_2 axial surface. It was suggested in Chapter 2 that the large apparent thickness of this unit could be due to doubling up by recumbent folding. The Kilbotn Schist lies structurally lower than the F_2 axial surfaces on Sørvikfjell, and at a similar level to the Harstad antiformal syncline. It thus possibly contains the southeastward continuation of the axial surface of the Harstad syncline. Furthermore, the quartzite at Kvitnes (see Salangen Group, Introduction, and Figure 25) is in the core of a NW-closing antiform, wrapped around by Kilbotn Schist. If the lithologic correlations suggested in Chapter 2 are correct, this fold is probably a further continuation of the Harstad antiformal syncline, extending its axial surface nearly 20 km across the strike.

The Kilbotn Schist underlies the Salangen Group marbles south of the Storvann Fault, which are in turn structurally overlain by the Storvann Group and Precambrian granite without significant repetition passing southward. This requires that all the rocks exposed in the area south of the Storvann Fault and north of Gausvik are in the inverted limb, or structural equivalent thereof, of the Kanebogen anticline.

The remaining possible major F_2 axial surface is recognized in the southeast corner of the map area near Fjelldal. Salangen Group marbles with intercalated schists, intensely flattened and boudinaged, are bracketed by thrust contacts with the Austerfjord Group above and Precambrian granite (Lodingen Granite) below. This relationship is equally well-explained by either interleaving by D_2 thrusting of the Salangen Group and basement rocks, or by recumbent folding of the D_1 Tjeldsund Thrust under the D_2 Austerfjord Thrust (see Plate II B, Section E-E'). The latter interpretation is favored because the rocks strongly affected by D_1 further north have generally tended to fold rather than fault during D_2. However, the critical relationships are concealed beneath Tjeldsund or further west.

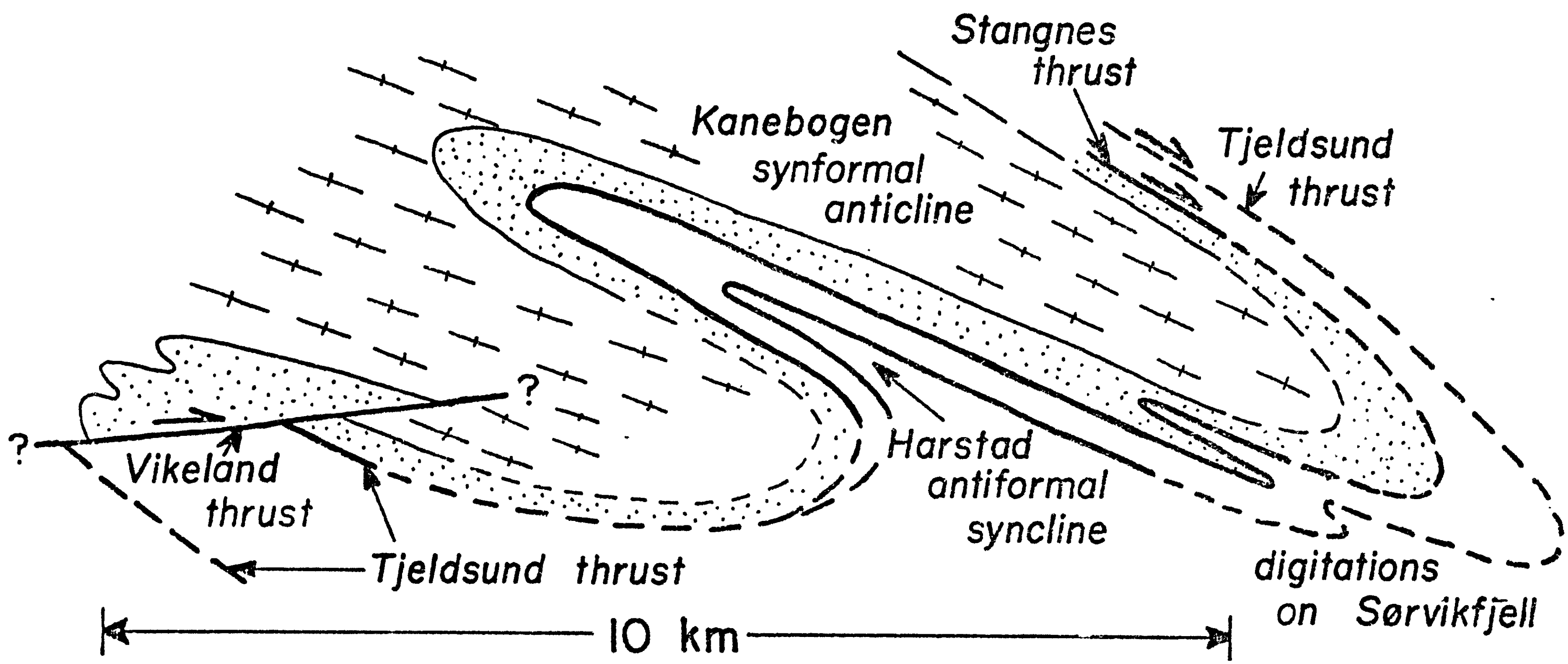
The position of the hinge (or pinch-out) of this belt of Salangen Group rocks is shown on Plates I and III to be under Tjeldsund. A second possibility exists. The schist and marble poorly exposed on the small hill east of Kongsvik are not typical Austerfjord Group lithologies (with which they have been grouped on Plate I). These could represent the nose of the band of Salangen Group rocks. The more conservative interpretation was shown on the map because the extent and quality of exposures encourage caution; however, if this second alternative is correct, it would increase the amplitude of this fold by several times over that shown on Plates I, II, and III.

The amplitude of the major F_2 folds is large but poorly constrained. Simple measurement of present minimum hinge to hinge distances using the basement/cover contact as a datum yields amplitudes of 5 to 10 km (Figure 31). These are in fact minimum reasonable estimates. Although F_2 structures were strongly refolded in later deformational events, the axial direction of F_3 is at a large angle to the F_3 slip line and hence probably near the intermediate strain axis (see below). Consequently, the post-F_2 length change in a NW-SE direction across most of the area (excluding locales of strong F_4 folding) is probably minor.

As shown in Figure 31, the Kanebogen anticline is the structurally highest F_2 fold recognized in the study area. It need not have been the structurally highest fold formed in F_2 in this area. Reconnaissance by the author east of the study area, where structurally higher units are preserved in the core of the Haafjell synform (Figure 2), indicates that F_2 recumbent folding is probably present at higher structural levels as well.

Figure 31.

Schematic composite profile of F_2 fold nappe geometry



A long-known problem of the Scandinavian Caledonides is the observation that early fold axes are mainly oriented transversely to the trend of the mountain belt (Kvale, 1953). This has stimulated considerable discussion leading to a variety of mechanisms being proposed to explain this fold axial orientation. These include initial formation of fold axes at low angles to the transport direction (Hansen, 1971), progressive rotation by continued finite strain into the transport direction (Bryant and Reed, 1969; Hakkinen, 1977), and tectonic transport parallel to the trend of the orogen (Olesen, 1971). On east Hinnøy, no such problem appears to exist; major early fold axes are oriented mainly parallel to the trend of the mountain belt. The relatively modest finite strain values recorded in the Harstad Conglomerate may be related to this; there may have been inadequate post -folding strain to rotate the fold axes into the transport direction. This would support Hakkinen's preferred interpretation that the transverse early fold axes in the Austerfjord area were produced by rotation by continued finite strain.

Minor Folds

Minor folds of F_2 age are commonly observed in the metasedimentary rocks of east Hinnøy. They are isoclinal and similar in geometry. Amplitude to wavelength ratios vary from about 3 to very large values. Fold hinges are sometimes completely isolated by limb attenuation so that no vergence can be determined.

Although good three-dimensional exposures of early fold axes with determinable asymmetry are not common, enough folds were measured to allow a slip line analysis by the separation angle method of Hansen (1971). The F_2 fold axis data are plotted stereographically in Figures 32A to E. Some subareas included no minor F_2 fold axis measurements, while in others only a few measurements were made, so that data from more than one subarea have been plotted on the same stereogram. A wide variety of axial orientations are represented, but vergence of the folds is consistently to the south and east. The variability of axial trends allows determination of the separation angle, which is the angle between domains of opposite fold vergence on a stereographic plot, within which the transport direction lies.

A separation angle only can be determined within the (common) axial plane of the folds analyzed. Consequently, for each subarea a visual best

Figure 32: Structural analysis of F_2 folds.

A - E: Equal area plots of F_2 fold measurements

F: F_2 slip line determination. Great circle segments are separation angles defined by each subarea, as labeled. Numbers in parentheses are number of measurements which were used to define each separation angle. Shaded area at lower right is the sector of overlap of all individual subarea separation angles with compatible vergence. This area thus contains the transport direction (slip line) defined by this analysis; this slip line is thought to represent the actual transport direction during F_2 folding (see text for discussion).

Figure 32A.

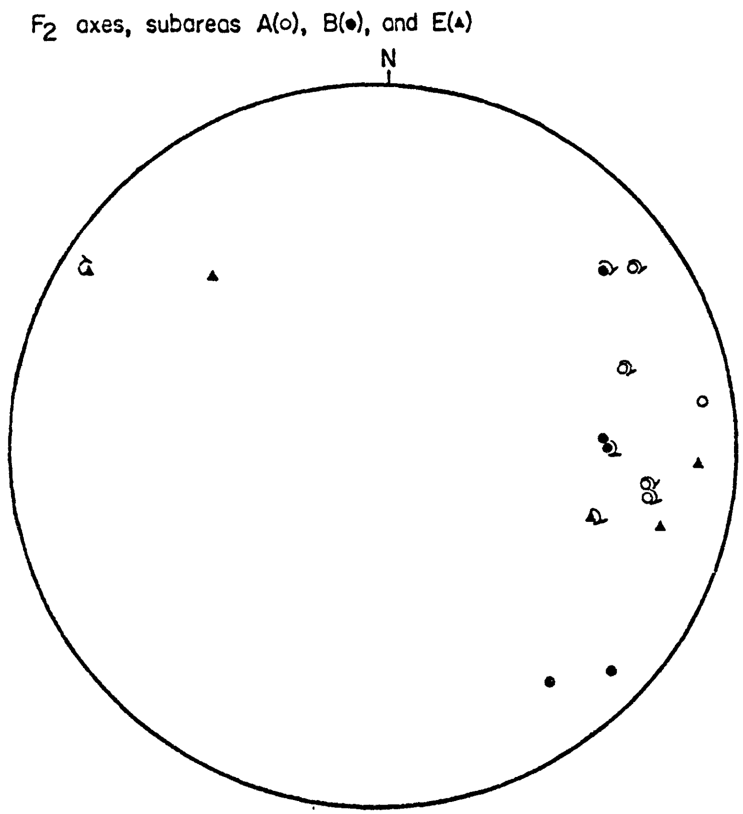


Figure 32B.

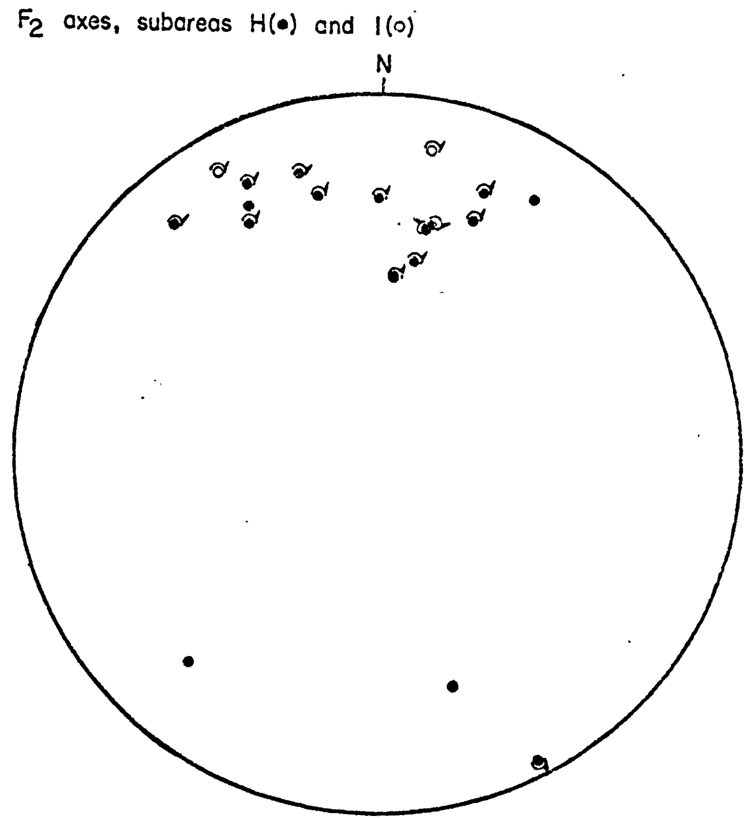


Figure 32C.

F_2 axes, subarea J

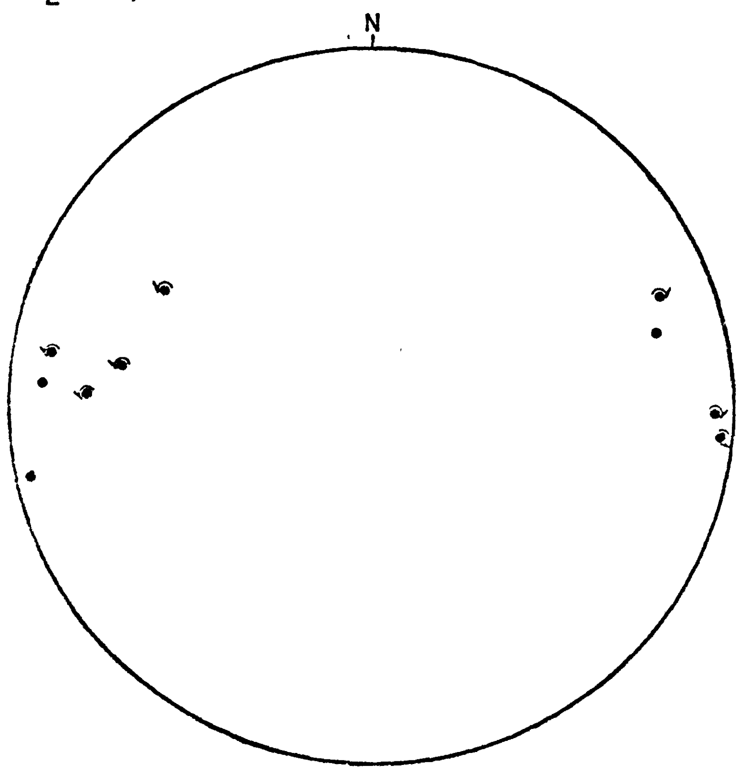


Figure 32D.

F_2 axes, subarea K

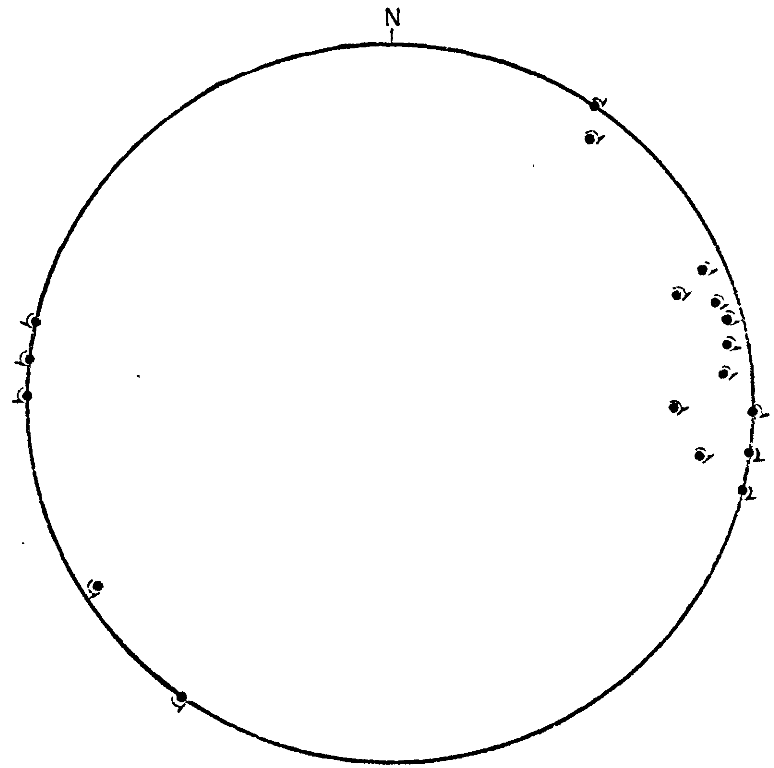


Figure 32E.

F_2 axes, subareas L(o), M(Δ), and Q(•)

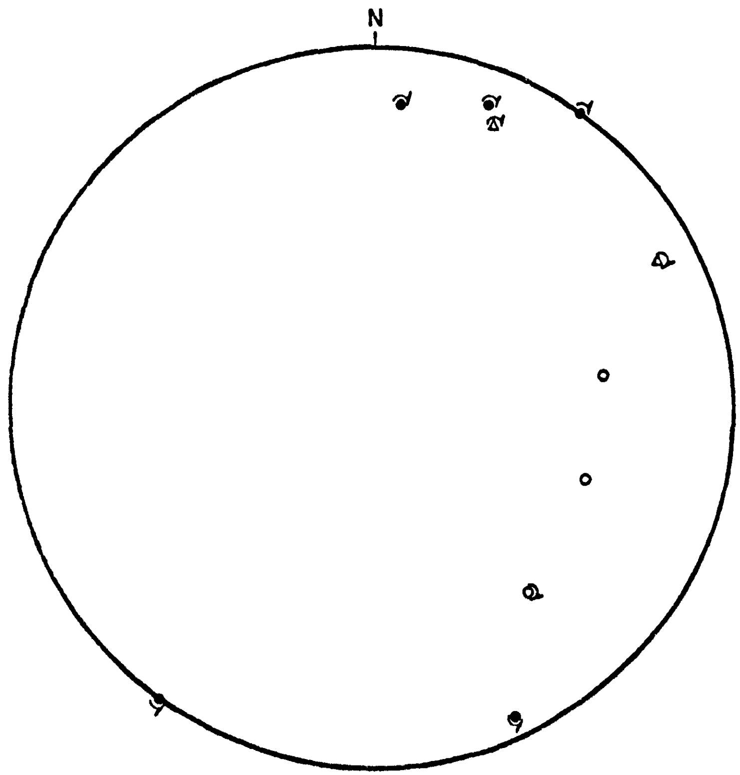
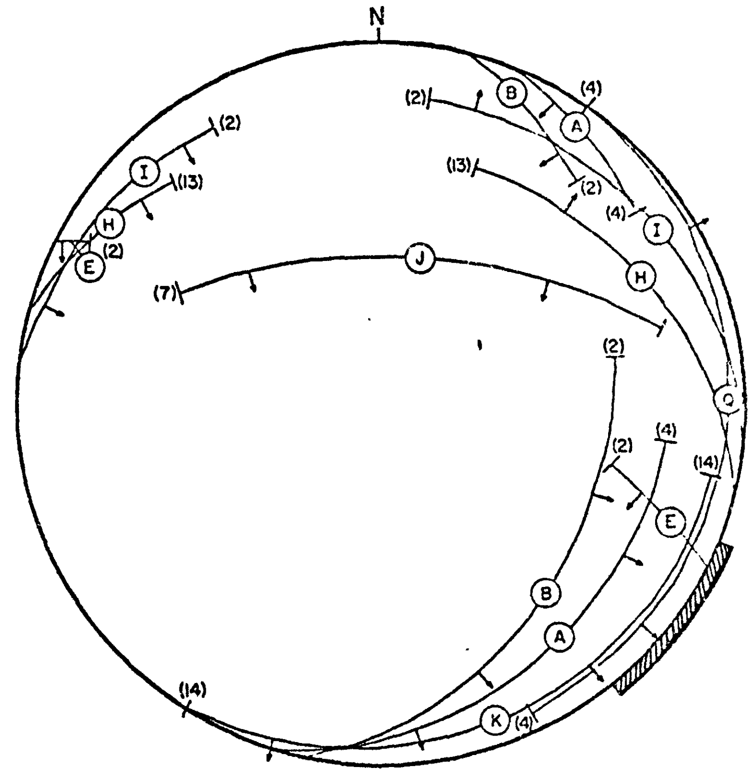


Figure 32F.

F_2 slip line determination



fit great circle was drawn through the fold axial measurements to approximate a mean S_2 orientation. A separation angle for each subarea was determined (Figure 32F); because the variability within each subarea is not large, these individual separation angles are wide, and individually not very informative.

Comparison of the subareas yields more precise results. The plunges of F_3 and F_4 folds are generally low, except where F_4 locally refolds F_3 (see below). This implies that the surface deformed by these folds was sub-horizontal prior to F_3 and F_4. Thus, the most appropriate plane in which to compare the different individual domains is the horizontal. The deviations from horizontal of the great circle segments on Figure 32F are due mainly to F_3 folding, which is asymmetrical and typified by long flat limbs and short steep to inverted limbs. The measurements plotted all lie in low -dipping, upright limbs except subarea J (the apparent steep dip of subarea E is an artifact of having only two measurements). Measurements from subarea J come from the overturned limb of the large F_3 antiform east of Storvann(S). Consequently, to restore the great circle segments to horizontal, all subareas but J were rotated the short direction to the primitive; the great circle segment from subarea J was rotated the long direction (southward).

There are two sectors where all subarea separation angles overlap, one in a NW-SE direction and one NE-SW. However, as can be seen in Figure 32F, the vergences of folds are not compatible in the NE-SW direction. It is thus considered that separation angle analysis of F_2 minor folds constrains a transport direction within the shaded arc on Figure 32F, with southeastward vergence. The validity of this analysis is reinforced by the lack of any conflicting measurements; all the F_2 fold axis measurements are consistent with this slipline orientation.

An important consideration is to which deformation this separation angle applies. A separation angle may develop in three ways: (1) Due to areally variable slip between planar volume elements, the strain trajectories locally vary to produce a variety of fold axial orientations (the tundra slide model, Hansen, 1971); (2) the surface folded may not have been strictly planar, or the superposed strain field not spatially homoaxial, leading to a variety of orientations of intersections between the folded surface and the XY-plane (plane of flattening) of the strain ellipsoid; this will lead to variable fold axial orientation for purely geometrical reasons; and (3) the superposition of a later, heterogeneous finite strain that rotates pre-existing homoaxial folds into a variety of orientations. In the first two cases, the separation angle has kinematic significance for the event which produced the folds. In the third, the separation angle applies to the kinematics of a later, superposed deformation.

In order to resolve this question, slipline determinations of the F_3 and F_4 fold phases were attempted using the Weiss (1959) method of folded lineations (Figures 51, 54). These determinations are discussed below with regard to the late fold phases, but it is concluded that neither of these sliplines bears any similarity to the separation angle defined here.

Boudinage of schist and amphibolite within marble, believed to be a result of stretching in the limbs of F_2 folds, has been recognized locally. The boudins are generally not well exposed in three dimensions, but three measurements were made in the rocks of the Salangen Group infold west of Fjelldal (Figure 33). These measurements are consistent with the elongation direction inferred by the F_2 separation angle.

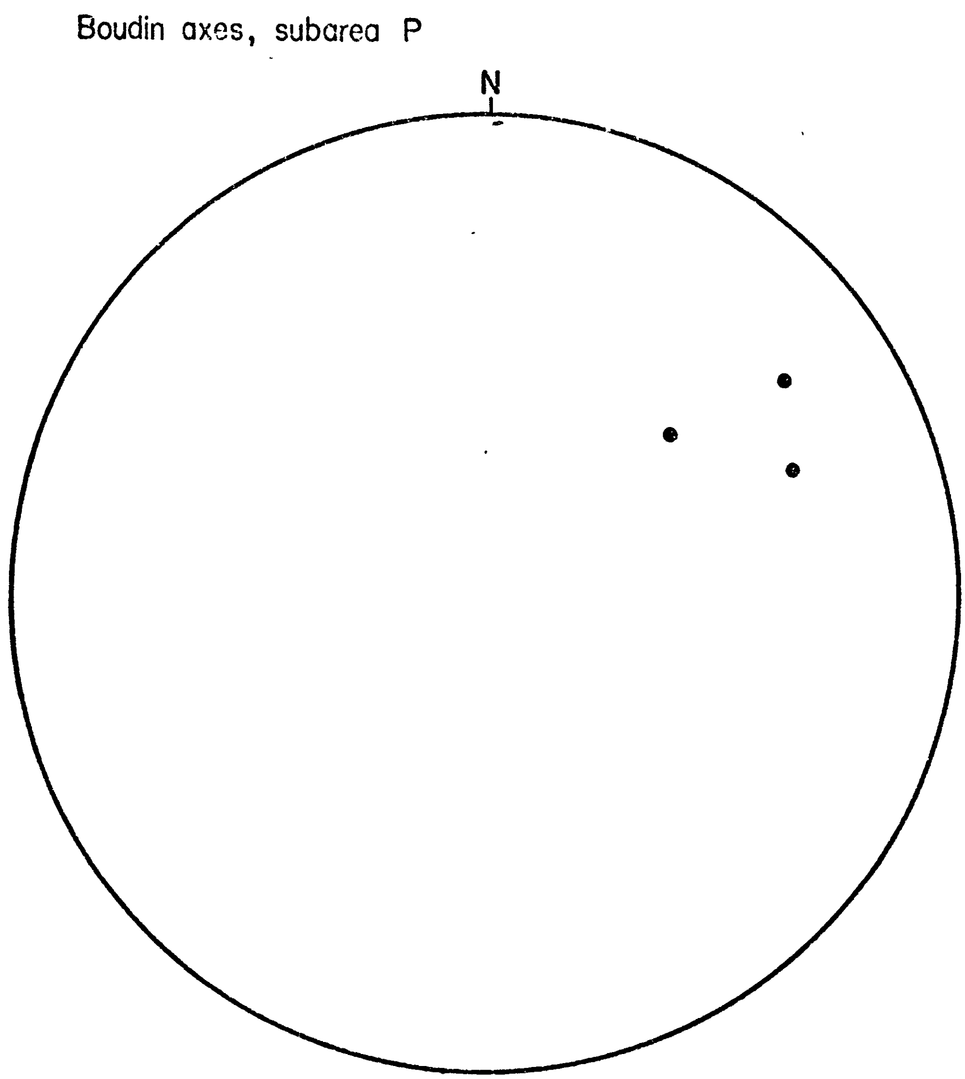
It is believed that this F_2 separation angle reflects the kinematics of the D_2 deformation. The preferred mechanism by which the variable F_2 axial orientation developed is the second listed above. Since folds axes within single subareas are fairly consistent, the heterogeneities required would be on a rather broad scale, a likely occurrence in superposed deformations. This obviates the need to appeal to the tundra slide model of Hansen (1971), which is difficult to relate mechanically to deformation mechanisms in amphibolite facies metamorphic terrains.

Thrust Faults

Two thrust faults of D_2 age have been mapped in the study area: The Austerfjord Thrust, named and described by Hakkinen (1977) from a type area at the western edge of Plate I; and the Vikeland Thrust, a somewhat similar structure at the head of Kvaefjord, especially well-developed on the west end of Vikelandsfjell. The possibility exists that these structures are the same thrust and connect below the surface, but this cannot be demonstrated so local terminology is retained.

These thrusts are in general characterized by zones of mild to very intense high temperature mylonitization from a few meters to hundreds of

Figure 33.



meters thick. Along the Austerfjord thrust, the mylonitic fabric is commonly well developed in the upper plate, while mylonitic effects die out rapidly downward in the lower plate. Similar relationships also occur at the Vikeland thrust, where, on Vikelandsfjell, more than a hundred meters' thickness of mylonite gneiss overlies non-mylonitic but structurally dismembered metasedimentary rocks of somewhat uncertain affinity. As noted by Hakkinen with regard to the Austerfjord thrust, these surfaces are synmetamorphic in their development. No retrograde metamorphism is recognized along the contact, and there is no contrast in metamorphic grade across it.

Unlike the D_1 thrusts, these are not considered to be fundamental tectonic boundaries juxtaposing once widely separated terrains. They are rather thought to be the response of the massive basement rocks to the deformation which produced recumbent folding in the cover rocks and their immediately subjacent basement. The transition between these two styles will be discussed with regard to the kinematic model for D_2.

Austerfjord Thrust

The Austerfjord thrust as described by Hakkinen (1977) placed Archaean gneisses on top of Middle(?) Proterozoic Austerfjord Group metasedimentary rocks and the 1400 Ma Lødingen Granite which intrudes it. Hakkinen thus viewed this thrust as the juxtaposition of two distinct basement terrains of different ages and as a thrust of potentially major transport.

The contact is readily recognizable in areas where the Austerfjord Group is present, due to the large contrast in erosion resistance of the massive granite gneiss of the upper plate and softer metasedimentary rocks in the lower plate. On the west limb of the F_3 antiform at Austerfjord (directly SW of the present study area), the Austerfjord Group pinches out and the thrust juxtaposes Precambrian granite against Precambrian granite, making it much more difficult to trace in the field.

It is not always clear that the actual granite/metasediment contact is the main locus of movement (see also Chapter 2, p. 56). At certain locations along the Austerfjord thrust (e.g., 0.5 km west of the summit of Tverrfjell), the granite gneiss directly above the Austerfjord Group quartzite and amphibolite is only mildly foliated for a few hundred meters upward. The granite gneiss then grades into fine-grained blastomylonite gneiss, suggesting the main movement was localized within the granite gneiss at a higher structural level.

In the present study, the Austerfjord thrust was traced southeastward from the limits of Hakkinen's mapping (the west side of Plate I) to and across Tjeldsund to Fjelldal where it merges with other Caledonian structures and cannot be confidently traced further as a separate structure. Relationships in the Fjelldal area are of vital importance for understanding of the Austerfjord thrust for three reasons: (1) they strengthen previously weak arguments that the thrust is a Caledonian structure; (2) they show that the thrust cannot be a surface of major tectonic transport; (3) they demonstrate that the thrust post-dates the emplacement of the Caledonian thrust nappe stack and is thus a D_2 structure.

On a small island in the intertidal foreshore area 2 km west of Fjelldal, the Austerfjord thrust is exposed placing Precambrian granite gneiss (probably Gullesfjord Gneiss) in a small klippe on top of Austerfjord Group amphibolite and quartzite. The rocks on both sides of the contact are strongly mylonitized for several meters away from the thrust. Structurally underneath the Austerfjord Group rocks (contact not exposed) are marbles and subordinate schists belonging to the Salangen Group. Eastward, the granite gneiss appears with some of its Storvann Group cover relatively intact, but inverted, presumably by recumbent F_2 folds. Little of the basal quartzite remains present, so that it is suspected some displacement has occurred along this contact. The Storvann Group rocks overlie Salangen Group marbles. Plate IIB, section E-E', depicts these relationships. It cannot be determined whether the Austerfjord thrust truncates the Tjeldsund thrust, so that the Storvann Group/Salangen Group contact on the east side of the synform is the Austerfjord Thrust in the strict sense, or if the Austerfjord Thrust reactivates the older thrust. The critical relationships are cut out by the Fjelldal Fault. However, it is clear that the Salangen and Storvann Groups are involved in the movement of the Austerfjord thrust. Prior to the mapping of this area, the Austerfjord thrust was known only to involve rocks older than 1400 Ma; the only hard evidence for a Caledonian age of movement was a single $^{40}Ar/^{39}Ar$ age of 390 Ma from biotite from the foliation along the thrust in the Austerfjord area (J. Sutter, pers. comm. to Hakkinen, 1977). The Caledonian age of the Austerfjord Thrust is now considered well-established.

South of the Fjelldal Fault, the Tjeldsund Thrust overrides a large lens of parautochthonous Storvann Group rocks on Ramboheia. Since the Fjelldal Fault is not a major fault, it can be safely stated that the Salangen Group rocks overlying this lens of Storvann Group are in virtual structural continuity with the Salangen Group rocks underlying the the Austerfjord Group 2 km west of Fjelldal. The granite gneiss on which the Storvann Group rests is almost certainly Lødingen Granite, one of the characteristic lower plate units of the Austerfjord thrust in its type area. Consequently, it is held that the Storvann Group rocks on Ramboheia belong to the lower plate of the Austerfjord thrust.

However, the Storvann Group rocks at Fjelldal are in the upper plate of the Austerfjord Thrust, as are all the Storvann Group rocks of east Hinnøy. Consequently, it appears that the same rock units are present on both sides of the thrust, leading to the conclusion that the Austerfjord Thrust is not a surface of major displacement.

This conclusion is supported by three independent lines of evidence. Griffin and others (1978) report that at the western margin of the Lødingen Granite, 30 km west of the Fjelldal area, the Lødingen Granite intrudes rocks of the Lofoten granulite terrain. A Rb/Sr whole rock age of 1380 $\pm$ 80 Ma from the granite in this area was reported by Griffin and others (1978). This age is statistically identical to the 1415 $\pm$ 80 Ma age obtained by P. Taylor (pers. comm. to Hakkinen, 1977) from the Austerfjord area. This clearly supports the conclusion that the Austerfjord thrust is not a thrust of large displacement since the relations require the thrust to die out completely to the southwest. E. Tveten (written comm., 1977) has reported just such a disappearance of the thrust from field observations on southwest Hinnøy, although tracing a granite on granite thrust fault is a difficult pursuit. Finally, the possible correlation of the Melaa Granite of the present study with the Lødingen Granite suggests another match of rock units across the thrust.

It has already been argued that the Salangen Group rocks comprise a far-travelled nappe. It was Hakkinen's (1977) suggestion that this and other nappes of metamorphic rocks may have rooted at the Austerfjord thrust, with the Lofoten terrain riding at the top of the pile. This suggestion becomes untenable if the Austerfjord thrust is not a major tectonic boundary; the allochthons must root to the west of the Lofoten terrain. This has implications for the timing of the Austerfjord Thrust. Salangen Group rocks are involved in Austerfjord thrust movements, but were already emplaced as far-travelled nappes. Consequently, the Austerfjord thrusting must post -date the emplacement of the thrust nappes and is a post-D_1 structure. Since F_3 folds deform the Austerfjord thrust and fabrics related to it, the thrust predates F_3.

Strict synchroneity of the Austerfjord Thrust with F_2 recumbent folds of other parts of the study area is difficult to prove, but two lines of permissive evidence suggest this is true. Hakkinen found that his F_1 isoclinal fold phase and movement on the Austerfjord thrust were essentially inseparable in time. The thrust is (rarely) folded by his F_1 folds, but the mylonitic fabric of the thrust and the axial planar schistosity of the folds are concordant and continuous with one another. Consequently, the deformations are essentially synchronous. If the most prominent and latest phase of Caledonian isoclinal folding in the Austerfjord area can be correlated with a similar phase, F_2 in this study, several kilometers to the east in the present map area, the Austerfjord thrust belongs to the D_2 of the present study.

A second argument is based on spatial considerations. Between Harstad and Storvann(S), a set of large F_2 recumbent folds represent an uncertain but significant amount of shortening of both cover and basement. These folds plunge under a large, mildly deformed slab of Precambrian basement to the east of Storvann(S). When the equivalent structural level reappears on the other side of the broad F_3 synform in which the basement slab rests, only the Austerfjord thrust and a thin lens or nose of Salangen Group rocks lies between upper and lower massive blocks of basement granite. It appears that the same strain, distributed through a pile of fold nappes in the northeastern part of the area, has been concentrated at a single horizon as a thrust fault in the southwestern part of the area. This implies synchroneity of F_2 folding and movement on the Austerfjord thrust, as well as suggesting a kinematic model for D_2 as a whole. Completion of this kinematic model is deferred until the Vikeland thrust, the shear zones at Middagstind, and the D_2 penetrative fabrics have been discussed.

Vikeland Thrust

The Vikeland thrust is marked by a zone of mylonitization developed in the Precambrian granite on the western shoulder of Vikelandsfjell, dipping gently east. From there it cuts southward to the east shore of Straumsbotn, and reappears, dipping westward, on the south shore of Kvaefjord west of Straumsbotn. North of Vikelandsfjell, the thrust disappears under Quaternary sediments.

At Kvaefjord, the mylonite zone along the thrust is a fine-grained schistose mylonite gneiss with a prominent stretching lineation bearing S64E, developed from Precambrian granite gneiss. This mylonitic fabric is locally deformed by F_3 chevron folds, demonstrating the pre-D_3 age of the thrust. In the lower plate near the mouth of Straumsbotn, quartzite, marble and quartz-garnet schist of the Storvann Group are exposed along the shoreline. Since the Storvann Group rocks south of Vikelandsfjell lie in the upper plate of the thrust, like the Austerfjord thrust the Vikeland thrust can not be a tectonic boundary of major displacement.

Along the east shore of Straumsbotn, a quartz-garnet schist, similar to those in Storvann Group, is interlayered with micaceous and feldspathic paragneisses of the Kvaefjord Group. Exposures lack good continuity so that the relationships are interpreted somewhat hesitantly, but it is considered that these rocks mark a zone of tectonic interleaving along the thrust. Between the east shore of Straumsbotn and Vikelandsfjell, the thrust is difficult to trace due to poor exposure, but lithologic repetitions suggest more slivering. The large F_2 antiform to the east is clearly rootless, so that the tectonic contact must at least approximate the relations shown on Plate I and Plate IID, section I-I'.

On the west shoulder of Vikelandsfjell, a zone more than 100 m thick of mylonitic granite gneiss is developed above a more or less chaotic zone of Storvann Group lithologies plus calcareous schist, pelitic schist, and marbles reminiscent of the sliver zone on the east shore of Storvann(S). Lithologies can scarcely be traced from one outcrop to the next. It is concluded that this is probably the zone of D_1 slivering reappearing in the lowest recumbent synform of the F_2 fold stack. It has probably been further dismembered by slivering associated with the Vikeland thrust.

Relative timing and geometry of the F_2 folding and Vikeland thrust are uncertain. No retrograde metamorphism appears associated with the Vikeland thrust; it appears to be synmetamorphic. On a large scale, it truncates the F_2 recumbent antiformal syncline south of Vikelandsfjell, but if folding and thrusting were closely associated in time, such truncations would be a natural consequence of the deformation. A weak to moderate penetrative schistosity, dipping gently east parallel to the thrust, crosses at a small angle an earlier schistosity (presumed to be S_1) in the upper plate. Stereographic plots of these fabrics (Plate IV, subarea B, S_1, and S_2) indicate both fabrics have experienced similar structural histories since their formation.

The Vikeland thrust is thus assigned a late D_2 age on the basis of similar transport direction (based on the stretching lineation) to F_2, similar timing relative to metamorphism and F_3, but its low angle truncation of F_2 structure. This is consistent also with its similar style to the Austerfjord thrust. It would appear that the transition from fold to thrust tectonics in D_2 may be in part accomplished in the manner apparent at Vikelandsfjell, with folds forming and then being cut by thrusts and transported in the later part of the same deformation.

Shear Zones at Middagstind

In the contact hornfels on the east and southeast sides of the Middagstind Quartz Syenite, commonly at the contact, and to a lesser extent within the pluton itself, zones of intense foliation are developed. These vary from a few cm to a few tens of meters in width, in one case reaching several hundred meters wide. The zones were not mapped individually in this study, but numerous structural attitudes were taken in the shear zones, and their general characteristics recorded. In the contact zone, the shear zones are characterized by weak to strong hornblende alignment (Figure 34). A hornblende lineation within the foliation is common. Where the shear zones are localized at the contact, the otherwise orthopyroxene -bearing contact hornfels is retrograded to amphibolite facies. In the syenite, the shear zones are only a few centimeters thick and very sparse, except for a 100 m-thick zone of mylonitic augen gneiss that is developed on the south side of the pluton. No lineation was recognized in the shear zones within the syenite. The foliation is defined by biotite alignment and feldspar elongation in the small shear zones. The fabric is developed by cataclasis, and forms through-going mylonitic bands in the augen gneiss (Figure 35).

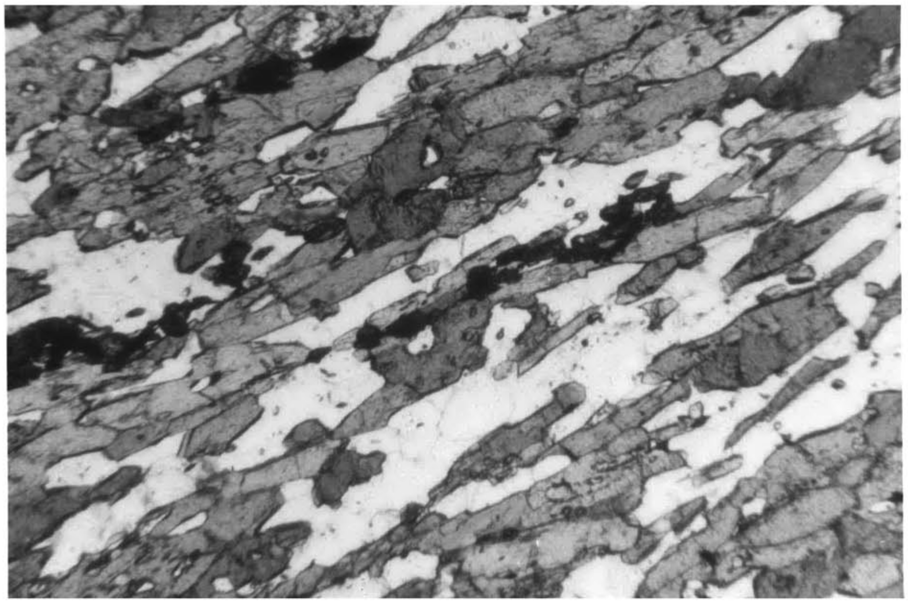


Figure 34: Photomicrograph of sheared hornfels.

Foliated hornfels (amphibolite) from sheared contact of the Middagstind syenite pluton (cf. Figure 10). Plane polarized light, 75X.

Figure 35: Mylonitized Middagstind Quartz Syenite.

A: Hand specimen (cf. Figure 7). Actual size.

B: Photomicrograph. Porphyroclasts are perthitic microcline, commonly with similar complex exsolution patterns observed in undeformed syenite (cf. Figure 8). Crossed polars, 75X.

Figure 35A.

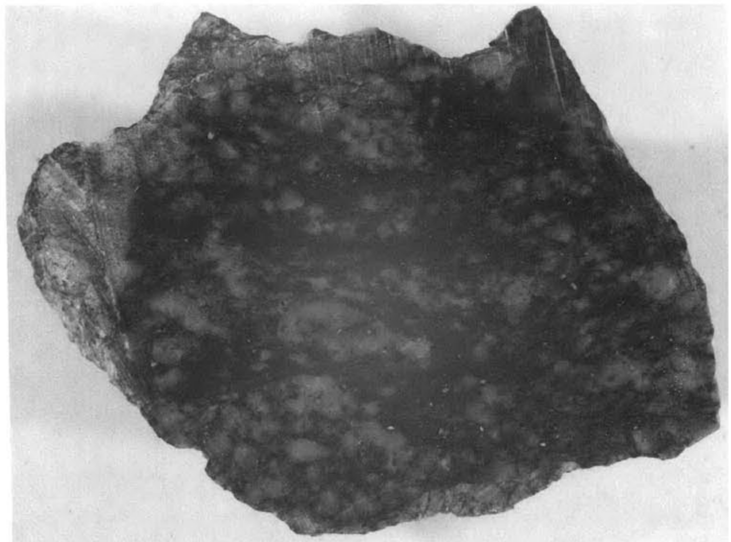
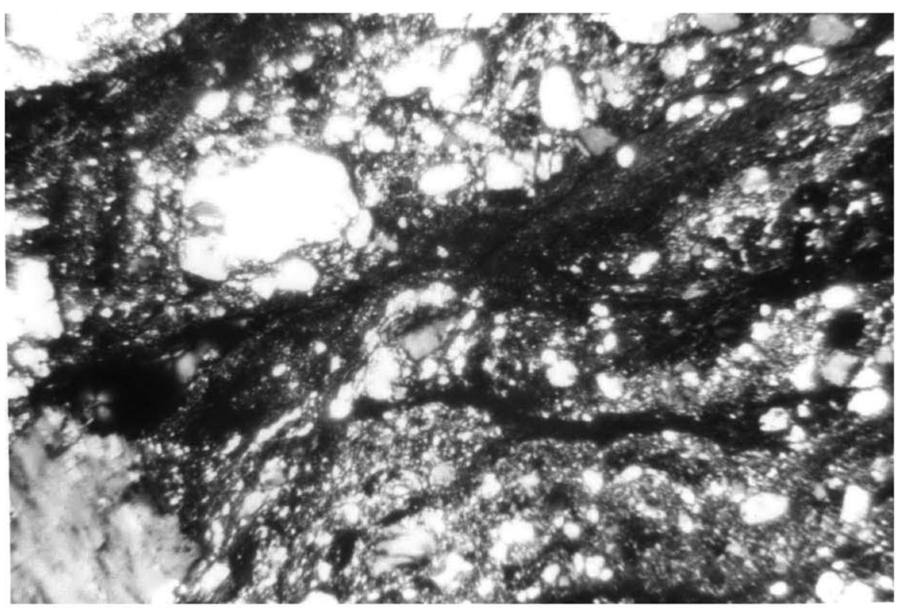


Figure 35B.



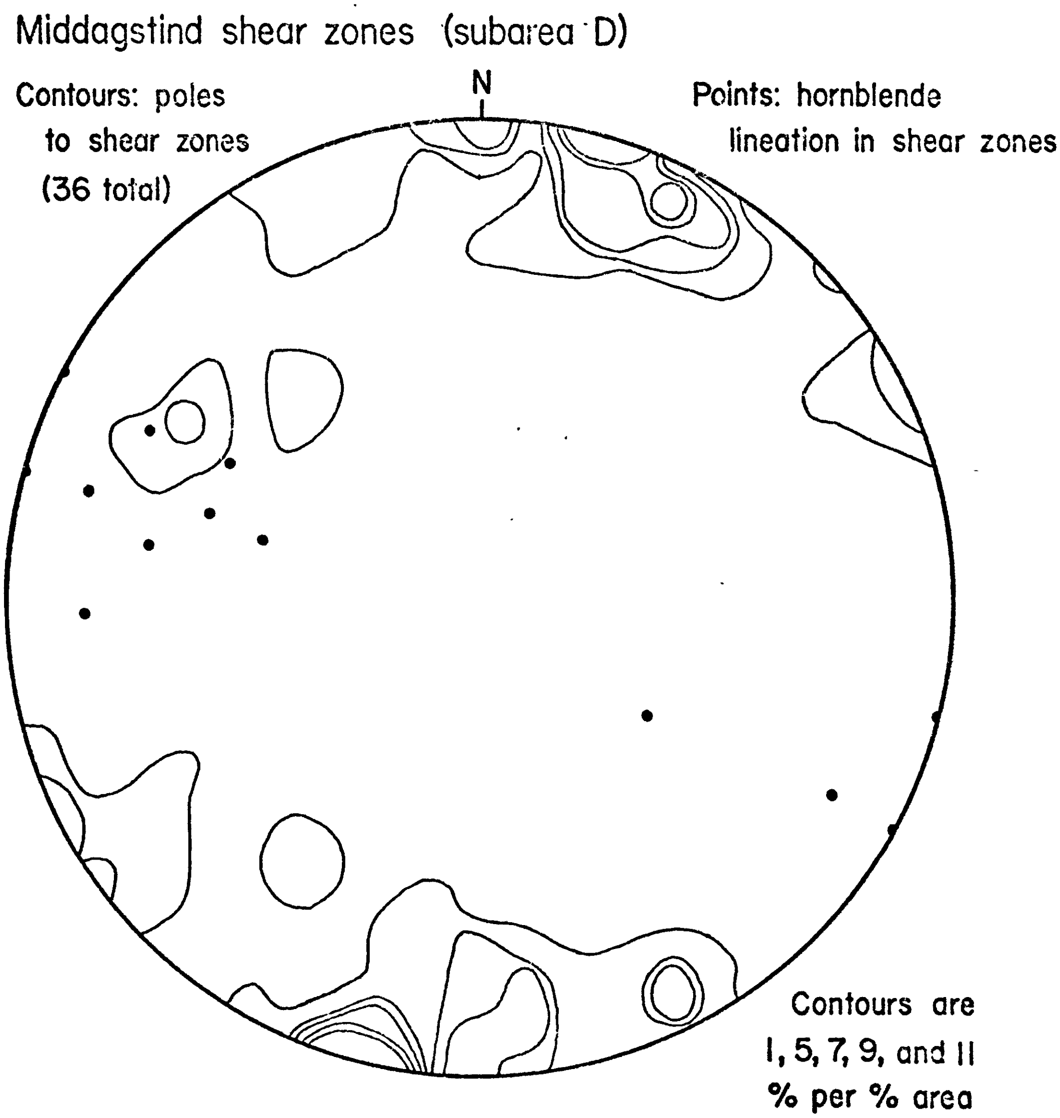
In Figure 36, poles to shear zones are contoured, and measurements of the associated hornblende lineation are plotted. The dominant shear set is subvertical, striking N70W, and the hornblende lineation generally has a low plunge within this plane. These data suggest strike-slip motion along the shear zones. Unfortunately, no adequate markers were recognized to allow net slip to be determined.

The similarity of the apparent kinematic direction of these shear zones to that inferred for D_2 folds, boudinage and thrust faults together with the evidence that the zones formed at amphibolite facies conditions, leads to the interpretation that these are D_2 structures. These shear zones may comprise an important link in understanding the large scale kinematics of D_2 deformation on east Hinnøy (see page 160 below).

Fabrics

Two main fabric elements developed during D_2: a schistosity, S_2, and a sporadic amphibole lineation, L_{2a}. Rarely, a stretching lineation is also preserved. The importance of S_2 schistosity has been difficult to determine. Small scale early folds generally show foliation passing through their hinges to form an intersection lineation parallel to the fold axis, and the generally isoclinal form of F_2 folds suggests transposition of S_1 has occurred. However, two lines of evidence indicate S_2 development may be more limited: (1) The Stangnes amphibolite, although clearly strongly folded by F_2, never shows overprinting of the mylonitic D_1 emplacement fabric by a second pervasive schistosity (a later S_3 or S_4 cleavage is rarely present). In the Harstad antiform (F_2), the amphibolite, though strongly folded, does not develop a new axial planar schistosity; S_1 can still be traced around the fold. (2) Some thin sections of Storvann Group schists suggest new mica crystallization did not always occur with F_2. Figure 65A shows a set of folds that fold the primary schistosity with only incipient mica growth parallel to the axial surfaces. The folds are overgrown by plagioclase porphyroblasts, which are in turn truncated by the S_3 spaced cleavage (see below). It is possible that these folds belong to a third pre-porphyroblast fold phase, otherwise unrecognized, but this seems unlikely. As a consequence, the folds are considered to belong to F_2, indicating that even in micaceous rocks S_2 may have limited importance. However, due to the isoclinal form of the

Figure 36.


Middagstind shear zones (subarea D)
Contours: poles to shear zones (36 total)
N
Points: hornblende lineation in shear zones
Contours are 1, 5, 7, 9, and 11 % per % area

F_2 folds, where the distinction of S_1 and S_2 is uncertain they can probably be assumed to be parallel.

The L_{2a} amphibole lineation is present occasionally in the Stangnes amphibolite, commonly in the Hesjevann assemblage amphibolite, and occasionally as aligned tremolite needles in marbles. It is assumed that this lineation has the same kinematic significance as a stretching lineation: that is, preferred growth of the long axis of amphibole crystals parallel to the maximum finite elongation. This is supported by approximate coincidence of the measurements of amphibole and stretching lineations (Figures 37 and 38).

Figure 37 plots amphibole lineations from all subareas. The attitudes cluster in an east-west direction, with mainly low plunges. Since S_2 had a subhorizontal orientation prior to F_3 (see discussion under F_2-Minor Folds), L_2 must also have been sub-horizontal after its formation. F_3 and F_4 folds are both of similar types, so that on stereographic projections, folded lineations describe great circle paths which pass through the slip lines of the folds. Since both F_3 and F_4 have steeply plunging slip lines (see below), reorientation by F_3 or F_4 folding of the amphibole lineations would produce steep plunges. It is thus considered that those amphibole lineations with steep plunges have been reoriented by late folds; low plunging lineations are only weakly reoriented and preserve their approximate original orientation.

The orientation of the L_2 hornblende lineations is not considered significantly different from previously described kinematic indicators of D_2 deformation, and is considered to reinforce the interpretation of the D_2 transport direction.

Rarely, a stretching lineation defined by elongation of quartz and feldspar grains is preserved in quartzites and granite gneisses. Measurements of this strerching lineation are plotted in Figure 38. It is similar in orientation to the amphibole lineation, but bears slightly north of east. This may be a reflection of the small data set, or may indicate the lineation records a combination of superposed strains. With regard to the latter possibility, it is noteworthy that the pebble elongation in the Harstad conglomerate is also ENE.

Development of the Tectonite Fabric in the Basement

Caledonian schistosities are developed in the basement rocks only in

Figure 37.

Summary of L_{2a} amphibole lineations, east Hinnøy

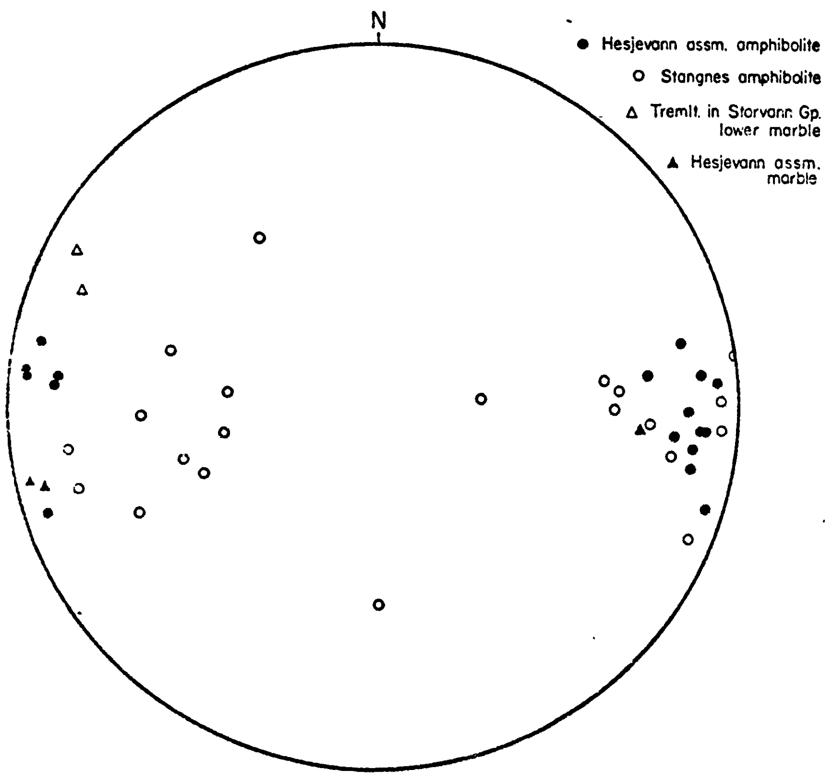
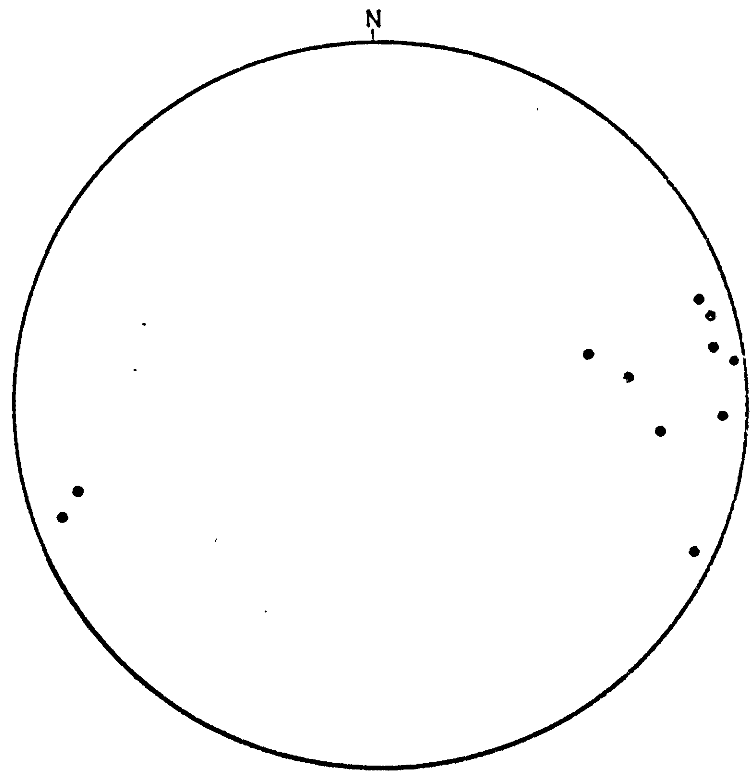


Figure 38.

Summary of stretching lineations, east Hinnøy



areas spatially close to contacts with metasedimentary rocks. A strong mylonitic fabric is present along the Austerfjord thrust where the Gullesfjord Gneiss overrides the Austerfjord Group, but dies out at structurally higher levels, and also downward into the Lødingen Granite below the thrust. The basement rocks near the stratigraphic contact with the Storvann Group are intensely foliated, but generally not mylonitic; this fabric similarly weakens and disappears away from the contact. The exposures of the basement/cover contact on the east shore of Storvann(S) and the roadcuts at Tjeldsund may form a transition between these two, the basement being very strongly foliated with local mylonitic bands. The transitional character of the fabric in the basement at Storvann(S) is consistent with the kinematic model described in the next section.

The progression from Precambrian granite unaffected by the Caledonian schistosity in the Hesjevann area (Figure 39), to similar rocks on Finnslettheia completely reworked by the Caledonian (Figure 42), seems to be controlled by the extent of retrograde metamorphism. The earliest signs of Caledonian tectonization are crude alignment of biotite and minor subgrain development in feldspars (Figure 40). As tectonization progresses, microcline begins to be replaced by muscovite, and plagioclase by clinozoisite. This suggests an addition of water to produce new hydrous (retrograde) mineral phases. Significant recrystallization of quartz and feldspar can lead to grain size reduction, yielding an augen gneiss composed of a fine-grained matrix of quartz, feldspar, micas, and clinozoisite surrounding porphyroblasts of relict igneous perthitic microcline and saussuritized, polysynthetically twinned plagioclase (Figure 41). In strongly reworked granite gneisses (Figure 42), the relict igneous feldspars disappear. Feldspars are completely re-equilibrated: microcline is non-perthitic, and plagioclase is rarely twinned, contains no clinozoisite inclusions, and is in textural equilibrium with adjacent phases. Muscovite and clinozoisite become essential phases of the rock (up to 10%), occurring as discrete, stable grains.

The extent of retrograde metamorphism described above may be controlled by the accessability of the water needed for the progress of the reactions. The spatial relationship between metasedimentary rocks and foliation in the basement granite suggests that the prograde metamorphism of the metasedi-

Figure 39: Photomicrograph of undeformed Melaa Granite.

Note random orientation of micas and and preservation of primary textures in feldspars (such as the Carlsbad twin in the large microrline grain at upper right). Crossed polars, 30X.

Figure 40: Photomicrograph showing incipient fabric development in Melaa Granite.

Note alignment of micas, incipient mylonitic bands and mortar texture in feldspars. Crossed polars, 30X.

Figure 39.

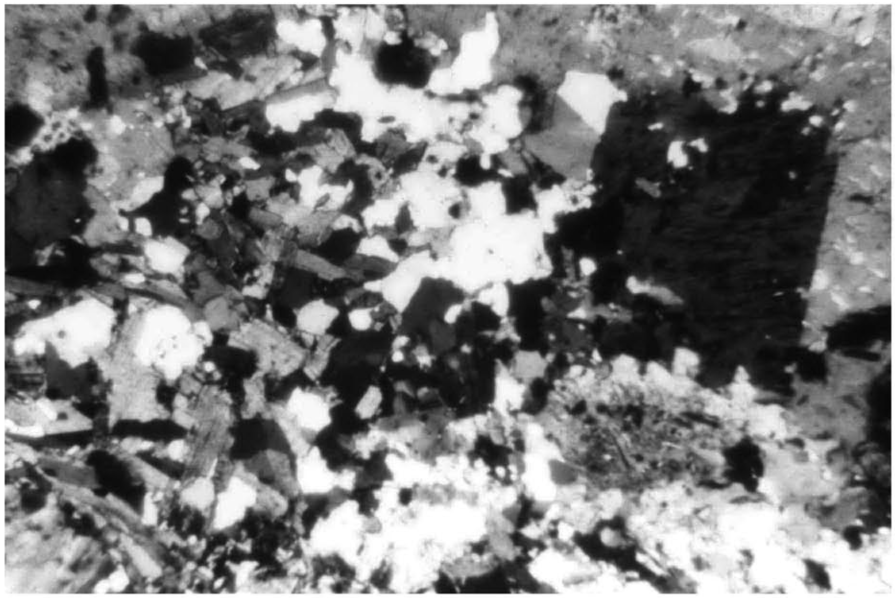


Figure 40.

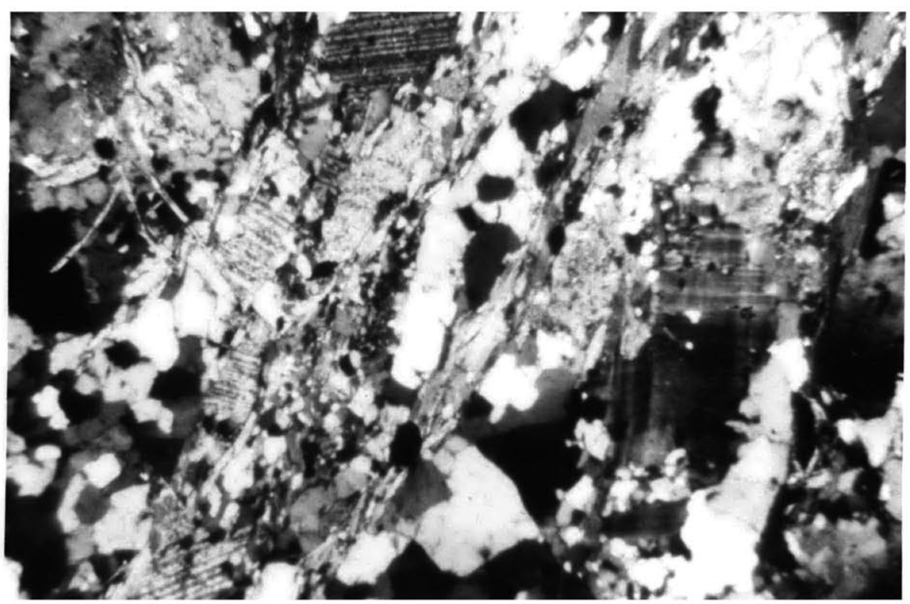


Figure 41: Photomicrograph of augen gneissic Melaa Granite.

Porphyroclasts of primary feldspars are set in a matrix of quartz, feldspar, biotite, clinozoisite, and muscovite. Crossed polars, 30X.

Figure 42: Photomicrograph of thoroughly recrystallized Melaa Granite.

Plagioclase is no longer saussuritized, microcline non-perthitic, and clinozoisite and muscovite occur as distinct grains in textural equilibrium with other phases.Crossed polars, 30X.

Figure 41.

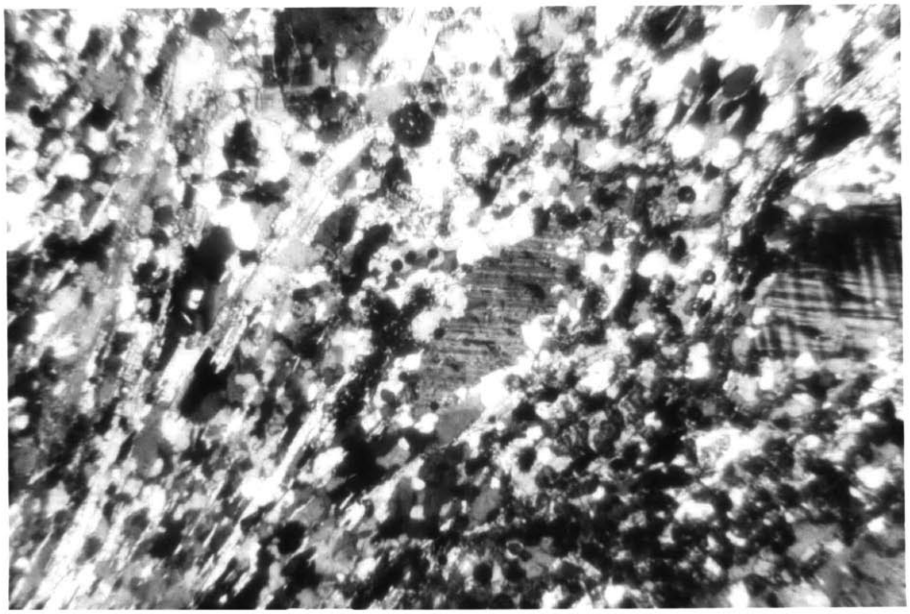
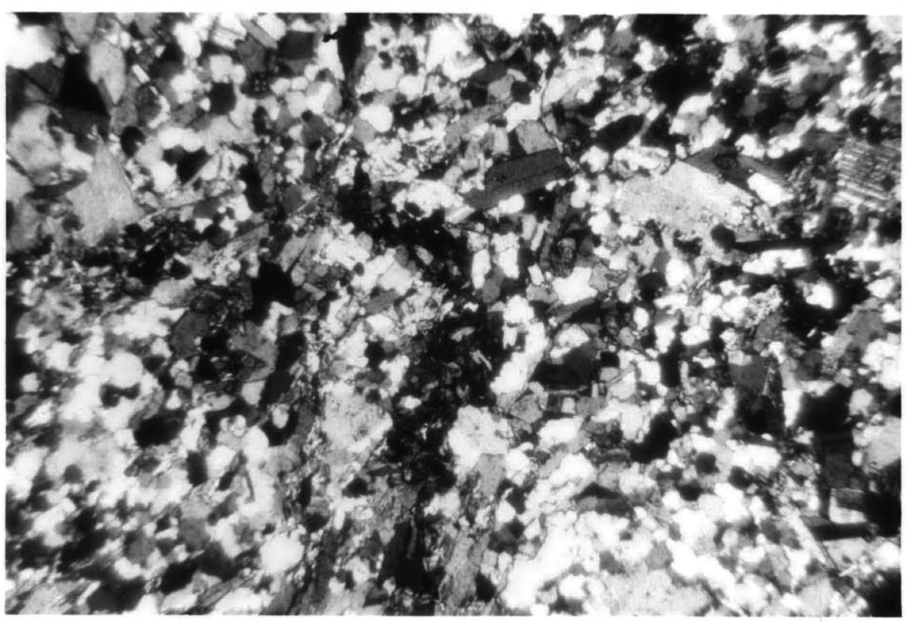


Figure 42.



mentary rocks may have been the primary source of this water.

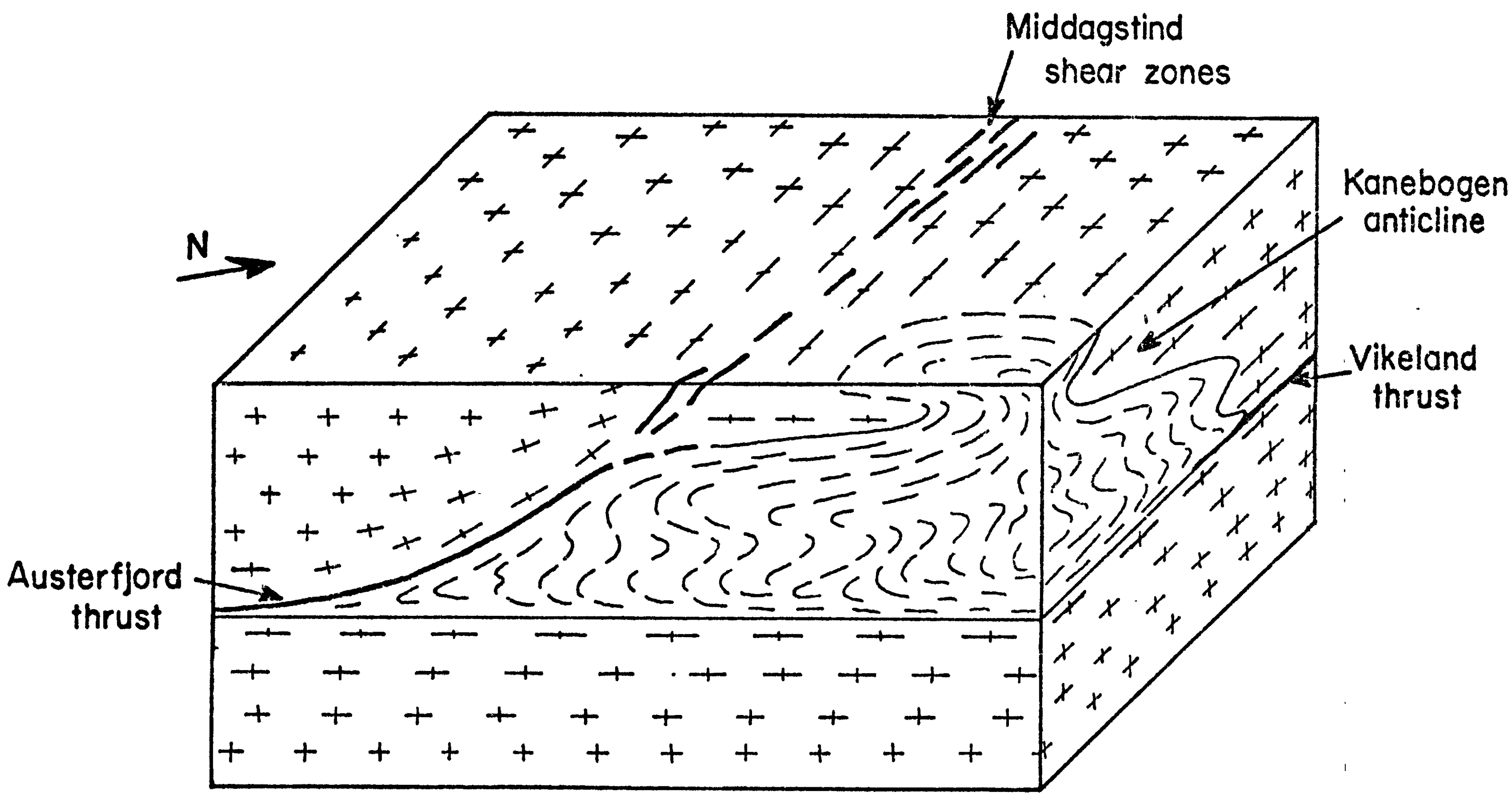
If this is correct, a further speculation is suggested. The Austerfjord thrust is well developed where the Austerfjord Group is present, but appears to die out southward where it must break through granite. It may be that the thrust nucleated on the Austerfjord Group pendants as a result of their prograde metamorphism resulting from tectonic burial by the D_1 nappe stack. This led to the limited thrusting of the basement to deform the overlying nappes and cover in a fashion similar to that described in the next section. The absence of a Caledonian foliation in the Middagstind syenite is not understood. Given the evidence for thorough Caledonian retrograde metamorphism in the syenite (Chapter 4), a Caledonian fabric would be expected on the basis of the arguments above. The accommodation of strain by discrete shear zones instead of penetrative flow in the area of the syenite could result from the different compositions of the rocks involved, or the position of these rocks within the overall kinematic scheme described in the next section. No satisfactory conclusion can be reached based on present knowledge.

Kinematic Model for D_2 Deformation

A kinematic model for D_2 deformation must incorporate a range of deformation styles including recumbent folding, steep ductile shearing, and low angle thrust faulting. It should account for the spatial relationships between these different deformation styles, and relate this to the kinematic indicators described in the preceding pages. I believe the model shown in Figure 43 satisfactorily accomplishes these objectives.

The essential feature of the model is the lateral transition from recumbent folding to thrusting over a relatively short distance. This is accomodated in two ways. First, the Austerfjord thrust is thought to laterally "ramp", and grade into a zone of dispersed strain accommodated by recumbent folding. This relationship is indicated by the change in thickness of the cover rocks underneath the upper plate of the Austerfjord thrust from the Kongsvik and Fjelldal areas in the south to the Storvann(S) area in the north. As noted in the Fjelldal area, the Austerfjord thrust may reactivate or cut out the Tjeldsund thrust. Northward toward Storvann(S), the strain is accommodated by a combination of distributed deformation in the infolded

Figure 43: Large scale geometrical model for D_2 on east Hinnøy. The Austerfjord thrust passes laterally into the F_2 recumbent fold sets of the northeast part of the study area. The transition is accomplished by a combination of lateral ramping, spreading of the zone of deformation, and ductile tear faulting along steep-dipping shear zones.



nappes and local mylonitic zones in the basement. Further north, recumbent folding has become the dominant mechanism of shortening, distributing the deformation through a considerable thickness of rock.

Second, ductile tear faulting along the vertical shear zones at Middagstind may also form a transition between the two structural styles. The amount and direction of movement in the thrusted basement is nearly uniform, but the movement field is very non-uniform in the adjacent domain of folding. The discrepancies are greater the further back from the leading edge of the thrust sheet one goes. The vertical shear zones may have developed in response to shear stresses in the vertical plane developed as a result of this geometrical incompatibility of the two deformation styles.

The cause of this lateral change in structural style is probably to be found in the Storvann Group. Southward along the basement/cover contact, the autochthonous cover has largely been stripped from the basement (directly northward, the mapping lacks adequate detail to evaluate this). In this part of Norway, only on east Hinnøy is there a large thickness of autochthonous cover preserved following D_1 nappe emplacement. The reason for its preservation can only be speculated on, but may be connected with pre-D_1 lower plate structures. It is worthy of note that the Stangnes Group pinches out not far from the place where the Storvann Group becomes very thin, supporting the possibility that a fundamental basement structure is present which controlled the geometry of nappe emplacement.

The mechanical effect of this apparently well-attached cover may thus have been to force the basement slab detached along the Austerfjord Thrust to ride up over its cover locally, causing it to become intimately involved with the cover in a more ductile deformation style.

D_3: Cross Folding

General Characteristics

The third phase of Caledonian deformation, D_3, is characterized by upright to overturned tight folds, with axes trending NW-SE to E-W. The fold geometry ranges from chevron to similar to, less commonly, concentric. The folds are dominantly overturned to the southwest. Fold amplitudes range from a few centimeters to more than a kilometer, but are dominated by minor folds of several tens of centimeters to several tens of meters amplitude, and major folds of 500-1000 meters amplitude. A secondary effect of this deformation within the basement south and west of Storvann(S) is the development of mylonitic ductile shear zones ranging from a few meters to a hundred meters or more wide.

The effects of this fold phase are most strikingly displayed on Plate I in the Finnslettheia/Sørvikfjell and Voldstadheia areas. However, analysis of structures indicates that little of the study area was entirely unaffected by this deformation.

Major Folds

There are four major F_3 folds recognized in the study area. The westernmost fold is a synform which runs from the southeastern corner of the map across Tjeldsund to the Melaa Vannene area. It probably continues as a very open fold beyond this point, but distinction of Caledonian and Precambrian folds and schistosities in the Middagstind area is uncertain. To the west of this fold is an upright antiform probably also of F_3 age, which exposes the Austerfjord thrust in the half-window which closes at Austerfjord. This fold was mapped by Hakkinen (1977) but is not shown on Plates I and III. To the northeast is an overturned antiform which is traceable across Voldstadheia to Storvann, and across the west shoulder of Finnslettheia. The connection of this axial trace to the open antiform at Straumsbotn is somewhat speculative, but not unreasonable.

In Gustavson's cross-section across the trace of the latter antiform (1972, p. 23, Fig. 20), he shows the area of Salangen Group marbles as synformal, making the basement terrain to the south antiformal. Fabric relations at the east shore of Storvann and along Tjeldsund demonstrate that the reverse is true; the S_3 cleavage consistently dips shallowly to moderately north while the S_1 schistosity dips steeply north at the contact. Thus, the belt of steep dip is the steep limb of a fold couple with the antiform to the north and the synform to the south (see Plate IIB, D-D'; Plate IID, I-I'). The interpretation of the basement region to the south of Storvann as synformal is reinforced by the fact that structures in the Salangen Group, which can be traced around this closure (in a general way) to Fjelldal, plunge northward under the basement at Fjelldal (see Plate IIC, G-G' and H-H', which are drawn parallel to the plunge).

Two more broad F_3 structures are present farther to the northeast. The rocks north of Voldstadheia to Finnslettheia and northward toward Torskvatsfjell are broadly synformal. The synform is not well defined, but can be recognized as a gross structural depression on Plate IIA, B-B' and C-C', and Plate IID, K-K'. The antiform to the north is much more clearly recognizable in outcrop as a re-entrant of Storvann Group rocks into the basement terrain directly south of Storvann(N). F_4 folding begins to be important north of here, and the identification of macroscopic F_3 folds becomes difficult. The northern F_3 antiform is refolded by F_4 folds so that, on the south side of Blåafjell the fold locally plunges as much as 70° NW. This refolding is responsible for the rather stubby appearance of the map pattern of what is a rather tight fold (see Plate IV, section K-K').

Mesoscopic Folds

Within the cover rocks and much of the adjacent strongly foliated basement, mesoscopic F_3 folds are very abundant. The style of F_3 folds changes markedly from one lithology to another. In schists and quartzites, the F_3 folds are kink-like or chevron folds, usually sharp-hinged and straight-limbed, with long subhorizontal limbs and short, steep limbs (Figure 44). Folds in the Stangnes amphibolite are geometrically similar to those in the schists, but lack the axial plane cleavage common in the micaceous rocks.

In marbles the F_3 folds take on a tighter and more sinusoidal form (Figure 45). Axial plane cleavage is less developed, and may be entirely absent.

The strongly foliated Precambrian granite near the contacts with the metasedimentary cover rocks folds much like the cover rocks, but the style tends more toward concentric geometry, grading toward chevron folds (Figure 46). Surfaces of slip along the foliation in general have not been observed, suggesting that a flexural slip mechanism (e.g., Ragan, 1973, Chapter 8) may be inappropriate for formation of these folds.

Because earlier folds already had a dome-and-basin geometry, F_3 folds interfere with these to produce both Ramsay Type 3 (Figure 47A) and Type 1 (Figure 47B) interference patterns. Type 3 patterns dominate both at outcrop scale and on map scale (Plate IIA, C-C'). Type 3 interference patterns

Figure 44: F_3 chevron folds in Storvann Group quartzite.

Note long subhorizontal limbs and short, steep to overturned limbs. View is westward, of outcrop on southern edge of summit area of Finnslettheia.

Figure 45: F_3 folds in marbles.

A: Sinusoidal similar folds in impure dolomitic marble of the Salangen Group. Outcrop is on northeast side of Voldstadheia.

B: Tight similar folds in upper calcite marble of the Storvann Group. Outcrop is 1 km southwest of Voldstadheia.

Figure 45A.

Figure 45B.

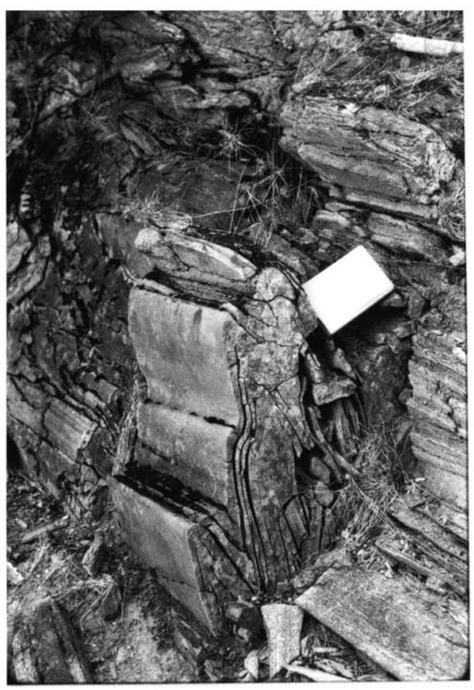


Figure 46: F_3 folds in pre-Caledonian granite gneiss.

Note near-concentric geometry, including cusp in core. Shoreline exposure is near Fjelldal, a few meters from contact with the Storvann Group.

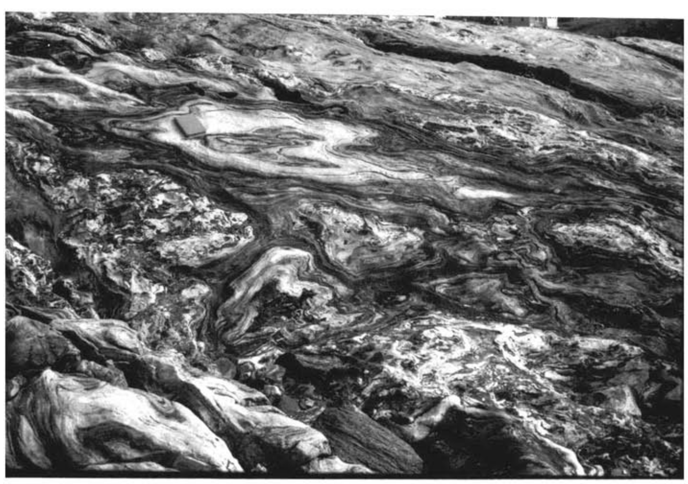
Figure 47: Fold interference patterns involving F_3.

A: Ramsay type 3 interference pattern in vitreous and micaceous quartzite of the Storvann Group, exposed in a roadcut along the west side of Tjeldsund. Early isoclinal hinge is probably F_2 in age.

B: Ramsay type 1 interference patterns in impure marble of the Salangen Group, east shore of Storvann(S). The very complex geometry of these domes and basins is probably due to interaction of F_1, F_2, and F_3.

Figure 47A.

Figure 47B.



result from folds with reasonably similar axial directions but very different movement directions (Ramsay, 1967), a condition consistent with the geometrical properties deduced for these fold phases independently.

Shear Zones in the Gullesfjord Gneiss

As the early Caledonian fabrics become less prominent in the basement, mesoscopic F_3 folds all but disappear. Some weak cleavage development remains, but confident distinction of different phases of Caledonian fabric development from Precambrian fabric becomes difficult. However, south of the Melaa Vannene, where the large F_3 synform becomes overturned, spatial adjustments in the core of this fold probably were accommodated by the formation of shear zones in the gneiss.

The shear zones are mainly composed of a brown schistose rock formed by grain size reduction of the granite and mobilization of quartz to produce veins which are often large and abundant. Thickness of the zones can range from 10 cm to several hundred meters, but most commonly they range from 50 to 200 m. The shear zones often weather to areas of flaggy rock piled like fallen dominoes (Figure 48). The zones may end abruptly, with dilatant voids accommodated by large knots of vein quartz (up to several meters across). No stretching lineations have been recognized in these zones, but intersection lineations with pre-existing fabrics are common, and are subparallel to the orientation of F_3 fold axes.

Though the zones are schistose, they do not appear mylonitic in the field. In thin section, the texture is one of cracked and mechanically twinned feldspar porphyroclasts in a matrix of fine-grained quartz, white mica, epidote, and biotite (Figure 49). Micas are kinked, and undulatory extinction is common in quartz. The rocks appear to have deformed under conditions where relatively little recrystallization accommodated strain.

The orientations of the shear zones are plotted in Figure 50. Their orientation is equally compatible with a relationship to D_2 or D_3 in this portion of the study area (cf. Plate IV, subareas L and M). However, the clearly post-metamorphic nature of the zones and the similarity of orientation of the intersection lineations to F_3 fold axes argue for a D_3 age.

Penetrative Fabrics

Two main fabric elements developed during F_3 folding: a weak to strong spaced cleavage, S_3, and a generally strong intersection lineation between

Figure 48: Broad shear zone in Gullesfjord Gneiss, Tverrfjell.

Figure 49: Photomicrograph of microcracked feldspar porphyroclast from a D_3 shear zone in the Gullesfjord Gneiss. Crossed polars, 75X.

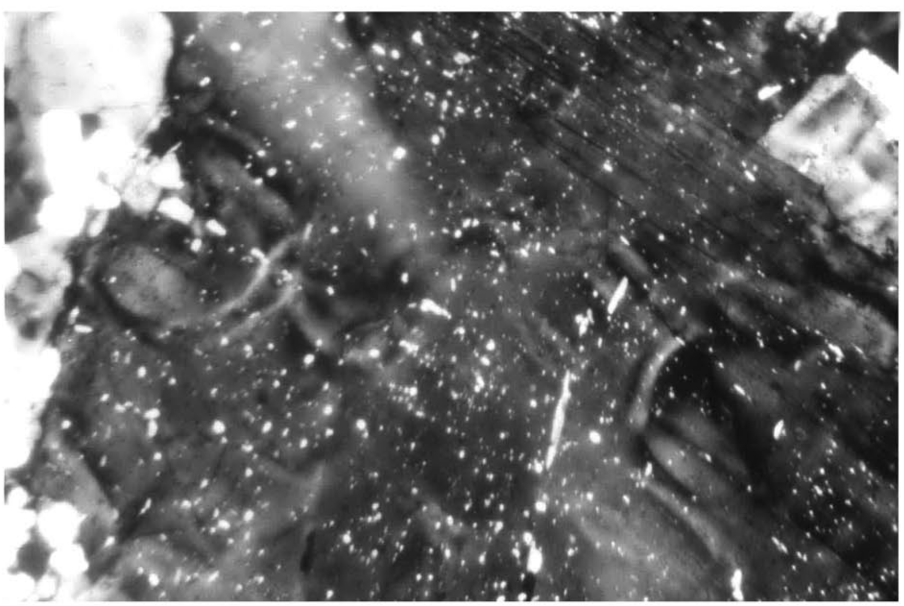
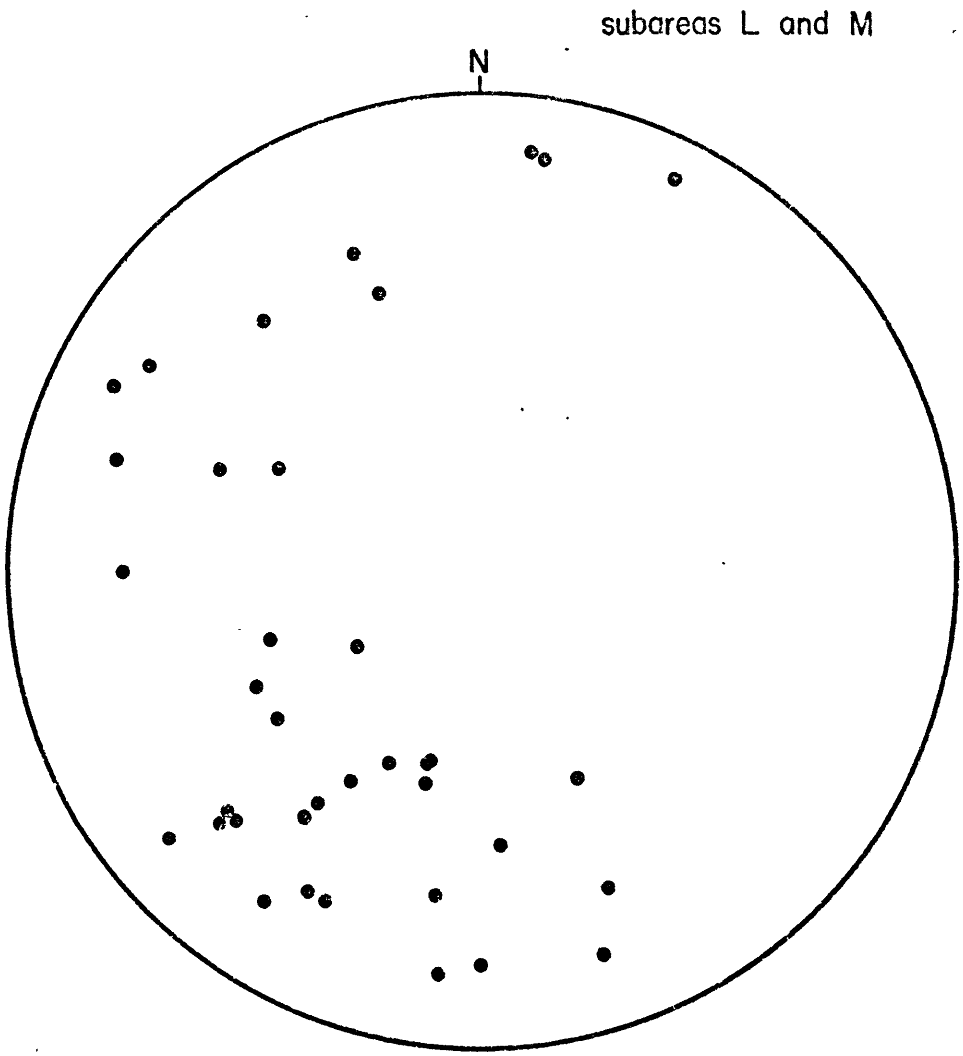


Figure 50: Equal area plot of shear zones in Gullesfjord Gneiss.



S_1 or S_2, and S_3 (often present even when S_3 is poorly developed), $L_{1/3}$ or $L_{2/3}$. The S_3 cleavage is best developed in micaceous schists, where it can become the dominant planar fabric in outcrop. This is especially true for the very micaceous Kilbotn Schist, where it has not been possible to determine anything conclusively about the geometry of D_1 or D_2 structures due to the strong textural overprint of late fold phases.

Marbles seldom develop a good S_3 fabric, presumably due to their low mica content. The mica present is generally rotated into parallelism with S_3, and some minor flattening of carbonate grains may develop, but usually it is difficult to measure the S_3 fabric in marbles. This has fortunate consequences for mapping the alternating schists and marbles of the Storvann Group, because one can always find S_1 in the marbles and S_3 in the schists. It is less fortunate in the Salangen Group marbles, where schists are rare. The relative paucity of S_3 measurements in subarea J (see Plate IV) is in part a result of this difficulty.

S_3 is seldom well-developed in amphibolites. This is presumably due to the absence of micas and quartz which are the main phases involved in formation of the S_3 fabric. By contrast, some of the granitoid rocks develop S_3 as a prominent secondary biotite foliation crossing the main schistosity. This is especially true in the shoreline exposures near the basement/cover contact at Tjeldsund. Here a strong S_3 cleavage in the basement granite crosscuts the steeply-dipping S_2 fabric, which is axial planar to isoclinal infolds of the Storvann Group basal quartzite and the basement.

The most prominent lineation throughout most of east Hinnøy is produced by the intersection of S_3 with an earlier schistosity. It is defined by the edges of micas in one foliation exposed on the other foliation surface. Since in this orientation the micas appear elongate, careful examination is required to distinguish this lineation from a stretching fabric. However, close parallelism with mesoscopic F_3 fold axes reinforces the interpretation of this lineation as an intersection.

The S_3 cleavage formed by a combination of solution and recrystallization processes. In thin section, cleavage traces are discrete surfaces, spaced a few mm apart, along which earlier grains are bent and truncated (e.g., Figures 64, 65A and B, 66A and B). New growth of chlorite, biotite, and muscovite has occurred along these surfaces, but grain truncations such as that of the plagioclase porphyroblast in Figure 17A suggest solution removal also may have occurred.

Structural Analysis

The measurements of S_1 and S_2, S_3, $L_{1/3}$ and $L_{2/3}$, and F_3 fold axes from subareas E, F, H, I, J, K, L, and P plotted on Plate IV primarily reflect F_3 deformation. Areas A, B, C, M, and Q also show F_3 effects, but have been overprinted by F_4 and/or F_5 and will be discussed later.

The following general characteristics about F_3 are drawn from these fabric data: (1) F_3 folds are nearly homoaxial, bearing mainly WNW-ESE, and have low plunges, mainly less than 20°; (2) S_3 axial planes dip moderately to steeply northward, consistent with southward vergence of the folds; (3) S_1 and S_2 poles form fairly complete great circle girdles, reflecting tight folding (with exceptions discussed below); and (4) the S_1/S_2 girdles generally contain point maxima reflecting moderate north -dipping upright fold limbs and steep south- or north-dipping fold limbs. The point maxima are particularly well-developed in the data from subareas J and P.

The consistent axial orientation and low plunges of the folds have two major implications. As discussed previously, the S_1/S_2 schistosities prior to F_3 must have been nearly planar and horizontal. Also, the homoaxial character of the folds indicates an absence of either heterogeneous rotational strain in the fold axial plane or heterogeneous slip between structural levels to produce a variety of axial orientations. The Hansen (1971) separation angle method of slipline determination clearly can not be used effectively here.

The bimodality of S_3 poles in some areas (e.g., subarea F) is not well understood. It is not due to poor choice of subarea boundaries, because both orientations are often observed on a local scale. It may result from slight F_4 folding or from fanning of the S_3 cleavage around fold hinges. Neither of these possibilities would predict specifically a bimodal distribution, however.

Individual variations of S_1/S_2 patterns result from a variety of local causes. In subareas I, K, and P, the primary problem is the paucity of data, leading to very "noisy" looking diagrams with similar overall characteristics to the general pattern. The north-plunging point maximum in subarea F results from that area being dominated by the steep south limb of a large F_3 antiform. The north limb of this fold was intensely redeformed by F_4, and so was treated separately (subarea C). The point maximum from subarea L has a similar origin: only data from the western limb of the large F_3 synform which dominates the southwestern portion of the study area are included.

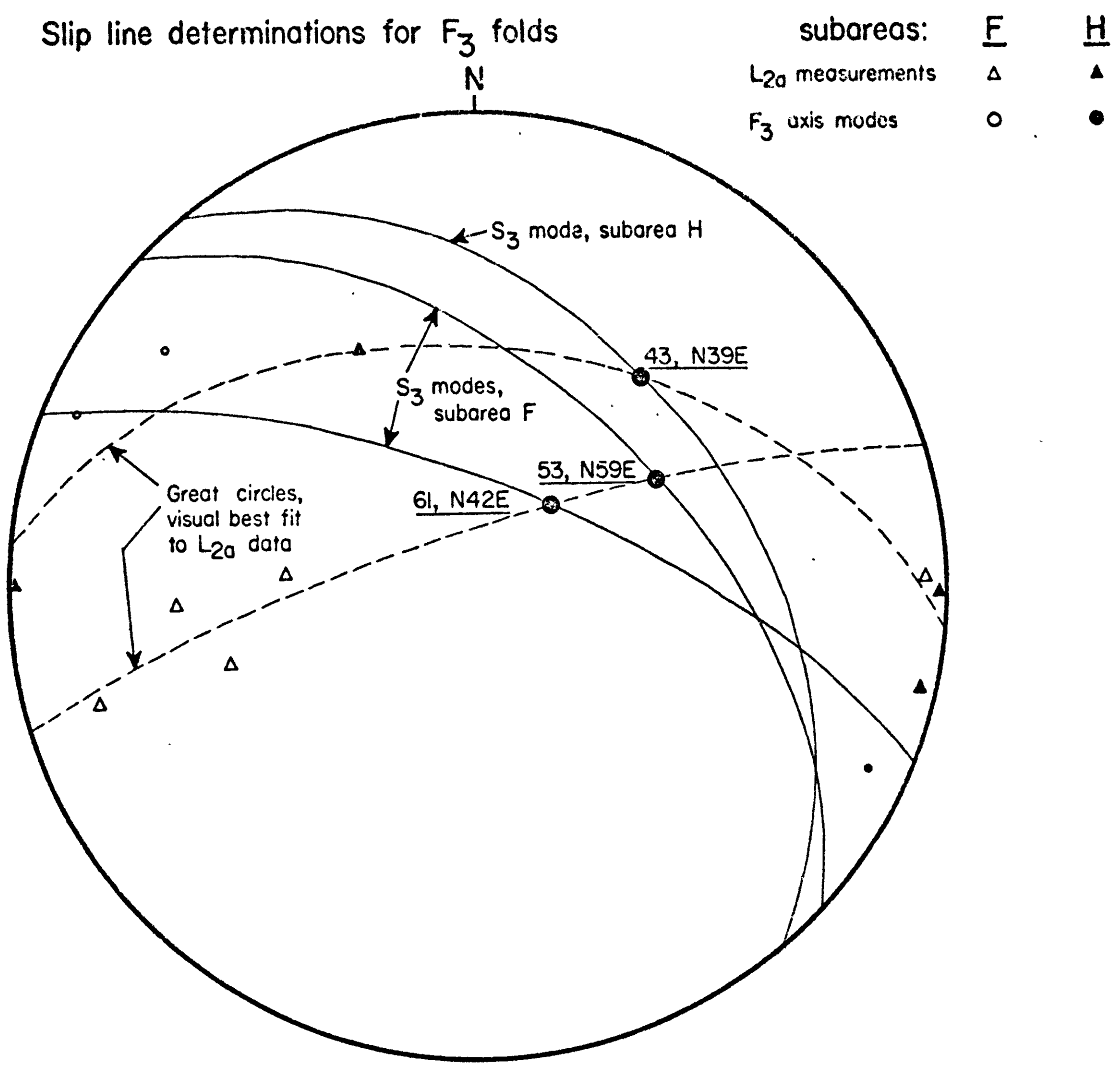
The best available method for slipline determination for F_3 is that of Weiss (1959). In similar folds, pre-existing lineations trace out great circle paths on a stereographic plot. The intersection of such great circles with the fold axial surface is the slip direction of the folding, roughly equivalent with the maximum finite elongation. Measurements of L_2 reoriented by F_3 are sparse, only defining approximate great circle orientations in subareas F and H. The precision of the determinations is further reduced by the bimodality of S_3 in subarea F.

The approximate F_3 slip line determinations are shown in Figure 51. Although the scatter is considerable, it is safe to conclude that the slip direction was steep and mainly southwest directed. The variability may in fact reflect variations in the strain field, folding being by definition a heterogeneous strain phenomenon in the first place. The orientation of this slip direction is interesting for two reasons. First, it is clearly different from the separation angle determined for F_2 axes, reinforcing the interpretation that the F_2 separation angle is not a result of later superposed deformation. Second, the direction of slip bears subparallel to the trend of the orogenic belt, implying these folds are an expression of tectonic transport parallel to the orogen. A possible setting for this will be considered in Chapter 6.

F_3 Fold Styles and Mechanical Behavior of Lithologies

The variability of F_3 fold styles described above presumably reflects different mechanical response of the lithologies at the ambient P/T conditions during D_3. Johnson and Honea (1975) concluded on the basis of experimental and theoretical studies that linear mechanical responses (e.g., Newtonian behavior) yield continuous, sinusoidal fold forms. To generate inherently discontinuous fold forms such as concentric and chevron folds, a nonlinear response is required, with plastic yielding occurring where shear stress is concentrated at fold hinges. Concentric and chevron folds

Figure 51: (Lower hemisphere equal area projection)



will develop in layered media with elastic/plastic rheology, and either negligible strength between layers (flexural slip mechanism), or infinite contact strength. Whether concentric or chevron folds are developed appears to be largely a function of the strain, with chevron folds resulting from greater shortening.

In response to layer-parallel stress, media with finite contact strength produce kink bands rather than regular fold forms, as a result of local layer-parallel yielding along the contacts (Honea and Johnson, 1976). An important aspect of kink folds is that unidirectional kinks, produced by superimposed layer-parallel simple shear, have senses of vergence (i.e., rotation of steep limb) opposite that of the superposed simple shear (Reches and Johnson, 1976). Hence, southward-overturned kink folds would result from a northward vergent shear couple. This would mean a slip line determined from such folds would not reflect the characteristics of the overall strain field.

The sinusoidal form of the F_3 folds in the marbles implies that the marble had no significant yield stress under the ambient conditions (probably greenschist facies; see Chapter 5). Thus, one can infer that under these conditions and at the appropriate (unknown) strain rate, carbonate rocks respond as viscous or pseudoviscous (power law?) materials. The concentric folds in the Precambrian granite gneisses suggest that at the same P/T and strain rate conditions, the granite is a plastic material. The homogeneity of the material and lack of slip surfaces in outcrop suggests that these concentric folds did not result from a flexural slip mechanism, but rather as a result of no yielding at contacts.

The sharp-hinged folds typical of F_3 in quartzite, schist, and amphibolite could be either interpreted as chevron folds or kink folds. Ramberg and Johnson (1976) have noted that these fold styles are difficult to distinguish in the field but should have opposite senses of overturning in the same strain field. The fact that these sharp-hinged F_3 folds have the same vergence as other F_3 folds indicates that these are true chevron folds and not kinks. This has two consequences: (1) the apparent elongation direction determined in Figure 51 refers to the overall kinematics of D_3 deformation, and (2) the mechanical response of these rocks to D_3 was plastic similar to the basement gneisses, but due to perhaps lower yield strength, the cover rocks shortened further to form chevron rather than concentric folds.

The development of the axial planar cleavage in F_3 (and F_4) folds may be intimately related to the phenomenon of yielding at the fold hinges. As noted in Chapter 2 with regard to the fabric in the calc-silicate facies of the Storvann Group lower calcite marble, it is common for S_3 to be stronger at the hinges of the F_3 folds than in the limbs. Presumably, cleavage formation begins as a result of layer-parallel stress of the layer. It may have a positive feedback aspect. Since the cleavage-forming reactions involve hydration (Chapter 4), cleavage formation will be enhanced by better accessability of water where it has already developed. The cleavage itself probably increases the permeability of the rock. If the cleavage reflects a significant portion of the strain in the rock, this positive feedback could be the reason for the plastic yielding in the fold hinges inferred from the fold geometry, and the reason for increased cleavage intensity at the fold hinges.

D_4: Upright Folds

General Characteristics

The fourth Caledonian deformation phase, D_4 , is much like the third, comprising upright to overturned folds, but with a NE-SW axial orientation. It was probably associated closely in time with F_3, but post-dates F_3 on two lines of evidence: (1) locally, mesoscopic folds of both fold phases are recognized in a single outcrop: the NW-SE folds (F_3) are folded around the NE-SW ones (F_4); (2) $L_{1/3}$ and poles to S_3 describe great circle paths on stereographic plots from subareas where both fold phases are present (see below), suggesting that D_3 fabrics are folded by F_4. The relative chronology suggested here is presented with reasonable confidence.

F_4 is more locally developed than F_3, appearing primarily in the northeastern part of the study area and to a lesser extent in the southeastern corner. The fold styles developed are much the same as those seen in F_3. Vergence is variable, as are axial surfaces, leading to generally steep but widely scattered sliplines. On the south side of the Stangnes peninsula, southeast vergent F_4 folds are often sheared out along low-dipping mesoscopic thrust surfaces, while the F_4 folds west of Middagsfjell (south end of J-J', Plate 11D) are north vergent.

The intersection of F_3 and F_4 folds to form domes and basins is especially noticeable in the area south of Middagsfjell (subarea C in Figure 28). The interference patterns are well-marked on Plate I by the contacts of the quartz-garnet schist, upper calcite marble, and pelitic schist units of the Storvann Group. F_3 folding was not very intense, since changes in plunge of F_4 fold axes are modest (20-30°). Very open northeast-trending folds also appear to produce open F_3/F_4 domes and basins in the Fjelldal and Ramboheia areas, such as the small basin which preserves the tiny klippe of Gullesfjord Gneiss on the foreshore island 2 km west Fjelldal. Again, this is an area where F_3 was relatively mild, mainly producing only a broad open synform.

It appears likely that the intensity of F_4 folding is inversely proportional to the extent of F_3 "corrugation" of earlier S-surfaces. Strong F_3 folding yielded a configuration too stiff for D_4 movements to deform.

F_4 folds interfere with earlier structures as well to produce some interesting patterns. Figure 52 shows a Ramsay type 3 interference pattern between F_2 and F_4 folds. Figure 53 shows the effects of F_4 folding on D_2(?) boudins on the south side of Stangnes peninsula. The boudins have been shoved back together, locally imbricated, and buckled. The maximum shortening direction of D_4 was thus nearly coaxial with the earlier elongation. This would predict a steep slipline for F_4 folds; it also suggests that the D_4 shortening component may be important in the composite finite strain history recorded in the pebbles of the Harstad Conglomerate.

Structural Analysis

Important F_4 effects appear in the stereographic plots of subareas A, B, C, and M. F_4 and F_3 structures were not plotted separately because the separation of the two by their field characteristics (other than orientation) is difficult. However, the bimodality of axial orientations present in all these areas, especially subareas A and B, suggests two fold sets, and the patterns displayed by subarea C strongly support this.

Like F_3, the F_4 folding is nearly homoaxial, with axes that trend NE-SW to ENE-WSW, and plunge at generally low angles. S_1 and S_2 plots show in general two partial girdles, of varying strength depending on the relative strength of the fold phases. Subarea C, dominated by F_4, has little indication of the NE-SW girdle characteristic of F_3 folds, subarea A

Figure 52: Interference of F_2 and F_4 folds.

Ramsay type 3 interference pattern between F_2 and F_4 folds in the Stangnes Amphibolite, 2 km west of Kanebogen.

Figure 53: D_4 imbrication and buckling of D_2(?) boudins.

Boudins are of amphibolite in calcite marble within the sliver zone under the Stangnes thrust, exposed on the shoreline east of Kanebogen.

Figure 52.

Figure 53.

shows subequal (weak) development of both, while subarea M shows mainly F_3 folds in the distribution of S_2 poles.

Distribution of S_3 and S_4 poles is generally complex in these areas, and difficult to interpret. However, those in area C are very suggestive. Several point maxima line up along a great circle girdle at a large angle to the F_4 axis. This is suggested to reflect folding of the S_3 axial plane cleavage around F_4 folds. The poles to a folded foliation behave like a folded lineation on a stereographic plot, so that a Weiss (1959) slipline analysis can be used. The broader point maximum falling slightly west of this girdle is the S_4 cleavage from the area west of Middagsfjell where F_4 folds are systematically north vergent. Thus, intersecton of the girdle of S_3 poles and the S_4 axial plane yields a determination of the slip direction for these F_4 folds (Figure 54).

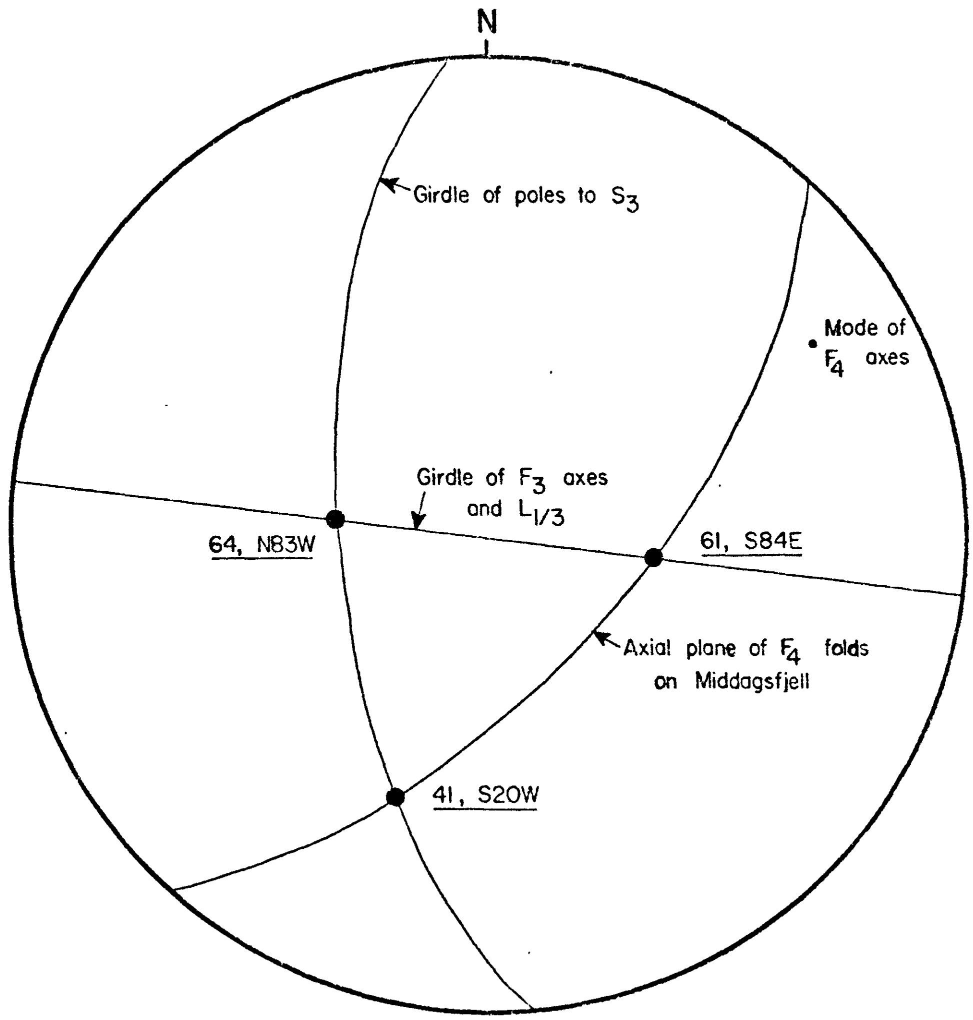
Closer examination of the plot of F_3 and F_4 axes and associated intersection lineations from subarea C suggests an additional, though less accurate, way to determine F_4 sliplines. A shallow-plunging secondary point maximum, oriented in a WNW-ESE direction, reflects F_3 folds. Steeper plunging point maxima define a crude vertical great circle girdle (the girdle could not be confidently inferred in the absence of independent evidence of refolding of F_3 by F_4). This "girdle" is thought to reflect folding of F_3 axes by F_4. The intersections of this folded lineation with the S_3 girdle and the S_4 axial plane geometrically should reflect slip directions of F_4 folds. Agreement of the three intersections is poor, in part due to the limited data, and in part due to the variations in fold vergence noted above (also recognizable on Plate IID, L-L'). More detailed analysis of a larger data set are required to improve understanding of the kinematics of this fold phase.

F_5: Minor Folds and Crenulation

Local minor folds and crenulations form the last gasp of Caledonian folding, recognized in three areas; near Harstad (subarea A), in the Kilbotn Schist, and at the southeast corner of the study area (subarea Q). In the Kilbotn Schist, a NW-trending, steep crenulation cleavage locally cross-cuts the S_3 spaced cleavage, producing extensive retrograde chlorite and disrupting the earlier fabric on a microscopic scale. In subareas A and Q, a NNW-trending set of upright folds on the scale of a few centimeters

Figure 54.

Slip line determinations for F_4 folds, subarea C



to a few meters is present, tentatively correlated with the latest crenulation-forming event at Kilbotn. The vergence of the NNW-trending folds near Harstad is eastward; those on Kvitfjell in the southeast corner of the map area verge west. No major structures of this generation are recognized, and it is considered to have minor significance in the Caledonian structural evolution.

Cenozoic(?) High-Angle Faults

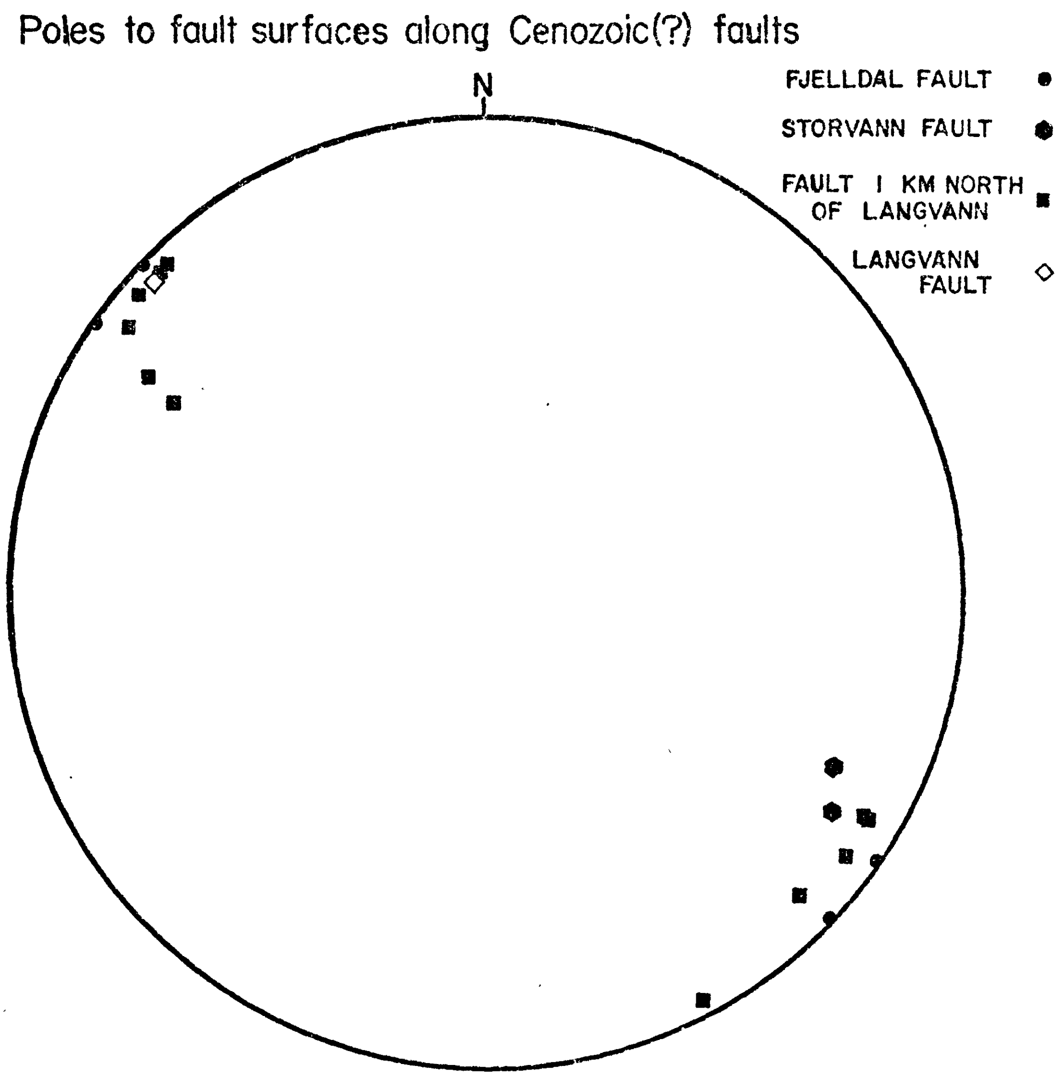
General Characteristics

A set of subvertical faults, trending N30E to N50E, have cut the rocks in the eastern part of the study area into a series of NE-trending blocks. The number and continuity of faults is significantly greater than previously recognized (cf. Gustavson, 1972, 1974a, b, c). Many of the high-angle fault juxtapositions have been previously explained by folds or thrusts. For example, the Straumsbotn Nappe of Gustavson (1972) was largely based on interpretation of the Langvann Fault as a thrust. Although the actual faults are only rarely exposed, contacts, folds, and fabrics are abruptly truncated along the traces inferred on the map. The major faults (Langvann, Storvann, Astafjorden and Fjelldal Faults) are topographically expressed as linear valleys or fault-line scarps.

Faults or secondary shear surfaces closely associated with them are exposed in four localities: (1) the long but minor fault north of Langvann is exposed 1 km NNW of Langvann; (2) subsidiary faults along the Langvann fault are exposed 0.5 km west of Langvann, where the fault places Storvann Group quartzite against Precambrian granite; (3) subsidiary shears in Salangen marbles along the Storvann Fault are exposed at the northeast corner of Storvann(S); and (4) several areas of subsidiary shearing along the Fjelldal Fault are exposed west of Fjelldal. Slickensides or other mesoscopic movement indicators have not been observed. Figure 55 plots measurements of fault surfaces from these localitites, emphasizing the steep attitude and consistent strike of the faults, aside from the minor E-W splays off the Storvann fault on the south side of Sørvikfjell.

The only well-exposed fault is the minor fault north of Langvann. Where exposed it cuts massive Precambrian granite so that the magnitude of displacement on this fault is unknown, though it is probably small further east

Figure 55.


Poles to fault surfaces along Cenozoic(?) faults
N
FJELLDAL FAULT
STORVANN FAULT
FAULT I KM NORTH OF LANGVANN
LANGVANN FAULT

where it only slightly offsets the basement/cover contact (Plate 1). Rocks along the fault have been intensely fractured (fracture spacing about 2 to 10 cm), in a vertical zone several meters wide. Local open space fillings of drusy quartz are present. No gouge or fault breccia were recognized. The very brittle style suggests movement occurred at low temperature and pressure, completely post-dating Caledonian events.

Vertical subsidiary fault surfaces are exposed along the Langvann Fault within the granite north of the main fault contact. F_3 folds and all previous structures in the quartzite are clearly truncated along the contact. From the ridge west of Langvann, the eastward trace of the fault is clearly recognizable, trending perfectly straight regardless of topography. Clearly, the steep dip of the shear surfaces here (86°N) applies to the main fault as well.

At Storvann, the Salangen Group marbles immediately south of the concealed fault are broken by abundant minor subvertical shears trending N35E and about 1-2 cm apart. The fractures are in part healed by secondary calcite fillings. Southwest-dipping compositional banding is offset right-laterally; where F_3 minor fold axis can be matched across shear surfaces, they are offset by north side down dip-slip. Attitudes of penetrative fabrics in the sheared zone are not significantly different from those south of the zone, indicating that no significant bending or rotation was associated with faulting.

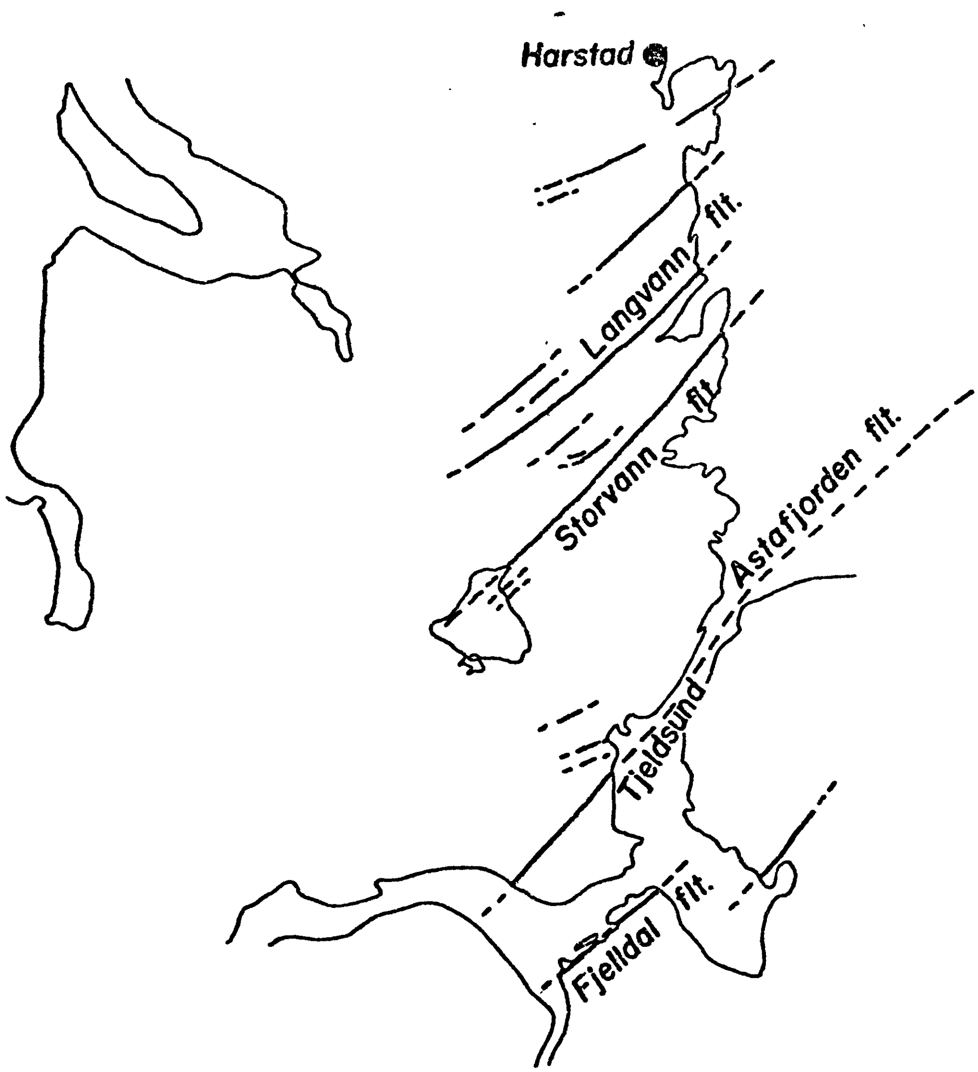
The exposures along the Fjelldal Fault are similar to the others described above: intense vertical fracturing with local open-space fillings of vein quartz. One minor fault exposed in the granite gneiss directly north of the main fault trace separates a vertical quartz vein one meter in a left-lateral sense. Hence, at least some minor strike-slip movement occurred in this area.

Regional Pattern and Sense of Movement

The major difference in the pattern of high-angle faulting shown in Figure 56 from the interpretations of previous workers is the arrangement of faults around Tjeldsund. Vogt (1942) identified the Astafjorden fault as a NE-trending dip-slip fault truncated to the southwest by the NNE-trending Tjeldsund fault. Both of these faults are entirely concealed by water. The Tjeldsund fault was mapped as curving at its south end and then terminating,

Figure 56: Map pattern of Cenozoic(?) high angle faults in the vicinity of the study area.

Note the uniformity of orientation in this interpretation (cf. Gustavson, 1974a).



since no evidence of the fault could be found on land along its projected trace. Vogt estimated right-lateral strike-slip of 3.5 km on the Tjeldsund fault while Gustavson (1972), adopting Vogt's overall interpretation, estimated 10 km of net slip.

There are several objections to these interpretations. Only planar features offset by the fault were considered by Vogt and Gustavson, thus determination of net slip requires assumption of the net slip direction. No independent data exist regarding the direction of slip. The surfaces being matched have been very complexly folded, so that small dip-slip separations can produce large surface separations of units. Finally, it is geometrically difficult to terminate a strike-slip fault with 10 km of movement over a distance of a few kilometers without a trace.

The present study indicates that all the faults trend NE, and consistently have NW side down dip-slip movement. The faults in general are not present in the western part of the map area. The reasons for these views are detailed below.

Three kilometers of right-lateral separation of the basement/cover contact occurs along the Langvann fault north Finnslettheia. Only 3 km along the strike, the vertically-dipping Stangnes Group rocks are scarcely separated across the fault. Consideration of the geometry of section C-C' (Plate 11A) suggests the basement/cover contact was probably only a few hundred meters above the topographic surface at Sørvikfjell opposite this contact north of the fault. Hence, northwest side down dip-slip best explains the relationships along this fault.

The major F_3 overturned antiform which crosses the Storvann fault is scarcely offset by it, although different rocks and structural units are juxtaposed. Tracing of the Storvann Group rocks around the antiform south fo the fault indicates they intersect the fault a kilometer or more above the present topographic surface (depending on fold geometry), indicating northwest side down movement.

On the north side of the Fjelldal Fault, Precambrian granite, its Storvann Group cover, and the structurally underlying Austerfjord Group structurally overlie the Salangen Group marbles. South of the fault these units are absent above the marbles. This relationship again indicates northwest side down dip-slip movement.

Northwest side down dip-slip movement along a simple extension of the

Astafjorden fault also best explains the relationships across Tjeldsund. As developed in the section of F_2 recumbent folding (see above), the extensive area of Salangen Group marbles south of the Storvann fault lies in the inverted limb of the Kanebogen recumbent anticline. It was also suggested that the quartzite and schist at Kvitnes on the east side of Tjeldsund are in the core of the underlying Harstad recumbent syncline. The Salangen Group marbles of the inverted limb would thus project above the rocks at Kvitnes. Hence, a slight southward bend of the Astafjorden fault to run through Tjeldsund, with NW side down dip-slip, would satisfy these relationships without requiring the existence, and abrupt disappearance, of a separate strike-slip fault in Tjeldsund.

The proposed southward continuation of the Astafjorden fault through the valley NW of Sandtorg is suggested for three reasons: (1) this is a prominent NE-trending valley discordant to surrounding bedrock structure; many other similar valleys to the north follow high-angle faults; (2) the F_3 fold axial traces do not readily match across the valley. Considering that the F_3 folding is complex and overturned on the north side and open and apparently simple on the south side, a simple one-for-one matching of the fold axes can not readily be made; and (3) continuation of the Astafjorden fault through this valley allows minimal deviation from the overall regional fault pattern.

While many of the faults have been shown to terminate westward, Tjeldøy to the south of the study area has not been mapped in comparable detail, so that it is uncertain whether the Astafjorden and Fjelldal faults extend beyond the western limits shown here.

Timing

Local control of the timing of movement of these faults is very poor. Their brittle style and total disregard for older structures indicate these faults post-date Caledonian uplift and cooling, dated at 360 Ma by Rb/Sr biotite/whole rock studies (Chapter 5). The only materials unaffected by the faults are Quaternary sediments. An early Cenozoic age is preferred on the basis of regional considerations, discussed in Chapter 6, but no dogmatic approach is justified at this time.

CHAPTER 4: CONDITIONS AND TIMING OF METAMORPHISM

Introduction

Metamorphic assemblages of both Precambrian (Svecofennian) and early Paleozoic age are present on east Hinnøy. The grades of both metamorphisms are mainly in the amphibolite facies, so that in the pre-Caledonian basement rocks, determining from which of the two periods of metamorphism any particular mineral assemblage derives is not always straightforward. The earlier metamorphism, roughly 1830 to 1700 Ma in age (Griffin and others, 1978), developed granulite facies mineral assemblages regionally in rocks to the south and west of the study area (Heier, 1960; Heier and Compston, 1969; Griffin and others, 1978; see Figure 3). Indications have been found in this study of local granulite facies conditions on east Hinnøy during this metamorphism, but amphibolite facies assemblages are typical.

Caledonian metamorphism on east Hinnøy developed lower to middle amphibolite facies mineral assemblages of the Barrovian or kyanite/sillimanite type (Miyashiro, 1961), characteristic of intermediate P/T conditions. Cover rocks are entirely recrystallized; no primary textures have been recognized other than the pebbles in the Harstad conglomerate. The basement rocks, not far out of equilibrium with the superposed Caledonian P/T conditions, are re-equilibrated to varying degrees. This probably depended on the availability of water from an external source; the grade of Precambrian metamorphism was probably somewhat higher, so that re-equilibration involved mainly retrograde hydration reactions.

Two types of retrogression of the Caledonian amphibolite facies mineral assemblages occurred. The late fold phases F_3 through F_5 developed semi-penetrative cleavages which involved partial retrogression of the amphibolite facies minerals. A later local chloritization and epidotization of both basement and cover rocks, of unknown age or origin, is patchily developed across the area. No fabrics are related to this late retrogression, nor has it been possible to spatially relate its development to any specific structures.

The mineral reactions included in the discussion below are undoubtedly far simpler than those which actually occurred in the rocks. However, these remarks are based solely on petrographic observations of a reconnaissance nature with respect to metamorphic petrology. Consequently, the simple reactions are written to show the general form of the actual reactions the author believes took place, and to allow some estimate of the metamorphic conditions to be made. A more precise analysis would be justified only with more extensive sampling and chemical analyses of the mineral phases involved.

Svecofennian Metamorphism

Petrographic evidence indicates that both amphibolite and granulite facies mineral assemblages were developed in the Svecofennian metamorphism on east Hinnøy. Evidence for granulite facies conditions in the vicinity of the Middagstind syenite is both direct and indirect. The innermost part of the mafic hornfels developed along the east and south sides of the pluton locally bears the mineral assemblage hypersthene + hornblende + labradorite + opaque, with only slight retrograde effects (Figure 10). This is a typical assemblage of the lower temperature part of the granulite facies where some hydrous phases (in this case hornblende) are still stable. In particular, this assemblage is on the high temperature side of the reaction:

hornblende + almandine + quartz + hypersthene + plagioclase + H_2O (1)

which is one of those suggested by De Waard (1965) to define the granulite facies. The coexistence of plagioclase and orthopyroxene also puts a rough maximum of 8 to 10 kb on the metamorphic pressure (Winkler, 1973); at higher pressures, these phases react to form garnet + clinopyroxene + quartz, the anhydrous equivalent of the left side of reaction (1).

Within the Middagstind pluton itself, no orthopyroxene has been observed. However, the textures in Figure 9 suggest the biotite + quartz meshes are pseudomorphic after an earlier phase. These pseudomorphs are in part rimmed by ferrohastingsite. These rims often occur as multiple isolated, optically continuous grains, suggesting they are relicts of an earlier assemblage.

The Middagstind syenite is part of a regionally extensive suite of calc-alkaline to alkaline, mainly intermediate plutons intruded during the latter part of the Svecofennian metamorphism (Heier, 1960; Green and Jorde, 1971; Griffin and others, 1975, 1978; Malm and Ormaasen, 1978). These plutons are nearly all orthopyroxene-bearing, having crystallized within a terrain undergoing regional granulite facies metamorphic conditions. Amphibole rims on pyroxenes are common in these rocks. The biotite + quartz meshes are thus thought to be pseydomorphic after primary orthopyroxene, with the original amphibole rims in part still intact.

The arguments for amphibolite facies metamorphic assemblages being preserved from the Svecofennian are more circuitous since the Caledonian metamorphism is demonstrably of similar grade. Other than in the inner part of the contact zone of the Middagstind Quartz Syenite, no orthopyroxene is observed, nor any textures suggestive of pseudomorphs thereof. The rocks in the vicinity of the syenite pluton can be demonstrated to have little or no Caledonian fabric on the basis of the truncation of fabrics by the 1726 Ma old pluton. The static retrogression of these wall rocks to form dispersed grains or granular aggregates of epidote is probably a Caledonian event. However, the primary amphibolite facies mineralogy of these rocks is unlikely to be Caledonian. For this to be the case, the rocks would have had to have undergone a thorough static crystallization mimetic after the Precambrian schistosity, followed by a partial static, non-mimetic recrystallization. For the rocks to have recrystallized this much in response to Caledonian events, yet never developed a Caledonian fabric, seems unlikely. Consequently, the primary amphibolite facies assemblages in these rocks are considered to be Precambrian.

The basement rocks of Hinnøy are generally not of appropriate compositions to develop diagnostic, low variance mineral assemblages. The most useful rocks are the hornblende diorite and lower grade parts of the contact hornfels which develop the assemblages:

hornblende + plagioclase (oligoclase-andesine) $\pm$ biotite $\pm$ quartz

actinolite + grunerite + plagioclase + biotite

These assemblages are characteristic of the amphibolite facies but give little information about metamorphic conditions beyond this.

The Svecofennian metamorphism of the basement on east Hinnøy was thus mainly in the amphibolite facies, but reached granulite facies conditions locally where intruded by the Middagstind Quartz Syenite. The presence of such "contact" metamorphism within a regional metamorphic terrain finds an analogy in the effect of the Bergell granitic complex in eastern Switzerland (e.g., Ernst, 1973). The upward transfer of heat by the rise of the Bergell body stretched out the isotherms on the east end of the Ticino thermal dome to produce contact-like metamorphic zones which are continuous with the widespread regional metamorphic zones of southern Switzerland. Similarly, the rise of the Middagstind syenite pluton above the regional granulite facies boundary probably stretched the isotherms upward into a local dome, producing contact-like granulite facies effects in the regional amphibolite facies rocks.

Caledonian Metamorphism

Amphibolite Facies ("Main Stage")

Cover Rocks

All of the rocks of the structural cover on east Hinnøy were thoroughly recrystallized by Caledonian events. A considerable range of compositions is observed within these rocks. However, there are relatively few occurrences of highly aluminous pelites or basic rocks, or of siliceous dolomites, which typically generate low variance mineral assemblages capable of tightly constraining metamorphic conditions.

The typical mineral assemblage of pelites is garnet + biotite + muscovite + quartz, which only indicates a minimum of upper greenschist facies conditions. However, associated basic rocks (see below) indicate the amphibolite facies was reached throughout the area, so that the absence of key minerals such as kyanite reflects the modest alumina content of the rocks rather than a lower grade of metamorphism (Figure 57). Three localities, (1) the power station roadcuts of Narvik Group rocks 1 km west of Kilbotn, (2) the Storvann Group pelitic schist exposures on the east shore of Storvann(S), and (3) shoreline exposures of Storvann Group quartz-garnet schist at Fjelldal (Figure 58), include mineral assemblages more restrictive of metamorphic conditions. The assemblages, by locality, are:

(1) kyanite + garnet + biotite + muscovite + quartz
(2) kyanite + sillimanite + garnet + biotite + muscovite + quartz
(3) kyanite + staurolite + garnet + biotite + muscovite + quartz

The first of these assemblages lies in a divariant field of P/T space bounded by the reactions (Figure 61):

$$\text{staurolite + muscovite + quartz = kyanite + garnet + biotite + } H_2O \quad (2)$$

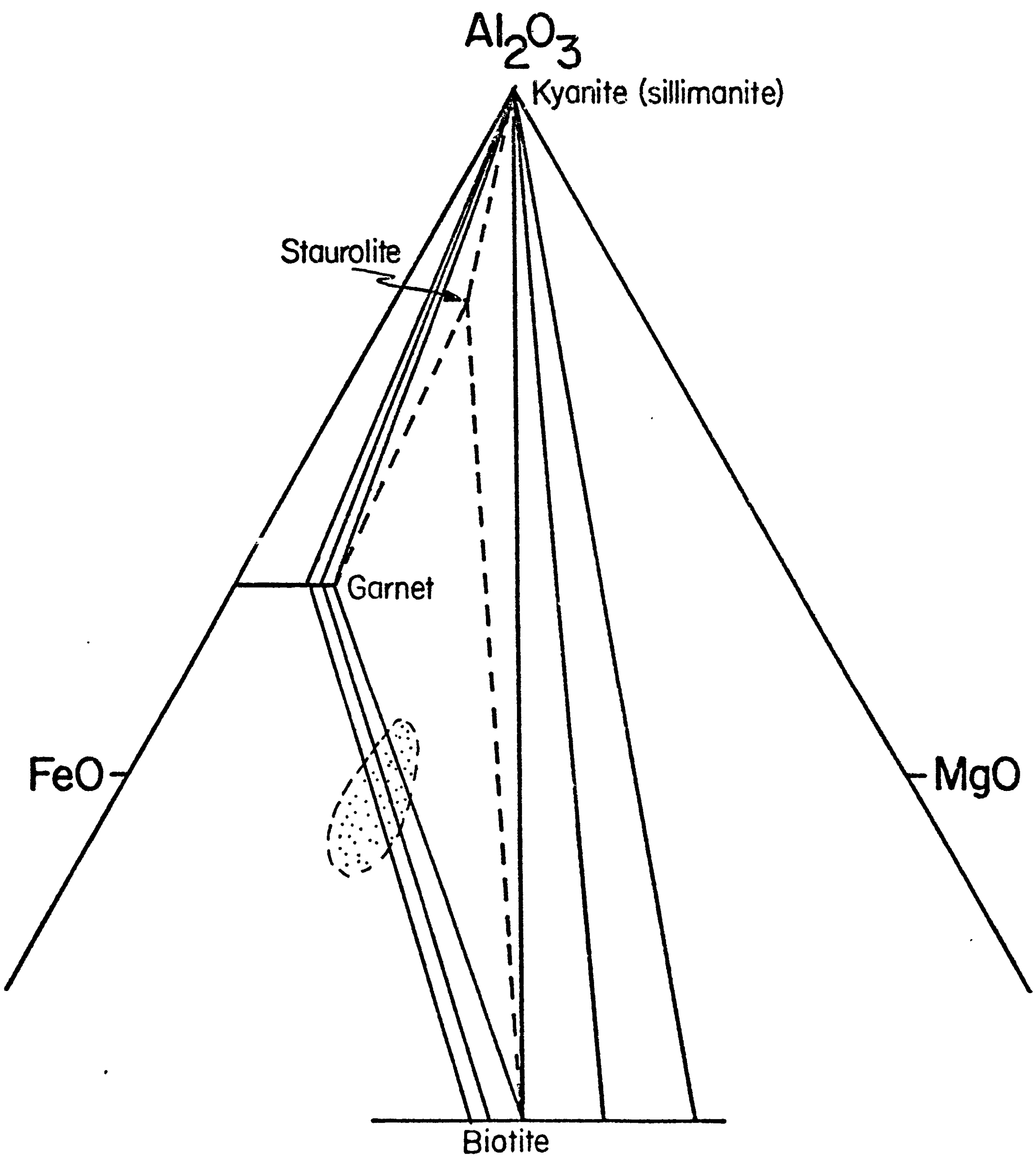
$$\text{kyanite = sillimanite} \quad (3)$$

$$\text{muscovite + quartz + } H_2O \text{ = kyanite + melt, or} \quad (4)$$

$$\text{muscovite + quartz = kyanite + K feldspar + } H_2O \quad (4a)$$

Figure 57: Thompson AFM diagram (Thompson, 1957) for metapelites of east Hinnøy.

Phase compatibilities shown are schematic rather than quantitative. Dashed tie lines from staurolite to kyanite, garnet, and biotite are present at the Fjelldal locality only. Sillimanite is known only from one locality. Stippled region indicates the inferred range of bulk rock compositions typical of pelitic compositions in this area.



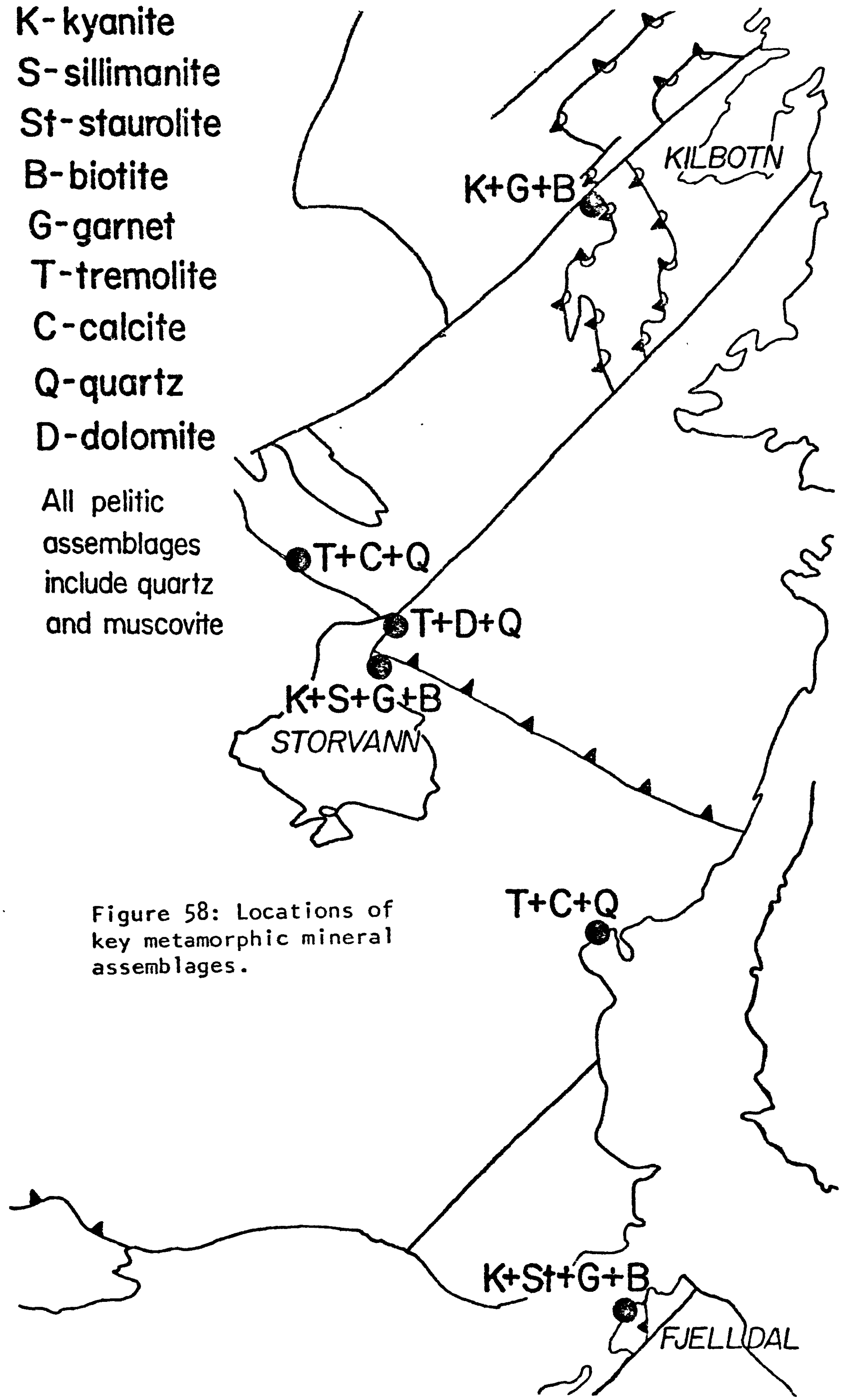


Figure 58: Locations of key metamorphic mineral assemblages.

Carmichael (1978) places the invariant point where reactions (2) and (3) intersect at 4.8 kb and 540° C, which thus are minima for the conditions which produced assemblage (2). The intersection of reactions (3) and (4), which are appropriate to water-saturated conditions, is at 750° C and 9 kb (Mueller and Saxena, 1977). Since melting is nowhere observed, this intersection provides an upper bound on metamorphic temperature. If the system was not water-saturated, reaction (4a) may extend up to intersect reaction (3), instead of reaction (4). This intersection would occur at lower temperature and pressure than (4a); its precise location would depend on the actual value of μ_{H_2O}. Consequently, the upper bound set by water-saturated conditions applies to water-undersaturated conditions as well.

The appearance of fibrolitic sillimanite at locality (2) (see also Chapter 2, p. 77), indicates an increase in metamorphic grade over assemblage (1). The temperature must have been near that of reaction (3), since kyanite and sillimanite coexist in the specimen. The appearance of sillimanite without the breakdown of the assemblage muscovite + quartz, and without partial melting, indicates that the geothermal gradient intersected reaction (3) below its intersection with reaction (4) or (4a), so that this places a maximum of 9 kb upon the metamorphic pressure.

The occurrence of staurolite with kyanite, garnet, and biotite at locality (3) may result from one of two causes: (a) the assemblage is genuinely univariant, corresponding to $P/T/\mu_{H_2O}$ conditions coinciding with reaction (2); or (b) the concentration of a minor element such as Zn, for which staurolite has a strong affinity, is high enough to increase the number of components required to describe the system, making assemblage (3) divariant. Without an analysis of the staurolite, this ambiguity cannot be resolved. In either case, the presence of staurolite suggests a somewhat lower grade of metamorphism than assemblage (1), since even if staurolite is stabilized by Zn, the rock must have been near the conditions at which reaction (2) occurs in the absence of Zn.

Basic rocks are distributed throughout the study area, bearing the assemblage

blue-green hornblende + oligoclase/andesine + clinozoisite + sphene $\pm$ garnet

which is the definitive mineral assemblage for the almandine amphibolite facies of Turner (1968). The assemblage is consistent with conditions inferred from pelites, but is itself not very restrictive of metamorphic conditions. A para-amphibolite from the Kilbotn schist south of the Storvann fault bears the mineral assemblage

hornblende + actinolite + garnet + quartz + calcite + clinozoisite + plagioclase

The hornblende/actinolite solvus is not calibrated at pressures as high as 5 kb, so that these data do not further constrain metamorphic conditions at present.

Although most of the carbonate rocks in the study area are calcite -rich and do not generate diagnostic mineral assemblages at these conditions, the assemblage

tremolite + calcite + quartz (+ white mica)

is widespread if sparsely developed (Figure 58). The absence of diopside (with one exception discussed below) puts a maximum on metamorphic temperature defined by the reaction:

tremolite + calcite + quartz = diopside + CO_2 + H_2O (5)

which has a temperature maximum in T/X_{CO_2} space at X_{CO_2} = 0.75. At 1 kb the temperature maximum is at 540 ± 50° C (Winkler, 1973). Linear extrapolation to 5 kb by the Clapeyron equation using thermodynamic parameters from Robie and Waldbaum (1968) and Robie (1966) increases the temperature maximum to approximately 625° C (see Figure 61).

Possible phase compatibilities for siliceous marbles on Hinnøy are shown in Figure 59. The reaction that separates the two diagrams,

talc + calcite = tremolite + dolomite + CO_2 + H_2O (6)

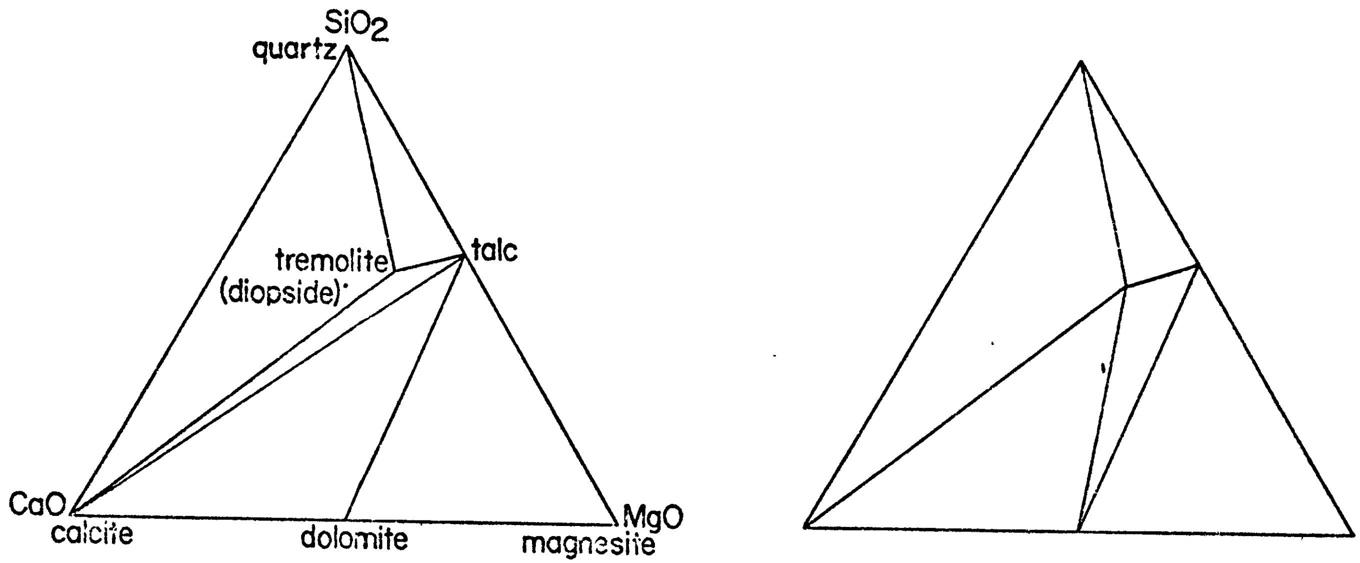
proceeds to the right (Figure 59A to 59B) at X_{CO_2} = 0.94-0.98 at 1 kb and temperatures within the stability of tremolite + calcite + quartz (Winkler, 1973). The assemblage tremolite + dolomite + quartz has been observed at Storvann(S) (Figure 58); talc has not been observed in any specimen. Consequently, the phase relationships shown in Figure 59B appear to be correct for east Hinnøy, implying high X_{CO_2} in the metamorphic fluid of marbles on east Hinnøy. Reaction (6) intersects reaction (5) mear its maximum on a T-X_{CO_2} diagram. Consequently, the maximum metamorphic temperature set by reaction (5) is essentially unaffected by which side of reaction (6) the mineral assemblages lie.

Diopside is present in trace amounts in a calcareous schist from the east shore of Storvann(S). Although the sensitivity of reaction (5) to

Figure 59: Phase compatibilities of siliceous marbles of east Hinnøy.

A: X_{CO_2} < 0.94 - 0.98

B: X_{CO_2} > 0.94 - 0.98



Phase compatibilities shown in Figure 59B are favored (see text), implying high X_{CO_2} in volatile phase.

X_{CO_2} is moderate because both volatile species occur on the same side of the reaction, low enough values of X_{CO_2} (less than about 0.2) will stabilize diopside at temperatures 50° or more below the maximum of the curve. Consequently, this change in mineralogy probably does not reflect a difference in metamorphic grade, but rather a different fluid composition.

Basement Rocks

The textural effects of the Caledonian dynamothermal metamorphism on the granite gneisses of the basement have been discussed in Chapter 3 (p. 151). The mineralogical effects include the breakdown of plagioclase to more sodic plagioclase plus clinozoisite, and formation of muscovite at the expense of K feldspar, by the reaction:

anorthite (in plagioclase) + microcline + water = clinozoisite + muscovite + quartz (7)

Because all the phases involved in this reaction can stably coexist in a granitic composition at a range of P/T conditions, the reaction represents no change in fundamental mineral compatibilities. As a consequence, the progress of the reaction does not necessarily require a change in P/T conditions of equilibration. The one vital requirement for the progress of reaction (7) is addition of water to the system. This further reinforces the interpretation presented in Chapter 3 that the availability of water was the main controlling factor in the development of a Caledonian overprint in the pre-Caledonian granite gneisses.

The Middagstind Quartz Syenite has been thoroughly retrograded to an amphibolite facies assemblage. Griffin and others (1978) suggested that widespread retrogression of the granulites in Lofoten reported by Griffin and Heier (1969) occurred at 1100 to 1000 Ma and was unrelated to Caledonian events. However, on east Hinnøy there is no structural, petrographic, or geochronologic evidence of an event at 1000 Ma, and the preferred interpretation is that retrogression of the Middagstind syenite was primarily a Caledonian event.

Two mineral reactions are thought to be primarily responsible for the Caledonian retrogression:

hypersthene + K feldspar + H_2O = biotite + quartz (8)

ilmenite + clinopyroxene (clinoamphibole ?) + O_2 = sphene + magnetite + quartz (9)

Reaction (8) is implied by the interpretation of the biotite + quartz meshes in Figure 10 as pseudomorphs of orthopyroxene. Reaction (9) is basically the oxidation of ilmenite to magnetite and sphene to produce the observed rimming of opaques by sphene in Figure 10. The Ca in sphene may have come from either the ferrohastingsite now present as relicts (?) in the rock, or from clinopyroxene now completely consumed by the reaction. It is worthy of note that the higher Ca/(Mg + Fe) ratio in clinopyroxene is more compatible with a balanced reaction. However, without an analysis of the amphibole, the interpretation remains uncertain.

Caledonian effects in the syenite thus include hydration and oxidation, yielding a mineral assemblage stable under amphibolite facies conditions.

The orthopyroxene-bearing hornfels of the contact aureole of the Middagstind syenite shows incipient retrograde metamorphism. The opaque phase is thinly rimmed by garnet, which is in turn surrounded by granular aggregates of plagioclase (Figure 60). This suggests a reaction of the form

plagioclase + magnetite (?) = garnet

Without mineral analyses, it is not possible to satisfactorily balance this reaction.

Summary of Caledonian "Main Stage" Metamorphism

The conditions of the Caledonian thermal peak inferred for rocks on east Hinnøy are shown graphically in Figure 61. The temperature is considered to have been between 550° and 600° C on the basis of phase equilibria in pelitic and calcareous compositions. Pressure is constrained between 4.8 and 9 kb, indicating intermediate crustal depths.

No concrete suggestion of significant variation in Caledonian metamorphic grade on east Hinnøy has been recorded. This contradicts the relations depicted in Gustavson's (1966, Fig. 2) metamorphic map, which shows most of east Hinnøy at garnet grade (upper greenschist facies), with only one small area at staurolite or kyanite grade. Evidence from all rocks examined in this study supports the conclusion that the entire study area experienced amphibolite facies conditions (kyanite grade) during the Caledonian orogeny. The absence of key indicator minerals in most rocks appears to be compositionally controlled, rather than reflecting conditions of metamorphism.

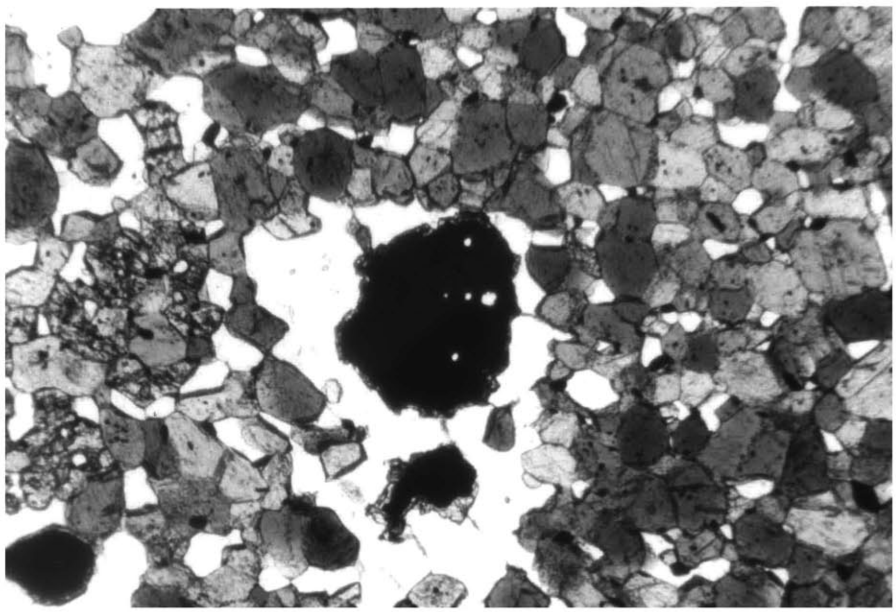


Figure 60: Photomicrograph of rim of garnet(?) on opaque mineral in hypersthene hornfels.

Plane polarized light, 75X.

Figure 61: Summary of constraints on conditions of Caledonian main stage metamorphism from mineral phase equilibria.

Stippled region: range of allowed conditions
Ruled region: range of conditions favored by author.

Abbreviations as in Figure 58, except:
- D - diopside
- A - andalusite
- V - vapor ($H_2O \pm CO_2$)
- L - silicate liquid

Sources of phase boundaries:

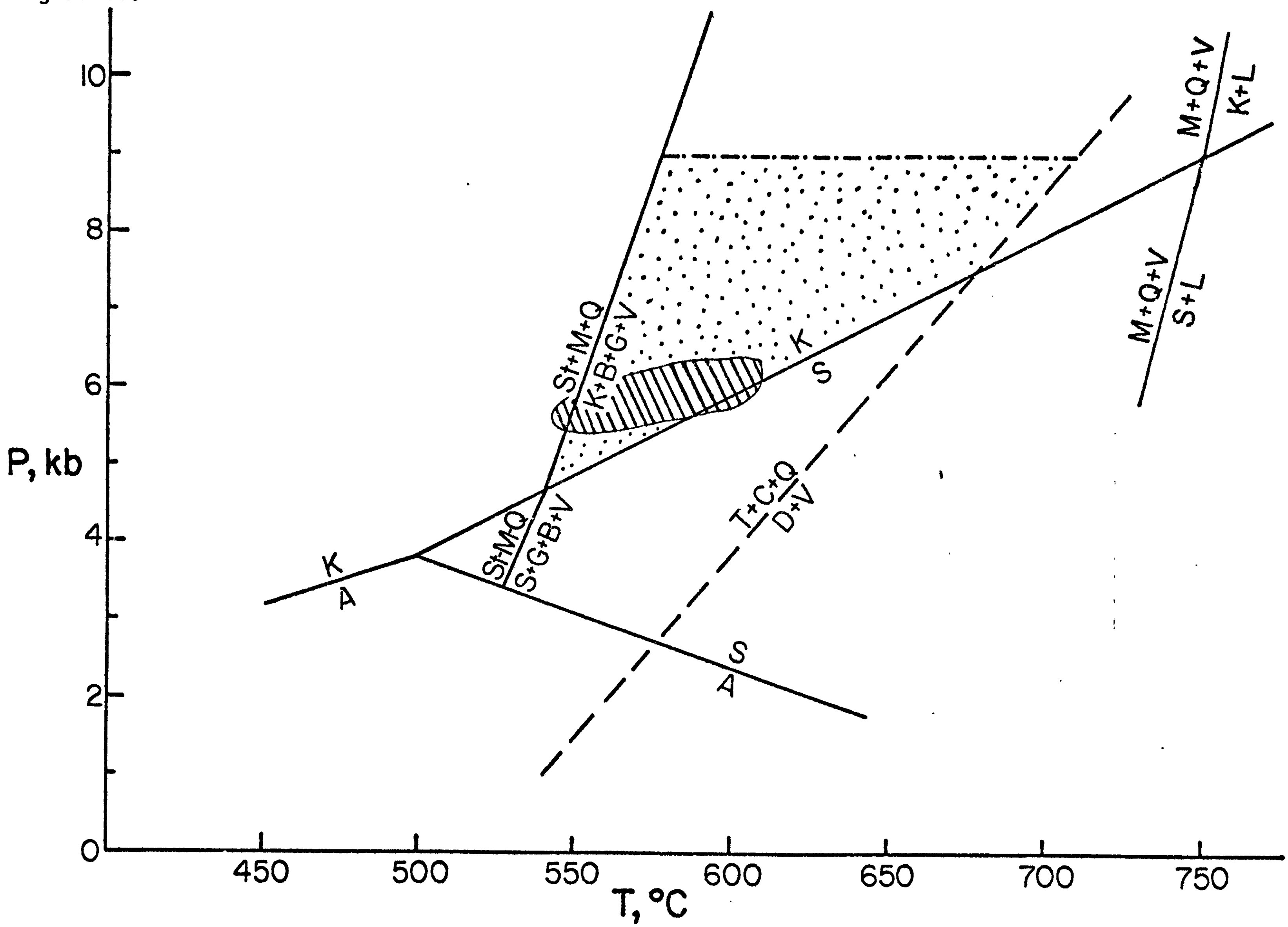
Al_2SiO_5 system: Holdaway and Lee, 1971

Staurolite breakdown: Carmichael, 1978

Muscovite + quartz breakdown: modified from Storre and Karotke (1971) to be compatible with Holdaway and Lee (1971)

Tremolite + calcite + quartz breakdown: Winkler, 1973; extrapolated in P/T space by the author.

Figure 61.



10
8
6
4
2
0
P, kb
450
500
550
600
650
700
750
T, °C
St+M+Q
K+B+G+V
St+MQ
S+G+B+V
K
A
K
S
S
A
T+C+Q
D+V
M+Q+V
K+L
M+Q+V
S+L

Retrograde Metamorphism

Two types of retrograde metamorphism can be distinguished on east Hinnøy. The more important is a result of deformation extending beyond the metamorphic peak (Figure 62; see next section). The late deformations D_3, and D_4 where present, began to develop new mineral assemblages replacing the amphibolite facies minerals along spaced cleavage planes S_3 and S_4. In pelites, the new mineral assemblage includes

chlorite + biotite ± muscovite + epidote + quartz

Whether garnet was stable in any composition at this time is not clear; the garnets produced by amphibolite facies metamorphism are invariably retrograded to chlorite ± epidote, but this may be a function of composition. No new garnet growth related to this event has been recognized, however.

In basic rocks, the assemblage

chlorite + biotite + epidote + calcite + quartz

is developed along the spaced cleavages. In carbonate rocks, tremolite is commonly altered to a fine-grained aggregate of talc? + quartz + carbonate. In marbles lacking tremolite, no affects attributable to this event are present except an occasional slight flattening of carbonate grains into an incipient solution (?) cleavage. The fact that these marbles, strongly folded in F_3 (see Plate IIB, D-D'), show such limited effects is also supportive of the interpretation that F_3 folds formed at still somewhat elevated temperatures, since it implies the carbonate grains could readily recrystallize to accomodate strain.

The mineral assemblages from D_3 and D_4 are characteristic of the greenschist facies. A temperature around 350 to 400° C is most likely, but the high variance of these assemblages does not allow quantitative constraints to be placed on this.

The late crenulation in the Kilbotn Schist, S_5, further retrogrades all earlier assemblages, replacing biotite with more chlorite, and growing some new biotite and muscovite. The essential mineral compatibilities are unaltered. To a first approximation, the continuous reaction (taken from Miyashiro, 1973, P. 208-9)

phengite + chlorite = muscovite + biotite + quartz + H_2O (10)

appears to have shifted toward the left, but no discontinuous reaction, where new mineral compatibilities are established, has been reached.

There is no indication that the retrogression associated with D_3

through D_5 represents a separate metamorphism. More likely the late deformations occurred during cooling after the thermal peak.

Local, strong retrogression is patchily developed in the study area. Parts of the contact aureole on the east side of the Middagstind syenite have been strongly epidotized, producing both disseminated fine-grained granular epidote and quartz-epidote-magnetite vein fillings. In the field this was at first thought to be part of the contact metamorphic effects of the syenite pluton. However, similar local retrogression occurs 2 km south of Torskvatsfjell in Kvaefjord Group rocks, in Storvann Group quartz-garnet schist on the south side of the summit area of Finnslettheia, and in restricted patches throughout east Hinnøy. Consequently, the Precambrian age of the Middagstind syenite precludes any relationship between it and this retrograde metamorphism. No alternative can be forwarded at this time.

Timing of Caledonian Metamorphism Relative to Deformation

Figure 62 summarizes the timing of deformations relative to an approximate thermal history. Evidence was presented in Chapter 3 ("Stangnes Thrust") that the earliest Caledonian structures recognized on east Hinnøy, D_1, probably developed under amphibolite facies conditions. In essence, the mylonitic fabric at the basal thrust contact of the Stangnes Group involves synkinematic growth of the amphibolite facies mineralogy (hornblende, oligoclase, clinozoisite), and has not experienced noticeable overprinting by later fabrics or post-kinematic annealing.

Evidence for growth of garnet synchronous with early schistosity development (generally S_2) is common in more aluminous schists, but lacking in less pelitic compositions. Garnet in the Storvann Group pelitic schist includes an earlier schistosity (S_1) as linear quartz inclusion trails (Figure 63). These continue into an outer zone with sparser inclusions which define curving trails, indicating synkinematic growth. Since S_3 cleavage truncates this internal fabric at a high angle (Figure 64), the curvature of the trails indicates growth synkinematic with D_2.

Snowball inclusion trails in garnets are also locally preserved in the Kilbotn Schist and the garnetiferous paragneisses of the Kvaefjord Group. However, it is not clear in these examples whether the garnets grew synkinematically in D_1 or D_2.

Kyanite in the Narvik Group pelitic schist and gneiss at Kilbotn grew

Figure 62: Plot of metamorphism versus deformation.

Mineral abbreviations for grade in pelites are same as in Figure 58, except Ch - chlorite.

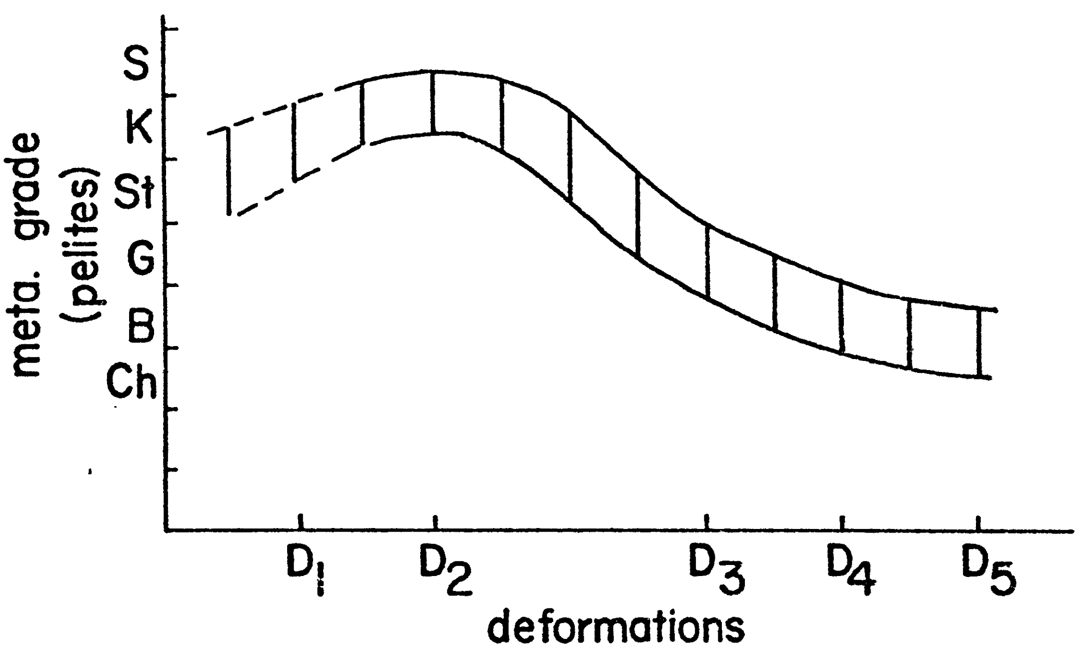


Figure 63: Photomicrograph of curving inclusion trails in garnet from Storvann Group pelitic schist.

Plane polarized light, 30X.

Figure 64: Photomicrograph of relationship between internal and external foliations of garnet porphyroblast.

Same specimen as Figure 63. Crossed polars, 13X.

Figure 63.

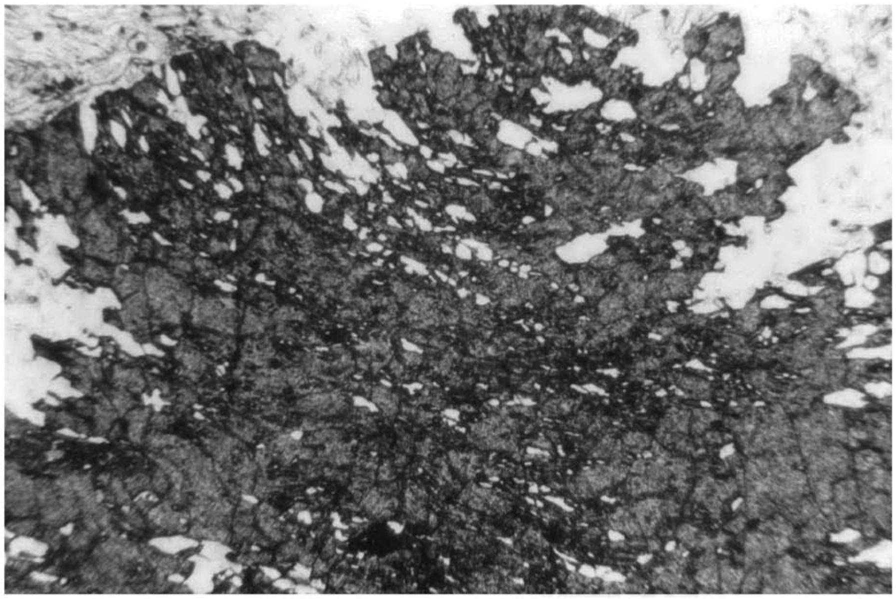
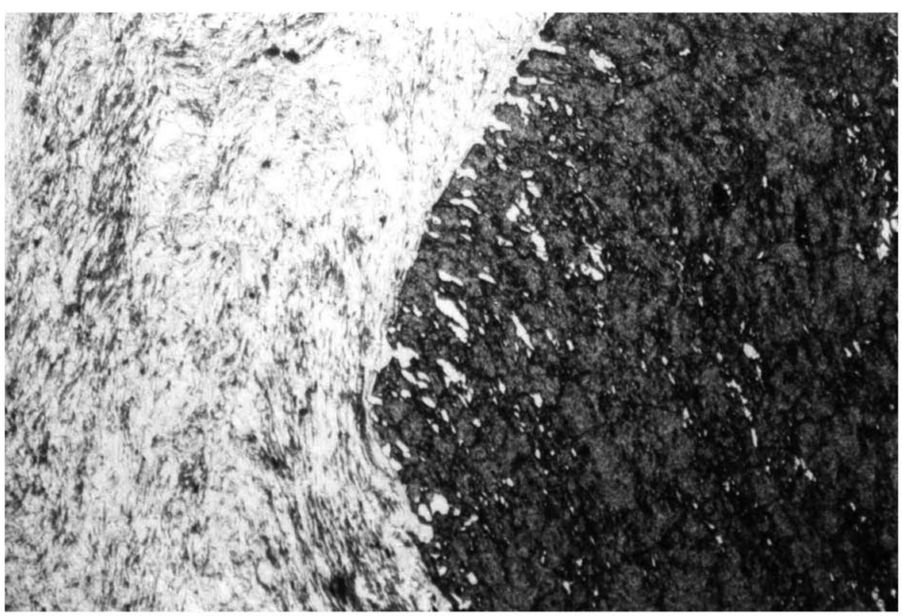


Figure 64.



parallel to the schistosity (S_1 parallel to S_2) ; it is uncertain whether this results from synkinematic growth during early deformations or static mimetic growth following D_2. Kyanite from the other localities mentioned in the previous section has been too strongly retrograded, and the host rocks too strongly overprinted by late cleavage formation, to determine timing of kyanite crystallization relative to early deformations.

Abundant static porphyroblast growth occurred after D_2. In the Storvann Group schists, randomly-oriented hornblende and plagioclase porphyroblasts overgrow D_1 and D_2 fabrics (Figure 65). In the calcareous schists of the sliver zone at the base of the D_1 nappe sequence, large poikiloblastic plagioclases overgrow the early schistosity and F_2 folds which deform it. In the Kilbotn schist, the garnet and hornblende porphyroblasts preserve the only evidence of a schistosity earlier than S_3 (Figure 66). The internal foliation preserved in the porphyroblasts is essentially always planar, indicating post-kinematic porphyroblast growth. The hornblende is now commonly somewhat elongate parallel to S_3, the late spaced cleavage, but there is no indication of crystallographic preferred orientation. Instead, this appears to result from truncation of the hornblende grains by S_3 cleavage surfaces to leave relicts which are elongate in the younger foliation (Figure 66B). The earlier schistosity is well preserved in the garnet amphibolite boudins within the Kilbotn Schist. The absence of quartz and micas, which are the dominant phases participating in the development of the late cleavages, apparently made this lithology resistant to D_3 through D_5 effects. Consequently, the hornblende still is aligned both dimensionally and crystallographically to define a strong schistosity which is locally overgrown, without rotation, by small garnet porphyroblasts.

All of the porphyroblasts in the rocks described above are retrograded where they are intersected by S_3 cleavage planes. Retrograde products are chlorite $\pm$ epidote in the case of garnet and hornblende, muscovite + quartz $\pm$ chlorite in the case of kyanite and staurolite, and sericite + epidote in the case of plagioclase (Figures 64, 65, and 66). Thus, the main porphyroblast growth appears to have occurred under static conditions between D_2 and D_3.

It cannot be categorically stated that the metamorphic temperature was essentially unaltered from at least late D_1 until after D_2. Although this

Figure 65: Photomicrographs of porphyroblast/matrix fabric relationships in quartz-garnet schist of the Storvann Group.

A: Plagioclase: The albite-twinned porphyroblast grew as a replacement of white mica which defines the early phase microfolds seen here. The porphyroblast thus post-dates these folds(F_2). It is in turn truncated by late solution(?) cleavage seen running subhorizontally through the upper part of the photo. Crossed polars, 75X.

B: Hornblende: The porphyroblast includes elongate quartz inclusion trails which preserve S_1 (or S_2?). The early schistosity is also preserved by biotite grains in the upper left corner of the photo. Both the porphyroblast and internal foliation were then truncated, by later spaced cleavage surfaces, here seen trending from upper left to lower right at a high angle to the early schistosity. Plane polarized light, 30X.

Figure 65A.

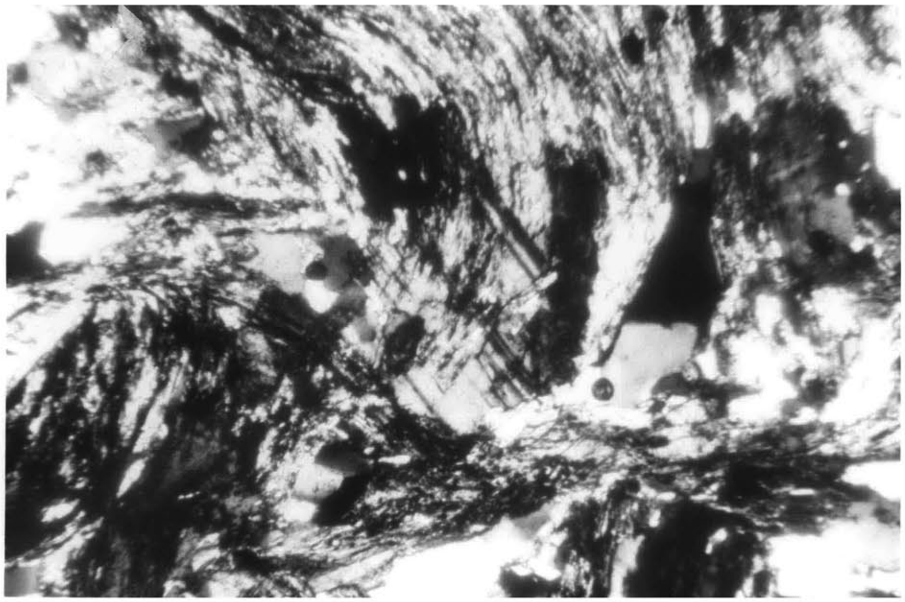


Figure 65B.

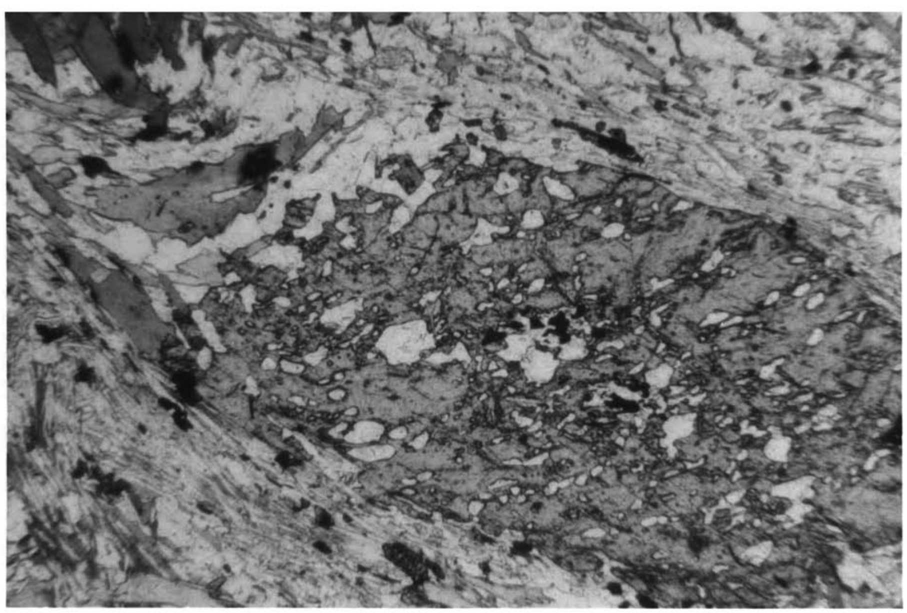


Figure 66: Photomicrographs of porphyroblast/matrix relationships in Kilbotn schist.

A: Garnet: Poorly defined linear inclusion trails in central zone of larger prophyroblast preserve early schistosity. Inclusion-free rim is interpreted as interkinematic (post-D_2, pre-D_3). Later spaced cleavage trace truncates garnet at high angle to earlier fabric. Plane polarized light, 75X.

B: Hornblende: Elongate quartz grains included in the hornblende porphyroblast (dark grain in center of photo) preserve earlier foliation. Hornblende elongation now seen is in large part due to truncation by strong S_3 cleavage defined by large biotite grains (subvertical in photo). Plane polarized light, 30X.

Figure 66A.

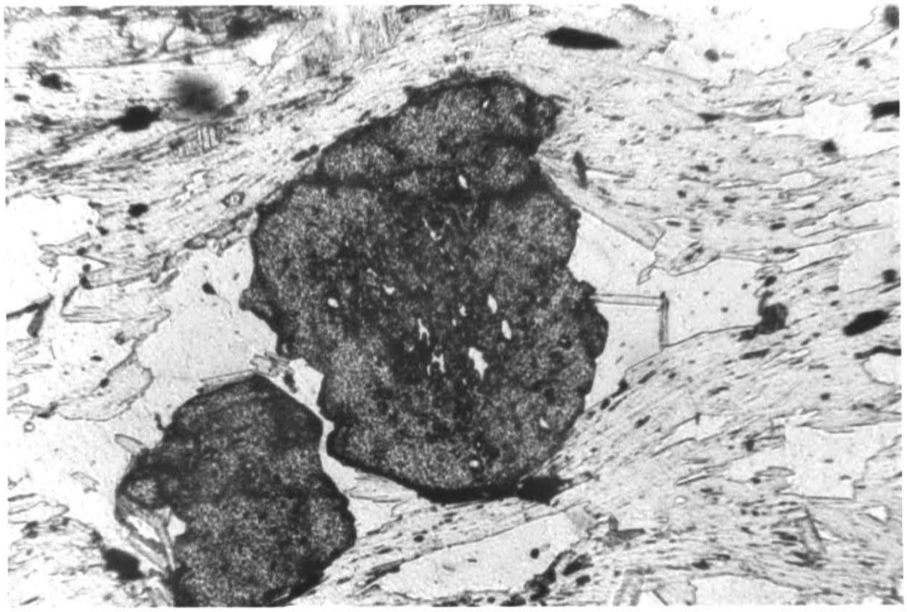
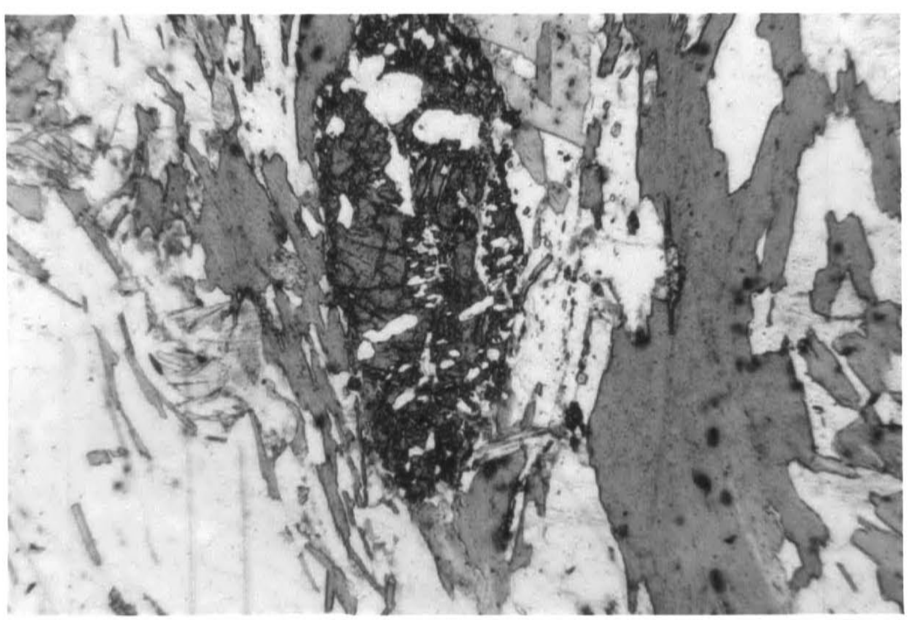


Figure 66B.



is the simplest interpretation consistent with the facts, it cannot be ignored that evidence of synkinematic growth of key minerals during D_2 is almost completely lacking. The possibility exists for a drop in temperature before or during D_2 time followed by a temperature increase after D_2, leading to the major period of porphyroblast growth. On the other hand, it cannot be inferred that the peak conditions of metamorphism occurred during the main growth of porphyroblasts. Porphyroblast growth may have resulted when tectonic grain size reduction ended rather than at the time of highest temperature. The key mineral assemblage, kyanite + garnet + biotite, in part developed post-D_2 (the garnets locally overgrow S_2), but probably began to develop earlier, since the kyanite is elongate in the schistosity.

As developed in the previous section on the late retrograde metamorphism, the late structural events seem to have occurred after the temperature had fallen below the amphibolite facies. The progressive retrogression of more and more biotite to chlorite (especially in the Kilbotn Schist) suggests that these deformations occurred in a regime of continuously decreasing temperature, presumably after tectonic stacking had ended, and uplift and erosion had begun.

CHAPTER 5: GEOCHRONOLOGY

Introduction

Prior to this study, no isotopic dating had been attempted of rocks on east Hinnøy. Consequently, it was considered worthwhile to attempt to establish the rudiments of a temporal framework for some of the rock-forming and structural events recognized in the field study. Rb/Sr geochronology of selected rocks was undertaken with the following objectives: (1) to determine ages for some of the basement rocks in order to better evaluate the relationship of the basement terrain of east Hinnøy to the Lofoten terrain to the west; (2) to provide constraints on the ages of some of the cover rocks; and (3) to provide constraints on the time of Caledonian metamorphism on east Hinnøy. The use of the Rb/Sr decay system on whole rocks of igneous origin also allows some constraints to be placed on the petrogenesis of these rocks.

The units chosen for study as whole rocks were the Melaa Granite and the Middagstind Quartz Syenite of the pre-Caledonian basement complex, and the Ruggevik Tonalite Gneiss from the allochthonous Stangnes Group. The selection of the units from the basement was a result of two considerations: (1) both units show clear intrusive relationships with other rocks of the basement complex so that minimum ages for other units are indirectly established as well as dating the actual plutonic rocks sampled; and (2) both of these bodies include undeformed or only slightly deformed portions. Deformation probably enhances mobility of Rb and Sr, especially when accompanied by recrystallization of biotite and feldspar which contain the bulk of these elements in the rock. Consequently, undeformed rocks are considered likely to have their whole rock isotopic systematics less disturbed by metamorphic effects. The Ruggevik tonalite was chosen for study as the only volumetrically significant granitoid body within the Caledonian allochthons exposed on east Hinnøy. It was hoped that this would yield a minimum age for the formation of the protoliths of the Stangnes amphibolite which it intrudes. Unfortunately, the whole rock systematics of this body are badly disturbed, and no meaningful whole rock age can be derived from the data presently available.

Biotite separates were prepared from samples of the Middagstind syenite and the Ruggevik tonalite to provide a minimum age for Caledonian metamorphism on east Hinnøy and to compare the time of cooling from widely separated portions of the study area.

Sample Collection

Samples were collected mainly in the course of lithologic and structural mapping, so that the number and size of specimens was perhaps smaller than that typical of a concerted geochronologic study. Samples were 3 to 5 kg blocks, trimmed of weathered material as much as possoble while on the outcrop. The standard field criterion for sample freshness was the author's inability to further crack the specimen with repeated blows of a 3 pound sledge. Petrographic examination appears to justify this criterion.

Sample distributions for the three bodies are shown in Figures 67, 68, and 69. Geographical spread of sample locations was a primary goal in an attempt to maximize spread of Rb/Sr ratios. This procedure may have compounded the difficulties encountered with the Ruggevik tonalite, since the petrographic variability (see Chapter 2) encountered in this unit may result from a composite origin, with variable $^{87}Sr/^{86}Sr$ initial ratios. However, the possibility that samples of the tonalite collected on a smaller scale will yield better results has not yet been tested.

Sample Preparation

Samples were split into several subequant pieces about 5 cm on a side, and all weathered-appearing material removed with a diamond saw (including along any cracks, which occasionally were not noticed during sample collection). The pieces were washed with water, and then the entire specimen was crushed in a jaw crusher and plate mill (steel crushing surfaces) to maximum grain size of approximately 2 mm. A "representative" 50 to 100 grams of this material was selected for powdering by removing several small scoopfuls of the crushed material. Powdering was done in a steel shatterbox for 2 to 3 minutes, until the powder appeared fine and uniform. Powders were stored in glass jars or polyethylene bags.

Biotite Separations

Four biotite separates were prepared, two from the Middagstind syenite and two from the Ruggevik tonalite. On the the separations from the tona-

Figure 67: Sample locations, Middagstind Quartz Syenite.

Area of syenite stippled. Sample 16D is the hypersthene bearing hornfels of the contact aureole.

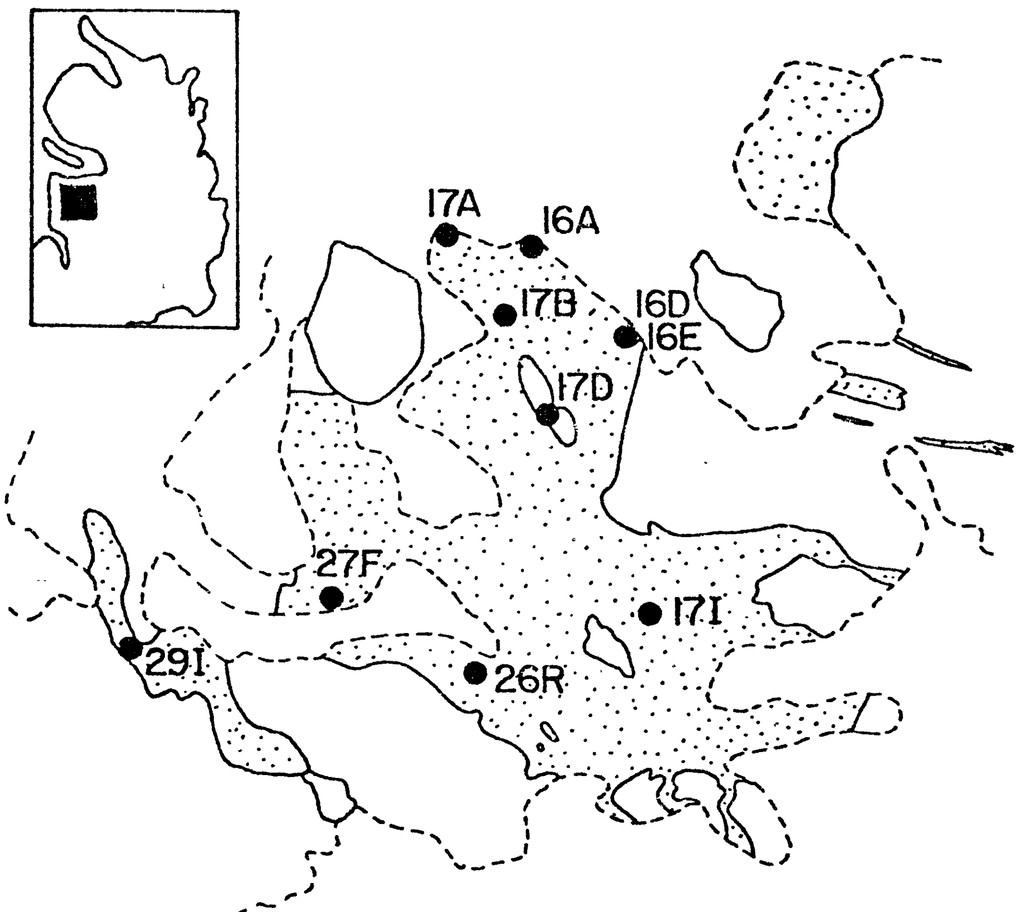


Figure 68: Sample locations, Melaa Granite.

Area of granite stippled.

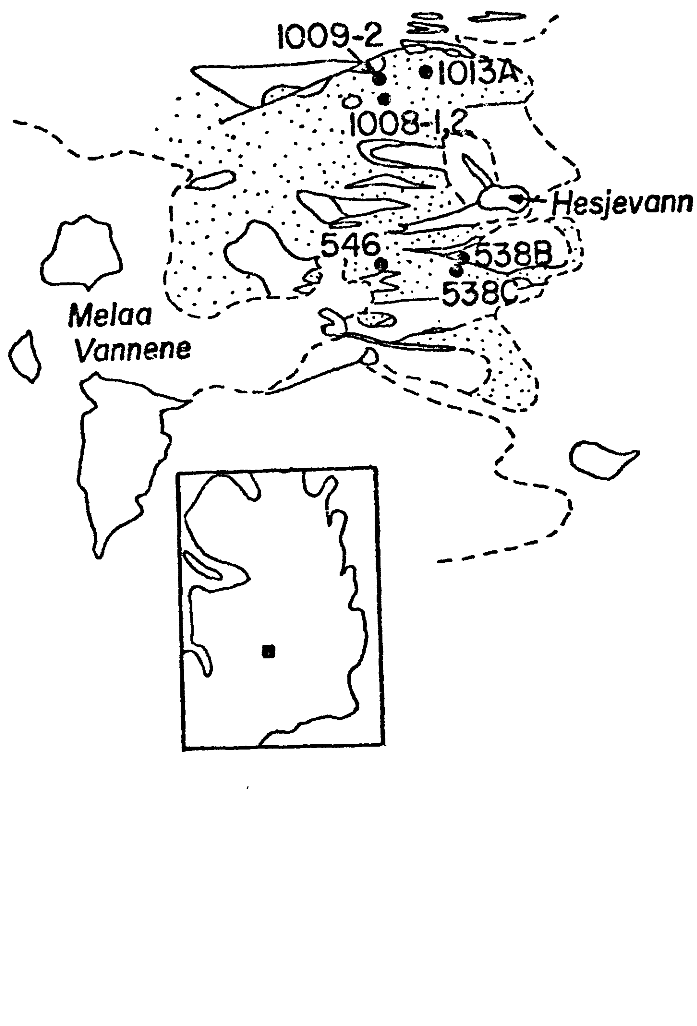
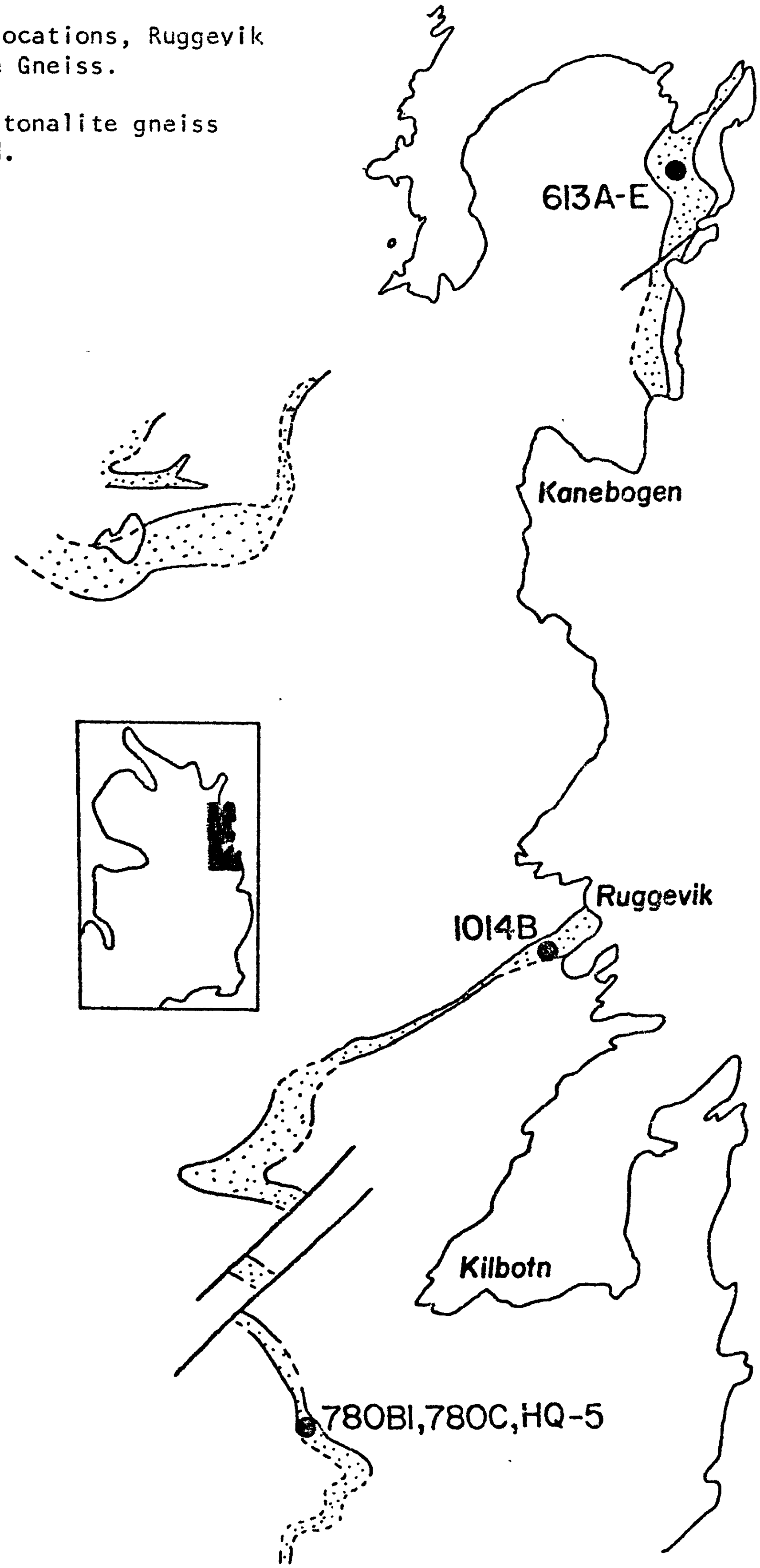


Figure 69: Sample locations, Ruggevik Tonalite Gneiss.

Area of tonalite gneiss stippled.



lite sample 780C, was done using heavy liquids (bromoform) and a magnetic separator by B. Nilssen at the Mineralogisk Museum in Oslo. Grain size of this sample was approximately 100 mesh. Some further hand-picking and concentration by shaking on filter paper yielded a reasonable degree of purity (>95%).

The other three separates, from samples 26R and 27F of the syenite and HQ-5 from the tonalite, were prepared by the author at M.I.T. The crushed material was sieved for grains smaller than 35 mesh and larger than 60 mesh, from which individual biotite grains were hand-picked. The only significant impurity in these separates is quartz in the syenite specimens, which is often intimately intergrown with biotite (see Chapter 2). However, quartz should have no effect beyond dilution of the sample, and a high degree of purity (>99%) is believed to have been achieved.

The lack of significant difference in age results (see Table 6) from the separates prepared two different ways suggests that no problems with mineralogical purity, chemical contamination, or bias due to different grain sizes, were introduced by the biotite separation procedures.

All separates were washed twice each with acetone and ethanol in an ultrasonic cleaner before dissolving. Chemical procedures were identical to those used for whole rocks (see Appendix).

Analytical Techniques

Isotopic analyses were performed in two different laboratories. The whole rock analyses of the Melaa Granite and Ruggevik tonalite samples were done at the Mineralogisk Museum of the University of Oslo, Norway. The Middagstind syenite whole rocks, two Ruggevik tonalite whole rocks (one a duplicate of a powder analyzed in Oslo), and all the biotite separates were analyzed at M.I.T. Chemical and filament loading procedures differ somewhat between the two laboratories; these are detailed in the Appendix. Slightly different results were obtained for $^{87}Sr/^{86}Sr$ of the sample run at both laboratories (.70520 at Oslo, .70505 at M.I.T.); the values barely overlap at the 2σ confidence level. Since high precision of initial $^{87}Sr/^{86}Sr$ was not a primary focus of this study, precise cross calibration was not attempted (all values are normalized to a value of $^{87}Sr/^{86}Sr$ = .70800 for Eimer and Amend $SrCO_3$). In age calculations, values from only one laboratory were used. This additional small source of uncertainty, not explicitly expressed, must be born in mind when interpreting initial $^{87}Sr/^{86}Sr$ calculated from whole rock data. The source of the difference in the analyses from the two laboratories is uncertain, but is assumed to be systematic so that relative values of analyses are accurate, and hence age calculations are unaffected.

Mineralogisk Museum

Rb and Sr concentrations were determined by X-ray fluorescence spectroscopy (XRF), using techniques developed by S. Jakobsen modelled after Pankhurst and O'Nions (1973). The XRF-analyses for both Rb and Sr of most of the Melaa Granite samples, and Rb analyses of three of the Ruggevik tonalite samples, were checked by isotope dilution using a mixed ^{87}Rb-^{84}Sr spike. Agreement of the techniques was consistently within the $\pm$ 5% accuracy estimated for XRF and $\pm$ 1% estimated for isotope dilution (2σ confidence level), with the exception of Sr for one sample (1009-2), which contains less than 10 ppm Sr. The isotope dilution values are considered more reliable, especially at low concentrations, and are used where available. Rb and Sr concentrations determined in Oslo are listed in Tables 4 and 5, annotated with regard to the technique used to obtain the value listed.

Sr isotopic compositions were determined on spiked and unspiked samples on a Micromass MS30 spectrometer, using precedures modelled after Pankhurst and O'Nions (1973; see also Appendix below). Analyses and analytical precision are given in Tables 4 and 5.

M.I.T.

Rb and Sr concentrations were determined by isotope dilution. Analytical precision is considered better than $\pm$ 1% in all cases, and probably close to 0.1% for Sr for which a mass fractionation correction was made, using $^{86}Sr/^{88}Sr = 0.1194$ as a normalizing standard.

Sr isotopic compositions were determined on spiked samples on NIMA-B, a 9 inch-60° solid source mass spectrometer built at the Carnegie Institution of Washington and now in use at M.I.T. Results are listed in Tables 3, 5, and 6.

In general, precision of the analyses at M.I.T. is superior, but greater difficulty was experienced in obtaining a stable ion beam during

Sr runs. Both of these differences are probably a consequence of two differences in procedure: (1) the use of perchloric acid rather than nitric acid as an oxidizing agent in the M.I.T. chemical procedure; the perchloric acid is difficult to remove completely from the sample, and if present, has disastrous effects upon performance in the spectrometer; and (2) the use of phosphoric acid in Oslo as a loading medium as opposed to Ta_2O_5 at M.I.T. The phosphoric acid seems to promote ionization of Sr somewhat more consistently, but tends to give a noisier signal.

Results and Discussion

Middagstind Quartz Syenite

The Rb/Sr whole rock data from the Middagstind Quartz Syenite (Table 3) are plotted on an isochron diagram ($^{87}Sr/^{86}Sr$ vs. $^{87}Rb/^{86}Sr$) in Figure 70. Linear regression of these data using a computer program modelled after York Model II (York, 1966; Brooks and others, 1972), gives an age of 1726 $\pm$ 31 Ma with initial $^{87}Sr/^{86}Sr$ of 0.7065 $\pm$ 0.0003 (2σ confidence level). While the points define a reasonably good linear array, departures from the best fit line exceed analytical error and must be attributed to geological causes. The most likely source of scatter is the Caledonian thermal overprint. While no apparent deformation affected any of the samples analyzed, mobility of fluids during Caledonian time, perhaps as an intergranular phase, was adequate for water to permeate the intrusion to the extent that all primary pyroxene was replaced by biotite + quartz (Chapter 4). Consequently, there is good reason to believe that some movement of Rb and Sr also occurred, accounting for the mild scatter of the data.

Both the age and initial ratio indicate that the Middagstind Quartz Syenite is a product of the period of granulite facies intrusions (mainly mangerite) widely developed to the west of Hinnøy in the Lofoten terrain. Griffin and others (1978) report ten Rb/Sr whole rock ages from intermediate to acidic plutons of similar overall chemistry ranging from 1805 to 1695 Ma, and initial $^{87}Sr/^{86}Sr$ ranging from 0.7030 to 0.7051. While the initial ratio obtained here is above the range reported by Griffin and others, most of these initial ratios lie above probably inferred mantle evolution paths of $^{87}Sr/^{86}Sr$ at 1700 Ma (e.g., Faure, 1977), suggesting

TABLE 3: Rb/Sr whole rock analyses from the Middagstind Quartz Syenite. See Figure 67 for sample locations.

Sample	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr \pm 2\sigma$
17A	52.3	243.2	0.6013	0.72053 ± .00005
17D	47.1	252.1	0.5402	0.72142 ± .00006
16A	58.8	247.8	0.6707	0.72146 ± .00006
17B	54.4	244.9	0.6279	0.72183 ± .00009
27F	54.4	219.0	0.7021	0.72414 ± .00006
16E	57.5	185.8	0.8946	0.72848 ± .00008
26R	50.9	137.2	1.051	0.73466 ± .00006
17I	73.7	100.3	2.046	0.75632 ± .00008
29I	80.0	57.6	3.96	0.80137 ± .00006

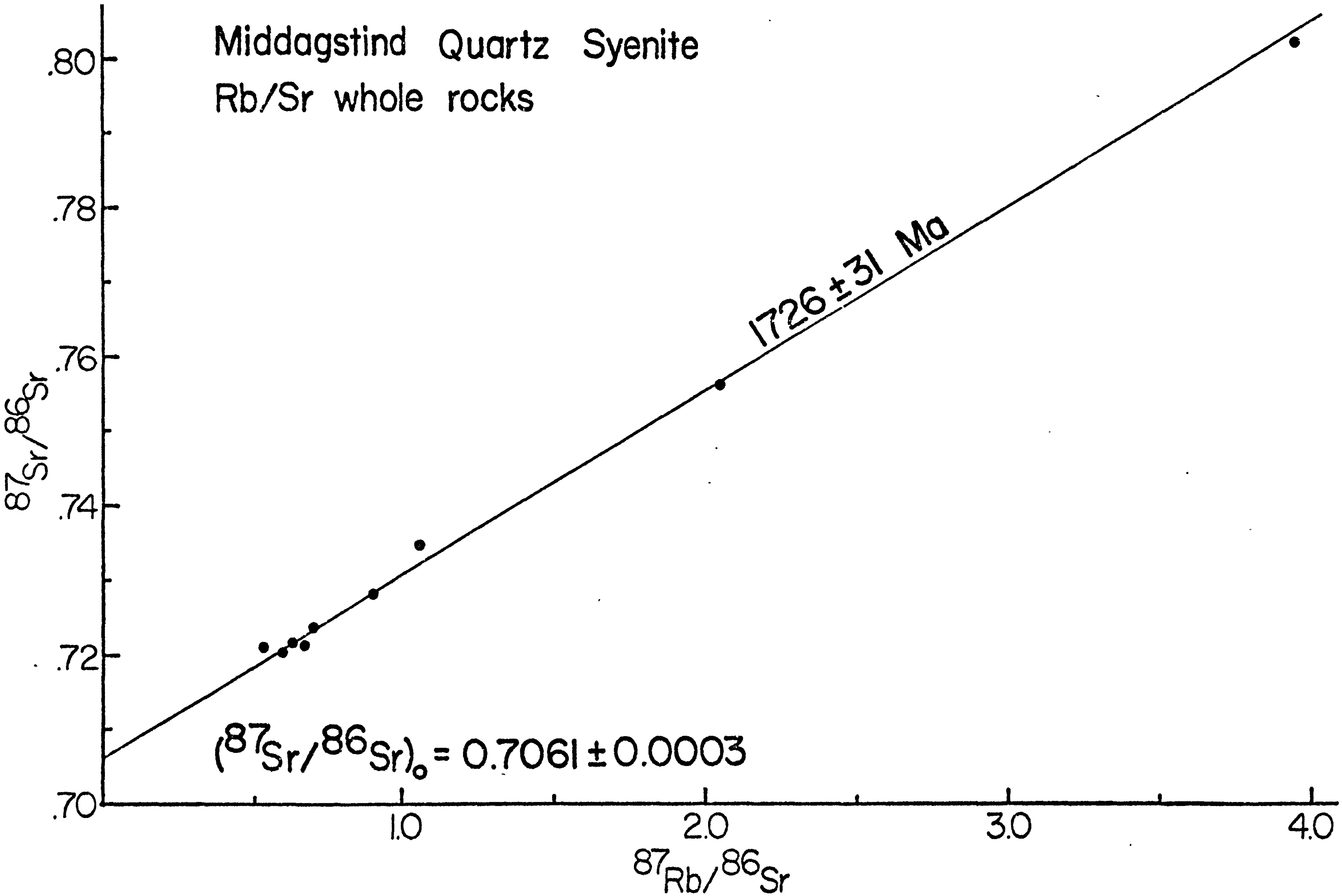
TABLE 4: Rb/Sr whole rock analyses from the Melaa Granite. See Figure 68 for sample locations.

Sample	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr \pm 2\sigma$
1013A	87	219	1.15	0.74250 ± .00010**
538C1	108*	128*	2.44	0.76656 ± .00014
538C2	109*	133*	2.38	0.76731 ± .00013
1008-1	100	21.9*	13.4	1.00876 ± .00007
1008-2	102*	20.2*	15.31	1.01440 ± .00007
1009-2	94.5*	8.20*	36.4	1.65201 ± .00012
546	110*	5.34*	68.8	2.24652 ± .00014
538B	117*	4.7	81.3	2.14219 ± .00015**

* Determined by isotope dilution; other concentrations in Table 4 determined by X-ray fluorescence.

** Unspiked run.

Figure 70.


Middagstind Quartz Syenite
Rb/Sr whole rocks
1726 ± 31 Ma
$(^{87}Sr/^{86}Sr)_o = 0.7061 \pm 0.0003$
$^{87}Sr/^{86}Sr$
.80
.78
.76
.74
.72
.70
1.0
2.0
3.0
4.0
$^{87}Rb/^{86}Sr$

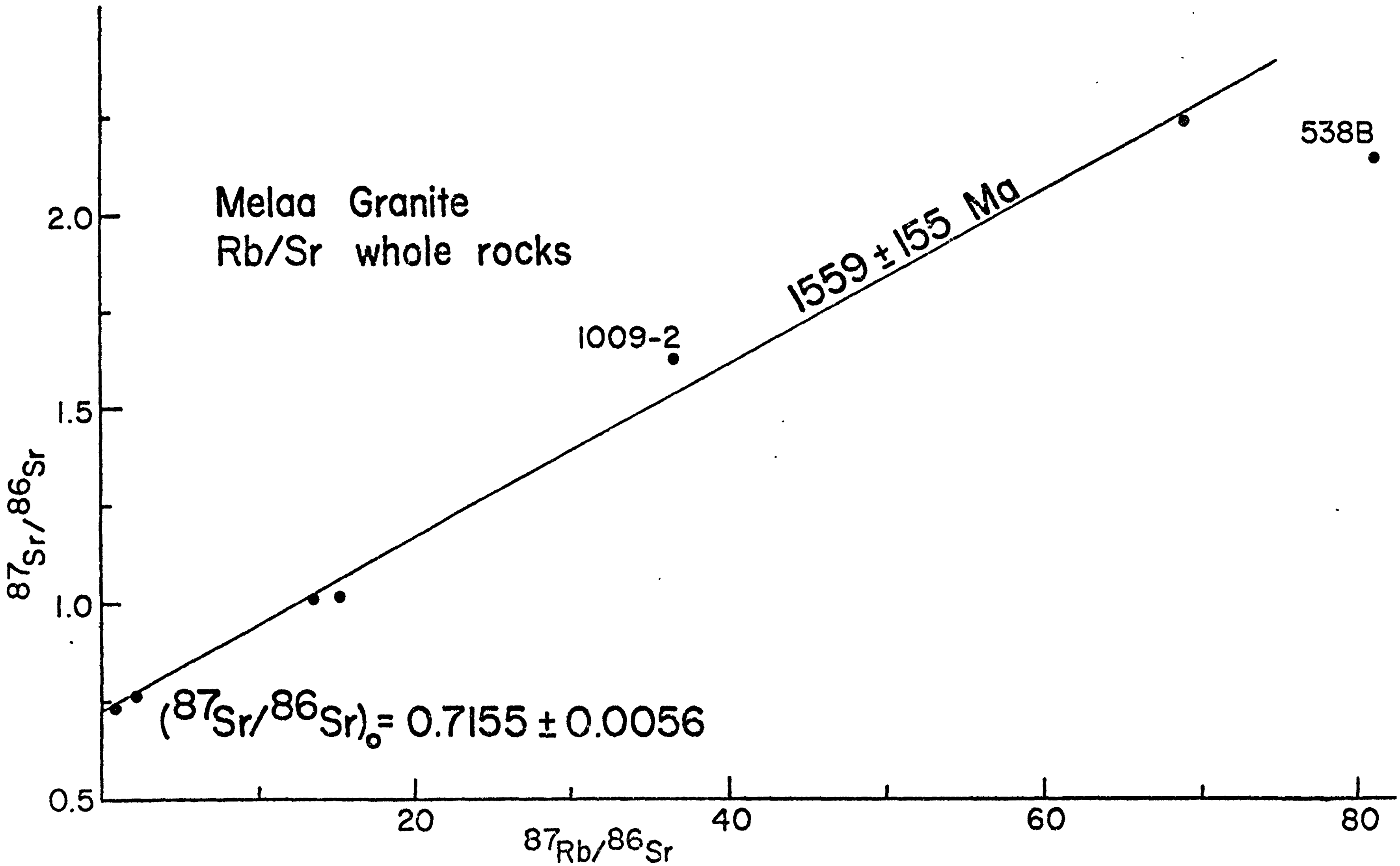
involvement of some continental crust in the magma genesis of the entire intrusive suite (Griffin and others, 1978). As a consequence, the slightly higher initial $^{87}Sr/^{86}Sr$ determined for the Middagstind syenite does not significantly weigh against this correlation. It is also possible that the isochron has been rotated slightly by metamorphic overprinting, consequently raising the apparent initial ratio.

It is believed that this age reflects the time of igneous crystallization of the Middagstind syenite rather than closure after metamorphic homogenization of Sr. The reasonable uniformity of both initial $^{87}Sr/^{86}Sr$ and ages now determined for eleven plutons of the suite is good: nine of the ten ages in Griffin and others (1978) have error ranges which overlap that presented here. Furthermore, the Proterozoic supracrustal (metavolcanic?) rocks of Langøy and Austavaagøy gave an only slightly older Rb/Sr whole rock age of 1830 $\pm$ 35 Ma (Griffin and others, 1978) which those authors considered to reflect homogenization of Sr in the granulite facies metamorphism. It is not clear why the Rb/Sr systems should have closed so early in the supracrustal rocks when the petrography of the mangerite suite, younger on the basis of both isotopic and field criteria, indicates that granulite facies conditions persisted until at least 1700 Ma. The author as a result prefers the interpretation that the age obtained from the supracrustal rocks reflects the time of formation of these rocks, rather than their metamorphism. The Proterozoic supracrustals are intruded by rocks of the mangerite suite on Flakstadøy, thus indicating that the mangerites must have initially formed after 1830 Ma. Consequently, the 1800 to 1700 Ma ages are considered to reflect the time of emplacement of the mangerite plutonic series, consistent with the interpretation of Griffin and others (1978).

Melaa Granite

Rb/Sr whole rock data from the Melaa Granite are plotted on an isochron diagram in Figure 71. The best-fit regression line (York model I, Brooks and others, 1972) shown does not include sample 538B, which came from a 0.5 m thick aplite dike cutting wall rock amphibolite. The dike, containing only about 5 ppm Sr, was highly vulnerable to contamination by less radiogenic Sr from the wall rocks. It is thought that the dike rock exchanged Sr with its surroundings during Caledonian metamorphism, account-

Figure 71.


Melaa Granite
Rb/Sr whole rocks
1559 ± 155 Ma
1009-2
538B
$(^{87}Sr/^{86}Sr)_o = 0.7155 \pm 0.0056$
2.0
1.5
1.0
0.5
20
40
60
80
$^{87}Sr/^{86}Sr$
$^{87}Rb/^{86}Sr$

ing for its position off of the linear trend. All other samples, showing a smaller degree of disturbance, were included in the regression, which yields an age of 1559 ± 155 Ma with initial $^{87}Sr/^{86}Sr = 0.7155 \pm .0056$.

The large error in the age reflects greater disturbance of the isotopic systematics than in the Middagstind syenite. This is consistent with the fact that all but one of the samples (1013A) has some amount of tectonite fabric development. Two possible interpretations of this age are considered: (1) the igneous crystallization of the Melaa Granite was earlier, perhaps as part of the petrographically similar Gullesfjord Gneiss, which gave a Rb/Sr whole rock age of 2660 ± 120 Ma (Griffin and others, 1978), but has been updated by Sr homogenization during metamorphism; or (2) the age approximately reflects the igneous crystallization of the Melaa Granite.

The high initial ratio determined here is not compelling evidence for metamorphic updating. The Lødingen Granite, which may be equivalent to the Melaa Granite (see below), has an initial $^{87}Sr/^{86}Sr$ of $0.7118 \pm .0015$ (Griffin and others, 1978), which overlaps the error range of the Sr initial ratio determined for the Melaa Granite in this study. Being true granites, both could result from anatexis of preexisting crust, thus explaining their high content of radiogenic Sr.

Field observations incline the author to favor the second interpretation. North of Hesjevann (Plate 1), there is an area of nearly one square kilometer of fabricless, undeformed Melaa Granite (Figure 41). If this were equivalent to the Gullesfjord Gneiss, it has survived undeformed several phases of Precambrian deformation accompanied by high amphibolite (sillimanite grade) to granulite facies metamorphism, which succeeded in penetratively deforming all of the equally massive granitoid rocks of west Hinnøy (Hakkinen, 1977). No known pre-Svecofennian rocks from Lofoten escaped the Svecofennian deformation; only the later intrusions of the mangerite suite (such as the Middagstind Quartz Syenite) were undeformed. Consequently, an age of crystallization younger than about 1750 Ma seems very likely.

The age determined for the Melaa Granite in this study, 1559 ± 155 Ma, overlaps error ranges of the two age determinations of the Lødingen Granite on Hinnøy (1415 ± 80 Ma, P. Taylor, pers. comm. to Hakkinen, 1977; 1380 ± 80 Ma, Griffin and others, 1978), though considerably older.

This may reflect Rb loss from some whole rock systems. Sample 1009-2, which falls above the regression line, was the most strongly foliated sample analyzed. Because the foliation provides channelways for water, weathering is a problem in foliated samples, and this specimen was found somewhat softer than others during preparation. Consequently, Rb loss by weathering of K feldspar to kaolinite may have occurred in sample 1009-2. If this point is discarded, and the remaining points are regressed together, the result is an age of 1465 ± 100 Ma with initial $^{87}Sr/^{86}Sr$ of 0.7177 ± .0017, considerably closer to the Lødingen Granite determinations.

Consequently, the Melaa Granite is considered to be middle Proterozoic in age and distinct from the Gullesfjord Gneiss. It is possible that it is correlative with the Lødingen Granite, though separated from it by the Austerfjord thrust. Further isotopic age and geochemical studies of these granites are needed to better evaluate this possibility.

Ruggevik Tonalite Gneiss

The Rb/Sr whole rock data collected in Oslo from the Ruggevik tonalite are plotted on an isochron diagram in Figure 72. Three observations about the data are significant. First, the Rb/Sr ratios of the rocks are uniformly low, so that the evolution of Sr isotopic composition with time has been small; consequently, even in the absence of a well-constrained age, some limits can be put on the initial Sr isotopic composition of the systems. This is especially true since the Sr concentrations are high, ranging from 490 to 1250 ppm; large amounts of Sr must be transferred to make major changes in isotopic composition of this unit.

Secondly, the systems are disturbed enough to destroy any convincing isochron relationship. The samples are petrographically variable. This may reflect a composite nature to the intrusion, with internal variability of initial Sr isotopic composition. Thus, some of the scatter may be primary. However, the more likely explanation is that Caledonian events have been adequate to move Rb and Sr on a scale larger than the sample size. The rocks are strongly penetratively deformed and completely recrystallized in Caledonian events.

Linear regression of the data (York Model I) yields an "age" of 720 ± 220 Ma with initial $^{87}Sr/^{86}Sr = 0.7045 \pm .0003$. The significance of the age is questionable at best. It is of the right order to correspond to

TABLE 5: Rb/Sr whole rock analyses from the Ruggevik Tonalite Gneiss. See Figure 69 for sample locations.

Sample	Rb(ppm)	Sr(ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr \pm 2\sigma$
613A	15.8*	830	.055	0.70525 ± .00010
613B	9.28*	600	.045	0.70520 ± .00006
613C	24.6*	1246	.057	0.70471 ± .00011
613D	38.9	774	.143	0.70552 ± .00010
613E	18.0	490	.106	0.70555 ± .00012
780B1	54.7	1052	.150	.070647 ± .00008
780C	39.1	1217	.093	0.70521 ± .00011
1014B	25.9	1039	.072	0.70526 ± .00008

* Determined by isotope dilution: other concentrations by XRF.

- -

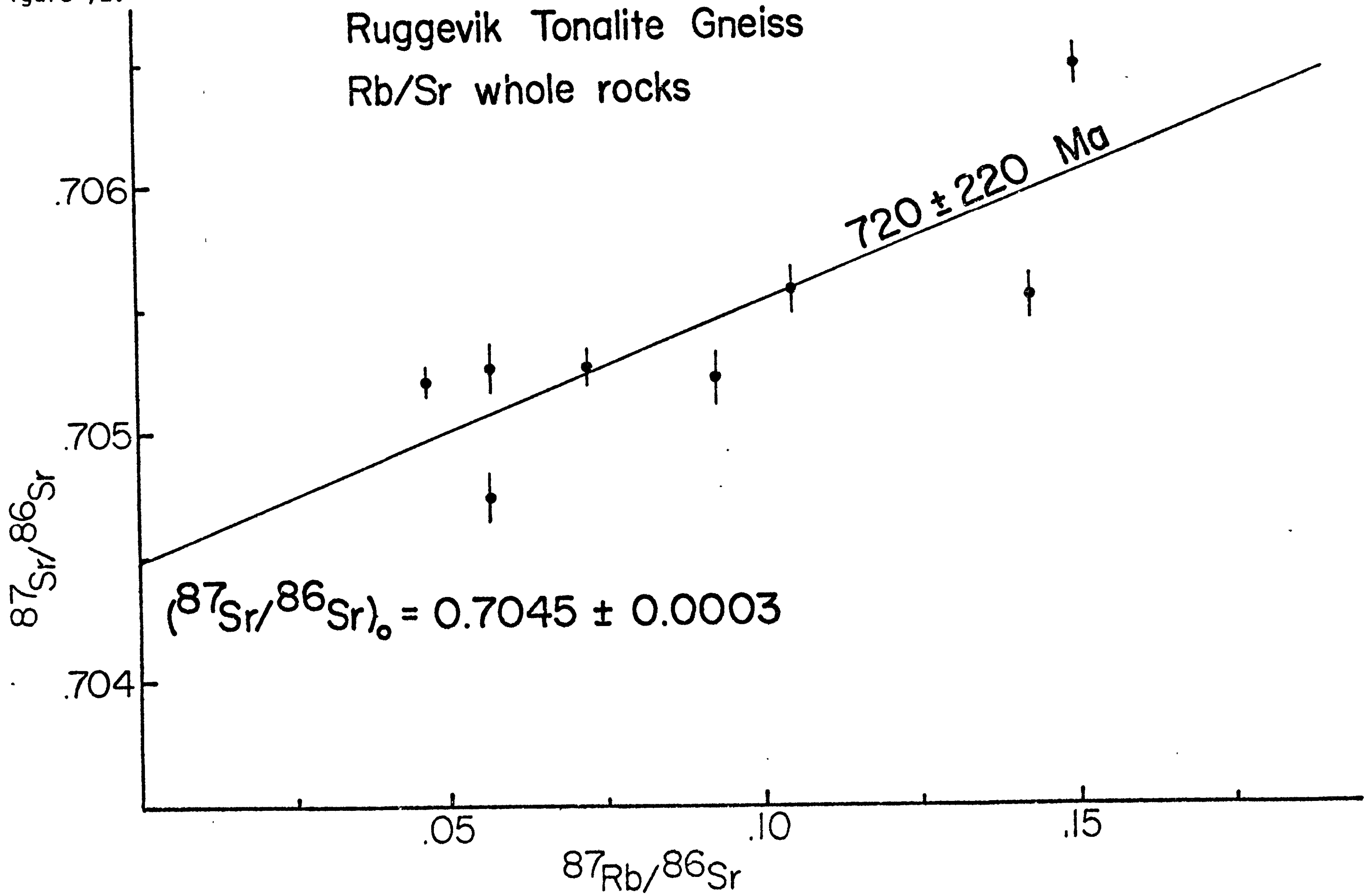
Ruggevik tonalite whole rock analyses done at MIT (all by isotope dilution)

HQ-5	30.5	1207	.0730	0.70476 ± .00004
780C	41.1	1241	.0934	0.70526 ± .00006

TABLE 6: Rb/Sr biotite analyses from the Middagstind Quartz Syenite and Ruggevik Tonalite Gneiss.

Sample	Rb(ppm)	Sr(ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr \pm 2\sigma$	Age(Ma) ± 2σ
26R	203	9.89	59.4	1.03442 ± .0005	361 ± 3
27F	227	5.41	121.4	1.32051 ± .0006	347 ± 4
HQ-5	290	27.9	30.0	0.85760 ± .00015	358 ± 4
780C	311	66.4	13.5	0.77109 ± .00010	362 ± 5

Figure 72.


Ruggevik Tonalite Gneiss
Rb/Sr whole rocks
720 ± 220 Ma
.706
.705
.704
$^{87}Sr/^{86}Sr$
$(^{87}Sr/^{86}Sr)_o = 0.7045 \pm 0.0003$
.05
.10
.15
$^{87}Rb/^{86}Sr$

the early stages of the development of the Iapetus ocean, the closure of which resulted in the Caledonian orogeny. However, given the data this is more speculation than interpretation.

The calculated initial $^{87}Sr/^{86}Sr$ puts a reliable upper bound on the $^{87}Sr/^{86}Sr$ of the rock at the time of crystallization, since any disturbance is likely to increase rather than decrease the value obtained. While clearly too high to argue for a source in depleted, ocean ridge type mantle, it is very low for granitic rocks developed in Precambrian continental crust (c.f. Kistler and Peterman, 1973; Armstrong and others, 1977; Early and Silver, 1973). The most likely setting for the formation of the Ruggevik tonalite, on the basis of the relatively non-radiogenic Sr isotopic composition, the lithology itself, and the high Sr concentrations, is in an island arc, with or without some continental basement. Slightly radiogenic Sr isotopes are typical of island arc settings (e.g., Faure, 1977, p. 116). As noted in Chapter 2, tonalitic plutons are typically developed in convergent margin arcs in either oceanic or very young continental crustal basement.

The high bulk Sr content of these (and most) tonalites is an unresolved petrogenetic problem. Any residual feldspar in the source region, or any feldspar crystal fractionation, will prevent Sr concentration in the magma from approaching the 1250 ppm Sr commonly observed, given reasonable crystal/melt partition coefficients and bulk Sr concentration in the system (1000-1500 ppm Sr at a maximum). Certainly, given the Sr concentrations typical of continental crustal rocks (e.g., the rocks of the Lofoten terrain typically range 100-500 ppm Sr, according to analyses by Griffin and others, 1978), it is essentially impossible to generate high Sr tonalites from typical continental crust even by total melting of the feldspar component in the source. Consequently, it seems more likely that high Sr tonalites are generated from melting of Sr-rich mafic materials; nonetheless, an eclogite facies mineralogy (plagioclase absent) seems to be necessary.

Biotite Ages

Four biotite separates, two from the Middagstind Quartz Syenite and two from the Ruggevik Tonalite Gneiss, were analyzed for Rb/Sr isotopic systematics with three objectives: (1) to determine if the Middagstind

Syenite, given its lack of Caledonian fabric, had suffered Caledonian thermal effects; (2) to place a younger time limit on Caledonian metamorphism; and (3) to compare the time of closure of the same mineral system from two areas separated by about 10 km to evaluate the uniformity of post-Caledonian cooling history in the study area.

The data, recorded in Table 6, are plotted in Figure 73. Whole rock data from tonalites used here were collected at M.I.T., including a new analysis of sample 780C, in order to eliminate the inter-laboratory analytical problem noted above. The ages calculated from these data, 358 $\pm$ 4, 362 $\pm$ 5, 361 $\pm$ 3, and 347 $\pm$ 4 Ma are considered to show satisfactory consistence. The lower age of sample 27F is not understood, since it is not significantly different in grain size from the other syenite, 26R, and was collected only 1 km away.

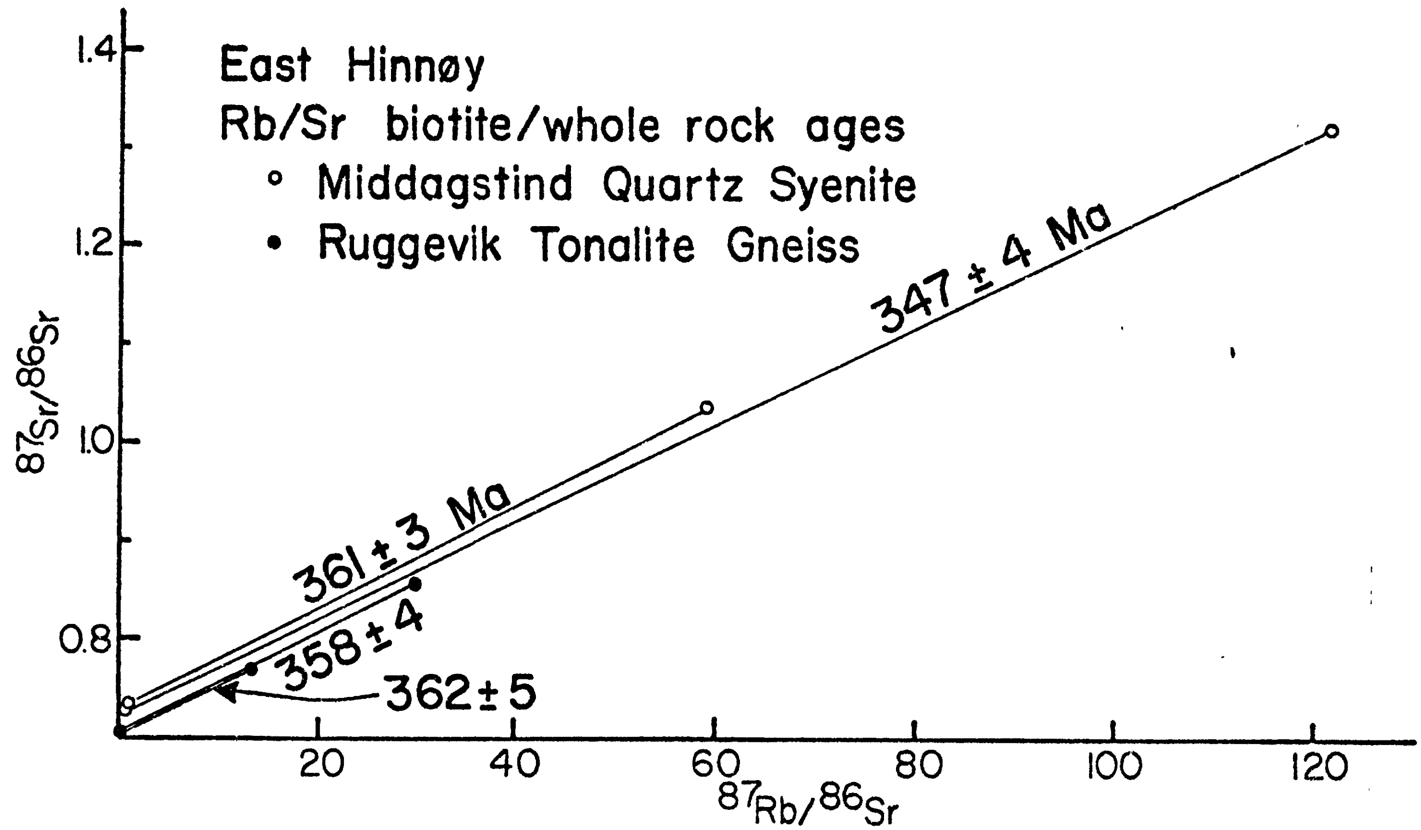
The late Devonian ages from the biotites in the Middagstind syenite indicate that these basement rocks experienced temperatures above a minimum of about 280° C during Caledonian events (Hodges and others, 1980). The interpretation that the shear zones at Middagstind, which formed at amphibolite facies conditions, developed in Caledonian D_2 (see Chapter 3) implies considerably higher temperatures were reached. These isotopic data constitute an independent check that elevated temperatures did develop in the pre-Caledonian basement at Middagstind during Caledonian time.

Similarly, the relatively low closure temperature of the biotite Rb/Sr system calculated by Hodges and others (1980) implies the actual thermal peak of Caledonian metamorphism was significantly earlier. This is consistent with stratigraphic observations further south in the Scandinavian Caledonides (Nilsen, 1973; Gee, 1975) which indicate that Caledonian deformation was completed in early Devonian time. These ages thus appear to reflect post-kinematic uplift of the area. The consistency of these ages over the breadth of the study area, and their further consistency with K/Ar mica ages 60 km to the south (Hodges and others, 1980), suggests this uplift was essentially a rigid, isostatically controlled process rather than one that involved significant deformation.

Deformation and the Disturbance of Rb/Sr Whole Rock Systems

Compston and Raaheim, along with others, in empirical studies of the resetting of Rb/Sr whole rock systems by metamorphism (e.g., Roddock and

Figure 73.



East Hinnøy
Rb/Sr biotite/whole rock ages
° Middagstind Quartz Syenite
• Ruggevik Tonalite Gneiss
347 ± 4 Ma
361 ± 3 Ma
358 ± 4
362 ± 5
1.4
1.2
1.0
0.8
$^{87}Sr/^{86}Sr$
20
40
60
80
100
120
$^{87}Rb/^{86}Sr$

Compston, 1977; Raaheim and Compston, 1977; Gabrielsen and others, 1979), have emphasized the importance of deformation and fluid mobility in the movement of Rb and Sr (especially Sr). In the present study, no evidence of resetting of whole rock systems has been observed, but disturbance of the systems varies from mild to intense. The degree of disturbance of Rb /Sr systematics in the plutonic rocks studied on east Hinnøy appears to be grossly proportional to the intensity of Caledonian deformation, but not related in a simple way to fluid mobility. For instance, water permeated the entire Middagstind syenite to produce thorough retrograde metamorphism without seriously disturbing the Rb/Sr systematics on the whole rock scale.

It is possible that the actual mechanism of movement of volatile components, e.g., diffusive versus infiltrative movement (Korzhinsky, 1952), differs depending on the presence or absence of accompanying penetrative deformation. The dominant transport mechanism under static conditions (e.g., grain boundary diffusion?) may not significantly affect mobility of non-volatile components such as Rb and Sr. By contrast, Rb and Sr may move much more rapidly if a fluid phase carrying non-volatile solutes is moved infiltratively through a rock. This latter mechanism might operate when a rock is deforming penetratively, even if interconnected pore space is not present, since rock volumes are displaced with respect to one another. Thus, fluid bearing Rb and Sr might be displaced through the rock in response to deformation at a rate much larger than simple diffusion could transport the Rb and Sr.

Alternatively, the change in apparent mobility of Rb and Sr may be a result of enhancement of solid state diffusion by plastic deformation of mineral grains. Introduction of additional dislocations by deformation will greatly enhance the rate of volume diffusion by crystal defect-controlled mechanisms.

It is not certain which of the above effects is more important in the enhancement of mobility of Rb and Sr by deformation. However, it seems certain that static metamorphism alone, regardless of fluid movement, does not adequately affect mobility of non-volatile species like Rb and Sr to strongly affect isotopic systematics on the whole rock scale.

CHAPTER 6: REGIONAL TECTONICS

Tectonic Context of East Hinnøy

Introduction

The Scandinavian Caledonides comprise the eastern half of a two-sided orogen, extending 1500 km from southwestern to northwestern Norway, and eastward into western Sweden. The structural vergence is dominantly toward the Precambrian Baltic Shield. The western half of the orogen is preserved in east Greenland, the Cenozoic opening of the Norwegian Sea having split the orogen longitudinally along its axis (Talwani and Eldholm, 1977; see Figure 2). In east Greenland, Caledonian structural vergence is mainly westward (Henriksen and Higgins, 1976). Dewey (1969) suggested that the presence of allochthonous rocks of oceanic affinities in Norway indicated that the Scandinavian Caledonide orogen formed as a result of closure of a proto-Atlantic ocean (the Iapetus ocean of Harland and Gayer, 1972), culminating in the collision of the Greenland and Baltic cratons in Siluro -Devonian time. In Scandinavia, the collision caused thrusting of large sheets of metasedimentary and metavolcanic rocks from oceanic and continental margin settings, together with varying amounts of pre-Caledonian continental basement, onto the Baltic craton. Subsequent uplift of coastal Norway exposed the Baltic continental basement underlying the allochthons. Consequently, the Caledonian allochthons in Scandinavia are essentially enormous klippen in a synformal region between the Baltic shield and the coastal basement exposures ("western gneiss terrain") of Norway (Gee, 1975a).

Locally, transverse upwarping has produced a series of windows and half windows through the nappe pile (Figure 1). The Rombak window, one of the largest of these, appears directly to the southeast of the study area (Figure 3), exposing (par-)autochthonous granite gneisses of the Baltic craton in a position underneath the Caledonian nappe stack.

Position Within the Collisional System

Burchfiel and Davis (1968) noted that in two-sided orogenic belts, the sense of vergence of one of the thrusted terrains is the same as that of the subduction boundary ("synthetic" thrusting), while the other has the opposite sense of vergence ("antithetic" thrusting). Thus, the location and sense of subduction during convergence and collision is fundamental to understanding the tectonic implications of relationships observed in the Scandinavian Caledonides, since it is vital to know where within the collisional system specific relationships developed in order to contribute to understanding of the system as a whole.

Gee (1975a) has shown that in the central Scandinavian Caledonides, the surface trace of the subduction zone must presently lie west of the present Norwegian coastline. The present study has demonstrated that this relationship is also correct at the latitude of Hinnøy (see below, and Chapter 3, "Austerfjord Thrust").

Several lines of evidence suggest that this subduction was westward under Greenland, making the thrusting of the Scandinavian Caledonides synthetic to the subduction.

A feature typical of convergent plate margins is the presence of a magmatic belt, commonly of calc-alkaline affinites, in the overriding plate. No igneous rocks of Caledonian age are known to intrude the Baltic continental basement anywhere in the Scandinavian Caledonides; all Caledonian intrusions occur in allochthonous rocks and were probably emplaced before the allochthons reached their present position upon Baltic continental basement. By contrast, Siluro-Devonian magmatism is widespread, though of undetermined volumetric significance, in the east Greenland Caledonides (Coe, 1975; Steiger and others, 1979). Consequently, it appears that if there was a continental margin magmatic belt, it was in east Greenland.

Subduction produces a depressed thermal gradient in the downgoing plate (e.g., Oxburgh and Turcotte, 1970), resulting in relatively high P metamorphism (e.g., Ernst, 1971). Metamorphism in the Scandinavian Caledonides is mainly of the intermediate pressure, kyanite/sillimanite type of Miyashiro (1961). However, the presence of the mineral assemblage kyanite + garnet + biotite (Chapter 4) indicates that the metamorphism is of a relatively high pressure variety of the kyanite/sillimanite type, corresponding to "bathozone 5" of Carmichael (1978). Furthermore, there are possible indications of true high P metamorphism in the western gneiss terrain of Norway. Brueckner and Griffin (1980) reported five Sm/Nd garnet-clinopyroxene ages of 407 to 447 Ma from eclogite bodies in the western gneiss terrain of Norway. These authors interpret the metamorphism to have occurred in situ within the gneisses, thus indicating high P metamorphism may have occurred in the Baltic continental basement in this area in Caledonian time.

In the Trondheim region (central Scandinavian Caledonides) rocks of oceanic petrochemical affinities (Gale and Roberts, 1972, 1974) occur at the highest preserved structural levels of the orogen in Scandinavia. Horne (1979) has presented evidence that these rocks comprise an island arc which was underthrusted synthetically by the Baltic craton in Silurian time.

Consequently, the evidence consistently supports the view that the rocks now exposed in the Scandinavian Caledonide orogen were deformed as a synthetic thrust terrain developed in the underthrusted plate of a collisional orogen. The structural development recorded in the study area primarily records the effects upon the downgoing continental crust and its associated cover rocks during the collision of two continental masses at a convergent plate boundary.

Tectonic Model for the Caledonian Orogeny in the Lofoten-Rombaken Transect

Regional Comparison of Deformational Sequences

The segment of the Scandinavian Caledonides from the Lofoten terrain on the west across to the (par-)autochthonous gneisses of the Baltic craton exposed in the Rombak window is termed for convenience the Lofoten-Rombaken transect. The present study area on east Hinnøy is located near the middle of this transect (Figure 74), at the contact between the Lofoten terrain (western gneiss terrain) and the major Caledonian allochthons. Figure 75 summarizes and compares the sequences of Caledonian structural events from this and other structural studies in the area of the Lofoten-Rombaken transect (Hakkinen, 1977; Tull and Hodges, work in progress; Olesen, 1971) and suggests correlations of events. Designations in the following discussion (D_1, D_2, etc.) will be those used for events on east Hinnøy (see Chapter 3), except where otherwise noted.

The earliest event in the study area, the emplacement of the allochthons upon the Baltic craton, appears to be expressed on west Hinnøy only by early mesoscopic isoclinal folding in the Austerfjord Group (F_0 of Hakkinen,

Figure 74: Locations of structural studies to date in the Lofoten-Rombaken transect. Ha: Hakkinen, 1977; B: this study; T: Tull and Steltenpohl, unpublished; Ho: Hodges, in progress; O: Olesen, 1971.

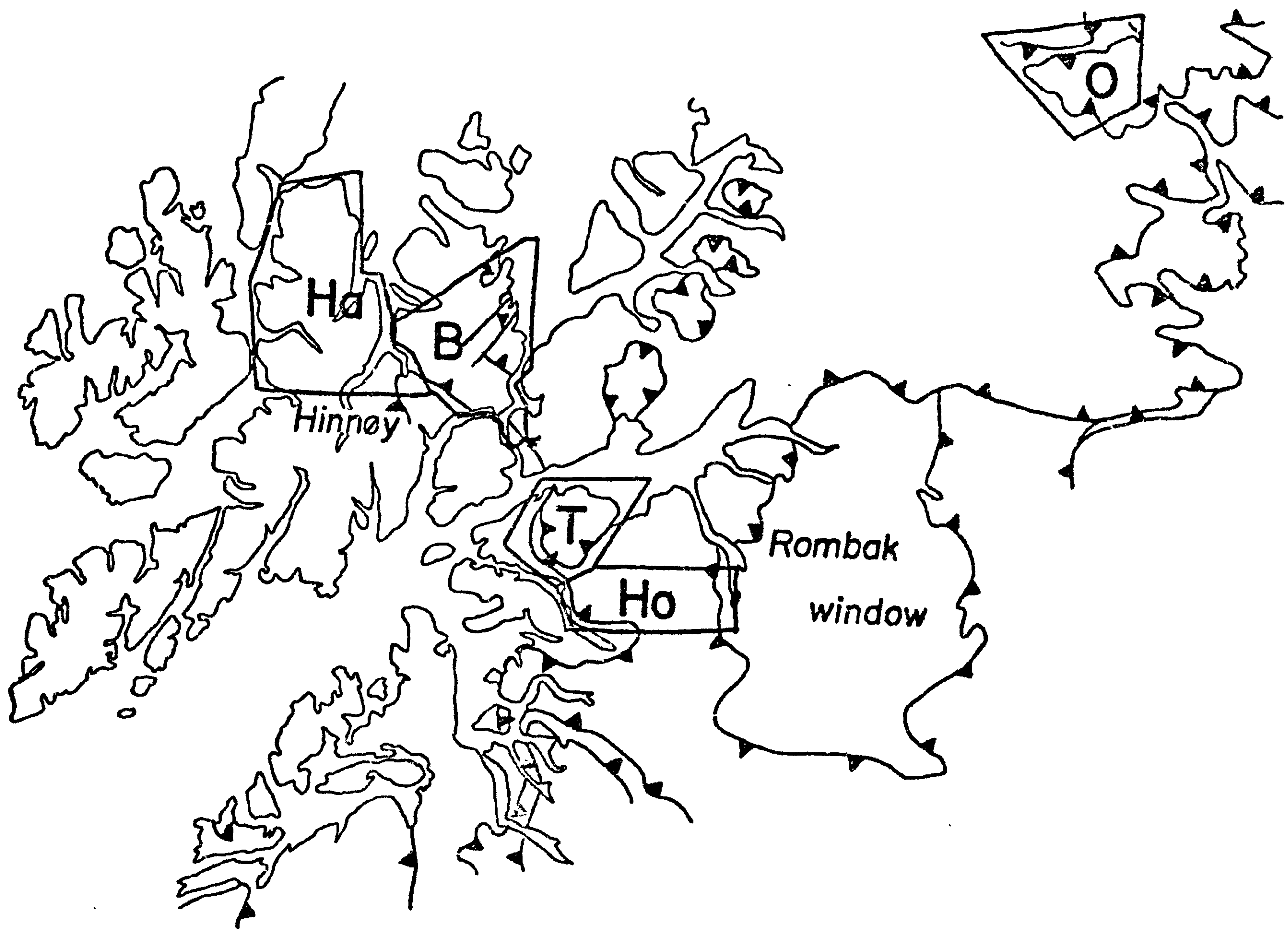
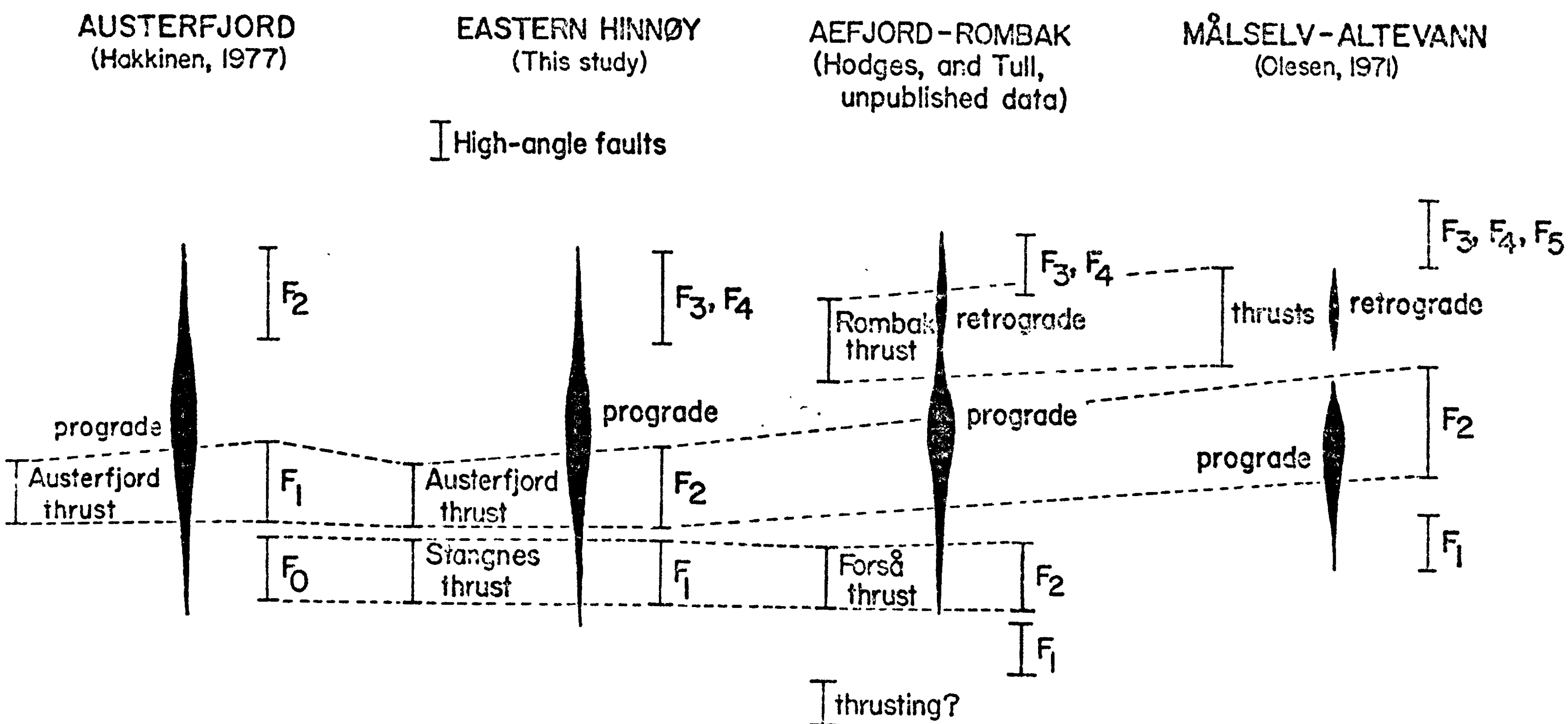


Figure 75: Comparison of structural/metamorphic histories from studies in the Lofoten–Rombaken transect (see figure 74 for locations).

Width of strip in middle of each diagram is expression of grade of metamorphism. Dashed lines indicate suggested correlations of events.



1977). Tracing in reconnaissance structures of D_1 age from east Hinnøy to Hodges' study area at the Tysfjord culmination indicates that the emplacement of the allochthons upon the western gneiss terrain was roughly synchronous in the two areas. Hodges (pers. comm., 1979, 1980) has reported evidence of earlier folds and thrusts which developed elsewhere and are allochthonous along the D_1 thrusts. No equivalent of these older events has been recognized on east Hinnøy.

As discussed in Chapter 3, movement on the Austerfjord thrust and the F_1 fold phase of Hakkinen (1977) are both considered to belong to D_2 of the present study. No clear evidence of structures attributable to this deformation has been reported from the area of the Tysfjord culmination.

Correlation of early events described by Olesen (1971) from the Maalselv area to those described here is speculative, given the large intervening area which has been mapped only in reconnaissance. It is likely that Olesen's F_2 fold phase, which produced south vergent recumbent folds on a scale of kilometers under kyanite grade metamophic conditions, is in some way related to the present D_2 deformation, but need not be strictly synchronous with it. This would imply that the S_1 fabric and rare F_1 folds of Olesen's study may have formed during the initial emplacement of the nappes and correlate with D_1 on Hinnøy. In the Maalselv area, porphyroblast growth occurred before the completion of F_2, while porphyroblasts grew mainly after the D_2 folding on Hinnøy. Whether this records diachroneity of deformation, metamorphism, or (most likely) both, is not certain.

At least one period of thrusting is present in the Tysfjord and Maalselv areas that has no apparent equivalent on Hinnøy. The thrusts that rim the Rombak window, and the thrusts which bound Olesen's upper and lower nappes in the Maalselv area are included in this event. These thrusts post-date Caledonian amphibolite facies metamorphism, retrograde mineral assemblages to greenschist facies along the thrust zones, and truncate earlier structures (Gustavson, 1966; Olesen, 1971; author's own observations north of Narvik). The inclusion of all these thrusts in a single phase is a result of ignorance. When the area between the north edge of the Rombak window and the Maalselv area is better studied, diachroneity in late thrusting is likely to be substantiated.

The absence of the post-metamorphic thrusts west of the Rombak window requires these thrusts to root into the western gneiss terrain basement, such as Gee (1975a, b) shows for several thrusts in the central Scandinavian Caledonides. Some authors (e.g., Nicholson and Rutland, 1969) have assumed equivalence of the post-metamorphic thrusts of the eastern areas to synmetamorphic thrusts further west. In the Lofoten-Rombaken transect, the post-metamorphic thrusts clearly retrograde the same metamorphic mineral assemblages which formed during the synmetamorphic thrusting. There is little question the two events are distinct in time. Consequently, where the late thrusts dip westward at the edge of the Rombak window, these thrusts must pass downward into the basement, making the western gneiss terrain allochthonous in this event.

Correlation of late fold phases on a one-to-one basis on this scale is probably not realistic and has not been seriously attempted in this study. Milnes (1974) has discussed the local variability of late fold development in the central Pennine Alps. The local nature of F_4 and F_5 on east Hinnøy has been noted in Chapter 3. Late folds in general represent the buckling and further thickening of the tectonic stack in response to continued convergence after the ending of nappe tectonics. The structural complexity of the rocks being deformed at this stage is obvious from the complexity of the preceding events. Corresponding complexity in the mechanical response of this system to stress is likely. Consequently, significant differences in late stage strain history from one location to another should be expected.

Generalized Tectonic Model

Figure 76 illustrates the regional tectonic model hypothesized for the Lofoten-Rombaken transect on the basis of present knowledge, in the form of NW-SE "cartoon" cross-sections. Figure 76A illustrates the emplacement of the allochthons upon the Baltic craton, including the Lofoten terrain. Autochthonous cover rocks (Storvann Group) are locally preserved, but otherwise have been stripped and transported eastward near the leading edge of the imbricate nappe stack. The allochthons were emplaced under amphibolite facies conditions. It is not clear at what time the subjacent pre-Caledonian basement of the Baltic craton reached thermal equilibrium with the overriding nappes, but the amphibolite facies metamorphism and basement involvement in the ductile deformation during D_2 indicates that thermal equilibrium was at least approached by D_2 time. No evidence of significant thermal in-

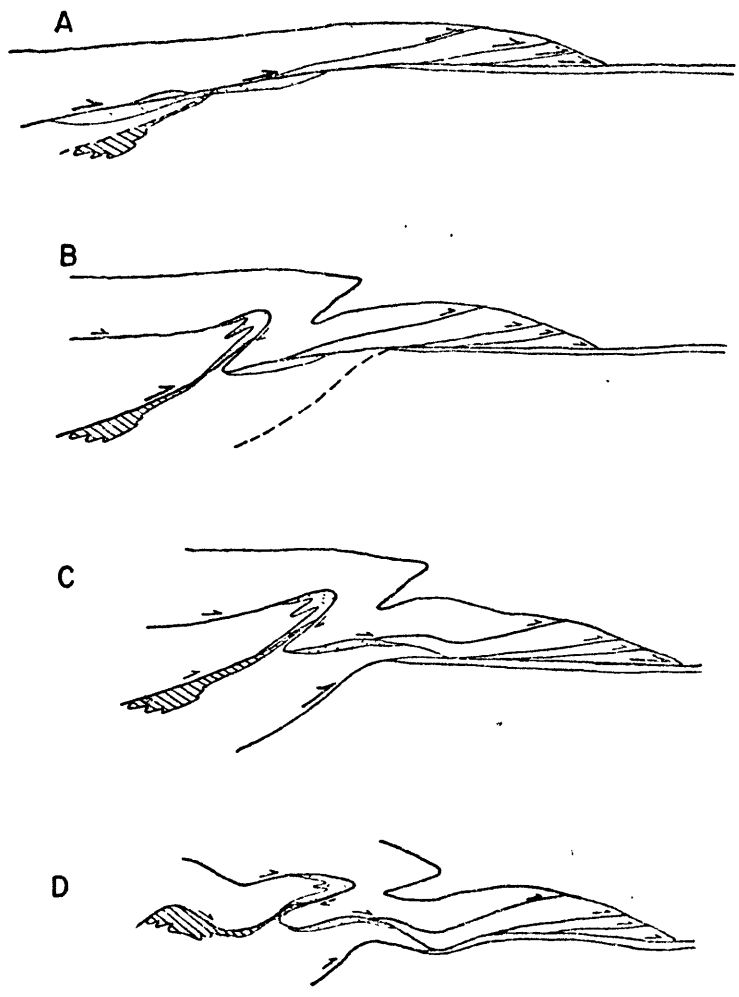
Figure 76: Inferred Caledonian structural/tectonic development of the Lofoten-Rombaken transect.

A: Emplacement of allochthons as imbricate thrust sheets. Note local preservation of autochthonous sedimentary cover (Storvann Group - stippled).

B: Failure of basement along relatively weak Austerfjord Group rocks (ruled), forming Austerfjord thrust and recumbently folding cover and basement.

C: Further thrusting of basement and its stuctural cover along late metamorphic thrusts.

D: Refolding of structural stack.



version was preserved.

Figure 76B shows the faulting of the Lofoten basement terrain to produce the Austerfjord thrust and the resulting recumbent infolding of the autochthonous and allochthonous cover rocks with the basement. This produces local inversion of the nappe sequence, and brings the Lofoten terrain to a higher structural position. It is important to note that this structural elevation of the basement post-dates the emplacement of the nappes and does not imply a major structural discontinuity between the Lofoten terrain and the Baltic craton. The primary subduction zone into which the major nappes root must lie west of the Lofoten block and consequently west of the modern shoreline of Norway.

Figure 76C shows the further eastward translation of the nappe complex together with the western gneiss region along the late metamorphic thrusts. Further imbrication of the eastern portion of the nappe terrain occurred at this time, but its geometrical relationship to earlier structures is unknown. Note that the rooting of the late thrusts into the basement implies ramping of the thrusts, which in turn leads to formation of a basement -cored ramp anticline in the upper plate. The uplift of the western gneiss terrain in coastal Norway may thus be as a ramp anticline produced by the rooting into the basement of regionally extensive late thrusts, a geometry directly analogous to recent interpretations of the Alleghenian thrusts of the central and southern Appalchians of the U.S. (Harris, 1979; Cook and others, 1979), and to the interpretation of the external massifs of the Swiss Alps presented by Hsü (1979).

Figure 76D shows schematically the refolding of the nappe terrain by late deformation phases. Although these folds are probably primarily the expression of continued continental convergence, strike-slip movements may also be expressed at this stage as well (see discussion of F_3 below).

In terms of the plate tectonic context introduced in the previous section, Figures 76A-C can be viewed as outward-migrating imbricate thrusting along the upper surface of a rigid, downgoing slab. This is directly analogous to the development of accretionary prisms at oceanic subduction zones (e.g., Seely, Vail, and Walton, 1974), or the formation of marginal decollement thrust belts (e.g., Bally, Gordy, and Stewart, 1966; Harris and Milici, 1978). Much of the structural development in Norway occurred at elevated temperatures, so that the second-order features of the structural style are the results of penetrative ductile flow rather than brittle slip along discrete fault planes. However, the large scale geometry is strikingly similar. In foreland thrust belts, earlier thrusts are deformed by the ramp anticlines of later thrusts breaking through underneath. The D_2 recumbent folding of the Caledonian nappe pile on east Hinnøy, while accentuated by superposed ductile strain, can be viewed as having been initiated by the folding of the nappe-emplacing D_1 thrusts around the ramp anticline of the Austerfjord thrust. The uplift of the western gneiss terrain has already been noted to have probably resulted in the same way. It is equally likely that the uplift of the basement to form the structural windows and transverse culminations within the nappe terrain (e.g., the Rombak window) is a consequence of the rooting into the basement of the most external thrusts of the Scandinavian Caledonides to form ramp anticlines along the central axis of the belt.

Nappe Transport Distances

Constraints on the distances the nappes in the Lofoten-Rombaken transect have traveled are very poor. No rocks can be matched from the autochthon to any of the allochthons, or between the allochthons. Given this limitation, it is impossible to demonstrate that presently vertically stacked rock bodies were once side by side, although this was probably true. Any attempt to quantitatively unstack the nappes in the Lofoten -Rombaken transect at present would thus be purely speculative. However, the total amount of movement is likely to have been large.

The composition of the Storvann Group metasedimentary sequence, i.e., quartzose terrigenous clastic rocks with laterally continuous carbonate horizons, implies a platformal or miogeoclinal depositional environment. However, without fossil control or some other means of estimating the thickness and stratigraphic continuity of the Storvann Group, one cannot distinguish between these two possibilities. As a consequence, no inferences can be made about the position of east Hinnøy relative to the pre-Caledonian continental edge. Thus, one can only say the edge of the continent was located somewhere west of the modern limits of the Lofoten terrain.

Arguments were given in Chapters 2 and 5 for an ensimatic origin for the Stangnes Group. This puts the minimum transport distance of these rocks at 60 km, their distance from the present limit of the Lofoten terrain.

However, an oceanic basin of some width presumably existed between the Baltic craton and the island arc (?) in which the Stangnes Group formed. Hodges (written comm., 1980) has found evidence that the Narvik Group pelites, which structurally underlie the Stangnes Group, may have been deposited in such an oceanic basin. Consequently, the distance travelled by the Stangnes Group may have been many times greater than the 60 km minimum estimate.

In western Sweden where stratigraphic control exists, Gee (1975a, b) has documented as much as 500 km of structural overlap by Caledonian thrust faults, an estimate which may be conservative. Given that the rocks investigated represent the closure of an oceanic basin, there is every reason to believe that hundreds of kilometers of total nappe transport occurred in the Lofoten-Rombaken transect as well.

Timing of Nappe Emplacement

The "classic" Caledonian orogeny in southern Scandinavia is placed in Siluro-Devonian time on the basis of stratigraphic data (Gee, 1975a). However, in recent years Sturt and various coworkers (Sturt and others, 1967, 1975, 1978) have emphasized the importance of the "Finnmarkian phase" of the Caledonian orogen, a period of complex deformation and plutonic activity of Cambro-Silurian age recognized in the allochthonous rocks of Finnmark (northernmost Norway). These have also presented stratigraphic evidence of pre-latest Ordovician (Ashgillian) deformation in allochthonous rocks south and east of Bergen (more than 1000 km from Finnmark), which they consider also to belong to the Finnmarkian phase (Sturt and Thon, 1976; Sturt and others, 1978). In Sturt's view (oral comm., 1979), the Finnmarkian phase was responsible for most of the penetrative deformation and medium to high grade metamorphism in the Scandinavian Caledonides. For instance, the events included in D_1 and D_2 of this study would be attributed by Sturt and coworkers to the Finnmarkian phase.

Two lines of evidence argue against this interpretation. All present geochronologic data from the Tysfjord transect support a Siluro-Devonian metamorphism. K/Ar systems from amphiboles from the Narvik Group near the Tysfjord culmination give ages of 430 to 450 Ma; calculations of the temperature of closure of the amphibole systems with respect to diffusion of Ar indicates that the minerals crystallized below their closure temperature, so that the age reflects the actual time of amphibolite facies recrystallization in these rocks (Hodges and others, 1980). J. Sutter (pers. comm. to Hakkinen, 1977) has determined an age 390 $\pm$ 6 Ma from biotite growing in the schistosity related to the D_2 Austerfjord thrust, using the $^{40}Ar/^{39}Ar$ stepwise release method. K/Ar mica ages from the Tysfjord area (Hodges and others, 1980), and Rb/Sr whole rock/biotite ages from the present study (Chapter 5) fall in the range 380 to 350 Ma. Consequently, a metamorphic peak in Silurian time, with cooling occurring through the Devonian, is indicated.

Sturt and coworkers have not considered the possibility that the deformations attributed to the Finnmarkian phase may have occurred far from the Baltic craton and never involved rocks of the Baltic continent. In neither Finnmark not the Bergen area have the pre-late Ordovician deformations been traced into rocks which were demonstrably part of the Baltic continent in pre-Caledonian time. For example, Sturt and others (1978, Fig. 6) show Finnmarkian phase intrusions cutting autochthonous Baltic crystalline basement, but no such relationship has been observed in the field.

From a more conceptual point of view, in pre-Silurian time the Iapetus ocean separated the Greenland and Baltic cratons, with at least one intervening island arc (Horne, 1979). Within this convergent system, deformation and metamorphism of ensimatic rock sequences would be expected to occur. These rocks would be later incorporated into the nappes emplaced on the Baltic continent during collision. The early-formed structures would have no contemporaneous counterparts on the Baltic continent, nor would they even be necessarily time correlative from one place to another.

Consequently, the deformation phases recognized in the present study area, which clearly do involve pre-Caledonian Baltic crystalline basement, are considered to be Siluro-Devonian in age. Earlier structures, allochthonous along Silurian thrusts, may be present in the Narvik Group (Hodges, pers. comm., 1979, 1980). The author remains skeptical that at present any relationship can be established between these structures and a unified laterally continuous Finnmarkian orogenic phase of Cambro-Ordovician age in the Scandinavian Caledonides.

Tectonic Significance of F_3 Crossfolds

The kinematics of the F_3 crossfolds are puzzling in the context of the overall structural picture. The commonly observed transverse orientation of early folds in the Scandinavian Caledonides has been satisfactorily explained by demonstrating that the direction of tectonic transport was at a small angle to the present fold axial orientation (Hansen, 1971; Bryant and Reed, 1969; see also Chapter 3, p. 132). Consequently, the transverse early folds are not in conflict with tectonic movements being primarily directed perpendicular to the trend of the orogen. The transverse orientation of F_3 folds in the present study cannot be similarly explained. The sliplines of F_3 folds are steeply inclined. The folds are nearly perfectly homoaxial, and the surface folded was subhorizontal prior to folding (Chapter 3). Consideration of this geometry leads to the conclusion that the maximum principal shortening was very likely subperpendicular to the axial orientation of the folds and consequently subparallel to the trend of the orogen.

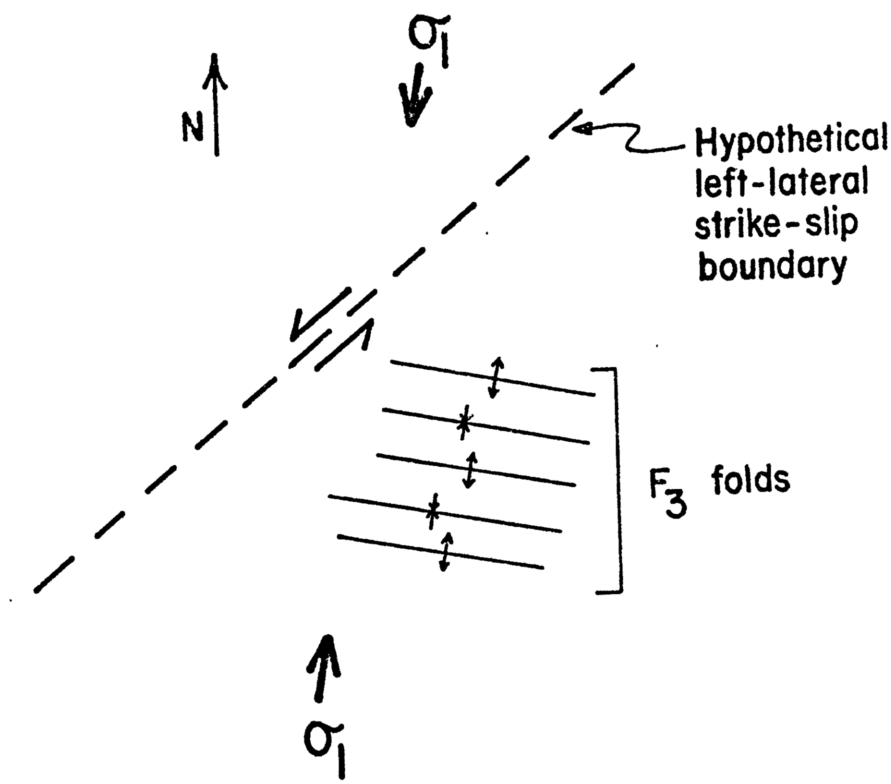
The F_3 folds thus record a major change in the orientation of the strain field in this area. One explanation of such a phenomenon would be a transition from simple convergence to "transpression" (Harland, 1971), a combination of strike-slip and convergent motion along a plate boundary. As shown in Figure 77, if the F_3 folds were produced by transpression, the sense of strike-slip would be left-lateral.

Several independent lines of evidence support the presence of late stage left-lateral movements in other parts of the Caledonide orogen. In western Norway, several basins of Devonian continental clastic sedimentary rocks are syn-depositionally bounded by both thrust and high-angle faults. Relations of these basins suggest that the high-angle faults are left -lateral strike-slip faults, possibly of major displacement (Nilsen, 1973; pers. comm., 1979).

In the British Caledonides, the Great Glen Fault has long been known as a major Devono-Carboniferous strike-slip fault. Although there is some disagreement regarding the magnitude and sense of displacement, most workers prefer left-lateral movement with a cumulative displacement of as much as 300 km.

In the northern Appalachians, paleomagnetic data from Upper Devonian continental redbeds indicate a Devonian paleolatitude for New England as much as 20° south of that for Upper Devonian rocks of the adjacent North American platform (Kent and Opdyke, 1978; Vander Voo and others, 1979).

Figure 77: Schematic map view of relationship of F_3 folds to hypothetical left-lateral strike-slip system.



This indicates the possibility of very large left-lateral strike-slip displacements in this southern part of the Caledonide orogen.

The suggestion that the F_3 crossfolds represent left-lateral transcurrent shear remains speculative in the absence of corroborative evidence for left-lateral displacements in the northern Caledonides. However, the kinematic pattern of the folds must somehow reflect the large-scale tectonic pattern; this possibility deserves further investigation.

Timing and Significance of the Cenozoic (?) High-Angle Faults

Due to the absence of local control on the timing of the high-angle faults in the study area (see Chapter 3), an age for the faults can only be suggested from regional considerations. Two periods of major high-angle faulting which post-date Caledonian metamorphism are known from Norway. The first is the Devono-Carboniferous strike-slip faulting mentioned above; the second is early Cenozoic in age (Oftedahl, 1960) and related to rifting of the Caledonian orogen which led to opening of the Norwegian Sea (Talwani and Eldholm, 1977).

While the orientation of the high-angle faults on east Hinnøy is compatible with either period of faulting, the NW side down dip-slip movement on the faults (Chapter 3) is more consistent with the faults being related to Cenozoic rifting. This is also consistent with the suggestion above that the middle Paleozoic strike-slip movements are expressed by F_3 folds on east Hinnøy.

The presence of relatively young high-angle faults in this vicinity is also corroborated by observations from Andøya, the island north of west Hinnøy (Figure 3). A small area of Upper Jurassic (Oxfordian) to Lower Cretaceous (Valanginian, and Hauterivian (?)) sandstones and shales is present near the north end of the island (Ørvig, 1960). A cross-section by Vogt (1905) reproduced in Ørvig (1960) shows vertical faults within and bounding the Mesozoic rocks.

Tectonic Implications

Geometry of the Nappe Sequence

Previous tectonic syntheses of various parts of the Scandinavian Caledonides have emphasized the importance of thrust nappe tectonics, but have assumed very simple geometrical relationships. Nicholson and Rutland (1969), Nicholson (1973), Gee and Wilson (1974), and Gee (1975a) assumed that the stacking of the Caledonian nappes occurred in a single period of thrusting, leading to a layer-cake stack of sheets. Sturt and others (1978) and Binns (1978) in synthesizing the tectonics of the northern Scandinavian Caledonides recognized the importance of multiple events during nappe emplacement, but retain simple, layer-cake geometrical models of the nappe sequence (Sturt and others, 1978, Figure 6; Binns, 1978, Figure 1).

The present study has documented inversions of the nappe sequence on the scale of several kilometers. Such evidence opens the possibility that much larger scale inversions may have occurred. Binns (1978) noted that the rock assemblages he designated nappes 5 and 7 were indistinguishable, including some rather unique lithologic types ("sagvandites"; see Binns, 1975, 1978). Binns thus inferred that the rocks of these nappes must have formed in close proximity despite their complete separation in his interpretation of the nappe sequence. Similarly, Gustavson (1966) noted a similarity between rocks of the Narvik Group, which lie structurally beneath the Salangen Group, and the Niingen Group, which overlies the Salangen Group.

I think the possibility cannot be neglected that these vertical repetitions of similar rock associations may represent the results of a second, superimposed major recumbent folding or thrusting event similar to that produced on a smaller scale by D_2 on East Hinnøy. Milnes (1974) and Milnes and Schmutz (1978) described large scale recumbent folding in the Pennine nappes of Switzerland which led them to completely reinterpret the vertical sequence of the nappes prior to the refolding. The same type of relationships seem to be appearing in the nappe terrain of north Norway. Multiple nappe-forming events overprinting one another, producing extremely complex large scale geometrical relationships, may thus be typical of intermediate levels of continental collision zones.

Basement Involvement, Decollement Tectonics, and the Deeper Structural Levels of Collisional Orogens

Hakkinen (1977) was the first to document the progressive disappearance of Caledonian structures into the pre-Caledonian basement (Lofoten terrain) of Hinnøy. The limited extent of this deformation is further documented in this study (Chapter 3). In order to account for this disappearance of Caledonian structures, Tull (1977) hypothesized that the Lofoten basement terrain might be a "zwischengebirge" (Kober, 1928; in modern tectonic terms, something like a microcontinent), and Hakkinen (1977) argued that the Lofoten block was a high level nappe. The present study has rendered both these hypotheses untenable by demonstrating that no major zone of shortening exists between the Lofoten terrain and the Baltic craton, and that the two were coextensive in pre-Caledonian time. Thus, the Lofoten block is exposed in a window through the nappe terrain, and the absence of Caledonian structures in much of the basement terrain represents a <u>downward</u> and possibly lateral disappearance of penetrative Caledonian deformation.

It has already been noted earlier in this chapter that there are important geometrical similarities between the structural style of the Lofoten -Rombaken transect and the decollement tectonics of foreland thrust belts. There is also a most important additional similarity: the deformation, within the amphibolite facies metamorphic core of the orogen, dies out downward into a rigid, non-involved basement. The upper surface of the pre-Caledoian crystalline terrain is strongly involved, to be sure. However, as discussed in Chapters 3 and 4, the development of a Caledonian structural overprint in the pre-Caledonian basement seems to be dependent on the availability of water from an outside source to permit retrograde metamorphic reactions to occur. At deeper crustal levels of the downgoing plate in a collisional orogen, there is no likely source of water; consequently, there is no metamorphic overprint and no new tectonite fabric, except perhaps in cases where solid-solid reactions can occur (e.g., the Caledonian (?) eclogite facies metamorphism reported from the Precambrian basement of western Norway by Brueckner and Griffin, 1980).

This has two major implications for the dynamics of collisional orogenic belts. Bally (1975) introduced the term "megasuture" to describe the portion of the surface of the earth where the lithosphere does not behave rigidly, but deforms in a ductile fashion. Bally included in the megasuture all areas where metamorphism or penetrative deformation occurs, since these imply non-rigid behavior. However, the results of this study indicate that precisely as in marginal fold and thrust terrains, metamorphic fold belts can be surficial phenomena, only involving the upper 15 to 20 km of the continental crust in penetrative deformation and recrystallization. At deeper levels, the lower crust and lithosphere still behave essentially as a rigid plate. Consequently, from the viewpoint of the dynamics of the lithosphere/aesthenosphere system, the plate tectonic assumption of rigid blocks with narrow boundaries may be accurate even <u>within</u> parts of broad orogenic belts such as the Scandinavian Caledonides. The broad zones of deformation, even where metamorphic conditions prevail, may be limited to the upper 20 km of the crust.

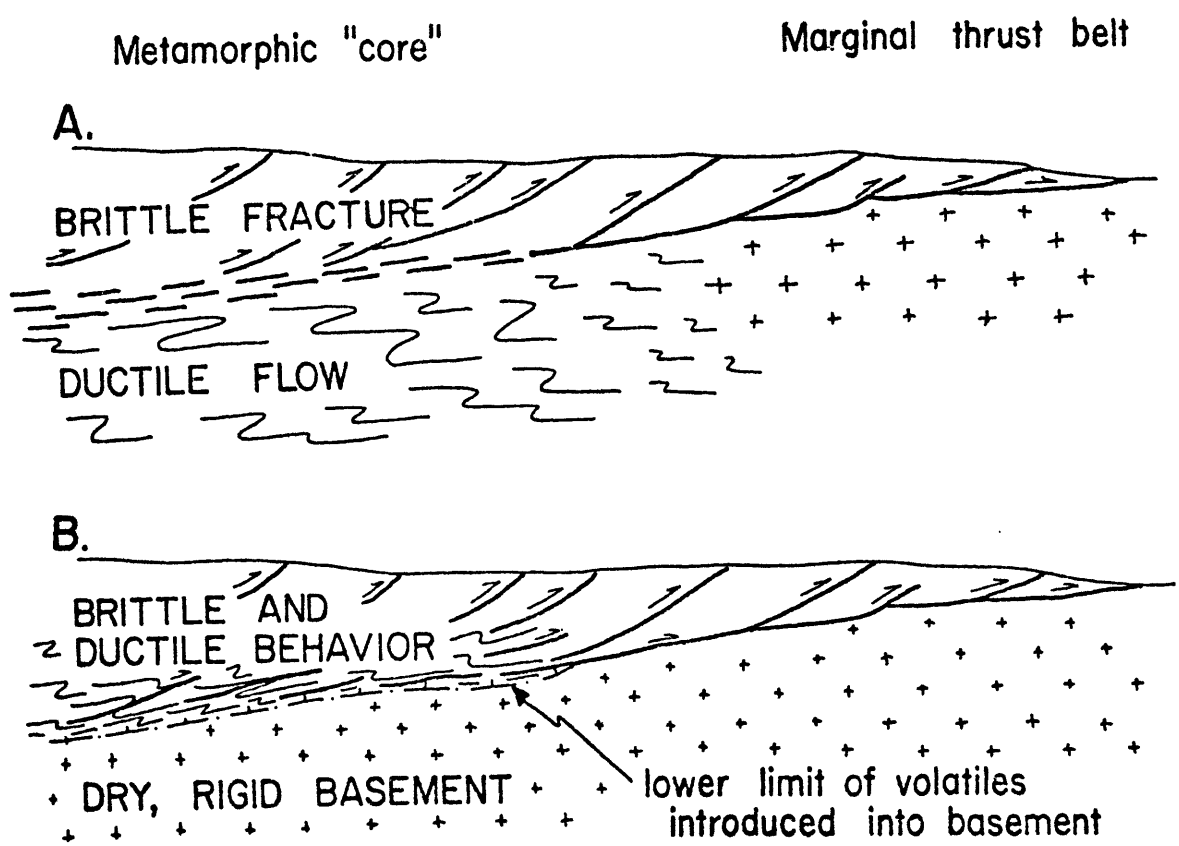
This may also illuminate a widespread problem in the structural geology of thrusted terrains. It is a common observation in orogenic belts around the world that crystalline thrust sheets involve rocks of only about the upper 15 km of the crust. This has been popularly attributed to a zone of brittle/ductile transition where brittle rocks above the transition are thrusted while rocks at higher temperatures and deeper levels flow penetratively (Figure 78A). This study suggests that essentially the opposite relationship may be the correct explanation in at least some cases. The level of basement detachment may be located in the metamorphic internal zone of an orogen by a level at which high strain gradients are produced, due to the limited extent of volatiles available to promote the recrystallization of basement rocks (Figure 78B). Below the detachment, the basement is not deforming ductilely; it may not deform at all. Once this level of strain concentration is established, it will propogate outward into lower temperature zones where it will be expressed as brittle thrusting rather than distributed shear. Displacement along this detachment will eventually bring crystalline rocks from the high temperature internal zone out into thrust contact upon the external zone.

This alternative model for the formation of basement-involved thrusts is considered to better explain the relationships in the Scandinavian Caledonides than the brittle/ductile transition model. It may prove inappropriate to basement-involved antithetic thrusting in Andean type orogens (cf. Armstrong and Dick, 1974; Burchfiel and Davis, 1975). However, this model may have general applicability to synthetic thrust terrains developed is collisional orogens. In particular, both Armstrong and Dick, and Burchfiel and Davis appeal to local high temperatures to cause detachment of crystalline thrust sheets. In a synthetic thrust belt developed in a collisional orogen, no external heat source, such as a magmatic arc, exists. Consequently, this model in which the level of detachment is chemically

Figure 78: Comparison of models for formation of crystalline thrust sheets.

A: Brittle/ductile transition model(e.g., Armstrong and Dick, 1974): High strain zone is localized by geometrical incompatibility of strain fields in brittle and ductile zones of deformation, leading to decollement at the transition between these structural domains. Note that this model implies downward extension of deformation and metamorphism to unspecified levels, perhaps beyond the Moho.

B: Volatile-controlled decollement model (this study): High strain zone is localized at limit of penetration of volatiles (mainly water) generated by prograde metamorphism of sedimentary rocks. Below this level, high temperature and stress are inadequate to deform crystalline basement in the absence of additional volatiles. This leads to decollement at the limit of volatile introduction, and produces a geometry directly analogous to that seen in marginal thrust belts.



controlled (by the availability of water), is more appropriate to the thermal structure of synthetic thrust belts.

APPENDIX: CHEMICAL PROCEDURES FOR ISOTOPIC ANALYSES

Mineralogisk Museum, Oslo

300 to 500 mg of powdered rock was decomposed in concentrated Merck Suprapur HF in a teflon beaker. The $^{87}Rb/^{84}Sr$ mixed spike solution, when used, was added during weighing prior to decomposition. Following decomposition, the sample was dissolved in concentrated HNO_3, then repeatedly in 6N HCl to drive off remaining HF and leave the cations as chloride salts. Samples were then picked up in 2.5N HCl and centrifuged in teflon tubes.

A two stage cation exchange procedure was employed. Stage 1 used large diameter columns (resin bed ca. 1 cm in diameter) and 2.5N HCl as solvent, leading to relatively rapid cation movement. Large volumes of elutions were collected (20-25 ml), so that Rb and Sr were concentrated with little loss but only moderate purity. Stage 2 used smaller columns (ca. 0.5 cm diameter) and 1.0N HCl, and smaller cation peaks were collected (4-5 ml). The elutions were collected in pyrex beakers and dried without further chemical procedures.

Rb and Sr samples were picked up with 0.1N HCl and evaporated onto Ta filaments using phosphoric acid as a loading medium. Rb samples were not conditioned prior to running in the mass spectrometer. Sr samples were conditioned in the source chamber of the mass spectrometer by heating for 1 minute at ca. 2.0 amps, and then for 2-3 minutes at about 1 amp.

M.I.T. (Center for Geoalchemy)

50 to 100 mg of powdered sample or mineral separate, with 1 gram of $^{87}Rb/^{84}Sr$ mixed spike solution, were dissolved in a mixture of Vycor distilled HF and $HClO_4$ in a covered teflon beaker. Following decomposition, the samples were dried and converted to chlorides by repeated dissolution in 6.2N Vycor HCl. The samples were picked up in 2.5N Vycor HCl and centrifuged in teflon tubes.

A single pass through cation exchange columns similar to Stage 2 in Oslo (see above) was used, but using 2.5N HCl. as the solvent. The Sr fraction of sample 291 (Middagstind Quartz Syenite) yielded poor analytical results using this procedure, so a second dissolution was passed twice through the cation exchange columns with identical procedures. This eliminated analytical problems.

Elutions from cation exchange columns were collected in teflon beakers. Alkali elutions were converted to sulfates by addition of a small amount of H_2SO_4 prior to drying. Sr elutions were oxidized with $HClO_4$ to remove any contamination by organic material from the cation exchange resin, then converted to $SrNO_3$ by repeated dissolution in HNO_3.

Alkalis were picked up in water and evaporated onto Ta filaments without the use of a loading medium. Alkali samples were not conditioned prior to analysis.

Sr samples were picked up in 0.1N HNO_3 and evaporated onto Ta filaments using Ta_2O_5 as a loading medium. Sr samples, except those from biotite separates, were conditioned by heating for one to two hours at 1.4 amps in an evacuated outgassing chamber. Biotite samples were run unconditioned. Sr isotopic composition was measured on biotite samples after the ^{85}Rb peak became indistinguishable from background. In some cases, an ^{87}Rb contribution to the mass 87 peak could be detected after the disappearance of a recognizable ^{85}Rb peak. This was expressed as a strong monotonic decline in mass 87/mass 86 ratio as ^{87}Rb continued to burn off of the filament. In these cases, these early analyses were discarded from calculations.

REFERENCES

Armstrong, R. L., and Dick, H. J. B., 1974, A model for the development of thin overthrust sheets of crystalline rock: Geology, v. 3, p. 35 -40.

Armstrong, R. L., Taubeneck, W. H., and Hales, P. O., 1977, Rb-Sr and K-Ar geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington, and Idaho: Geol. Soc. Amer. Bull., v. 88, p. 397-411.

Bally, A. W., 1975, A geodynamic scenario for hydrocarbon occurrences: Proc. 9th World Petroleum Cong., Tokyo, v. 2, p. 33-44.

Bally, A. W., Gordy, P. L., and Stewart, G. A., 1966, Structure, seismic data, and orogenic evolution of Southern Canadian Rockies: Bull. Can. Petrol. Geol., v. 14, p. 337-381.

Binns, R. E., 1975, A new sagvandite occurrence near Lyngenfjord, Troms, Norway, and the relationship between Troms sagvandites and the geological history of the area: Acta Borealia, A. Scientia, v. 30, p. 1-16.

Binns, R. E., 1978, Caledonian nappe correlation and orogenic history in Scandinavia north of lat 67°N: Geol. Soc. Amer. Bull., v. 89, p. 1475 -1490.

Boyle, A. P., Griffiths, A. J., and Mason, R., 1979, Stratigraphical inversion in the Sulitjelma area, Central Scandinavian Caledonides: Geol. Mag., v. 116, p. 393-402.

Brooks, C., Hart, S. R., and Wendt, I., 1972, Realistic use of two-error treatments as applied to rubidium-strontium data: Rev. of Geophys. Sp. Phys., v. 10, p. 551-577.

Brueckner, H. K., and Griffin, W. L., 1980, Caledonian Sm/Nd ages of Norwegian eclogites: Trans. Amer. Geophys. Union (EOS), v. 61, p. 389.

Bryant, B., and Reed, J. C., Jr., 1969, Significance of lineation and minor folds near major thrust faults in the southern Appalachians and the British and Norwegian Caledonides: Geol. Mag., v. 106, p. 412-429.

Burchfiel, B. C., and Davis, G. A., 1968, Two-sided nature of the Cordilleran orogen and its tectonic implications: Internat. Geol. Cong., 23rd, Prague, Rept. Sec. 3, Proc., p. 175-184.

Burchfiel, B. C., and Davis, G. A., 1975, Nature and controls of Cordilleran orogenesis, western United States: extension of an earlier synthesis: Amer. Journ. Sci., v. 275-A, p. 363-396.

Carmichael, D. M., 1969, On the mechanism of prograde metamorphic reactions in quartz-bearing pelitic rocks: Cont. Min. Petrol., v. 20, p. 244-267.

Carmichael, D. M., 1978, Metamorphic bathozones and bathograds: a measure of the depth of post-metamorphic uplift and erosion on the regional scale: Amer. Journ. Sci., v. 278, p. 769-797.

Coe, K., 1975, The Hurry Inlet granite and related rocks of Liverpool Land, East Greenland: Grøn. Geol. Unders., Nr. 115, 34 pp.

Cook, F. A., Albaugh, D. S., Brown, L. D., Kauffman, S., Oliver, J., and Hatcher, R. D., 1979, Thin-skinned tectonics in the crystalline southern Appalachians, COCORP seismic reflection profiling of the Blue Ridge and Piedmont: Geology, v. 7, p. 563-567.

De Waard, D., 1965, The occurrence of garnet in the granulite-facies terrane of the Adirondack Highlands: Journ. Petrol., v. 6, p. 161-191.

Dewey, J. F., 1969, Evolution of the Appalachian/Caledonide orogen: Nature, v. 222, p. 124-129.

Early, T. O., and Silver, L. T., 1973, Rb-Sr systematics in the Peninsular Ranges batholith of southern and Baja California: Trans. Amer. Geophys. Union (EOS), v. 54, p. 494.

Ernst, W. G., 1971, Do mineral parageneses reflect unusually high-pressure conditions of Franciscan metamorphism?: Amer. Journ. Sci., v. 270, p. 81-108.

Ernst, W. G., 1973, Interpretive synthesis of metamorphism in the Alps: Geol. Soc. Amer. Bull., v. 84, p. 2053-2078.

Faure, G., 1977, Principles of Isotope Geology: Wiley, New York, 464 pp.

Fleuty, M. J., 1964, Tectonic slides: Geol. Mag., v. 101. p. 452-456.

Foslie, S., 1941, Tysfjords Geologi: Norges Geol. Unders., Nr. 149, 298 pp.

Foslie, S., 1949, Haafjells-Mulden i Ofoten: Norges Geol. Unders., Nr. 174, 129 pp.

Gabrielsen, R. H., Naterstad, J., and Raaheim, A., 1979, A Rb-Sr study of a possible Precambrian thrust zone, Hardangervidda-Ryfylke Nappe Complex, southwest Norway: Norsk Geol. Tidss., v. 59, p. 253-263.

Gale, G. H., and Roberts, D., 1972, Palaeogeographical implications of greenstone geochemistry in the outhern Norwegian Caledonides: Nat. Phys. Sci., v. 238, p.60-61.

Gale, G. H., and Roberts, D., 1974, Trace element geochemistry of Norwegian Lower Palaeozoic basic volcanics and its tectonic implications: Earth and Plan Sci. Letters, v. 22, p. 380-390.

Gee, D. G., 1975a, A tectonic model for the central Scandinavian Caledonides: Amer. Journ. Sci., v. 275-A, p. 468-515.

Gee, D. G., 1975b, A geotraverse through the Scandinavian Caledonides - Østersund to Sulitjelma: Sveriges Geol. Unders., Ser. C, Nr. 717, 66 pp.

Gee, D. G., and Wilson, M. R., 1974, The age of orogenic deformation in the Swedish Caledonides: Amer. Journ. Sci., v. 274, p.1-9.

Green, T. H., and Jorde, K., 1971, Geology of Moskenesøy, Lofoten, North Norway: Norges Geol. Unders., Nr. 270, p. 47-76.

Griffin, W. L., and Heier, K. S., 1969, Parageneses of garnet in granulite facies rocks, Lofoten-Vesteraalen, Norway: Cont. Min. Petrol., v. 23, p. 89-116.

Griffin, W. L., Heier, K. S., Taylor, P. N., and Weigand, P. W., 1975, General geology, age, and chemistry of the Raftsund mangerite intrusion, Lofoten-Vesteraalen: Norges Geol. Unders., Nr. 312, p.1-30.

Griffin, W. L., Taylor, P. N., Hakkinen, J. W., Heier, K. S., Iden, I. K., Krogh, E. J., Malm, O., Olsen, K. I., Ormaasen, D. E., and Tveten, E., 1978, Archaean and Proterozoic crustal evolution in Lofoten-Vesteraalen, North Norway: Journ. Geol. Soc. Lond., v.135, p. 629-647.

Gustavson, M., 1966, The Caledonian Mountain Chain of the Southern Troms and Ofoten areas. Part I. Basement rocks and Caledonian metasediments: Norges Geol. Unders., Nr. 239, 162 pp.

Gustavson, M., 1972, The Caledonian Mountain Chain of the Southern Troms and Ofoten areas. Part III. Structures and structural history: Norges Geol. Unders., Nr. 283, 56pp.

Gustavson, M., 1974a, Berggrunnskart Narvik, 1:250,000: Norges Geol. Unders., Trondheim.

Gustavson, M., 1974b, Harstad. Beskrivelse til det berggrunns geologiske gradteigskart, N9 - 1:100,000: Norges Geol. Unders., Nr. 310, 36 pp.

Gustavson, M., 1974c, Ofoten. Beskrivelse til det bergrunns geologiske gradteigskart, M9 - 1:100,000: Norges Geol. Unders., Nr. 310, 36 pp.

Hakkinen, J. W., 1977, Structural geology and metamorphic history of western Hinnøy and adjacent parts of eastern Hinnøy, North Norway: Ph.D. thesis, Rice Univ., Houston, 161 pp.

Hansen. E., 1971, Strain Facies: Springer-Verlag, New York, 207 pp.

Harland, W. B., 1971, Tectonic transpression in Caledonian Spitsbergen: Geol. Mag., v. 108, p. 27-41.

Harland, W. B., and Gayer, R. A., 1972, The Arctic Caledonides and earlier oceans: Geol. Mag., v. 109, p. 289-314.

Harris, L. D., 1979, Similarities between the thick-skinned Blue Ridge anticlinorium and the thin-skinned Powell Valley anticline: Geol. Soc. Amer. Bull., v. 90, p. 525-539.

Harris, L. D., and Milici, R. C., 1977, Characteristics of thin-skinned style of deformation in the southern Appalachians, and potential hydrocarbon traps: U. S. Geol. Surv. Prof. Pap. 1018, 40 pp.

Heier, K. S., 1960, Petrology and geochemistry of high-grade metamorphic and igneous rocks on Langøy, northern Norway: Norges Geol. Unders., Nr. 207, 245 pp.

Heier, K. S., and Compston, W., 1969, Interpretation of Rb-Sr age patterns in high-grade metamorphic rocks, north Norway: Norsk Geol. Tidss., v. 49, p. 257-283.

Helland, 1897, Lofoten og Vesteraalen: Norges Geol. Unders., Nr. 23, 87 pp.

Henriksen, N., and Higgins, A. K., 1976, East Greenland Caledonian fold belt, in Escher, A., and Watt, W. S. [Eds.], Geology of Greenland: Grøn. Geol. Unders., Copenhagen, p. 182-246.

Hodges, K. V., Bartley, J. M., Krueger, H. W., and Hodges, L. R., 1980, K/Ar and Rb/Sr mineral ages and the cooling history of part of the northern Scandinavian Caledonides: Geol. Soc. Amer. Abs. with Progs., in press.

Holland, T. J. B., and Norris, R. J., 1979, Deformed pillow lavas from the central Hohe Tauern, Austria, and their bearing on the origin of epidote-banded greenstones: Earth and Plan. Sci. Letters, v. 43, p. 397-405.

Honea, E., and Johnson, A. M., 1976, Part IV. Development of sinusoidal and kink folds in multilayers confined by rigid boundaries: Tectonoph. v. 30, p. 197-239.

Horne, G. S., 1979, Melange in the Trondheim Nappe suggests a new tectonic model for central Norwegian Caledonides: Nature, v. 281, p. 267-270.

Hsü, K. J., 1979, Thin-skinned plate tectonics during Neo-Alpine orogenesis: Amer. Journ. Sci., v. 279, p. 353-366.

Jakobsen, S. B., and Wasserburg, G. W., 1978, Interpretation of Nd, Sr, and Pb isotope data from Archean migmatites in Lofoten- Vesteraalen, Norway: Earth Plan. Sci. Letters, v. 41, p. 245-253.

Johnson, A. M., and Honea, E., 1975, Part III. Transition from sinusoidal to concentric-like to chevron folds: Tectonophys., v. 27, p. 1-38.

Kent, D. V., and Opdyke, N. D., 1978, Paleomagnetism of the Devonian Catskill redbeds: evidence for motion of the coastal New England -Canadian Maritime region relative to cratonic North America: Journ. Geophys. Res., v. 83, p. 4441-4450.

Kistler, R. W., and Peterman, Z. E., 1973, Variations in Sr, Rb, K, Na, and initial $^{87}Sr/^{86}Sr$ in Mesozoic granitic rocks and intruded wall rocks in central California: Geol. Soc. Amer. Bull., v. 84, p. 3489 -3512.

Kolderup, C. F., 1898, Lofoten og Vesteraalens gabbro-bergarter: Bergens museums Aarbok VII.

Korzhinsky, D. S., 1952, The distinction between the infiltration and diffusion metasomatic columns in relation to the minerals of variable composition: Doklady Akad. Nauk S.S.S.R., v. 86, p. 596-600.

Kvale, A., 1953, Linear structures and their relation to movement in the Caledonides of Scandinavia and Scotland: Quart. Journ Geol. Soc. Lond., v. 109, p. 51-73.

Malm, O. and Ormaasen, D. E., 1978, Mangerite-charnockite intrusions in the Lofoten-Vesteraalen area, North Norway: Norges Geol. Unders.

Milnes, A. G., 1974, Post-nappe folding in the western Lepontine Alps: Eclog. Geol. Helv., v. 67, p. 333-348.

Milnes, A. G., and Schmutz, H.-U., 1978, Structure and history of the Suretta Nappe (Pennine zone, central Alps) - a field study: Eclog. Geol. Helv., v. 71, p.19-33.

Miyashiro, A., Evolution of metamorphic belts: Journ. Petrol., v. 2, p. 277-311.

Miyashiro, A., 1973, Metamorphism and metamorphic belts: George Allen and Unwin, London, 492 pp.

Moberg, J. C., 1908, Bidrag till kännadomen av de kambriska lagren vid Torneträsk: Sveriges Geol. Unders., Ser. C, Nr 212.

Mueller, R. F., and Saxena, S. K., 1977, Chemical Petrology: Springer -Verlag, New York, 394 pp.

Nicholson, R., 1973, The Scandinavian Caledonides, in Nairns, A. E. M., and Stehli, F. G., [Eds.], The Ocean Basins and Their Margins: Plenum Press, p. 161-203.

Nicholson, R., and Rutland, R. W. R., 1969, A section across the Norwegian Caledonides: Bodø to Sulitjelma: Norges Geol. Unders., Nr. 260, 86 pp.

Nilsen, T., 1973, Devonian (Old Red Sandstone) sedimentation and tectonics of Norway: A. A. P. G. Mem. 19, p. 471-481.

Oftedahl, O., 1960, On supposed marginal faults and the oblique uplift of the land mass in Cenozoic time: Norges Geol. Unders., v. 208, p. 351 -357.

Olaussen, S., 1977, Palaeozoic fossils from Torms, Norway: Norsk Geol. Tidss., v. 56, p. 457-459.

Olesen, N. Ø., 1971, The relative chronology of fold phases, metamorphism, and thrust movements in the Caledonides of Troms, North Norway: Norsk Geol. Tidss., v. 51, p. 355-377.

Ørvig, T., 1960, The Jurassic and Cretaceous of Andøya in northern Norway: Norges Geol. Unders., Nr. 208, p.344-350.

Oxburgh, E. R., and Turcotte, D. L., 1970, Thermal structure of island arcs: Geol. Soc. Amer. Bull., v. 81, p. 1665-1688.

Pankhurst, R. J., and O'Nions, R. K., 1973, Determination of Rb/Sr and $^{87}Sr/^{86}Sr$ ratios of some standard rocks and evaluation of X-ray fluorescence spctrometry in Rb-Sr geochemistry: Chem. Geol., v. 12, p. 127-136.

Ragan, D. M., 1973, Structural geology. An introduction to geometrical techniques (2nd Ed.): Wiley, New York, 208 pp.

Raaheim, A., and Compston, W., 1977, Correlations between metamorphic events and Rb-Sr ages in metasediments and eclogite from Western Tasmania: Lithos, v. 10. p. 271-289.

Ramberg, I. B., and Johnson, A. M., 1976, Part V. Asymmetric folding in interbedded chert and shale of the Franciscan Complex, San Francisco Bay area, California: Tectonophys., v. 32, p. 295-320.

Ramsay, J. G., 1967, Folding and Fracturing of Rocks: McGraw-Hill, New York, 568 pp.

Reches, Z., and Johnson, A. M., 1976, Part VI. Asymmetric folding and monoclinal kinking: Tectonophys., v. 35, p. 295-334.

Robie, R. A., 1966, Thermodynamic properties of minerals, in Clark, S. P. [Ed.], Handbook of Physical Constants: Geol. Soc. Amer. Mem. 97, p. 437-458.

Robie, R. A., and Waldbaum, D. R., 1968, Thermodynamic properties of minerals and related substances at 298.15°K (25.0°C) and one atmoshere (1.013 bars) pressure and at higher temperatures: U. S. Geol. Surv. Bull., No. 1259, 256 pp.

Roddock, C., and Compston, W., 1977, Sr- isotopic equilibration: a solution to the paradox: Earth Plan. Sci. Letters, v. 34, p. 238-246.

Seely, D. R., Vail, P. R., and Walton, G. G., 1974, Trench Slope Model: in, Burk, C. A., and Drake, C. L. [Eds.], The Geology of Continental Margins: Springer-Verlag, New York, p. 249-260.

Smith, R. E., 1968, Redistribution of major elements in the alteration of some basic lavas during burial metamorphism: Journ. Petrol., v. 9, p. 191-219.

Steiger, R. H., Hansen, B. T., Schuler, Ch., Bär, M. T., and Henriksen, N., 1979, Polyorogenic nature of the southern Caledonian fold belt in East Greenland: an isotopic study: Journ. Geol., v. 87, p. 475-495.

Storre, B., and Karotke, E., 1971, An experimental determination of the upper stability limit of muscovite + quartz in the range 7 - 20 kb water pressure: Neues Jahrb. Mineral. Monats., p. 237-240.

Strand, T., 1960, Cambro-Silurian stratigraphy, in, Holtedahl, O. [Ed.], Geology of Norway: Norges Geol. Unders., Nr 208, p. 163-165.

Sturt, B. A., Miller, J. A., and Fitch, F. J., 1967, The age of alkaline rocks from west Finnmark, northern Norway, and their bearing on the dating of the Caledonian orogeny: Norsk Geol. Tidss., v. 47, p. 255 -273.

Sturt, B. A., Pringle, I. R., and Roberts, D., 1975, Caledonian nappe sequence of Finnmark, northern Norway, and the timing of orogenic deformation and metamorphism: Geol. Soc. Amer. Bull., v. 86, p. 710 -718.

Sturt, B. A., and Thon, A., 1976, Discussion: The age of orogenic deformation in the Swedish Caledonides: Amer. Journ. Sci., v. 276, p. 385 -390.

Sturt, B. A., Pringle, I. R., and Ramsay, D. M., 1978, The Finnmarkian phase of the Caledonian orogeny: Journ. Geol. Soc. Lond., v. 135, p. 597-610.

Talwani, M., and Eldholm, O., 1977, Evolution of the Norwegian-Greenland Sea: Geol. Soc. Amer. Bull., v. 88, p. 969-999.

Thompson, J. B., Jr., 1957, The graphical analysis of mineral assemblages in pelitic schists: Amer. Mineral., v. 42, p. 842-858.

Tull, J. F., 1972, The geology and structure of Vestvaagøy in Lofoten, North Norway: Ph. D. thesis, Rice Univ., Houston, 149 pp.

Tull, J. F., 1977, Geology and structure of Vestvaagøy, Lofoten, North Norway: Norges Geol. Unders., Nr. 333, 59 pp.

Turner, F. J., 1968, Metamorphic Petrology: McGraw-Hill, New York, 403 pp.

Van der Voo, R., French, A. N., and French, R. B., 1979, A paleomagnetic pole position from the folded Upper Devonian Catskill redbeds, and its tectonic implications: Geology, v. 7, p. 345-348.

Vogt, J. H. L., 1905, On Andøens Jurafelt, navnlig om landets langsomme nedsynken under juratiden, og den senere haevning samt gravforkastning: Norges Geol. Unders., Nr 43.

Vogt, T., 1909, Om eruptivbergarterne paa Langøen i Vesteraalen: Norges Geol. Unders., Nr. 21, p. 198-213.

Vogt, T., 1918, Geologiske studier langs den østlige del av fjellkjeden i Tromsø amt: Norsk Geol. Tidss., v. 4, p. 260.

Vogt, T., 1922, Bidrag til fjellkjedens stratigrafi og tektonik: Geol. Foren. Forh., v. 44, p. 714.

Vogt, T., 1942, Trekk av Narvik - Ofoten traktens geologi: Norges Geol. Unders., Nr. 21, p. 198-213.

Vogt, T., 1950, Map sheet Narvik, 1:100,000: Norges Geol. Unders., Trondheim.

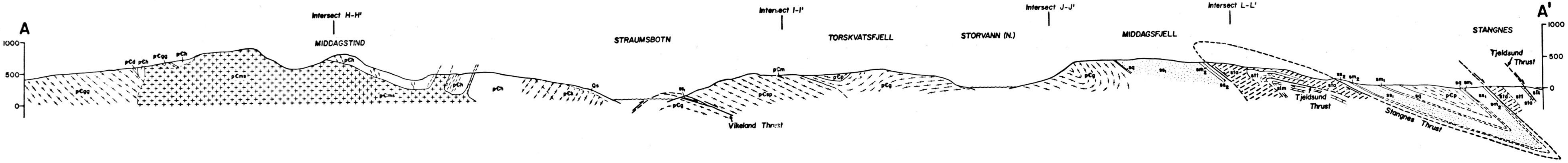
Weiss, L. E., 1959, Geometry of superposed folds: Geol. Soc. Amer. Bull., v. 70, p. 91-106.

Wilson, M. R., and Nicholson, R., 1973, The structural setting and geochronology of the basal granitic gneisses in the Caledonides of part of Nordland, Norway: Journ. Geol. Soc. Lond., v. 129, p. 365-387.

Winkler, H. G. F., 1973, Petrogenesis of Metamorphic Rocks: Springer-Verlag, New York, 334 pp.

York, D., 1966, Least-squares fitting a straight line: Can. Journ. Phys., v. 44, p. 1079-1086.


STRATIGRAPHIC COLUMN
Qs Quaternary sediments, undifferentiated
LOFOTEN TERRAIN
STORVANN GROUP[a]
CAMBRIAN?
sq2 Pelitic schist: garnet-two mica-(kyanite) schist, often rusty-weathering; locally slightly calcareous
sm2 Calcite marble: gray marble with buff-colored bands, slightly micaceous
sq1 Quartz-garnet schist: quartz-rich garnet-mica schist, minor quartzite intercalations
sm1 Calcite marble: gray micaceous marble, locally graphitic or pyritiferous
sq: Micaceous quartzite: two mica-bearing quartzite, with cm-scale layers of vitreous quartzite
sq1: Quartz-biotite schist: dark gray schist, locally garnetiferous
Nonconformity
PRECAMBRIAN
pCg pCgf Melå Granite[a]: medium- to coarse-grained granite and augen gneiss
pCgf: aplitic masses gradational with pCg
AUSTERFJORD GROUP[b]
pCaq Quartzite: white vitreous sericite quartzite
pCaa Amphibolite: calcareous amphibolite, with intercalations of buff-colored marble
pCas Garnet-muscovite schist
pCam Tremolite marble: buff to greenish calcite marble with tremolite porphyroblasts
pCab Biotite schist: fine- to medium-grained quartz-biotite schist, in part graphitic
pCah Hornblende schist: hornblende-biotite schist and schistose amphibolite
pCaf Quartzofeldspathic schist
pCau Schist and amphibolite, undifferentiated
Tectonic contact
pCms Middagstind Quartz Syenite[a]: quartz syenite, with metamorphic coronas of opaque-sphene-biotite-quartz-ferrohastingsite-garnet
pCh Hornfels: hornblende and hypersthene hornfels of contact aureole of Middagstind Quartz Syenite; distinction from pCa to south difficult, contact here arbitrarily located
KVAEFJORD GROUP[a,b]
pCsp Semipelitic paragneiss: two mica gneiss and schist, in part garnetiferous
pCfg Fine-grained quartzofeldspathic gneiss
pCp Hornblende paragneiss: hornblende-clinopyroxene-quartz gneiss, in part calcareous
pCc Granite-pebble conglomerate
pCcs Calcareous schist
pCk Kvaefjord Group, undifferentiated
HESJEVANN ASSEMBLAGE[a,b]
pCa Amphibolite: massive amphibolite, in part biotite-bearing
pCw Calc-silicate rock: milky white, fine-grained massive tremolite-clinozoisite-quartz rock
pCs Schists: quartz-muscovite, garnet-mica, and quartz-biotite-magnetite schist
pCm Marble: buff-colored calc-silicated calcite or dolomite marble
pCq Quartzite: sericite quartzite, feldspathic quartzite; rare quartz-pebble conglomerate
Contact ambiguous
pCgg Gullesfjord Gneiss: granitic to granodioritic orthogneiss, locally difficult to distinguish from Melå Granite
pCmg Archaean(?) migmatitic gneiss
pCd Hornblende diorite
a New name proposed in this study.
b Order of units listed has no age significance.
CALEDONIAN ALLOCHTHONS (in order of tectonic stacking)
SALANGEN GROUP[b]
SILURIAN?
slk Kilbotn Schist[a]: garnet-two mica schist with abundant quartz segregations; locally hornblende-bearing with pods of garnet amphibolite
slsa Pelitic schist and amphibolite: garnet-muscovite-(hornblende) schist and garnet amphibolite with large (2-5 cm) garnets
slb Buff-colored marble: micaceous (phengitic), in part dolomitic, marble; some cm-scale amphibolite stringers
slp Pink-banded calcite marble
slm Calcite marble: gray marble, in part micaceous; rarely tremolite-bearing
slc Harstad Conglomerate: polymict, matrix-supported metaconglomerate; clasts: fine-grained granitoids, quartzite; matrix: quartz-mica schist and calc-schist
sls Schist, undifferentiated
Thrust contact
STANGNES GROUP[a]
HADRYNIAN?
stt Ruggevik Tonalite Gneiss[a]: biotite tonalite gneiss, commonly hornblende-bearing, occasionally muscovite-bearing
sta Stangnes Amphibolite[a]: banded and schistose amphibolite, rarely garnetiferous
Thrust contact
NARVIK GROUP
CAMBRO-SILURIAN?
ng Pelitic gneiss: garnet-kyanite-two mica gneiss and schist with augen-shaped quartz-feldspar aggregates; injected with synkinematic aplitic and pegmatitic dikes; minor amphibolite bodies
Thrust contact
?
cs Calc-schist, marble, minor amphibolite; affinities uncertain
sz Sliver zone: tectonically interleaved schist, marble and amphibolite; extensive boudinage, no cataclasis or retrograde metamorphism
Thrust contact
(LOFOTEN TERRAIN)
HARSTAD
STANGNES
KVAEFJORD
VIKELAND
STRAUMSBOTN
SALENFJELL
MIDDAGSFJELL
STORVANN
RUGGEVIK
KILBOTN
SØRVIKFJELL
SØRVIK
FINNSETHEIA
KEIPFJELL
REVSNES
AUSTERFJORD
HESJEVANN
STORVANN (S.)
VOLDSTADHEIA
RUNDFJELL
TVERRFJELL
SJETNESKARTIND
KONGSVIKTIND
SAETERTIND
GAUSVIK
SANDTORG
KONGSVIK
TJELDSUND
FJELLDAL
RAMSTAD
RAMSUND
RAMBOHEIA
NOT MAPPED IN THIS STUDY
5° 20' E.
68° 46' N.
5° 30' E.
68° 46' N.
5° 40' E.
68° 46' N.
5° 50' E.
68° 50' N.
5° 50' E.
68° 40' N.
5° 30' E.
68° 34' N.
5° 40' E.
68° 34' N.
STUDY AREA
ATLANTIC OCEAN
Arctic Circle
SWEDEN
NORWAY
PLATE I
GEOLOGIC MAP OF EAST HINNØY AND PART OF THE ADJACENT MAINLAND, NORTH NORWAY
GEOLOGY BY JOHN M. BARTLEY, 1977-1979
Topographic base modified from the "Harstad" 1:50,000 sheet, Norges Geografiske Oppmåling
CONTOUR INTERVAL = 60 METERS
All elevations in meters
SCALE:
0 0.5 1.0 2.0 km
N
EXPLANATION OF STRUCTURAL SYMBOLS
Contacts
Primary contact (dashed where inferred)
High-angle fault (dashed where inferred, dotted where concealed)
arrow shows dip
Thrust fault (dashed where inferred, dotted where concealed)
Overturned thrust fault, barbs on present upper plate (dashed where inferred, dotted where concealed)
Tectonic slide (dashed where inferred, dotted where concealed)
Folds and Fabrics
Closed symbols: pre- to synmetamorphic structures (D1 and D2)
Open symbols: late to post-metamorphic structures (D3 and D4)
Antiform, with plunge
upright
overturned
Synform, with plunge
upright
overturned
Strike and dip of schistosity
Crenulation or spaced cleavage
Shear zone (D3 or D4)
Mineral lineation
Stretching lineation
Intersection lineation (S2 and S3)
Line of cross-section
bend in section
Boudin axis
Minor fold axis
upright, or vergence indeterminate
verging fold (here to north)
Intersecting foliations, and resulting lineation



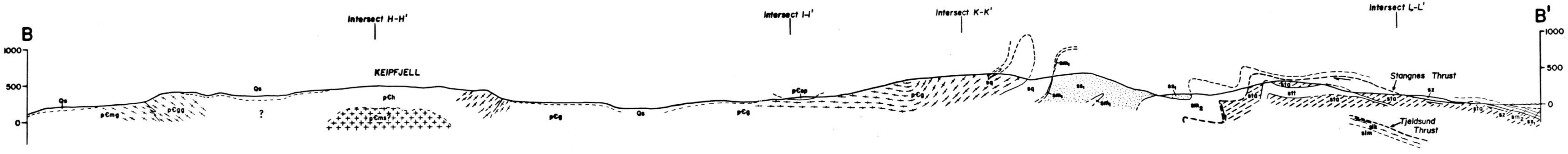
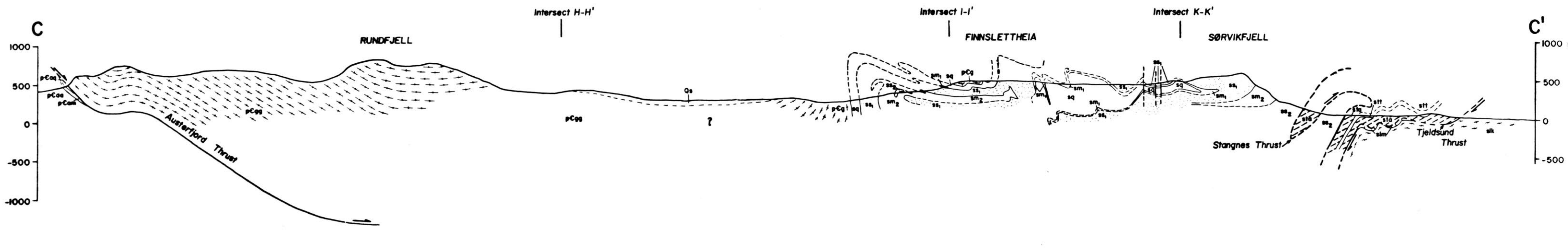
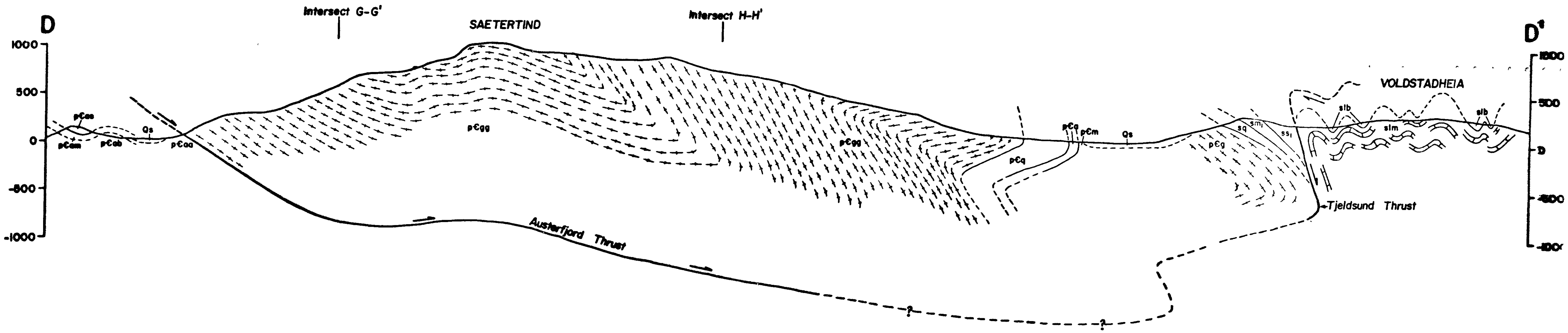


PLATE IIA
GEOLOGIC CROSS-SECTIONS OF EAST HINNØY

SYMBOLS AS IN PLATE I EXCEPT:

- Trace of main foliation in basement not Caledonian in all cases
- Compositional banding in Salangen Group marbles
- Shear zone
- Trace of s_1 in Kilbotn Schist (slk)





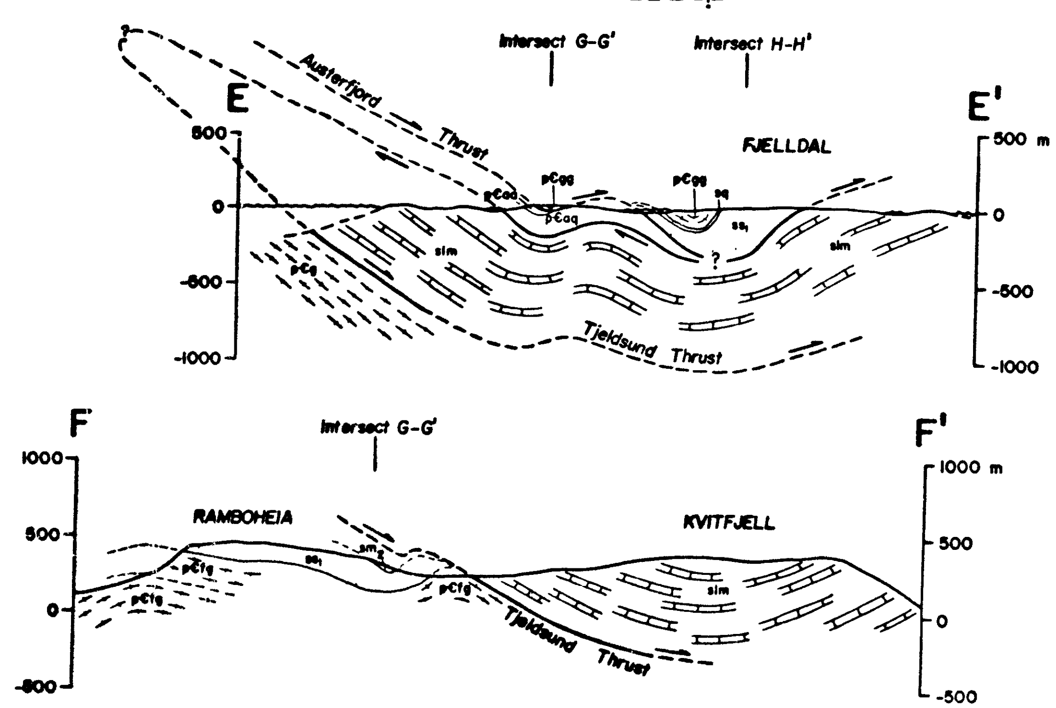


PLATE IIB
GEOLOGIC CROSS-SECTIONS OF EAST HINNØY

SYMBOLS AS IN PLATE IIA

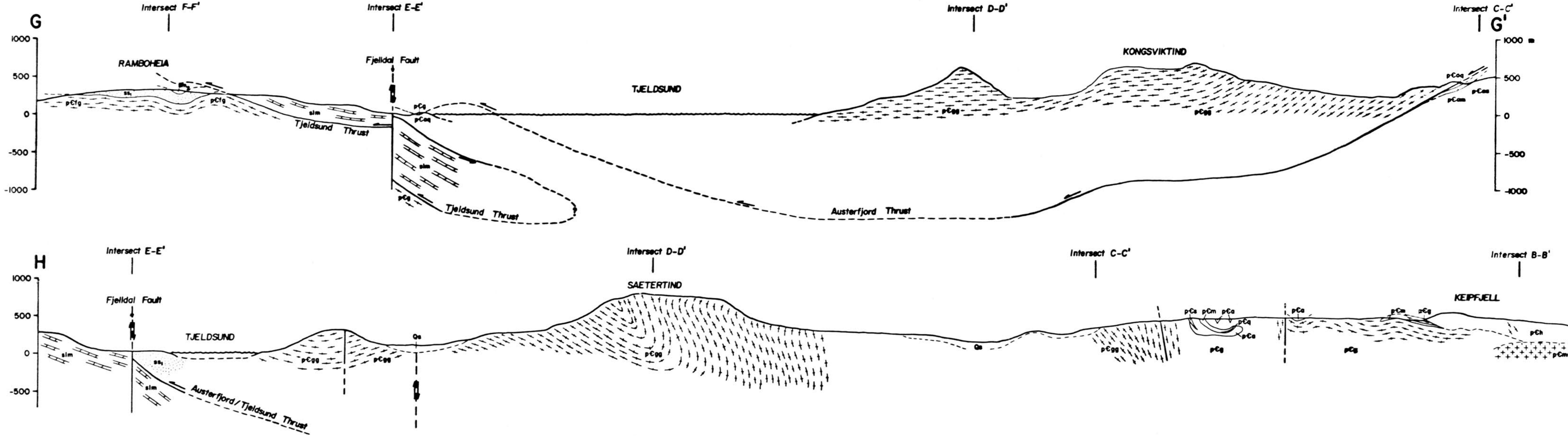


PLATE IIC
GEOLOGIC CROSS-SECTIONS OF EAST HINNØY

SYMBOLS AS IN PLATE IIA

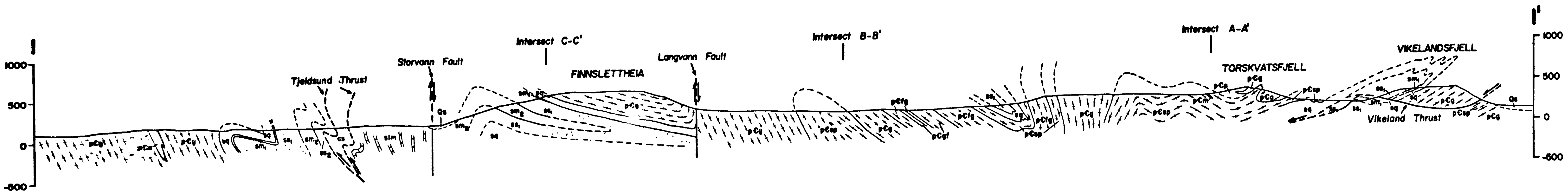
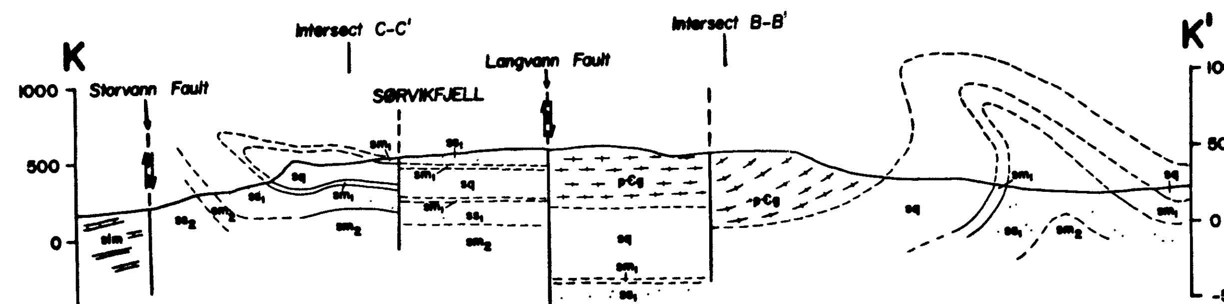
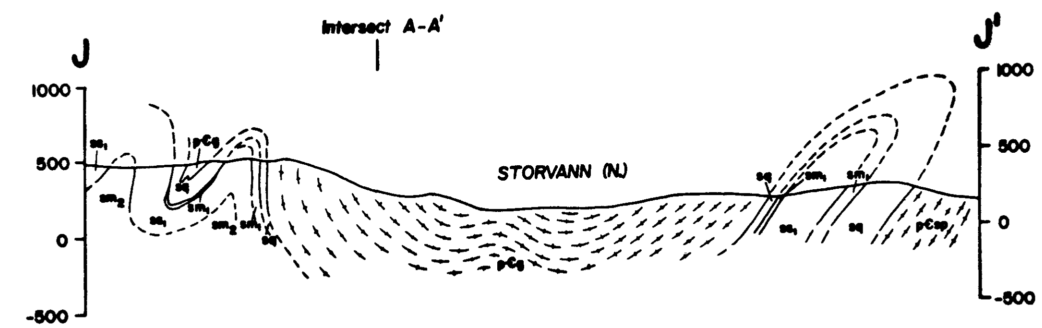
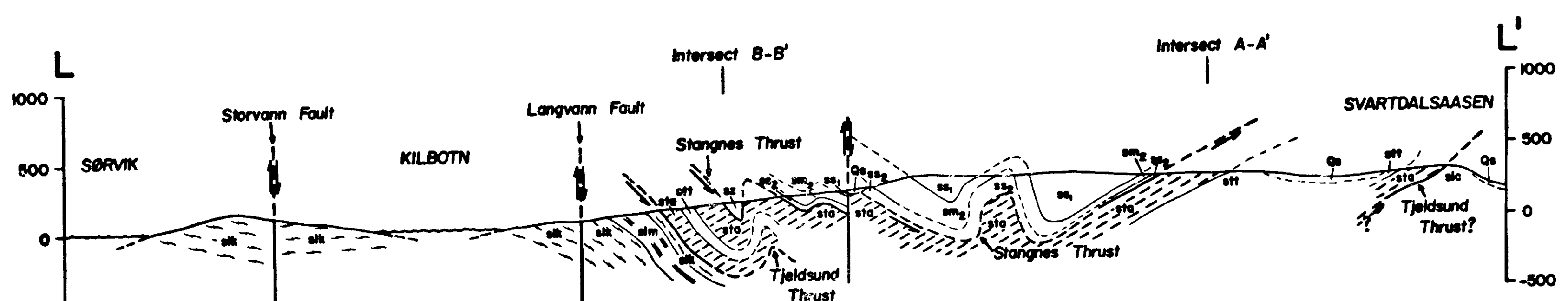


PLATE IID
GEOLOGIC CROSS-SECTIONS OF EAST HINNØY

SYMBOLS AS IN PLATE IIA







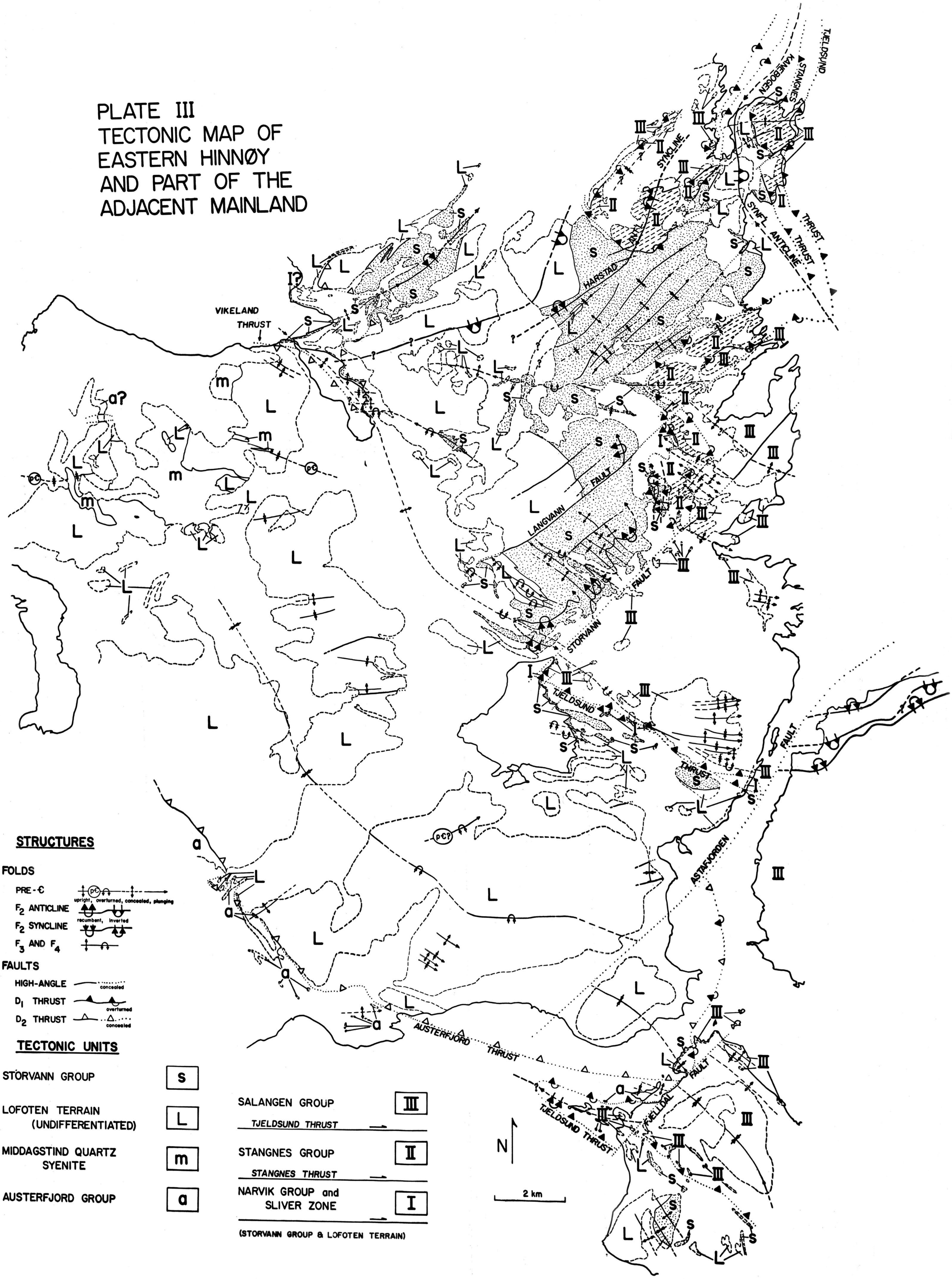


PLATE III
TECTONIC MAP OF
EASTERN HINNØY
AND PART OF THE
ADJACENT MAINLAND
STRUCTURES
FOLDS
PRE-Є
upright, overturned, concealed, plunging
F_2 ANTICLINE
recumbent, inverted
F_2 SYNCLINE
F_3 AND F_4
FAULTS
HIGH-ANGLE
concealed
D_1 THRUST
overturned
D_2 THRUST
concealed
TECTONIC UNITS
STORVANN GROUP
S
LOFOTEN TERRAIN (UNDIFFERENTIATED)
L
MIDDAGSTIND QUARTZ SYENITE
m
AUSTERFJORD GROUP
a
SALANGEN GROUP
III
TJELDSUND THRUST
STANGNES GROUP
II
STANGNES THRUST
NARVIK GROUP and SLIVER ZONE
I
(STORVANN GROUP & LOFOTEN TERRAIN)
N
2 km
VIKELAND THRUST
HARSTAD
SYNCLINE
LANGVANN FAULT
STORVANN FAULT
TJELDSUND THRUST
ASTAFJORDEN FAULT
AUSTERFJORD THRUST
TJELDSUND
KANEBOGEN
STANGNES
THRUST
ANTICLINE

PLATE IV: EQUAL AREA PLOTS OF FOLIATIONS AND FOLD AXES FROM EAST HINNØY

SEE TEXT, FIGURE 28, FOR SUBAREA LOCATIONS