SOME ULTRABASIC BODIES AND RELATED ROCKS
IN SUNNMØRE, SOUTH NORWAY

BY

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CHAPTER I. INTRODUCTION

A. THE GENERAL GEOGRAPHICAL AND GEOLOGICAL SETTING OF THE AREAS STUDIED.

The Sunnmøre district of South Norway is part of the basal gneiss region of N.W. southern Norway. Holtedahl (1944) coined the term 'basal gneiss' to describe this gneiss complex which passes conformably to the E. and S.E. into the main region of Eo-Cambrian (Sparaomite) and Cambro-Silurian metasediments of the Norwegian Caledonides (Fig. 1).

This gneiss complex was considered by early field workers to be Pre-Cambrian but is now generally recognized as being Caledonian in age (Holtedahl 1960, Neumann 1960, Strand 1961, and MacDougall and Green 1964).

Within the layered gneisses of this district there are numerous occurrences of peridotite and eclogite, principally occurring in a broad zone from Almklodalen (62°N, 5°30'E) near Øheim on the coast south of Ålesund, north eastwards to Tafjord (62°13'N, 7°26'E) at the easternmost extremity of Storfjord (Fig. 2).

This study has been concerned principally with the occurrences of peridotite, garnet peridotite and eclogite in the Tafjord district, a study initiated by O'Hara and Mercy (1963).

In this area these rock types are well exposed in an extremely rugged glaciated topography with a relief of up to 6,000 feet, so that access to certain key localities is extremely arduous and often rather hazardous. The area is covered with snow for most of the year hence limiting the effective field season to three months at the most, but on the higher ground to two months or even less. Even during the short summer season low cloud down to about 2,000 feet impedes access to most of the key localities, which lie above 3,000 feet, and restricts one to fjord-side and roadside exposures which are principally/
FIGURE 1

Schematic and simplified geologic-tectonic map of the southern part of the Scandinavian Caledonides, after Strand, 1961.
principally of the regional gneisses.

With an extremely high annual precipitation and an abundance of high mountain lakes, this area contains one of the oldest established networks of hydro-electric power stations in Norway. With the continued construction of dams and tunnels the catchment area and the power generated are being continually increased. These projects have necessitated the construction of minor roads within the Tafjord district, but there is no road access to the outside world. The precipitous nature of the fjord walls has rendered the building of a road into Tafjord an impossible task, particularly in view of the risk of rock avalanches. One particularly calamitous rock fall into the fjord earlier this century resulted in major loss of life by drowning among the inhabitants of the village of Tafjord.

Access to the village of Tafjord is therefore either by ferry along the fjord or on foot over the mountains from Grothli or Geiranger to the south.

Reported occurrences of garnet peridotite inclusions within the Norwegian peridotite masses were restricted to the Tafjord and Almklovdalen districts (Eskola 1921). However, in the course of this study another occurrence of this rock type has been discovered at Ugelvik on the island of Otterøy (62°45'N, 6°40'E) on the north side of Moldefjord in Romsdal, some 45 miles N.W. of Tafjord. It was decided to extend this present study to consider this important locality, despite the fact that it lies outside the main zone of peridotite occurrences. Although not so well exposed as the Tafjord occurrences this peridotite locality is much more accessible, being exposed on level ground at approximately sea level.

During the course of this study brief surveys were also made of the garnet/
garnet peridotite occurrences at Almklovtdalen studied by Lappin (1962), and of the large peridotite mass near Hornindal, Nordfjord (62° N, 6°30' E). At the latter locality peridotite with associated eclogite occurs about 10 kilometers N.E. of Hornindal on the north side of the road to Hellesylt, Sunnylvfjord.

B. THE REGIONAL GEOLOGY OF THE GNEISS COMPLEX - SUMMARY OF PREVIOUS WORK.

Broad outlines of the regional geology have been given in reviews by Holtedahl (1960) and Strand (1961). The basal gneiss complex is considered to be the highly deformed, metamorphosed and metasomatized basement of the Norwegian Caledonian orogenic belt, where rocks of the Pre-Cambrian basement and Caledonian geosynclinal sediments have been intimately welded together. Thus while inclusions of quartzite, limestone, lime silicate and pelite probably represent sediments of Caledonian age it is not unlikely that in the gneiss complex 'Caledonized' remnants of the original Pre-Cambrian basement comprise a significant portion. Such remnants of the basement are assumed to have been infolded and remetamorphosed in the Caledonian orogeny so as now to reflect principally Caledonian structures and grades of metamorphism.

This region is considered as marking a prominent culmination (Hernes 1957) in the Norwegian Caledonide basement, while to the south and south-east the crystalline nappes of the Sogn-Jotunheimen district are piled up in a strongly marked synclinoorium, and to the north large masses of relatively undeformed Cambro-Silurian sediments are preserved in the Trondheim synclinoorium.

Gjelsvik and Gleditsch have produced a highly generalised and in detail/
Figure 2. Geological Map of Sunnmøre and parts of Nordfjord, West Norway. (Gjelsvik 1951).

Note the broad zone of dunite outcrops extending from Almklövdalen on the north side of Nordfjord near Øheim (not marked) and Selje, north eastwards to Tafjord at the easternmost extremity of Storfjord (the major fjord in the central area of the map).
detail inaccurate geological map of the Sunnmøre district (Gjelsvik 1951), but one which is nevertheless useful in that it shows the most important peridotite outcrops (Fig. 2). According to Gjelsvik the gneisses are mainly coarse-grained granodioritic vein gneisses, partly augen gneisses, with discernable horizons of quartzite, limestone and lime silicate gneisses. He recognized two sets of fold movements, a major set of steeply folded synclines and anticlines - often partly overturned - folded about E-W axes along with minor coaxial parasitic folds, and a later set of minor cross folds with approximately N-S axes.

Lappin (1962) has performed a more detailed structural analysis of the gneisses of the Selje and Almklovdalen districts to the south-west, and has derived the following generalized succession of tectonic events:

1. Thrusting and formation of recumbent nappes.
2. Folding about E-W axes - regional fold axis trending E.N.E. and plunging 20°E.
3. Shallow flexural folding about N-S axes.
4. Late thrusting and nappe formation.
5. Posthumous folding.

Lappin considered the evolution of the gneiss complex viewed in the broad light of the Caledonian orogeny to be as follows... A series of pelitic, semipelitic, arkosic and psammite sediments with interbedded basic rocks and limestones were subject to two periods of metasomatism under amphibolite facies conditions. The first, essentially a Na2O metasomatism, was associated with, or succeeded by metamorphic differentiation as well as replacement. The second, K2O metasomatism, was associated with the waning of the tectonism and was characterized by positive growth of microcline. Shearing movements/
movements seem to have been important throughout the history of the complex.'

Muret (1960) has given an interesting structural interpretation of the gneisses of the Romsdal district just to the north-east of Tafjord, drawing analogies with the deep Pennine structures of the Alps. He recognized four main tectonic phases:

1. The piling up of several recumbent folds or nappes.
2. The folding of the nappes about E-W plunging axes.
3. The formation of a large culmination and associated N-S folds which are due to gliding from the culmination.
4. N-S fracturing as a final result of the culmination.

Muret considered the rocks of Sunnmøre and Romsdal as representing the structurally lowest levels of the Caledonian orogeny now exposed, and the major structures as being Pennine-type nappes with axes plunging at moderate angles eastwards.

He outlined one such major structure encompassing the Romsdal and Tafjord districts (Fig. 3) distinguishing within this nappe structure two principal structure zones:

1. A southern zone (Grotli, Lom, Lesja, Opdal) where the rocks show repeated overfolding and complex involutions, striking approximately N.W.-S.E. with a N.E. dip (the frontal, dorsal and central parts of the nappe).
2. A northern zone (Sylte, Andalsnes, Sundalsoyra) where the beds are highly inclined with a N.E.-S.W. strike and S.E. dip (the root zone of the nappe).

Muret also recognized three broad petrographic series as follows in downward sequence:

**Series of Pelitic Schists** - phyllites, mica schists, limestones, dolomites and quartzites.
Figure 3. Outline of major nappe structure encompassing the Romsdal and Tafjord districts as envisaged by Muret (1960).
Schistes du synclinorium de Trondheim

Sédiments Eocambriens et Cambro Siluriens

Gneiss de base

profil A

profil B

0 10 20 30 40 50 60 70 80 km

A

Opdal

Aurd

Lesiaskog

Vågå

Andalsnes

Grotli

Tafjord

B

phylites - limite gneiss micaschists

gneiss fortement inject, muscovite rare

gneiss d'injection

gneiss
quartzites. Greenschists and amphibolites (metamorphosed basic rocks) along with peridotites are also important members of this series, which is equated with the Cambro-Silurian Trondheim schists.

Series of Psammites - arkoses, conglomerates, schists and anorthosites with the characteristic presence of haematite. This series is equated with the Eo-Cambrian sparagmites.

Basal Gneiss Series - these are generally banded or bedded gneisses varying in composition from acid to basic although mostly dioritic, granodioritic, or granitic. Relicts of discordance and stratigraphic position indicate that this series is probably Pre-Cambrian, but has been involved with its sedimentary cover in the Caledonian deformation and metamorphism.

Muret considered that these three rock series have been involved in regional metamorphism and granitic injection before, during and after the emplacement of the nappes, but in general the metamorphic zones and zones of intensity of granitic injection are concentric about the root zone of the nappes.

O’Hara and Mercy (1963) have recognized a stratigraphic sequence within the layered gneiss complex in the Tafjord-Crotli-Geiranger triangle, which is broadly comparable with that suggested by Muret. They considered that the layered gneisses in general are probably a group of metasedimentary rocks plus minor hypabyssal and extrusive igneous rocks which have been folded, regionally metamorphosed under amphibolite facies conditions, and injected. They have pointed out that large scale recumbent folding about axes which plunge eastwards at moderate angles dominate the observed structures, and almost all linear structures plunge parallel to this axis. The axial planes/
planes of these folds strike N.N.E.-S.S.W. and dip E.S.E.

Hernes (1954) has described rocks of the Molde peninsula in Romsdal, which lie E.N.E. on strike with the Otterøy rocks, as consisting mainly of quartz-dioritic, granodioritic and granitic gneisses with broad bands of crystalline limestone and zones of eclogite, eclogite amphibolite, garnet amphibolite and amphibolite.

In review it appears likely that a deformation history of thrusting, nappe formation and two fold episodes has been operative throughout the basal gneiss region, and furthermore is broadly comparable with the deformation history of other parts of the Caledonian orogenic belt in Norway. One may cite examples of early folding about roughly E-W axes in south, west and north Norway, and posthumous folding about similar axes as late as the Lower Devonian. Also in most of these areas there are indications of later N-S folds, and this trend becomes dominant in the Bergen district. Clearly then the stress field which caused the folding, and probably also the thrusting and nappe formation, had a relatively constant orientation throughout much of the Caledonian front for a considerable period of time.

Available age dates for the gneisses are scant and often ambiguous. However, K-Ar dates of 385 and 405 million years on micas from gneisses of this region reported by Neumann (1960), a K-Ar date of 372 million years for biotite from gneiss, and Rb-Sr dates of 401 and 382 million years and a K-Ar date of 415 million years for phlogopite from altered eclogite inclusions in the gneiss (MacDougall and Green 1964), clearly indicate the effect of the Caledonian metamorphism on this region.

However, K-Ar dates of 590 and 582 million years for other micas from/
from the gneisses, and K-Ar dates of 980 and 1160 million years for amphiboles and 950 million years for phlogopite from altered eclogite inclusions in the gneisses may date a Pre-Cambrian event of about 1100 million years age (MacDougall and Green 1964), the lower ages being the result of varying argon loss during the Caledonian orogeny.

Therefore on both structural and age dating evidence the gneisses of the basal gneiss complex are considered to be predominately Caledonian in age, or at least to have been influenced appreciably by the Caledonian orogeny.

C. NOMENCLATURE

To avoid unnecessary confusion to the reader, the following list of definitions of the ultrabasic rock types described in the course of this study is given.

PERIDOTITE - general group name for olivine-rich ultrabasic rocks.
DUNITE - peridotite consisting essentially of olivine only (> 90%).
HARZBURGITE - peridotite consisting of olivine and significant amounts of orthopyroxene.
WEHRLITE - peridotite consisting of olivine and significant amounts of clinopyroxene.
LHERZOLITE - peridotite consisting of olivine and significant amounts of both ortho- and clinopyroxene.
WEBSTERITE - pyroxenite consisting essentially of orthopyroxene and clinopyroxene with negligible olivine.
ECLOGITE - rock type consisting essentially of a pyrope-rich garnet and an omphacitic clinopyroxene. Varieties of eclogite are qualified by/
by accessory minerals eg. quartz, kyanite, orthopyroxene and amphibole.

As garnet peridotite mineral assemblages are extensively described throughout this thesis, it has been found convenient to use the following standard abbreviations:

OL. - OLIVINE
OPX. - ORTHOPYROXENE
CPX. - CLINOPYROXENE
GNT. - GARNET

D. THE NATURE OF THE SPECIFIC PROBLEMS UNDER INVESTIGATION AND THE PROCEDEURE USED FOR TACKLING THEM.

The principal object of this study has been to elucidate the paragenesis of the garnet peridotites which occur within the peridotite bodies in the Caledonian gneisses of this region of Norway.

The general assemblage of the garnet peridotites (OL + GNT + CPX + OPX) is clearly distinctive of the eclogite mineral facies (Fig. 4), the join OL - GNT in magnesian compositions being confined to this mineral facies (Eskola 1920; O'Hara 1960a). Olivine-garnet assemblages in eulysites are far more ferriferous (Tilley 1936; Gjelsvik 1957) and are quite distinct from these garnet peridotite eclogite facies assemblages. Characteristic minerals of the eulysitic assemblages are fayalitic olivine, hedenbergite, iron hypersthene, almandine-rich garnet and apatite, together with a significant opaque ore fraction. Perthitic felspar also occurs in certain of the eulysitic assemblages from Tafjord (Gjelsvik 1957).
FIG. 4. ANHYDROUS MINERAL PARAGENESIS

GRANULITE FACIES

ECLOGITE FACIES

\[ \text{SiO}_2 \]

\[ \text{CO}_2 \]

\[ \text{CAO} \]

\[ \text{MGO} + \text{FeO} \]

\[ + \text{Albite} \]

\[ \text{SiO}_2 \]

\[ \text{CO}_2 \]

\[ \text{CAO} \]

\[ \text{MGO} + \text{FeO} \]

\[ + \text{Jadeite} \]

\[ \text{AN} \]

\[ \text{CPX} \]

\[ \text{OPX} \]

\[ \text{OL} \]

\[ \text{GNT} \]

\[ \text{KY, Sillt.} \]

\[ \text{SP} \]

\[ \text{OL} \]

\[ \text{GNT} \]

\[ \text{KY} \]

\[ \text{SP} \]

\[ \text{CO} \]

\[ \text{AL}_2\text{O}_3 \]
There are four basically different origins which might explain any random body of garnet peridotite exposed at the earth's surface. These are as follows:

1. Crystallization in situ to an eclogite facies assemblage.

2. Crystallization in situ to another mineral facies assemblage, but subsequently metamorphosed to an eclogite facies assemblage.

3. Crystallization at depth within the crust or mantle to an eclogite facies assemblage, with subsequent tectonic emplacement into its present position.

4. Crystallization at depth within the crust or mantle to another mineral facies assemblage, then subsequently both metamorphosed to an eclogite facies assemblage and tectonically emplaced into its present position.

Minor variations in origin within each of the four possibilities can be invoked to explain specific features of the mineralogy, chemistry and field relations of particular garnet peridotite occurrences, but these are all just variations on these four basic possibilities. For example, if particular garnet peridotites are assumed to have crystallised at depth to an eclogite facies assemblage and subsequently tectonically emplaced into their present position, then the following variations are possible. They may represent:

1. unmodified primary mantle peridotite material

2. mantle peridotite material modified by partial melting processes within the mantle - perhaps even the residual after the subtraction of a basaltic composition partial melt.

3. peridotite accumulates of a basaltic composition partial melt removed from its site of generation in the mantle, and crystallized to an eclogite facies assemblage within the lower regions of the crust or perhaps even in the upper mantle.
4. peridotites of either of these first three types modified by recrystallization to a lower temperature eclogite facies assemblage during transit to its present environment.

5. peridotite assemblages of any of the foregoing types modified by mechanical sorting processes connected with their tectonic emplacement.

However, whichever of these minor variations may have been operative, all demand the basic underlying hypothesis that the garnet peridotites have crystallized at depth and not in their present environment to an eclogite facies assemblage, and have subsequently been tectonically emplaced.

In an attempt to evaluate and either prove or eliminate each of the four basic possible origins, a detailed study of the occurrence of garnet peridotite at Kalskaret, Tafjord has been performed. The procedure used to tackle this problem has been fundamentally two-fold.

Firstly, a study has been made of the field relationships of this rock type with reference to the enclosing garnet-free peridotites and the adjacent gneiss complex. Of particular relevance here are detailed observations on the nature of the contacts between these rock types, any contact metamorphic effects, the mineral facies of the adjacent rock types, and the textural features of the garnet peridotites themselves. Such observations should indicate whether the garnet peridotites have crystallized in situ or have been tectonically emplaced into their present environment.

Secondly, a study has been made of the petrochemistry of the garnet peridotites, both with respect to their mineral compositions and the variation in bulk rock chemistry which has resulted in the presence of these particular rock types. Such studies should indicate whether the garnet peridotites crystallized initially to an eclogite facies assemblage, or whether they crystallized/
crystallized firstly to a different mineral facies assemblage and have only subsequently been metamorphosed to an eclogite facies assemblage. To this end the Kalskaret garnet peridotite mass has been sampled in detail and the samples analysed to show the extent of the chemical variation. Attempts have then been made to explain the origin of this compositional variation in terms of the igneous, metamorphic or mechanical sorting of a mineral assemblage of eclogite or other mineral facies.

Mineralogical data for the Kalskaret garnet peridotites (O'Hara and Mercy 1963, Mercy and O'Hara 1965 b) has been supplemented by further data, and the likely stable conditions of formation of such mineral assemblages discussed in terms of relevant high temperature and pressure experimental petrology studies.

The Almklovdalen garnet peridotites have been compared with those from Kalskaret on the basis of mineral and bulk rock composition data given by Lappin (1962), O'Hara and Mercy (1963), Mercy and O'Hara (1965 a,b).

Bulk chemical and mineral compositions of the Ugellvik garnet peridotites have been presented and compared with existing data on the other garnet peridotite occurrences, in particular with the Kalskaret and Almklovdalen occurrences.

Eclogite lenses within the gneisses and within the peridotites at certain rare localities (Kaldhussaeter, Tafjord and Hornindal, Nordfjord) have been compared chemically with the peridotites and enclosed garnet peridotites to see if there is any likelihood of any relationship between them other than one of similarity of mineral facies. These eclogite masses have also been compared petrographically and chemically with garnet amphibolite and amphibolite masses which also occur in the gneiss complex, in an attempt to confirm/
confirm or deny their chemical similarity as suggested by several Norwegian field workers (Gjelsvik 1952, Hennes 1953, and Kolderup 1960) and to choose between a metamorphic origin in situ as suggested by these workers, a metamorphic origin at depth followed by tectonic emplacement as suggested by Lappin (1962), O'Hara and Mercy (1963) and O'Hara and Yoder (1963), or any other possible origin (cf. Yoder and Tilley 1959, 1962).

Peridotite and eclogite are the most important rock types in modern hypotheses on the nature, mineralogy and chemistry of the rocks in the earth's upper mantle. Hence it is important to evaluate the significance of the Norwegian occurrences of these rock types as possible direct representatives or derivatives of the earth's mantle, in particular in terms of likely mantle compositions suggested by the chondrite meteorite model, and likely mineralogy inferred from high temperature and pressure experimental petrology studies and geophysical evidence.

If it is concluded that some or all of the Norwegian occurrences of these rock types represent mantle material, then they are obviously significant to problems connected with the genesis of basalt magmas as partial melts of mantle rocks and must be viewed in that light. They would then have to be capable of yielding partial melts of suitable composition.

E. NORWEGIAN PERIDOTTITES AND ECLOGITES - SUMMARY OF PREVIOUS WORK.

In a classic memoir Eskola (1921) reviewed the literature on the peridotites and eclogites of the basal gneiss region of southern Norway and listed most important occurrences.

Early field workers Reusch (1877), Brøgger (1880) and Vogt (1883) witnessing/
witnessing the generally conformable contacts and lack of evidence of intrusion of the peridotite masses in the gneiss complex, arrived at the conclusion that the peridotites were not intrusive in the gneiss, but were contemporaneous and of the same origin as the gneiss, whatever that origin might have been.

However, Eskola recognized the intrusive nature of the peridotites and considered that they were primary igneous rocks of slightly earlier consolidation than the gneiss complex itself.

Eskola recognized a two-fold subdivision of ecolgite occurrences:
1. as lenticular masses in the gneiss and
2. as bands or lenses in the olivine-rock.

He pointed out that the latter type of ecolgite, or more strictly garnet peridotite, is quantitatively very inconspicuous occurring as generally banded segregations of garnet and pyroxenes in the olivine rock, and shows chemical characteristics clearly pointing to a close genetic connection with the olivine rock, being poor in alkalis and rich in Cr$_2$O$_3$.

As for the ecolgites directly enclosed in the gneiss Eskola alleged that they are greatly variable in their chemical and mineralogical composition as well as in structure, differing from the segregations in the olivine rock in being almost free of Cr$_2$O$_3$ and containing considerable quantities of alkalis.

In summary Eskola concluded that the peridotites, ecolgites and anorthosites of this province were all igneous in origin and genetically related, being "...derived by a process of differentiation from one and the same magma." He considered that the peridotites and ecolgites formed as early differentiates from the magma, the enclosing layered acid and intermediate composition gneisses crystallizing under subsequent amphibolite facies conditions, and the consanguinous labradorite anorthosite masses solidifying still later.
Lappin (1962) as a result of his studies of the eclogites, dunites and anorthosites of the Selje and Almklovdalen districts concluded that each of these rock types has been tectonically emplaced as crystalline rocks in the banded gneiss complex. He has related their tectonic introduction to periods of tectonism in the gneisses and to the development of various gneissic types and has derived the following sequence of events:

1. Development of the banded gneiss complex.
2. Introduction of large eclogite bodies.
3. Boudinage of the larger eclogite bodies.
4. Introduction of anorthositic types.
5. Development of microcline augen gneisses etc. — and partial amphibolitization of the eclogites.
6. Introduction of the dunites in a late shear zone.

The dunites and associated eclogites were thought to have been derived from the mantle, the eclogites in the gneiss from the orogenically downbuckled crustal basaltic layer, and the anorthosites from a deep Pre-Cambrian complex.

O'Hara and Mercy (1963) similarly concluded from field studies in the Tafjord district that the peridotites with enclosed garnet peridotite layers, and the eclogite lenses in the gneiss were tectonically emplaced as crystalline rocks into the layered gneiss complex. They alleged that field relationships showed that the peridotite bodies could not be igneous rocks crystallized in situ, either as early differentiates from a magma which later gave rise to the layered gneiss complex — as suggested by Eskola (because the majority of that complex is derived from sedimentary rocks), or as separate intrusions/
intrusions of peridotite magma (because the contact relations show none of the features associated with true igneous rocks). They considered that the peridotites of the Kalskaret area of Tafjord were emplaced by more or less vertical movements in narrow zones after the formation of the regional recumbent folds, about N.N.E.-S.S.W. axes, and relatively late in the metamorphic history of the region.

From the consideration of their mineralogy and field relations, O'Hara and Mercy further concluded that the garnet peridotite inclusions within these peridotite masses probably represent somewhat modified mantle peridotite material (modified perhaps by the subtraction of a basaltic fraction) but they might represent completely recrystallized peridotites formed by igneous processes near the top of the mantle.

As regards the eclogite bodies in the gneisses they concluded that they were related neither chemically nor spatially in the field to the garnet peridotites, the only relationship being a similarity of mineral facies. It has been suggested (O'Hara and Yoder 1963) that such eclogites were probably formed by the metamorphic recrystallization of deep-seated gabbros, or crustal basic igneous rocks carried down in orogenic movements.

Mercy and O'Hara (1965,a) have concluded that the geochemistry of the garnetiferous peridotites and garnet-pyroxene rocks within the peridotites of Tafjord and Almkløvdalen does not compare closely with that of the eclogites in gneisses, nor with that of the suite of eclogites and garnet peridotites in kimberlite diatremes, nor with that of any extrusive magma type, nor with that of the products of igneous crystal accumulation and crystal sorting obtainable at low pressures where the C.I.P.W. norm closely represents the/
the actual minerals likely to be present.

Gjelsvik (1952) has described metamorphosed dolerites in the gneiss area of Sunnmøre. He has noted that large dolerite lenses, marginally amphibolitized, are concentrated north of a line from the mouth of Storfjord across Stordial to Åndalsnes, but that it is not impossible that some of the abundant amphibolites and eclogites of the southern part of Sunnmøre are altered dolerites. He has described the microscopic features in the stage by stage amphibolitization of the dolerites noting the production of garnet and amphibole from the original assemblage of olivine, diopside, plagioclase etc., and claiming that in certain cases garnet and diopside form a stable assemblage which he calls 'eclogitic.'

Thus Gjelsvik has concluded that the final products of the metamorphism of the dolerites are eclogite-like, amphibolitic rocks with crystalloblastic texture, suggesting that under extreme metamorphic conditions true eclogites are formed. As regards the age of the dolerites, he has noted that since they are regionally metamorphosed with generally pronounced secondary schistosity they cannot be younger than the last orogeny of the region (the Caledonian one), but the fresh appearance of many of them suggests a late orogenic origin.

Gjelsvik has also suggested that 

...'it is even possible that the dunites of Sunnmøre, most frequently situated to the south of the dolerite area, represent earlier derivatives of the dolerite magma.'

Hernes (1953) has described eclogite amphibolites from the Molde peninsula which he likewise considers as metamorphically upgraded basic rocks. He considered that with metamorphism the original assemblage of mostly an andesinic plagioclase and clinopyroxene, has been altered to a predominately garnet/
garnet-pyroxene (eclogite) assemblage. He has also recognized later retrogressive reactions, however, with the pyroxene altering to a pyroxene-plagioclase symplectite and the garnet developing amphibole-plagioclase kelyphitic rims.

As regards the origin of the basic layers in the gneiss complex Hernes considered that they perhaps represent Caledonian gabbroic or basaltic igneous rocks, but as they are associated with limestones in the Molde peninsula they may represent a scarn-like basic front related to the granitization of the gneiss formation.

Kolderup (1960) has also recognized the eclogite bodies in the gneisses as metamorphic in origin, differing he alleged from the garnet amphibolite masses only in degree of metamorphism. He has described some complex basic masses which occur in the four major synclines between Nordfjord and Sognefjord. The southerly masses are norites and olivine norites, the more northerly ones amphibolites, garnet amphibolites, eclogite amphibolites and eclogites. He has thus concluded that the eclogites and eclogite amphibolites are metamorphic derivatives of Caledonian gabbros and basalts. The amphibolitization of the eclogites was thought to have been connected with the granitization of the gneisses, but he did not really make it clear whether he considered the garnet amphibolites to be due to retrogression of the eclogites, or whether they represent a stage in the progressive metamorphism of the Caledonian basic igneous rocks into eclogites.

Schmitt (1964) has described such a varietal suite of eclogites in the Eiksund area of Sunnmøre, and one that contrasts strongly with the petrographic details reported in the literature for other field occurrences of eclogite studied in Sunnmøre, that one immediately suspects a laxity in his definition/
definition of the rock type 'eclogite'. Unfortunately the petrographic details of this work have not so far been published, to confirm or deny this criticism. Schmitt considered that the diversity of the eclogite associations and the specific character of each indicated that the eclogites were formed by the metamorphic crystallization of mafic rocks of varying parentage. He clearly considered that the eclogites reflect local metamorphic conditions, rather than the conditions at depth in the mantle or lower regions of the crust.

It should be stressed that while all the hypotheses reviewed here (other than Eskola's) to account for the origin of the eclogite bodies in the gneiss complex envisage a metamorphic origin, those of Gjelsvik, Hernes, Kolderup and Schmitt suggest a metamorphic origin in situ during the Caledonian orogeny while the hypotheses of Lappin, and O'Hara and Yoder envisage a deep crustal metamorphic origin and subsequent tectonic emplacement of the eclogites into their present environment.

MacDougall and Green (1964) have reported isotopic age determinations on minerals from certain Norwegian eclogites. Despite several ambiguities and some 'impossible' K-Ar ages for pyroclines (for example 8100 and 7350 million years m.y. - considerably older than current estimates of the age of the earth, about 4500 m.y.) considered to be the result of excess radiogenic argon in the pyroclines due to the existence of a high partial pressure of radiogenic argon at some time in the history of the eclogites, they felt justified in making certain conclusions about the metamorphic history of the eclogites.

It was considered that the K-Ar ages of 1850 and 1750 m.y. for subhedral hornblende in an enstatite eclogite pegmatite within gneiss at Grynivaag near Selje and eclogite within peridotite at Rødhaugen, Almklodalen, respectively,
respectively, were meaningful ages and represent the first stage of the retrogressive alteration of eclogite to amphibolite. As such it could represent the waning of the pressure-temperature environment responsible for the eclogite assemblages or a separate, later metamorphic event producing the amphibolite facies rocks. It was considered at least to set a minimum age on the eclogite crystallization.

Further phlogopite and amphibole, secondary after pyroxene and garnet, from eclogite in the gneiss gave K-Ar ages of 1160, 980, 950 and 415 m.y. and Rb-Sr ages of 401 and 382 m.y.

The authors have conceded that there is no unique interpretation of these results, but have suggested that they show that both the eclogites within the gneisses and the eclogite layers within the peridotites were partly affected by the Caledonian orogeny, but owe their original crystallization to a much earlier period (>1750-1850 m.y. ago) and their main secondary alteration also to an earlier Pre-Cambrian event, possibly in the range 950-1150 m.y.

Hence the concept that the eclogites are solely metamorphic products of the Caledonian orogeny is not consistent with their interpretation of the isotopic age data.

However, O'Hara and Mercy (1965) have stressed that in view of MacDougall and Green's (1964) demonstration of the mobility of radiogenic argon in the deep-seated metamorphic environment, one must consider all the K-Ar dates with suspicion unless they are substantiated by Rb-Sr dates. The only Rb-Sr dates agree with certain K-Ar dates in demonstrating that a significant metamorphic event occurred about 400 m.y. ago, namely during the Caledonian orogeny.
In this section it is intended to discuss the field relationships of the various occurrences of peridotite and eclogite observed during the course of this study.

Studies of the contact, metamorphic and structural relations of these rock masses relative to the enclosing gneiss complex are obviously critical to several problems relating to their origin and particularly their mode of emplacement into their present environment.

The dating of the intrusion of the peridotite and eclogite rock masses with respect to the folding and metamorphism of their enclosing country rocks depends first of all on a thorough megascopic and microscopic study of the gneiss complex as a whole to determine the chronological order of the tectonic and metamorphic phases which have affected it. Following this a study of the mineralogy (mineral growth) of the present assemblages and structural features of the particular rock types in question, should result in an accurate dating of their intrusion with respect to the metamorphic and fold phases of the gneiss complex as a whole.

A complete breakdown of the structural and metamorphic history and general stratigraphy of the gneiss complex in the Tafjord region is being attempted by Mr. H.K. Brueckner of Yale University, the results of whose work will be relevant to this present study.

However, attempts have been made to define the date and mode of emplacement of the peridotite and eclogite rock masses by close observations on their contact, metamorphic and structural relations with the immediately adjacent country rocks.
A. THE TAFJORD DISTRICT:

1. GENERAL DESCRIPTION.

The majority of field studies performed have been concerned with the peridotite and eclogite outcrops in the Tafjord district of Sunnmøre. The main peridotite outcrops in the Tafjord district are shown on Fig. 5. Within this district the areas which have been studied in greatest detail are those around the peridotite outcrops at Kalskaret, Kaldhussaeter, Ranukdalen and Viksvatn. Each of these areas is discussed in turn later.

Although it is difficult to outline any definite large scale fold structures in the gneiss complex of this district, it is likely that such fold structures do exist but have been complicated by refolding and perhaps by thrusting. It is apparent that large tracts of gneiss (particularly in the Kaldhussaeter-Viksvatn area) have foliation planes which broadly strike N-S and dip at moderate angles to the east, while intervening tracts of gneiss (in the Tafjord village - Kalskaret area, for example) have more steeply inclined foliation planes striking more approximately E-W.

It is considered that such variations in foliation plane disposition in the gneisses, outline large scale recumbent folds with axial planes which strike approximately N.E.-S.W., and dip S.E., and fold axes which plunge at moderate angles eastwards. (Fig. 6,a.).

These recumbent folds may not in fact represent the initial fold phase of the complex but possibly refold even more fundamental recumbent nappe structures, perhaps analogous with the deep Pennine nappe structures of the Alps. Moret (1960) outlines one such major nappe structure which he envisages encompassing a large area of gneisses in the Tafjord and Romsdal districts (Fig. 3). These early nappe forms are assumed to form the regional schistosity/
Map of the main peridotite outcrops in the Tarjord district.
FIG. 5

PERIDOTITE OUTCROPS

SCALE 1:50000
schistosity folded by the E-W axis folds described above.

However, as this area probably represents the structurally lowest levels of the Caledonian orogenic belt now exposed (Holtedahl 1960, Strand 1961, Muret 1960), the original schistosity in certain areas of 'basement' gneiss may be Pre-Cambrian in origin related to earlier folds which may remain in some places but have generally been rotated into alignment with the later Caledonian structures.

Fig. 7 stereonet A shows that the poles to the foliations of the gneisses in the Tafjord district remote from the peridotite outcrops lie in a well defined girdle indicating that the distribution is controlled by folds with axes plunging at moderate angles (30-40°) eastwards. Measurements of minor fold axes and lineations also demonstrate a broadly similar eastward plunge of most of the structures.

Although approximate E. or E.N.E. plunges of the minor fold and linear structures are predominant throughout most of the district, later N-S axis folds appear in the Viksvatn area, discussed shortly.

Minor fold structures show considerable complexities with minor shearing and thrusting, together with rodding and boudinage effects. In some cases minor folds with east plunging axes have such highly variable axial planes that the folding can only really be described as 'flow' folding. However, elsewhere it seems possible to distinguish three sets of such minor folds on the basis of the orientation of their axial planes and the presence or absence of an axial plane schistosity.

These fold types are, in chronological order (Fig. 6, b,c,d.):

a. Tight isoclinal folds with near vertical E-W striking axial planes and
FIG. 6

d. Form of major fold structures viewed down east plunging fold axes

Form of minor fold structures with east plunging axes

b.
c.
d.
on axial plane schistosity. The limbs of these folds are commonly sheared out, with isolated fold cores being quite common. These folds are clearly predominately shear folds with a considerable degree of axial plane movement having been involved in many cases.

b. Tight isolines with roughly N-S striking axial planes gently to moderately inclined eastwards, and with an axial plane schistosity.

c. Broad flexural folds with generally rather steeply inclined E-W striking axial planes and without an axial plane schistosity. These folds appear particularly commonly developed adjacent to the large peridotite masses especially in the Kalskaret region.

The rocks of the gneiss complex of this district are mostly a series of pelitic and semipelitic gneisses with occasional horizons of augen gneisses. Most mineral assemblages contain varying proportions of certain of the following minerals: quartz, perthitic alkali felspar, plagioclase (oligoclase-andesine), biotite, hornblende, almandine garnet, muscovite, chlorite, sphene, epidote, apatite and iron ore. The augen of the gneisses are generally of perthitic microcline.

Bands of calc-silicate gneiss are thin and rare, being characterized by concentrations of epidote and in rare instances diopside.

There are conspicuous horizons of quartzite within the gneisses. These provide suitable marker horizons for attempts to outline major fold structures in the area. However, considerable difficulties arise when one attempts to follow out the quartzite horizons, as access to much of the ground is extremely hazardous or impossible and furthermore the quartzites are often impersistent along the strike as they have been extensively folded, sheared and/
Figure 7. Poles to Foliations - solid circles
Fold Axes and Lineations - crosses
Stereonet A. Gneisses from the Tafjord district remote from the peridotite outcrops.
Stereonet B. Gneisses adjacent to the Kalskaret peridotite outcrops.
Stereonet C. Kalskaret peridotites.
and boudinaged. Nevertheless the association of quartzite close to the peridotite masses at Kalskaret, Kaldhussaeter, Ramukdalen and Viksvatn may indicate that they occur together in one prominent folded stratigraphic horizon (cf. O'Hara and Mercy 1963 Fig. 3).

The presence of quartzites, thin calc-silicate bands and the predominance of pelitic or semipelitic rock types indicates that most of the gneiss complex of this district, is metasedimentary in origin.

Rare rock specimens have been found containing small amounts of kyanite (Brueckner, personal communication) in a general assemblage of quartz + plagioclase + biotite + almandine garnet + kyanite + sphene + magnetite.

It is therefore considered that the gneisses belong predominantly to the kyanite-almandine-muscovite subfacies of the almandine-amphibolite facies (Turner and Verhoogen 1960).

In the area to the west of Kaldhussaeter Vatn the generally heterogenous assemblage of pelitic and semipelitic gneisses of most of the area is underlain by more homogeneous, only crudely foliated, coarse grained, acidic gneisses including some augen gneisses and Hills of granite. Odd fold structures of highly variable axes and axial planes (including N.N.W.-S.S.E. axis folds with axial plane schistosities not recognized elsewhere in the district), may be only partly 'Caledonised' Pre-Cambrian structures and the rocks of this area may represent a window of the Pre-Cambrian basement in the Caledonian metasediments.

It is considered that the Caledonian gneisses of the Tafjord district as a whole were developed as follows. A series of pelitic, semipelitic and quartzitic sediments with interbedded basic igneous rocks have been folded and metasomatically/
metasomatically injected, with regional metamorphism under almandine-
amphibolite facies conditions. As the augen gneisses can be shown to have been folded by all three minor east plunging fold types, the main injection probably occurred comparatively early on in the fold and metamorphic history, probably in association with the major recumbent nappe formation.

Peridotite, eclogite and anorthosite form conspicuous inclusions in the layered gneisses but their total volume forms only a very small proportion of the total, the amounts of eclogite being particularly subordinate.

Recognition of the fact that both the peridotites, with enclosed garnetiferous layers, and the eclogite lenses in the gneiss contain eclogite facies mineral assemblages \((\text{GNT} + \text{OL} + \text{OPX} + \text{CPX} \text{ and } \text{GNT} + \text{CPX} + \text{quartz})\), respectively and now occur in a regional gneiss complex of amphibolite facies grade, indicates that they are foreign and not indigenous to their present environment and requires an explanation as to their origin and mode of emplacement. It was this problem that it was hoped to solve by detailed observations on the occurrences of these rock types in the Tafjord district.

2. THE KALSKARET AREA.

There are numerous outcrops of peridotite in the Kalskare area, the col at about 3,000 feet above sea level, between Tafjord and Herdalen (Map I).

The important occurrence of garnet peridotite noted by Eskola (1921) lies almost on the eastern lip of the col overlooking Tafjord.

The main peridotite outcrop extends down from Kalskare to Onilsa Vatn/
Vatn with its contact approximately in the wall of the large cliff feature. However, much of the peridotite on the lower ground by Onilsa Vatn, and between Onilsa Vatn and the village of Tafjord is probably not in situ, but has slumped down from its original contact position against the cliff.

Outcrops of peridotite are dotted over the col itself with intervening tracts of gneiss between them. However, the contact relationships are generally poorly exposed. Peridotite also outcrops extensively above the col on the left hand side moving towards Herdalen, on a low ridge on the northern side of Jordhorn.

Plots of the poles to the foliation planes, together with fold axes and lineations, for the Kalskaret peridotites and adjacent gneisses are given on Fig. 7. Stereonets B and C, respectively. These show that there is a considerable measure of concordance between the peridotites and the gneisses, with the foliation plane disposition of both being related to fold axes which plunge eastwards at moderate to low angles. However, taken overall the foliation planes of the peridotites are slightly more inclined than those of the enclosing gneisses.

In ascending to Kalskaret from Onilsa Vatn the gneisses of the contact walls on both sides of the main peridotite mass clearly seem to have been dragged up into a near vertical disposition, suggesting an anticlinal structure. (Fig. 9c). This structure, similar to that of a salt dome, suggests that the peridotite mass has pushed its way up into the gneisses from below as a diapir, doming and dragging up the enclosing gneisses adjacent to it. It would therefore appear that, in this instance at least, the broad concordancy between the peridotites and the adjacent gneisses was the result of the mode of emplacement of/
of the peridotites as diapiric intrusions pushing up into the gneisses. Such a mode of emplacement would explain the generally slightly more steeply inclined disposition of the foliation planes of the peridotites compared with those of the enclosing gneisses (Fig. 7. Stereonets B & C).

Flexural folds with axes plunging at low to moderate angles eastwards and with vertical axial planes are common both within the peridotites and the contact gneisses. Such folding in the peridotites often takes the form of tight chevron folds, being best picked out by thin enstatite-chlorite veins.

Both the peridotites and the contact gneisses have been extensively sheared, the former being well foliated throughout with prominent zones of highly sheared 'platy' peridotites.

At one specific locality in the cliff contact wall about 500 feet above Onilsa Vann a small pod of peridotite can be observed clearly emplaced in the core of a small anticlinal structure in the gneisses. This is regarded as further evidence of the diapiric intrusive nature of the peridotites into the gneisses. At this locality the peridotite has no obvious original compositional banding but has developed an axial plane schistosity. The contact of the peridotite with the gneiss is well exposed and consists of at the most, 1 foot of altered tremolite-rich rock. The first 4 to 6 feet of the contact gneisses are very strikingly sheared (mylonitized) and very felsic, but beyond that they pass into more normal flaggy mafic gneisses.

The foliation planes of the peridotites on the Kalskaret col are highly variable in disposition, being related to fold structures with axes plunging at moderate to low angles eastwards, as are the foliation planes of the/
the gneisses. It is therefore apparent either that the peridotites and the
gneisses have been folded together subsequent to the emplacement of the peridotites
or that folding of the peridotites occurred virtually synchronously with their
emplacement in more or less vertical zones into previously formed or more likely
contemporaneously forming fold structures in the gneisses.

In view of the demonstration of the diapiric intrusive nature
of the peridotites in the Kalskaret area, the latter explanation appears the
more feasible. However, whichever of these alternative explanations represents
the actual case it is apparent that when the peridotites were intruded into
the gneiss complex, the stress system responsible for the repeated folding of
the gneisses about east plunging axes was still dominant.

A conspicuous horizon of quartzite, often with anorthosite structurally
below it, occurs on both sides of the Kalskaret col (Map I) probably outlining
a major fold structure in this area with the peridotites emplaced into its core
(O'Hara and Mercy 1963).

Anorthosite outcrops most strikingly high on the southern flank of
the col. It shows prominent cataclastic textures suggesting tectonic
emplacement. Large blue grey labradorite porphyroclasts, often several inches
in length, are considerably fractured and recrystallized to a mosaic of much
smaller (1-2 mm.) crystals. Approaching the well exposed upper contact, the
anorthosite is exceedingly cataclased and friable but right at the contact a
very finely banded and compacted anorthosite mylonite occurs. Along the contact
itself there is a thin band of amphibolite, and above that the gneiss is highly
brecciated with amphibole and epidote filling the open fractures. A specimen
of this Kalskaret anorthosite has been collected for analysis (T270).
In general the peridotites of the Kalskaret area are exceedingly fresh dunites, being relatively free of enstatite.

However, at the western end of the col overlooking Herdalen there is a small highly altered peridotite mass. Two specimens (T177 and T178) of this extensively chloritized and amphibolized peridotite have been analysed. At the lower contact of this mass plagioclase amphibolite passes downwards into plagioclase-garnet amphibolite, which has the texture of an amphibolitized eclogite (analysed specimen T183).

Prominent eclogite lenses also occur in the gneisses above this altered peridotite mass.

A highly foliated amphibolite lens, 5 feet wide at its maximum, occurs in a fold in one of the peridotite masses in the centre of the col. A specimen (T195) has been analysed in an attempt to determine whether it represents a highly retrograded garnet peridotite horizon indigenous to the peridotite, or a country rock type eclogite inclusion in the peridotite. Its field occurrence as a restricted and sharply bounded mass in the peridotite suggests that it represents a tectonically emplaced inclusion.

A prominent mass of garnet amphibolite occurs adjacent to one of the peridotite masses on the north side of the col. No actual contact with the peridotite can be seen, however. The texture and negligible foliation of this mass contrasts with that of the eclogite inclusions in the gneiss. A specimen has been analysed (T256).

The occurrence of garnetiferous peridotite within the Kalskaret peridotites is exceedingly restricted, forming an irregular lens about 160 feet in length and 40 feet at a maximum across the strike, conformably within the normal enstatite poor dunite. Neither GNT nor CPX occurs in the normal peridotites of/
of the Kalskaret area, nor for that matter of the Tafjord district as a whole. The form of the garnet peridotite outcrop is illustrated in Fig. 8. Strictly speaking garnet peridotites occur in two separate horizons separated by garnet free peridotite. However, a few thin horizons of peridotite with amphibole pseudomorphs after garnet in fact occur in the peridotite immediately below the main garnetiferous layers.

Laterally, particularly westwards, the garnet peridotites interdigitate with garnet free peridotites (Fig. 9a). Furthermore, the occurrence of a certain amount of olivine throughout most of the garnetiferous materials, suggests that they are in fact indigenous to the peridotites, despite their restricted occurrence.

The most marginal horizons show a certain degree of amphibolitization but generally they are remarkably fresh.

The majority of the garnet peridotite rock types contain appreciable amounts of GNT, CPX, and to a lesser extent OPX, with only relatively small amounts of OL. Broadly speaking, however, all gradations exist between an enstatite poor dunite and an almost OL and OPX free eclogite. Thin bands of normal peridotite assemblage (OL + minor OPX and chromite) occur interlayered with the garnetiferous rock types (Fig. 9,a).

Rock types with particular proportions of GNT, CPX, OPX and OL are often inconsistent laterally. This is sometimes caused by tight isoclinal folding (Fig. 9,b) but elsewhere particular layers just appear to lens out along strike. This latter effect is particularly well illustrated by the occurrence of garnet websterite (Fig. 8).

13 rock samples were collected across the garnet peridotite layers along section A-B, Fig. 8., and analysed to show the extent of the chemical variation/
FIG. 9

a. Interdigitation of dunite and garnet peridotite at Kalskaret

Layered garnet peridotites

Dunite

Scale approx. 1" to 3'

b. Folding in garnet peridotites

Scale approx. 1" to 2'

c. Section through inferred Kalskaret structure

Peridotite
variation which has resulted in the development of these garnetiferous rock types. The details of the exact horizon of collection, the mineral assemblage and the mode of each rock specimen are given in (Table No. 4 and Appendix III).

Another noteworthy feature is the occurrence of a corundum bearing gneiss just below the garnet peridotite bearing outcrop. This feature is discussed further in a subsequent general section on the contact relations of the peridotite masses.

3. THE KALDUSSAETER AREA.

The Kaldhussaeter peridotite mass outcrops just above the tree level on the steep western flank of Storfjellet, on the eastern side of Kaldhussaeter Vatn (Map II). It forms an elongate lens extending for about 1½ kilometers along the strike and has a maximum stratigraphic thickness of about 100-120 metres.

The foliation plane, minor fold axis and linear structure measurements in the peridotite and the immediately adjacent gneisses show a clear concordancy (Fig. 10. Stereonets A & B, respectively), the disposition of the foliation planes of both being related to axes plunging about 20-40° just north of east.

Despite numerous minor complexities, the foliation planes over much of the central part of this peridotite mass strike approximately N-S and dip at moderate angles (30-50°) eastwards, paralleling the general regional gneiss foliations over this area. In this central zone the peridotite is of two varieties, either a massive crudely foliated type or a highly sheared 'platy' type. The latter type appears to have developed through extreme cataclasis in certain prominent movement horizons, within the peridotite, which are parallel to the contacts.
Figure 10. Poles to Foliations - solid circles
Fold Axes and Lineations - crosses
Poles to Joint Planes - open circles
Axial Plane Schistosities of Folds - broken great circles.

Stereonet A. Kaldhussaeter peridotite mass
Stereonet B. Gneisses adjacent to the Kaldhussaeter peridotite mass.
Stereonet C. Main quartz eclogite lens within the Kaldhussaeter peridotite mass.
Stereonet D. Anorthosite horizon below the Kaldhussaeter peridotite mass.
At the two extremities of the lens, the peridotite occurs as large overturned folds with roughly E.N.E. plunging axes, the foliation planes being more steeply inclined but still broadly concordant with those of the enclosing gneiss. However, the overturning is much more pronounced in the peridotite, which appears to have been capable of a much higher degree of plastic deformation or laminar flow resulting in intense microfolding of the peridotite in places with the development of axial plane schistosities, notably in the mass at the southern extremity of the lens. (Plate 1). However, the plastic deformation appears to have been superceded by more brittle deformation effects, shearing and jointing. Thus shear planes are developed within the peridotite parallel to the contact round the large overturned folds, while sets of joint planes perpendicular to the fold axes have also been prominently developed in places (Fig. 10. Stereonet A). The gneisses adjacent to the peridotite are also highly sheared, their compact nature and striped appearance resembling mylonites.

The contact between peridotite and gneiss is well exposed in the stream bed at the southern end of the mass. Along much of the contact, peridotite and gneiss can be seen in juxtaposition separated by a reaction zone only about 1 foot in width, consisting of talc-magnesite serpentinitite against the peridotite and biotite amphibolite against the gneiss. Elsewhere, however, wider but irregularly developed and generally only crudely foliated lenses of mostly tremolitic amphibole (but with talc, magnesite, anthophyllite and phlogopite also) are developed near the contact within the peridotite and lenses of biotite and hornblende within the gneiss. The peridotite is also markedly more serpentinized along the contact.

Much of the peridotite mass consists of virtually monomineralic dunite. Hence there is often a conspicuous lack of any obvious primary mineral layering and/
and subsequently developed axial plane foliations are correspondingly ill-defined. Elsewhere, however, small amounts of enstatite in the peridotite help to define the primary mineral layering and subsequent metamorphic foliations are often defined by sparse secondary chlorite flakes.

Within the overturn at the northern end of the peridotite mass there is a conspicuous horizon of partly amphibolitized eclogite, about 35' thick at a maximum. (Map II). This eclogite mass has a sharp contact against the peridotite being retrograded at the contact to a well foliated hornblende schist.

This eclogite is clearly distinguishable from the garnet peridotite layers in the Kalskaret peridotite through its sharp tectonic contact, the lack of interdigitation or transitions between peridotite and eclogite, and its different mineralogical character - namely the presence of conspicuous amounts of both quartz and plagioclase and the lack of olivine or orthopyroxene. On the other hand, it is mineralogically and texturally similar to eclogite which occurs in abundance as boudins within the gneisses enclosing the peridotite mass.

This country rock type eclogite inclusion within the peridotite is generally poorly foliated but in places apparent segregation of its constituent primary minerals - garnet, clinopyroxene and quartz, and secondary minerals, - amphibole, plagioclase and clinocoidsite has produced notable mineral banding. Such mineralogical segregation is probably the result of the likely tectonic emplacement origin of the mass. In general, however, the eclogite is quite massive and relatively mineralogically homogeneous.
When traced downhill at its northern end this eclogite horizon thins rapidly from about 25' to 3' over about 10' laterally, being apparently constricted by the enclosing peridotite. However, about 250' below similar eclogite again occurs within the peridotite, as it also does in a small outcrop on strike to the east of the main eclogite outcrop (Map II). It appears that the eclogite horizon within the peridotite has been both folded and boudinaged within the peridotite.

To show whether this eclogite within the Kaldhussaeter peridotite mass is chemically comparable or not to eclogite which occurs as inclusions in the gneiss complex, chemical analysis has been performed on a sample (T271) consisting of an accumulation of several small specimens collected over its outcrop.

Conspicuous mafic compositional banding within the Kaldhussaeter peridotite occurs in a small isolated peridotite outcrop in the central portion of the mass at a height of about 3,600' (Plate 2B).

The layering consists primarily of a sequence of amphibole, chlorite, orthopyroxene and olivine rich layers. The layering sequence in the best exposed example is illustrated in Fig. 11. Specimens of the main 2' thick, amphibolite layer (T80) and of the only partly amphibolitized eclogite layer with conspicuous garnets (T81), were selected for analysis in an attempt to clarify whether these rock types represent altered garnet peridotite layers or altered country rock type eclogite inclusions as noted elsewhere in the Kaldhussaeter peridotite mass.

On the basis of their field occurrence as layers intimately interbanded with peridotite and the cataclastic texture of the amphibolite layers, it was considered that they represent amphibolitized garnetiferous peridotite/
FIG. II  MAFIC LAYERING WITHIN KALDHUSSAETER PERIDOTITE.

- DUNITE  
- CHLORITE BAND  
- DUNITE  
- ORTHOPYROXENITE  
- HARZBURGITE + AMPHIBOLITE  
- ORTHOPYROXENE + AMPHIBOLE  
- DUNITE  
- AMPHIBOLITE - T 80  
- DUNITE  
- AMPHIBOLIZED ECLOGITE - T 81  
- DUNITE  

SCALE APPROX. 1" TO 2"
peridotite layers indigenous to the enclosing peridotites.

Only some 15' above the mafic layering just described, there is a small isolated pod of apparent country rock type eclogite enclosed in the peridotite. This pod is about 3-4' thick and only exposed for about 10' along the strike. The eclogite of this pod is massive and only partly amphibolitized, mineralogically and texturally resembling the main country rock type eclogite inclusion in the peridotite and the eclogite inclusions directly enclosed in the gneisses. To confirm or deny this a specimen (T205) was collected for analysis.

There are abundant lenses of eclogite directly enclosed in the gneisses adjacent to the Kaldhussaeter peridotite mass, particularly in the gneiss crags directly above the peridotite outcrop.

The eclogite clearly occurs in lenses of boudin form (Plate 3A), the more massive eclogite having apparently behaved during tectonism in a more rigid fashion than the enclosing mafic and felsic gneisses. The boudins are generally of a small order of magnitude, commonly only a foot or less thick. However, several such small boudins can generally be noted along any particular eclogite bearing stratigraphic horizon. The largest eclogite boudin noted was some 5' thick and extended for about 20' along the strike.

The margins of the larger eclogite boudins are generally more amphibolitized than the central portions, and in certain cases the amphibolite rim can be seen to be discordance to the internal structure of the boudin, pointing to the occurrence of boudinage before amphibolitization. Furthermore, certain eclogite boudins have been observed to have been invaded by quartz veins, further evidence that the boudinaging occurred at least before the injection/
injection and amphibolitization of the gneiss complex had ceased.

As regards the country rock type eclogite inclusions within the peridotite one must consider whether there is any relationship between them and their enclosing peridotites other than one of a similarity of mineral facies (the peridotites are all recognized as eclogite facies assemblages as a consequence of the indigenous garnet peridotite layering in the Kalskaret peridotite mass). The fact that both rock types are clearly foreign to their present environment, belonging to the same mineral facies (eclogite) which is clearly distinct from the mineral facies (amphibolite) of the gneiss complex as a whole, and their close field association points to some genetic relationship between them. On the other hand, field evidence from the eclogite lenses directly enclosed in the gneisses tends to indicate that they were intruded and boudinaged before the completion of the injection and metamorphism of the gneiss complex, while the fresh and relatively unaltered appearance of the peridotites suggests emplacement after the main metamorphic phase. Furthermore the extensively amphibolitized nature of the eclogite mass within the peridotite contrasts with the fresh appearance of the enclosing peridotites, and suggests that the eclogite was exposed to amphibolitization before inclusion within the peridotite.

If, however, the eclogite is just a random tectonic inclusion in the peridotite, why are there no inclusions of other members of the gneiss complex in any of the peridotites of the Tafjord district? Perhaps the answer lies in the relative competencies of the members of the gneiss complex into which the peridotite was intruded.

The problem of the petrogenesis of these quartz eclogites is discussed/
discussed further in a subsequent chapter.

Structurally just below the Kaldhussaeter peridotite mass there is a prominent anorthosite horizon, some 200-300' thick. It mostly outcrops below the tree level and is hence difficult to follow, but over much of its outcrop it appears to be relatively constant in thickness and in mineralogy (labradorite + minor amphibole and/or epidote). Just to the south side of the Grovadalen river, however, it appears to rapidly lens out into typical layered mafic and quartz-felspathic gneisses of the gneiss complex. Elsewhere, in particular just above the Tourist Hut at the north end of Kaldhussaeter Vatn the anorthosite is interspersed with quartz-felspathic gneisses.

As in the Kaldhussaeter peridotite mass and its enclosing gneisses, the disposition of the foliation planes in the main quartz eclogite lens within the peridotite and the anorthosite horizon below the peridotite (Fig. 10, Stereonets C and D, respectively) is again controlled by fold axes plunging at low to moderate angles just to the north of east.

As a result of the apparent concordancy between the peridotites and the anorthosite with the enclosing gneisses it would appear likely that both these rock types had been intruded into the gneiss complex and later folded together with it. Only in the overturned folds at either end of the peridotite mass is there any hint of any discordancy to suggest diapiric intrusion of the peridotites as evidenced in the case of the Kalskaret peridotites.

4. THE RANUKDALEN AREA.
A large peridotite mass occurs in Ranukdalen overlooking from the south the unoccupied hamlet of Tajordsaeter. A complex of smaller peridotite bodies occurs slightly higher up Ranukdalen itself. (Fig. 5).

From a distance the lower peridotite mass appears as a classic example of a tectonic diapir (Plate 3B). It stands out as an upright dome-shaped mass with mostly near vertical foliations appearing to be discordant with the apparently flat lying surrounding gneisses. However, closer inspection shows that the contact gneisses also have steeply inclined foliations close to the peridotite mass and are therefore concordant to the latter. The flat lying appearance of the gneisses is produced by a prominent set of joint planes.

The poles to the foliations in this peridotite mass form only a very poorly defined girdle (Fig. 12, Stereonet A). However, this girdle together with a few minor fold axis and lineation measurements again indicates the influence of approximately E. or E.N.E. plunging fold structures controlling the structural disposition of the foliation planes in this peridotite mass.

The poles to the foliation planes in the upper Ranukdalen peridotites lie on a well defined girdle indicating a fold axis plunging rather steeply (70-80°) to the N.E. approximately (Fig. 12, Stereonet B). Almost all the foliation planes are highly inclined — often nearly vertical. Microfolding in this peridotite mass is extensive, the minor folds having axes plunging steeply to the N.E. and developing good axial plane cleavages and schistosities defined by chlorite flakes.

Structural measurements in the Ranukdalen gneisses likewise show the influence of approximately E. or E.N.E. plunging fold structures (Fig. 12 Stereonet/
Figure 12.

Stereonet A. Lower Ranukdalen peridotite mass.
- Poles to Foliations - solid circles
- Fold Axes and Lineations - crosses

Stereonet B. Upper Ranukdalen peridotites.
- Poles to Foliations - solid circles
- Fold Axes and Lineations - crosses
- Axial Plane Schistosities of Folds - broken great circles.

Stereonet C. Ranukdalen gneisses.
- Poles to Foliations - remote from peridotites - solid circles.
- adjacent to peridotites - open circles.
However, the foliation planes in the gneisses as a whole are generally less steeply inclined than those of the peridotite masses, while gneiss foliation planes near peridotites likewise appear on the average more steeply inclined than those remote from peridotite masses.

It is considered that the general concordant discordancy (as it might perhaps appropriately be called) of the peridotite masses to the enclosing gneisses is a result of their tectonic diapiric intrusion. As a result of their upwards emplacement, the foliation planes of the peridotites are more steeply inclined than those of the general regional gneiss complex, but adjacent to the peridotites the latter have been steepened to become essentially concordant to the margins of the peridotite masses.

It would appear, however, that the tectonic emplacement of the peridotites has been controlled by the same stress system responsible for the development of E. or E.N.E. plunging fold structures and lineations throughout the gneiss complex. Hence the disposition of the foliation planes in the peridotites shows the influence of this stress system. However, as pointed out for the Kalskaret peridotites, rather than the peridotites having been folded with their enclosing gneisses after their emplacement it would appear perhaps more likely that folding and tectonic emplacement were more or less synchronous events.

The most notable mineralogical feature of the Ranukdalen peridotites is the abundance of enstatite; harzburgite being by far the most common rock type in these masses. However, several features indicate that the enstatite in these rocks is mostly of secondary growth.

Evidence/
Evidence from enstatite poor peridotites from elsewhere in the Tafjord district and enstatite bearing garnet peridotites from Kalskaret, indicates that enstatite is the mineral most affected by the cataclasis associated with the tectonic emplacement of these rock masses. As a result, enstatites are generally of small grain size (>1 mm.), while larger relict porphyroclasts show remarkable strain features.

Some of the Ramukdalen peridotites appear to contain a certain amount of primary enstatite which is highly cataclased and takes up preferred orientations in microfolds.

However, in the Ramukdalen harzburgites acicular enstatite crystals occur up to a foot or more in length and fail to show appreciable deformation effects while the matrix olivine of these rocks has been highly cataclased and recrystallized, often with pronounced dimensional orientation. Such enstatites show only minor strain effects and little evidence of cataclasis. Furthermore many of them are of apparently random orientation and some can be observed growing over and including grains of the often pronounced dimensionally orientated olivine fabric. Enstatite and talc veins can also be seen to cut across the foliation planes of the peridotites in places. On the other hand, smaller enstatite crystals can often be seen bent and folded by the microcrewelations of the foliation, and sometimes orientated enstatites even form a crude lineation on the foliation surfaces.

The best explanation would appear to be that much of the enstatite in these rocks was of a late stage secondary growth generation in the original peridotite, growing mainly in the foliation planes but with random orientation of/
of the enstatite laths within these planes. Some, however, must have grown cross-cutting the foliation planes.

It would then appear that during the final stages of intrusion and folding of the peridotite masses certain of these enstatites have been folded within the foliation planes but many of the enstatites were too large and rigid to be reorientated during the folding so that enstatites which originally grew cross-cutting the foliation remained that way.

The occurrence of enstatite and talc veins cross-cutting the foliation planes of the peridotites further suggests a late secondary origin of the enstatite through alteration of the original peridotite.

5. THE VIKSVATN AREA

This area (Fig. 5) lies between about 4,250-4,500' above sea level and is covered for most of the year by appreciable snow. It is therefore only accessible at the height of summer and even then snow obscures many of the contacts between the exposed rock types.

The area just to the north of the lake itself is of particular interest because of the conspicuous development in that region of folds of a style not recognized elsewhere in the Tafjord district. These folds which are predominately flexural and disharmoncous in style have generally N-S axes, as distinct from the predominately E. or N.E. plunging fold axes and lineations recognized elsewhere. These folds do not develop an axial plane schistosity and can be clearly demonstrated to refold the earlier E. plunging fold axes and lineations. Fig. 13, Stereonet A shows that the refolded E. plunging lineations and fold axes/
Figure 13.  Poles to Foliations - solid circles
Fold axes and lineations - crosses
Axial Planes of N-S axis folds - broken great circles.

Stereonet A.  Viksvatn gneisses: refolded 1st fold (E-W axis) axes and lineations.

Stereonet B.  Viksvatn gneisses: measurements from zone of prominent N-S axis folding.

Stereonet C.  Viksvatn gneisses: direction and amount of plunge, together with shear sense, of fold axes of small scale disharmonic folds measured over a restricted vertical sequence.

Stereonet D.  Peridotite pod at northern end of Viksvatn: direction and amount of plunge, together with shear sense, of disharmonic folds.
axes broadly lie on a small circle indicating subsequent flexural folding about an approximately N.N.W.-S.S.E. axis.

On Fig. 13. Stereonet B, the poles to foliations in the gneisses in this zone of prominent N-S axis folding can be seen to lie on a girdle indicating a fold axis plunging at a low angle (20° or even less) just to the W. of N. However, although there is a maximum of fold axes plunging in this direction folds of similar flexural disharmonic style in fact show considerable variation in the direction of plunge of their fold axes. Several plunge at shallow angles to the S.E., while a few have more north-easterly plunges. In fact it can be seen that all these fold axes lie in a great circle which represents the main regional foliation (N.N.W.-S.S.E. strike and shallow dip to E.N.E.).

There is abundant evidence (viz. extensive shearing and mylonitization of quartzite, anorthosite, eclogite etc.) that there has been a considerable degree of tectonic movement in this area. It appears likely that most of this movement was concentrated along low angle thrust planes and that the N-S axis folds are essentially 'drag' folds associated with this movement.

One would expect that such folds would have axes approximately at right angles to the movement direction. Thus if the movement direction is assumed to be from the east up the dip of the regional foliation planes it is likely that most of the drag folds will have approximately N-S horizontal axes. However, in practice it appears that many of the fold axes will be tilted to plunge slightly to the N. or S., while a few will be offset even further and plunge at greater angles between N. through E. to S. An occasional fold axes may even plunge east parallel to the movement direction.
direction.

A comparable distribution of fold axis orientations in disharmonic folds in relation to the movement direction has been recognised in glaciers, salt domes andropy lava flows (personal communication R. Hansen and H. Brueckner, who suggested the analogy with the present case).

The orientation of minor drag folds on the limbs of a major flexural fold is illustrated in Fig. 14. If the direction and amount of plunge of minor drag folds together with their shear sense looking down plunge is measured on any one limb of such a major structure, then the movement direction responsible for the fold formation can be defined. It is obvious that the shear senses of the drag folds are reversed on the different limbs of the major fold structure.

Fig. 13 Stereonet C shows the direction and amount of plunge, together with shear sense, of fold axes of small scale disharmonic folds measured over a restricted vertical sequence (10-20') of gneisses in the Viksvatn area. It can be seen that these fold axes roughly define a great circle which, in fact, represents the regional foliation of the area. The different shear senses of those folds with plunges in the S.E. quadrant from those with plunges to the N.N.W. or in the N.E. quadrant, indicate that an approximately E-W movement direction was responsible for the generation of these folds. It might be thought that the shear senses of these folds on the stereonet indicate that the movement was in fact from west to east, down the dip of the regional foliation planes. However, the movement direction defined by the shear senses of the folds is only a relative feature indicating that in the narrow sequence of gneisses measured here, the upper layers have been moving to the east relative to the/
FIG. 14  IDEAL FLEXURAL FOLD WITH FORM OF MINOR DRAG FOLDS
the lower layers. The complex as a whole may have been moving
towards the west or the east.

It can be seen that the shear senses of the drag folds measured
in this case and the implied movement direction correspond to those
shown on the steeper limbs of the major fold structure (Fig. 14), if one
considers oneself looking north down the plunge of the major fold structure.
Measurements on the middle limb of the major fold structure would, however,
show opposite shear senses of the minor drag folds and hence the opposite
movement direction.

However, it is apparent from both cases that the principal
movement involved in the generation of these disharmonic drag folds was
approximately in the E-W plane. The apparent movement direction of any
particular rock horizon can only be defined relative to its underlying
or overlying rock horizons and hence will show reversals from different rock
horizons despite the fact that the movement direction of the rock complex
as a whole has been in only one direction.

Owing to the occurrences within this movement zone of peridotites
and eclogites, most likely on mineralogical grounds to have been intruded
from lower crustal levels, it is most conceivable that the principal
movement direction of the complex as a whole has been up the regional dip —
in other words from approximately east to west.

However, as the gneiss complex of the region as a whole is
considered as possibly marking a prominent culmination in the Caledonian
basement (Hernes 1957), the alternative possibility that the movement
involved has been essentially a gravity sliding of the mobile gneiss
complex cannot be entirely ruled out.
At one locality of considerable interest just to the north of Viksvatn, a small peridotite mass occurs in the core of a fold with anorthosite forming one of the enclosing folded rock horizons. A sketch of the relationships at this locality is given in Fig. 15. Unfortunately the complex relationships were somewhat obscured by snow at the time of observation. Nevertheless the following features of interest were noted.

Anorthosite, quartzite and gneiss occur folded together in a fold with an axis plunging about 25° just to the east of north, with peridotite apparently in the core of this fold. On the lower limb of the fold anorthosite overlies with slight angular discordance highly mylonitized quartzite and is also somewhat discordant to the overlying gneiss. Clearly there has been extreme movement along this horizon which perhaps marks an important thrust plane. The anorthosite, quartzite and gneiss all show intense approximately N-S axis microfolding.

There is also intensive small scale disharmonic folding within the small peridotite lens. The direction and amount of plunge, together with shear sense, of these folds is plotted in Fig. 13, Stereonet D. The majority of these folds have approximately N-S axes, mostly plunging at small angles to the north. However, in an effort to accurately define the movement direction from the reversal in shear sense of these folds, fold axes plunging to intermediate directions between north through east to south were particularly sought for and measured.

As with the plot of disharmonic folds in the gneisses (Fig. 13, Stereonet C) the fold axes of these folds lie in a great circle, in this case representing the principal foliation plane orientation in the peridotite mass.
Figure 15. Sketch of relationships around the small dunite pod at the northern end of Viksvatn.
The switch in shear senses of these microfolds similarly indicates that the movement direction responsible for the generation of these folds was approximately in the E-W plane.

Anorthosite is particularly abundant in the gneiss complex of the Viksvatn area. The anorthosite horizons are generally irregular in thickness along strike, thick horizons tens or even a hundred or more feet in thickness splitting up into a number of different horizons interlayered within mafic gneisses and amphibolites. Although the anorthosite masses are broadly concordant with the gneiss along strike, in detail they can often be seen to be discordant at their contacts with the gneisses. The internal foliation in the anorthosites can often be seen to be at an angular discordance to that in the gneiss, while different layers within the anorthosites sometimes also show angular discordancies. The anorthosites can also be observed to be brecciated in places, with small fractures separating slightly displaced anorthosite layers. The anorthosite is often intricately involved with the adjacent gneiss, sometimes interveined with it and often interfolded with it in N-S axis folds on both a major and minor scale. Angular blocks of anorthosite and cores of anorthosite interfolded with the gneiss can be observed torn off from the main anorthosite horizons.

Schieren of amphibolite within the anorthosites are generally rare. However, at the southern end of Viksvatn a large mass of highly retrogressed and extensively mylonitized eclogite occurs within the main anorthosite horizon. The eclogite and anorthosite have become intricately involved with much of the eclogite having been sheared out into thin schieren of amphibolite within the anorthosite (Plate 4A).

The anorthosites have clearly been affected by the N-S axis flexural folding/
folding on both a major and minor scale. On the other hand there is little evidence of indisputably earlier E-W axis folds with axial plane schistosities. Occasionally, however, crude lineations plunging at moderate angles eastwards are apparently developed on the foliation surfaces, and may represent first fold lineations.

In summary it appears that the tectonic history of this area has been to a large extent dominated by considerable late stage movements and the associated development of roughly N-S axis disharmonic drag folds. It is considered most likely that the movement was concentrated along low angle thrust planes, and that the principal movement direction of the gneiss complex as a whole was up the dip of the regional foliation, namely from east to west.

The thrusting and N-S axis folding can clearly be demonstrated to have been later than the prominent E.N.E. plunging fold structures and lineations ubiquitous throughout the Tafjord district.

At first it was considered that there might be some correlation between the thrusting and the intrusion of the abundant anorthosite masses which occur in this area. The anorthosites clearly have tectonic contacts with the adjacent gneisses while their cataclastic textures, minor displacement features and internal foliation discordancies all suggest that considerable movement has occurred within them.

The relationship of the anorthosites to the earlier E-W axis folding is uncertain on the evidence witnessed in the Viksvatn area. However, the occurrence of anorthosite and peridotite clearly folded about the earlier E-W axes in the Kaldhussaeter area, perhaps indicates the emplacement of these rock masses before the ensuing thrusting and N-S axis/
axis folding which has occurred in the Viksvatn area.

B. THE PERIDOTITE OCCURRENCE NEAR HORNINDAL, NORDFJORD

A large peridotite mass occurs about 10 kilometers north-east of Hornindal, mostly outcropping on the hillside on the north side of the road to Hellesylt, Sunnvlvfjord.

The most important feature of this peridotite mass is the occurrence within it of a thick lens of eclogite. This eclogite is texturally and mineralogically similar to eclogite which occurs as lenses in the gneisses adjacent to the peridotite and to the country rock type eclogite which occurs within the Kalskaret peridotite mass. This eclogite lens similar to the latter occurrence has a sharp contact with the peridotite being altered to amphibolite along the contact. It in no respect resembles the indigenous garnet peridotite layers found in the Kalskaret and Almklovâlen peridotites.

The peridotite itself shows a thin alteration band of chlorite, talc and enstatite at the well exposed top contact of the eclogite lens, followed by considerable enstatite enrichment in the peridotite immediately above the eclogite lens. There therefore seems to have been limited reaction between the eclogite and the peridotite, perhaps with transfer of a certain amount of SiO₂ to the peridotite.

Chemical analysis has been performed on an accumulated sample (H₄) of several small eclogite specimens collected from this eclogite lens in the peridotite. An eclogite sample (H₂) collected from a lens directly enclosed in the adjacent gneisses has also been analysed.
C. THE PERIDOTITE OUTCROPS IN THE AIKLOVDALEN DISTRICT.

The peridotite occurrences in this area have been studied by Lappin (1962) and are represented in Fig. 16. The arcuate form of the peridotite outcrops is intriguing and suggests that they occur principally in one major folded horizon. However, Lappin (1962) has concluded that the peridotites were probably intruded in a late shear zone after the regional fold episodes.

Garnet peridotites occur in all three major peridotite masses and may represent a folded and boudinaged original stratigraphic horizon within the peridotites (Lappin 1962).

As in the case of the Kalskaret garnet peridotites, these garnet peridotites form primary layers in the general garnet free peridotites - being interlayered with the latter, their lens like form probably being the result of the relative incompetencies of the garnetiferous and non-garnetiferous peridotites under the tectonic conditions responsible for the emplacement of these rock masses.

The intricate interlayering of garnetiferous and non-garnetiferous peridotites, particularly in the Lien mass, clearly indicates that the garnetiferous layers are indigenous to the peridotites and that they do not represent tectonically emplaced 'foreign' inclusions (viz. the country-rock type eclogite inclusions in the Kaldhussaeter and Hornindal peridotite masses).

As in the Kalskaret garnet peridotites, intense microfolding also renders particular layers impersistent laterally.

Bulk rock and mineral compositions for several Alklovdalen garnet peridotite/
Fig. 16. Almklovdaalen Peridotite Outcrops.

After Lappin (1962).

Peridotite

Garnet Peridotite Horizons

Eclogite Lenses in Gneiss.

KMS.
peridotite assemblages have been given by Lappin (1962), O'Hara and Mercy (1963), and Mercy and O'Hara (1965a,b).

Garnet peridotites occur most extensively in the Lien peridotite mass, where they are interlayered with garnet free peridotites over a zone about 200 metres wide. The following general sequence of rock types was recognized in upward succession from the northern contact.

1. Garnet free peridotites with prominent chlorite flakes.
2. Thin ribs of partly amphibolitized garnet poor peridotites and lherzolites within garnet free peridotites.
3. Prominent "eclogitic" horizon with an interlayered sequence of garnet (20-40%) and clinopyroxene (20-40%) rich rocks, namely garnet websterites and lherzolites, with only a little or no olivine and amounts of orthopyroxene subordinate to clinopyroxene.
4. Series of garnet lherzolites (garnet 10-20%, clinopyroxene 5-10%) with a few thin "eclogitic" bands.
5. Amphibolitized garnet poor dunites.
6. Prominent zone of chromite dunites - perhaps 10 metres thick.
7. Thick sequence of garnet free, chlorite rich peridotites.
8. Thin bands of garnetiferous dunites interlayered with chlorite peridotites.
9. Broad zone (perhaps 100 metres) consisting predominately of garnetiferous dunites (~5-10% garnet) but containing certain garnet and clinopyroxene rich bands, and also garnet free layers. In many of these rocks only amphibole pseudomorphs after garnet persist.
10. Garnet free peridotites again with prominent chlorite.

All the analysed mineral assembles from Lien - namely N69, N70, and N71 (O'Hara and Mercy (1963), Mercy and O'Hara (1965a,b)) , and 205a/
205a, 209 and 294 (Lappin 1962) - are of the more 'eclogitic' rock types. However, the fact that these particular rock types form only a minor fraction of the total sequence of garnetiferous peridotites should not be overlooked. A garnet poor peridotite with about 5-10% garnet as an average, is in fact the most abundant garnet bearing peridotite rock type. Unfortunately the analysis of just such a rock type from Lien is lacking at present.

There is just a very narrow horizon of rather amphibolitized garnet peridotites in the Grubse peridotite mass. One thin layer of garnet websterite was noted among typical garnet lherzolites in which the garnet is extensively amphibolitized and clinopyroxene appears relatively abundant.

In the Hellebust–Gudsal peridotite mass there is an important outcrop of garnet peridotites just behind the prominent Rodhaugen scarp feature. They are again interlayered with garnet-free peridotites but more restricted in occurrence than in the Lien mass. Of the garnetiferous rock types, garnet lherzolites with significant amounts of both olivine and orthopyroxene are most abundant, although a few thin layers of garnet websterites occur. However, true bimineralic eclogites (GNT and CPX only) are absent as they also are among the Lien and Grubse garnet peridotites. At all these localities there appears to be at least a little olivine or orthopyroxene, or both, even in the most garnet and clinopyroxene rich rock types.

However, a little to the north of the garnet peridotite occurrence just described there is in fact a lens of eclogite within the Hellebust–Gudsal peridotite mass. It is best exposed in a small stream section near Ekraemspeter farm. This rock type differs from all the other Almklodalen garnet peridotites in being a true eclogite consisting almost entirely of garnet/
garnet and clinopyroxene with only minor amphibole and a little accessory rutile. It contains neither olivine nor orthopyroxene. This is the rock type referred to as the Rødhaugen eclogite by Eskola (1921) and O'Hara and Mercy (1963). A sample (N75) has been analysed (Mercy and O'Hara, 1965a), while Eskola (1921) has provided analyses of its constituent minerals (garnet and clinopyroxene) and has calculated a bulk analysis.

This essentially bimineralic eclogite occurs in a layer about 4.6" thick with an amphibolitized margin against 'normal' garnet free peridotite. No intermediate garnet peridotite rock types occur at this locality nor is there any interlayering of the eclogite and peridotite. The eclogite seems to be confined to this one horizon and to be remarkably homogeneous in mineralogy within that layer. This contrasts with the remarkable heterogeneity in mineralogy of the garnet peridotites recognized elsewhere, which show considerable variations in the modal proportions of olivine, garnet, clinopyroxene and orthopyroxene over very small thicknesses.

The petrogenesis of the Rødhaugen eclogite will be considered in a subsequent chapter.

D. **THE GARNET PERIDOTITES OF UGELVIK, OTTERØY.**

Two small isolated occurrences of peridotite occur on the northern side of the island of Otterøy in Moldefjord (Fig.17): one occurrence in the village of Ugervik, the other about one kilometre north-east of the village.

The most distinctive feature of both peridotite masses when compared with those described in the Tafjord and Almklovdalen districts is the widespread/
FIG. 17

PERIDOTITE OUTCROPS

SCALE APPROX. 1:50000
widespread occurrence of garnet throughout much of the peridotite. The garnet peridotites at these other localities are remarkably restricted in occurrence within garnet free peridotites and often contain high modal proportions of garnet (20-40%) and pyroxenes (30-60%).

Here, however, peridotites with garnet or pseudomorphs after garnet probably comprise rather more than half the total volume of the peridotite masses, the modal proportions of garnet in the garnet peridotites averaging about 10% by volume or even slightly less. The modal proportions of the pyroxenes are also low, averaging between 6-10% by volume in the garnet peridotites.

Furthermore the general degree of serpentinization of the olivine in these peridotites is quite high, 50% or more, and quite distinctive from the generally remarkably fresh character of the olivines from the peridotites of Tafjord and Almklovdaln.

5 rock samples from the Ugelvik peridotites were selected for bulk rock and constituent mineral analyses. The modes of these rocks are given in Table No. 4. Of these rock samples, U2 and U16 are representative of the generally garnet-poor peridotites, and U3 of the garnet free peridotite. U8 represents a garnet peridotite with approximately maximum garnet content (~23%) of general occurrence. U6 is an unusual rock type being a garnet peridotite with an abnormally high concentration of clinopyroxene (~47%). This rock type is of very limited occurrence, however, occurring as a thin band about 1.6" thick at its maximum which can only be followed for about 20' laterally. Within this band there is microlayering with considerable variations in garnet content, but the predominate mineral is clinopyroxene.

At a few other localities layers of garnet peridotites, apparently composed/
composed mostly of garnet and clinopyroxene, have been observed but these are generally very thin (1" or less in thickness) and form a very insignificant fraction of the total volume of the peridotite masses.

The localities for the collection of analysed samples $U_2$, $U_3$, $U_6$ and $U_{16}$ are noted on Fig. 16. $U_8$ was collected from the other peridotite mass, to the north-east of the village of Ugelvik.

The peridotite rock types are generally quite massive and no highly sheared 'platy' peridotite horizons occur as in the peridotites of the Tafojord and Almklovaleden districts. The peridotite is often well jointed, however.

It is often difficult to ascertain the compositional layering, especially where the peridotite is non-garnetiferous. However, slight variations in the proportions of garnet in the garnetiferous layers often produces a clear compositional layering.

At first glance it was considered that it might be possible to map out broad zones of garnet peridotites with intervening non-garnetiferous zones. However, although broadly speaking such zones do exist, in detail it was found to be an impossible task to map them out because such zones are often impersistent laterally and predominately non-garnetiferous layers often contain thin isolated garnetiferous layers and vise versa. There are, however, certain zones where garnet peridotites predominate and others where they are generally absent.

At no locality is the actual contact between peridotite and gneiss exposed. However, the gneiss appears to be broadly concordant with the peridotites at the contacts, along which both appear to have suffered a greater than normal degree of shearing and deformation. Peridotite and/
Peridotite outcrop at the village of Ugelvik on the island of Otterøya.
and gneiss can be observed most closely in juxtaposition along the eastern margin of the north-eastern peridotite mass. Specimens of garnet peridotite from here show cataclastic textures broadly similar to those observed in the Kalskaret garnet peridotites, with straining, cataclasis, recrystallization and dimensional orientation of the constituent minerals.

Such cataclastic textures together with the sheared nature of the adjacent gneisses and lack of any recognizable high temperature thermal aureole around the peridotite masses, therefore point to the tectonic emplacement of these peridotite masses into their present position.

The gneiss complex adjacent to the Ugelvik peridotites consists predominately of layered acid gneisses with subordinate mafic gneisses. The latter typically occur as restricted lenses or boudins within the acid gneisses.

The typical acid gneiss assemblage is: large porphyroblasts of antiperthitic oligoclase + microcline + quartz + hornblende + biotite + iron ore + sphene.

A study of the basic gneiss assemblages indicates that the metamorphic grade was probably initially of granulite facies but later regressed to almandine-amphibolite facies.

The critical assemblages are:

1. Garnet + augite + hornblende + biotite + plagioclase + iron ore + rutile + sphene + quartz.
2. Garnet + augite + hornblende + hypersthene + plagioclase + biotite + iron ore + rutile + sphene + quartz.

In these assemblages the pale green augite typically has alteration rims of hornblende and the rutile of sphene. The biotite seems primary, however.
However, the primary assemblage garnet + augite + hypersthene + biotite + plagioclase + rutile + iron ore + quartz does not fall into any of the subdivisions of the amphibolite, pyroxene hornfels or granulite facies of Turner and Verhoogen (1960).

Strictly the garnet-clinopyroxene join is only stable in the silica deficient basic assemblages of the granulite facies except in iron-rich assemblages. The assemblage garnet + clinopyroxene + quartz is only stable in the eclogite facies, but then plagioclase is unstable. Furthermore microscopic examination of the gneiss assemblages in question indicates that texturally they do not resemble retrogressed eclogite facies mineral assemblages. The garnets do not have kelyphitic reaction rims and the clinopyroxene does not show a symplektite alteration to an intergrowth of hornblende and plagioclase, rather only a broad hornblende rim.

It is therefore considered that these mineral assemblages, although they probably show some degree of chemical disequilibrium, probably correspond most closely to the hornblende-granulite sub-facies of the granulite facies (Turner and Verhoogen 1960). Subsequently they have been partially retrograded to an almandine-amphibolite facies assemblage.

Highly retrograded eclogites are occasionally preserved in the gneiss complex. These occur generally as small lenses of boudin form, with extensive outer rims of amphibolite (Plate 4B). The primary mineral assemblage of these rocks appears to have been: garnet + clinopyroxene + orthopyroxene + phlogopitic mica + quartz + rutile. Commonly, however, the clinopyroxene is almost entirely altered to a fine-grained amphibole-plagioclase/
plagioclase symplektite and coarser grained amphibole forms kelyphitic rims round the garnets. These eclogites commonly contain appreciably more quartz than the garnet-pyroxene bearing granulites. Epidote and clinozoisite occur as secondary minerals in some of the altered eclogites.

The foliation planes in the gneisses strike approximately E.N.E.-W.S.W. and have steep to near vertical dips. A study of microfolding in the gneisses indicates that tight isoclinal folds with steep axial planes and near horizontal E.N.E-W.S.W. fold axes predominates. Such folds generally develop an axial plane schistosity.

Compositional banding in the peridotites is broadly conformable to the foliation planes in the adjacent gneisses, and likewise shows the influence of E.N.E.-W.S.W. axis folding. Prominent sets of joints in the peridotites can be interpreted as ac joint planes, perpendicular to the inferred fold axis (b) direction.

It is noticeable that the large porphyroblasts of antiperthitic oligoclase and microcline so diagnostic of the acid gneisses are highly deformed during the late fold and shear movements, being significantly so deformed in the sheared gneisses adjacent to the peridotite masses. This suggests emplacement of the peridotites after the main injection, metasomatism and metamorphism of the gneiss complex.

In summary the following seems to have been the most likely sequence of events relating to the emplacement of the Ugelvik garnet peridotites into their present environment.

1. Development of gneiss complex of broadly granulite facies grade from a series of psammites with minor interlayered pelites and perhaps basic igneous bodies.

2./
2. Tectonic emplacement of the peridotite masses into this complex although not necessarily at the present level. Possibly also tectonic intrusion of quartz eclogite masses into the gneiss complex at this period.

3. Partial retrogression of the gneiss complex to almandine-amphibolite facies grade, probably associated with an extensive period of injection and metasomatism—granitic veins, quartz pegmatite veins, and growth of felspar porphyroblasts.

Tectonic conditions involved a stress system with a near vertical movement direction and the development of tight isoclinal shear folds with near horizontal E.N.E.-W.S.W. fold axes.

4. Tectonic emplacement of the peridotite masses into their present position under the influence of the above stress system. The felspar porphyroblasts are highly sheared against the peridotites. In fact the whole complex is highly sheared even quite remote from the peridotite masses. There are also prominent zones of brecciation in the gneisses.

The peridotites may not have moved up very far in this last movement phase, but conversely they may have moved great distances.

The tectonic history of the gneiss complex seems to have been one of an initially highly mobile complex giving way with lowering of the intensity of injection and metamorphism to a cooler more rigid highly sheared complex.

E. GENERAL DISCUSSION ON THE CONTACT RELATIONS OF PERIDOTITE MASSES.

The Norwegian peridotite masses have generally concordant and sheared/
sheared contacts with the adjacent gneisses, with a distinct lack of any high temperature contact metamorphic effects.

As these peridotites are composed mostly of a forsteritic olivine ($\sim 6-8\%$ Fa) and thefusion point of such an olivine is extremely high ($\sim 1800-1890^\circ C$) one would have expected very marked thermal aureoles around these peridotite bodies if they represent igneous intrusions into their present environment. However, there is a marked absence in all cases observed of any such high temperature thermal aureole.

The lack of evidence of a high temperature intrusive origin is a familiar problem in the discussion of the petrogenesis of alpine-type peridotite masses in orogenic belts throughout the world. Turner and Verhoogen (1960) have given a good account of the controversy surrounding their origin.

Hess has for long been the chief exponent of theories envisaging the magmatic intrusion of the rock masses. He has advocated (1937, 1938) the existence of a low temperature hydrous ultramafic magma. However, the existence of a serpentinite magma has been discounted by Bowen and Tuttle (1949), and Kitahara et al (1966) on the basis of experimental evidence in the system MgO-SiO$_2$-H$_2$O, which indicates that no liquid forms at water pressures up to 30 kilobars and temperatures up to 900-1000$^\circ C$. Bowen (1928), ruling out the existence of a primary peridotite magma, proposed that the peridotites were intruded as a crystal mush lubricated perhaps by small quantities of intergranular magmatic liquid, or even water vapour, to impart the necessary degree of mobility.

Thayer (1960) has attempted to distinguish between alpine-type peridotites and truly stratiform gabbro-peridotite complexes on the basis of differences in form, structure, composition, texture and contact metamorphic/
metamorphic effects.

However, in recent years several orogenic peridotite masses have been described (cf. MacKenzie 1960; Green 1964a,b; Rothstein 1957,61; Smith and MacGregor 1960; Challis 1965a,b) which show some of the features of stratiform peridotites such as layering, accumulative textures and high temperature contact metamorphism. Green (1964a,b), MacKenzie (1960), Smith and MacGregor (1960) and Challis (1965a,b) describe the development of pyroxene granulites in high temperature thermal aureoles around the Lizard, Tinaquillo, Mount Albert and Red Hills peridotites, respectively—indicating temperatures of intrusion probably in the order of 800–1000°C.

The recognition of such high-grade thermal aureoles, is contrary to the classical observation that alpine-type peridotites have little or no associated contact metamorphism. However, the inferred temperatures of intrusion still indicate that these peridotites were far from being completely molten at their time of emplacement into their present environments. It appears likely that these peridotites represent magmatic accumulates crystallized initially at greater depths in the crust and subsequently tectonically emplaced into their present environment before they were completely cooled.

The problem still remains, however, as to the composition of the original magma from which the peridotites have recrystallized, namely whether a primary peridotite magma does exist or whether these peridotites are accumulates of a more basaltic magma. The latter possibility leaves the problem of the removal of the overlying portion (liquid or crystal accumulate) before the peridotite mass can be tectonically intruded (Green 1964a).

However, the hypothesis of Challis (1965a) that the Red Hills, Dun Mountain and/
and Red Mountain peridotites in New Zealand originally represented differentiated accumulates of deep level magma chambers associated with Permian spilitic volcanics and tuffs, appears quite convincing.

It is therefore possible that the only significant distinction between alpine type and stratiform peridotites may be their differing environments of crystallization.

Alpine type peridotites crystallize in active orogenic belts and are therefore liable to remobilization and tectonic intrusion into higher crustal levels. The proliferation of such peridotites in this type of environment perhaps indicates that such orogenic conditions favour the production of ultrabasic magma at depth, presumably by fusion of the uppermost mantle. To what extent cataclastic textures obliterate primary accumulative textures in these peridotite masses will depend on how far these masses have moved in a quasi-crystalline state. It is to be expected that such masses may or may not develop a significant thermal aureole in their final resting place depending on how completely they had cooled by the time of their final emplacement, and with their rate of emplacement with respect to their rate of heat loss.

Large peridotite masses in young orogenic belts (viz. Cuba, Puerto Rico and Philippines) may, however, have never had a magmatic origin in the crust but may represent altered sub-oceanic mantle material thrust into its present position, as suggested by Hess (1960a).

Stratiform peridotites, on the other hand, are of magmatic crystallization in stable crustal regions and hence have not suffered reintrusion and cataclasis. They display typical igneous crystallization features viz. accumulative textures, cryptic layering and compositional differentiation/
differentiation across the larger masses.

Returning to the case of the Norwegian peridotites, it is apparent that while their tectonic contacts and lack of any significant high temperature thermal aureole rule out igneous crystallization in situ, they do not rule out igneous crystallization at depth within the crust. The possibility must be considered that these masses have only moved short distances from their original site of crystallization just leaving their contact aureole behind as it were. However, the total absence of any apparent contact metamorphic effects tends to indicate that these masses were very cool at the time of their emplacement which, along with the extreme cataclastic textures witnessed in many of these peridotite masses, suggests considerable tectonic transport of these rock masses.

In the case of the Norwegian peridotites where the actual contact is exposed there is characteristically only a narrow alteration zone, a foot or so in width, resulting from limited reaction between the peridotite and the adjacent gneisses. Within this zone, the actual contact is thought to lie between a talc + magnesite + pale green tremolite-actinolite amphibole rock against the peridotite and a biotite + dark green hornblende rock on the country rock side. Elsewhere, somewhat wider lenses containing the same low temperature hydrous mineral phases occur spasmodically along the contacts. No development of new high temperature minerals, particularly pyroxenes, is ever evident in the contact gneisses.

An absence of quartz is generally conspicuous in the immediately adjacent gneisses, which are generally finely foliated felsic and amphibolitic gneisses. At two localities in the Tafjord district corundum was noted in the gneisses immediately adjacent to peridotite masses.

It/
It therefore appears that the strong chemical contrast between the peridotites and the adjacent gneisses has resulted in limited chemical transfer between them. In particular it would appear that minor transfer of SiO₂ into the peridotite margins has commonly occurred leaving behind a quartz-free felsic gneiss with occasional corundum.

The contact features noted between the Norwegian peridotites and their adjacent country rocks are similar to those described by Phillips and Hess (1936) as being the typical hydrothermal metamorphic differentiation effects between peridotite (or serpentinite) and siliceous country rocks. The limited extent of the contact alteration zone around the Norwegian peridotites suggests that the reaction occurred after the main injection and metamorphism of the gneiss complex. Furthermore the poorly foliated or unfoliated nature of the alteration zone indicates that the reaction occurred after the tectonic emplacement movements which have sheared the contacts.

F. TEXTURAL EVIDENCE SUGGESTING TECTONIC EMPLACEMENT OF THE PERIDOTITE MASSES.

Remarkable cataclastic textures are developed in the Tafjord, Almklovdalen and Ugelvik peridotites, in particular in pyroxene-rich garnet peridotites from these localities. (Plate 5).

Orthopyroxene is the mineral which shows the most remarkable deformation and cataclasis textures. With only a mild degree of deformation these grains appear to have developed very fine regular strain lamellae //⟨100⟩ and broader more irregular strain lamellae //⟨001⟩. However, with further straining the orthopyroxenes have become highly bent and deformed, with kink bands and fractures appearing. Further deformation has resulted in total cataclasis so that only a fine grained mosaic of orthopyroxene grains remains.
remains. Normally, however, a few highly deformed porphyroclasts remain showing clear signs of marginal cataclasis.

Clinopyroxene in these rocks shows similarly extensive straining, deformation and cataclastic textures.

The finely cataclased pyroxene grains appear to have recrystallized to strain-free grains.

The garnet in these rocks is often broken up into trains of smaller grains, and even appears in some cases to have recrystallized along planes of strain running through some of the larger crystals. Also neighbouring garnets are cataclased where in contact with each other.

In the garnet peridotites olivine generally shows less remarkable deformation textures than its coexisting minerals. Relict olivine porphyroclasts are rare, most of the olivine having been cataclased and recrystallized to a fine grained mosaic of strain free grains.

However, in some of the dunites from Almklovelden and Tafjord large strained and fractured olivine porphyroclasts up to several inches in length, have been noted. These porphyroclasts often have irregular strain lamellae and can be observed to have recrystallized both marginally and along planes of stress running through them to much smaller strain free olivine grains.

The olivine fabric within certain zones in these peridotite masses, presumably those zones along which internal movement has been most concentrated, shows striking dimensional and lattice preferred orientations (Plate 1). Although detailed petrofabric work on these rocks has not been performed in the course of this study, it is important to discuss the significance of the fabric of these rocks to hypotheses on the origin, in particular the mode of emplacement, of these rock masses.

A/
A certain degree of preferred orientation of olivine has been noted in peridotites from widely different environments - deformed orogenic peridotite masses, stratiform peridotite complexes, peridotite nodules in basalts and even in olivine basalt dykes. The interpretations of the origin of these olivine fabrics vary considerably and are the subject of a certain amount of controversy (c.f. Phillips (1938), Turner (1942), Brothers (1959), Battey (1960), Colée (1963) and others).

It is apparent that a number of different processes can produce preferred orientation olivine fabrics, namely normal igneous accumulation processes, flow in a liquid medium, laminar flow controlled by intragranular gliding and/or recrystallization in a quasicrystalline medium, or post intrusive deformation. The problem is to distinguish between the olivine fabrics formed by these different processes. The superposition of fabrics produced through a combination of these processes will obviously lead to complications. Thus the fabric of alpine-type peridotites crystallized at depth within the crust and subsequently tectonically emplaced and folded, may in fact be the superposition of three different fabrics and hence very difficult to interpret.

In the case of the Norwegian peridotites with ample evidence of extensive, straining, shearing and cataclasis a purely magmatic origin of the olivine fabric can clearly be ruled out. At the most only relicts of an original igneous accumulative fabric would be expected to survive the extensive deformation history of these peridotites.

The problem remains, however, to distinguish between the crystal orientation effects due to the forces of tectonic emplacement and those resulting from late foliation deformations. It therefore appears that the dating of/
of the emplacement of these peridotite masses with respect to the regional fold episodes recognized, on the basis of petrofabric analysis alone, should be approached with caution.

A definite relationship between the regional fold axis and the olivine lattice orientation in peridotite bodies has been recognized by several workers (Ladurner 1956, Battey 1960, Yoshino 1964), with preferred orientation of the (010) plane of the olivines in the foliation plane and their crystallographic c-axes parallel to the fold axis. However, the importance of laminar flow in a quasisolid medium in the development of preferred orientation fabrics in tectonically intruded peridotite masses has been stressed by Turner (1942) and Brothers (1959). Ernsr (1935) considered the fabric of an Almklovdalen dunite to be of deformational origin, noting the flattening of olivine grains parallel to the foliation and the elongation of some grains in the fold axis direction.

Lappin (1962) considered the olivine orientation in the Almklovdalen peridotites studied by him to be too variable to allow definition of a single mechanism of origin. However, he concluded that the fabric of most of these peridotites was probably the result of the tectonic emplacement origin of these rocks, moving through internal recrystallization.

To the present author it appears likely that for the most part the foliation in the Almklovdalen and Tafjord dunites is the result of laminar flow through internal recrystallization during tectonic emplacement of these rock masses in an essentially crystalline state. In highly sheared zones, however, the fissility is essentially a schistosity of cataclastic origin resulting from dimensional orientation of elongated fragments of ruptured olivine grains.

Demonstration of a close relationship between olivine orientation and/
and the regional axis of folding would point to the emplacement of these peridotites either prekinematically or synkinematically with the folding. With the clear demonstration that at least the peridotites of the Tafjord district have been folded, it seems likely that the present olivine orientation fabric of the these peridotites represents essentially one produced during the tectonic emplacement of these masses but somewhat modified by synchronous or later folding. The possibility of the existence of an original igneous fabric modified by the effects of later tectonic emplacement and folding must still be considered, however.

Cataclastic textures and preferred dimensional orientation fabrics are not so extensively developed in the peridotites at Ugelvik, as they are in those from Tafjord and Almklovdalen. However, such textures and fabrics are developed in certain specimens near the contacts of the Ugelvik peridotite masses. In these rocks the olivine no longer forms a uniform subhedral mosaic as in most of the Ugelvik peridotites, rather large strained olivine porphyroclasts can be observed recrystallizing marginally to smaller dimensionally orientated strain-free olivines. Also garnet and pyroxene grains show the effects of extreme deformation and cataclasis.

Textural evidence thus points to the tectonic emplacement of the Ugelvik peridotites, as well as those at Tafjord and Almklovdalen.

G. SUMMARY OF CONCLUSIONS

From the details, discussed in this chapter, of the field occurrences of the rock types under investigation it is considered that the following conclusions are justified.

1. Cataclastic textures, preferred orientation fabrics, the sheared nature
of their contacts and the lack of any recognizable high temperature thermal aureole all point to the tectonic emplacement of the peridotite masses studied and are contrary to Eskola's (1921) hypothesis of the igneous crystallization of these masses in their present environment.

2. The fresh nature of the Tafjord and Almklovdalen peridotites and the very limited extent of the hydrothermal contact alteration zone around them, suggests emplacement after the main injection and metamorphism of the gneiss complex.

3. In the case of the Tafjord peridotites, at least, the recognition of E-N-E. plunging fold structures in them indicates the likely influence during their tectonic emplacement of the same stress system which was responsible for repeated folding about the same axis in the gneiss complex. With the evidence of the diapiric intrusive nature of certain of the peridotite masses (Kalskaret, Ranukdalen) it appears most likely that this folding of the peridotites occurred virtually synchronously with their emplacement into contemporaneously forming fold structures in the gneisses, rather than the peridotites having been folded together with the gneisses subsequent to their emplacement.

Peridotite is also clearly affected by later N-S axes folding in the Viksvatn area, one of prominent late tectonic movement. Thus while Lappin (1962) has considered that the peridotites of the Almklovdalen district were emplaced in a prominent shear zone after the regional E-W axis and N-S axis folding, the evidence from the peridotites of the Tafjord district suggests emplacement somewhat earlier in the tectonic history but still late in the metamorphic and injection history of the gneiss complex.

4. The restricted occurrences of garnetiferous peridotites within the Kalskaret, and
and three of the Almklovldalen, peridotite masses are interlayered with the 'normal' garnet-free peridotites and appear indigenous to their enclosing host rocks.

5. The Rødhaugen eclogite occurrence is distinctive from the other garnet peridotite occurrences in being essentially bimineralic (without olivine or orthopyroxene), mineralogically homogeneous, and not interlayered with the peridotite which encloses it.

6. The occurrences of 'country-rock' type eclogite within the Kaldhuassaeter and Hornindal peridotites are contrary to Eskola's (1921) two-fold subdivision of eclogite occurrences in this region of Norway. These eclogite masses appear to represent tectonic inclusions in the peridotite and are clearly distinct from the garnet peridotite inclusions recognized elsewhere.

7. The widespread occurrence of garnetiferous rock types within the peridotite masses at Ugelvik, the generally low modal proportions of garnet and pyroxene in these rock types, and their highly serpentinized nature all render them somewhat distinctive from the other garnet peridotite occurrences.

8. Cataclastic textures, sheared contacts, minor marginal and internal discordancies, and the lack of thermal contact metamorphic effects all indicate the likely tectonic emplacement of the anorthosite masses in the Tafjord district. The influence of E.N.E. plunging fold structures in the Kalddhussaeter area, and N-S axis folding in the Viksvatn area suggest emplacement of the anorthosites more or less contemporaneously with the peridotites.

9. Tectonic emplacement of the eclogite lenses in the gneisses cannot be proved.
proved irrevocably from field evidence alone. Subsequent boudinage and the high degree of amphibolitization have obscured their contact relations. They were probably emplaced and boudinaged before the main injection and metamorphism of the gneiss complex, and hence represent earlier intrusions than the peridotites or anorthosites.

It should be noted that of the above conclusions - 1, 2, 4, and 9 are wholly consistent with the conclusions of Lappin (1962) and O'Hara and Mercy (1963), while conclusion 8 is partially consistent.
CHAPTER III. MINERAL PARAGENESIS IN THE PERIDOTITES

A discussion of the primary and secondary mineral assemblages of the peridotites, in particular the garnet peridotites of Kalskaret, Almklovdalen and Ugelvik, is undertaken in this chapter to elucidate their likely conditions of formation.

Mineralogical data for the Kalskaret garnet peridotites (O'Hara and Mercy 1963, Mercy and O'Hara 1965b) have been supplemented by further data on two other mineral assemblages (T153 and T96).

Mineral compositions of certain Almklovdalen garnet peridotite assemblages have been given by Lappin (1962), O'Hara and Mercy (1963), and Mercy and O'Hara (1965b).

Mineral compositions for analysed Ugelvik garnet peridotite assemblages (U₂, U₆, U₈ and U₁₆) are presented here and compared with the existing data on the other Norwegian garnet peridotite occurrences.

The new mineral composition data together with calculated structural formulae are given in Tables No. 1-3. The petrographic details of the analysed mineral assemblages are given in Appendix III, the modes of the assemblages in Table No. 4. Estimation of the proportion by volume of the minerals in many of these assemblages is difficult as a result of their cataclastic textures and the resultant variations in grain size.

In the last chapter it was noted how the widespread occurrence of garnetiferous peridotites at Ugelvik and their generally low modal proportions of garnet and clinopyroxene contrasts with the very restricted occurrence of garnet peridotites in the Tafjord and Almklovdalen districts and their generally high modal contents of garnet and clinopyroxene.

A study of the mineral composition data indicates that the constituent/
### Table No. 1

#### Olivines

<table>
<thead>
<tr>
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<tbody>
<tr>
<td>SiO₂</td>
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</tr>
<tr>
<td>TiO₂</td>
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<td>0.04</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>0.44</td>
<td>0.35</td>
</tr>
<tr>
<td>Cr₂O₃</td>
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<td>0.01</td>
</tr>
<tr>
<td>Fe₂O₃</td>
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<tr>
<td>FeO</td>
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<td>7.27</td>
</tr>
<tr>
<td>MnO</td>
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<td>0.09</td>
</tr>
<tr>
<td>NiO</td>
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<td>0.40</td>
</tr>
<tr>
<td>MgO</td>
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<td>52.51</td>
</tr>
<tr>
<td>CaO</td>
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<td>0.12</td>
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#### Orthopyroxenes

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<td>0.04</td>
</tr>
<tr>
<td>Al₂O₃</td>
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<td>0.35</td>
</tr>
<tr>
<td>Cr₂O₃</td>
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<td>0.01</td>
</tr>
<tr>
<td>Fe₂O₃</td>
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<tr>
<td>FeO</td>
<td>17.12</td>
<td>7.27</td>
</tr>
<tr>
<td>MnO</td>
<td>0.17</td>
<td>0.09</td>
</tr>
<tr>
<td>NiO</td>
<td>0.36</td>
<td>0.40</td>
</tr>
<tr>
<td>MgO</td>
<td>43.79</td>
<td>52.51</td>
</tr>
<tr>
<td>CaO</td>
<td>0.16</td>
<td>0.12</td>
</tr>
<tr>
<td>Total</td>
<td>100.00</td>
<td>100.00</td>
</tr>
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</table>

*SiO₂* By difference from 100.00%

*FeO* Total iron as FeO

#### Cation Proportions on the Basis of 4-Oxygens

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<td>0.001</td>
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#### Cation Proportions on the Basis of 6-Oxygens

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<td>0.245</td>
</tr>
<tr>
<td>Ti</td>
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</tr>
<tr>
<td>Al³⁺</td>
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<td>0.010</td>
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<tr>
<td>Cr³⁺</td>
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<td>-</td>
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<tr>
<td>Fe²⁺</td>
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<td>0.146</td>
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<td>Mn</td>
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<td>Fe x 100/Fe + Mg</td>
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<td>7.21</td>
</tr>
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Ca 0.85  0.63
Mg 85.50  92.11
Fe²⁺ 15.65  7.26
Fe x 100/Fe + Mg 18.0  7.30
### TABLE NO. 2.

<table>
<thead>
<tr>
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<td>0.07</td>
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<td>0.50</td>
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<tr>
<td>Fe₂O₃</td>
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<td>0.54</td>
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<td>K₂O</td>
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<td>TOTAL</td>
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</table>

*SiO₂ By difference from 100.00%

<table>
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<tr>
<th></th>
<th>ClinoPyroxenes</th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
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<tr>
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<td>T153</td>
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</tr>
<tr>
<td></td>
<td>U₈</td>
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**Jadeite calculated after formation of NaFe³⁺ and NaCr³⁺**
TABLE NO. 3.

<table>
<thead>
<tr>
<th>GARNETS</th>
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<th>T153</th>
<th>T271</th>
<th>U₂</th>
<th>U₆</th>
<th>U₈</th>
<th>U₁₆ ¹</th>
<th>U₁₆ ²</th>
</tr>
</thead>
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<tr>
<td>*SiO₂</td>
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<td>40.89</td>
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<td>40.05</td>
<td>41.75</td>
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<td>0.11</td>
<td>0.12</td>
<td>0.07</td>
<td>0.10</td>
<td>0.28</td>
<td>0.06</td>
<td>0.06</td>
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<td>Al₂O₃</td>
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<td>21.84</td>
<td>20.87</td>
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<td>20.98</td>
<td>20.71</td>
<td>20.68</td>
<td>20.43</td>
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<td>3.99</td>
<td>3.50</td>
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<td>1.93</td>
<td>1.90</td>
<td>2.02</td>
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<td>13.83</td>
<td>19.94</td>
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<td>6.95</td>
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<td>MnO</td>
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<td>0.50</td>
<td>0.44</td>
<td>0.50</td>
<td>0.42</td>
<td>0.52</td>
<td>0.49</td>
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<td>&lt;0.01</td>
<td>&lt;0.01</td>
<td>n.d.</td>
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<td>0.05</td>
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<td>n.d.</td>
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<td>MgO</td>
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<td>17.05</td>
<td>20.11</td>
<td>21.40</td>
<td>20.76</td>
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<tr>
<td>CaO</td>
<td>6.71</td>
<td>4.55</td>
<td>8.58</td>
<td>4.92</td>
<td>6.02</td>
<td>4.79</td>
<td>4.23</td>
<td>4.16</td>
</tr>
<tr>
<td>TOTAL</td>
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<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
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</tbody>
</table>

*SiO₂ By difference from 100.00%

<table>
<thead>
<tr>
<th>CATION PROPORTIONS ON THE BASIS OF 12-OXYGENS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Si</td>
</tr>
<tr>
<td>Ti</td>
</tr>
<tr>
<td>Al³⁺</td>
</tr>
<tr>
<td>Cr³⁺</td>
</tr>
<tr>
<td>Fe³⁺</td>
</tr>
<tr>
<td>Fe²⁺</td>
</tr>
<tr>
<td>Mn</td>
</tr>
<tr>
<td>Ni</td>
</tr>
<tr>
<td>Mg</td>
</tr>
<tr>
<td>Ca</td>
</tr>
<tr>
<td>Ca²⁺</td>
</tr>
<tr>
<td>Mg²⁺</td>
</tr>
<tr>
<td>Fe²⁺</td>
</tr>
<tr>
<td>Mn²⁺</td>
</tr>
</tbody>
</table>
constituent minerals of the Ugelvik garnet peridotites are chemically
distinct from those of the Kalskaret and Almklovdalen garnet peridotites
being more magnesium and chromium rich than the latter.

The primary mineral assemblages of the Kalskaret and Almklovdalen
garnet peridotites are therefore discussed together first, followed by a
discussion of the primary mineral assemblages of the Ugelvik peridotites,
which are then compared and contrasted with the former group.

A. PRIMARY MINERAL ASSEMBLAGES

1. TAFJORD AND ALMKLOVDALEN PERIDOTITES.

The primary mineral assemblage of the predominately garnet free
peridotites from both the Almklovdalen and Tafjord districts consists
mostly of olivine (generally > 90%) with subordinate orthopyroxene and
chromite. The Almklovdalen peridotites also contain small amounts
(about 1%) of clinopyroxene distributed throughout them. No clinopyroxene
is now apparent in the garnet free peridotites of the Tafjord district,
however. Nevertheless the low but significant content of CaO in some of
these peridotites suggests the original presence of clinopyroxene in
these rocks. The CaO is now contained in amphibole which along with chlorite
occurs as a generally sparse secondary mineral in these peridotites. O'Hara
and Mercy (1963) have given the analysis of amphibole from garnet free
Kalskaret peridotite, indicating that the amphibole is a tremolite with a
CaO content of 11.10 wt.%.

The degree of serpentinization of the olivine is generally very
low in the large peridotite masses, except locally along the margins.
### TABLE NO. 4.

**ESTIMATED VOLUME PROPORTIONS OF MINERALS IN ANALYSED SPECIMENS COLLECTED ACROSS THE KALSKARET GARNET PERIDOTITE LENS**

<table>
<thead>
<tr>
<th>SAMPLE NO.</th>
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<th>CPX</th>
<th>OPX</th>
<th>AMPHIBOLE</th>
<th>CLORITE</th>
<th>OPAQUES</th>
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<tbody>
<tr>
<td>833/T200</td>
<td>24</td>
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<tr>
<td>834/T266</td>
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<td>-</td>
<td>-</td>
<td>10</td>
<td>trace</td>
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<td>trace</td>
</tr>
<tr>
<td>835/T150</td>
<td>56</td>
<td>-</td>
<td>trace</td>
<td>trace</td>
<td>31</td>
<td>9</td>
<td>trace</td>
</tr>
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<td>836/T151</td>
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<td>10</td>
<td>-</td>
<td>4</td>
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<td>19</td>
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<td>trace</td>
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<td>-</td>
<td>2</td>
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<tr>
<td>838/T153</td>
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<td>53</td>
<td>5</td>
<td>10</td>
<td>-</td>
<td>2</td>
</tr>
<tr>
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<td>trace</td>
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<td>-</td>
<td>2</td>
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<td>841/T96</td>
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<td>trace</td>
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<td>842/T155</td>
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<td>trace</td>
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<td>54</td>
<td>30</td>
<td>4</td>
</tr>
<tr>
<td>843/T156</td>
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<td>-</td>
<td>40</td>
<td>18</td>
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### ESTIMATED VOLUME PROPORTIONS OF MINERALS IN ANALYSED UGBELVIK PERIDOTITES

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<th>U₆</th>
<th>U₈</th>
<th>U₁₆</th>
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<td>47</td>
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<tr>
<td>SERP. AFTER OPX</td>
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<td>-</td>
<td>-</td>
<td>-</td>
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<td>6</td>
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<td>CLINOPYROXENE</td>
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<td>47</td>
<td>10</td>
<td>-</td>
</tr>
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<td>1</td>
</tr>
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<td>1</td>
<td>-</td>
<td>15</td>
<td>23</td>
<td>6</td>
</tr>
</tbody>
</table>

**NOTE:** Mineral proportions are estimated to be accurate only to within ± 10-15% of the value given particularly in the case of the Kalskarre garnet lherzolites with pronounced cataclastic textures, highly variable grain size distribution and a general inhomogeneous distribution of the minerals.
Chemical and X-ray determination (O'Hara and Mercy 1963) of minerals from the Tafjord garnet free peridotites (olivines 6.6-8.1% Fa, orthopyroxenes 6.6-7.8% Fs.) and from the Almklovdalen garnet free peridotites (olivines 7-8.1% Fa, orthopyroxene 7.5% Fs) show the restricted range in composition of these minerals and their similar composition range from the two districts concerned.

A number of optical 2V determinations on olivines and orthopyroxenes from garnet free peridotites throughout the Tafjord district have been made. These determinations were made on the universal stage by direct measurement across the acute bisectrix. The compositional estimates from these determinations were olivine 4-8% Fa + about 2% Fa, and orthopyroxene 4-7% Fs + about 1% Fs. It is considered that these composition estimates further substantiate the restricted composition range of the olivines and orthopyroxenes from the garnet free peridotites.

With the appearance of garnet, the olivines and orthopyroxenes become significantly more ferriferous in composition and increasingly so with the increasing 'eclogitic' nature (garnet and clinopyroxene content) of the rock type (c.f. Table No.5).

The primary mineral assemblage of the garnetiferous peridotites is Ol + GNT + CPX + OPX + minor opaque ore fraction. The mineralogical heterogeneity of the garnet peridotites from these localities is quite apparent in the field. These rock types show considerable variations in the modal contents of Ol, GNT, CPX and OPX over very small thicknesses (c.f. Table No.4, also Mercy and O'Hara 1965a Table 1A). These variations are related to large variations in the bulk rock composition of closely adjacent rock samples.

The mineralogical data likewise shows considerable variations in mineral/
<table>
<thead>
<tr>
<th></th>
<th>LIEN, AIMKLOVDALEN SPECIMENS</th>
<th></th>
<th></th>
<th></th>
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<th>KALSKARET, TAFJORD SPECIMENS</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>N69</td>
<td>N70</td>
<td>N71</td>
<td>N23</td>
<td>N26</td>
<td>T96</td>
</tr>
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<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>Fa15.4</td>
<td>-</td>
</tr>
<tr>
<td>CPX</td>
<td>Fa11.2</td>
<td>Fa10.2</td>
<td>-</td>
<td>-</td>
<td>Fa16.0</td>
<td>-</td>
</tr>
<tr>
<td>CPX</td>
<td>Of10.8</td>
<td>Of9.7</td>
<td>Of11.5</td>
<td>Of16.6</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>CPX</td>
<td>Ca50 M47.7</td>
<td>Ca46 M51.3</td>
<td>Ca49 M48.5</td>
<td>Ca47.4 Mg48.0</td>
<td>Ca45.7 Mg49.2</td>
<td>Ca46.4 Mg49.3</td>
</tr>
<tr>
<td>GNT</td>
<td>Ca12 Mg66</td>
<td>Ca12 Mg65</td>
<td>Ca12 Mg64</td>
<td>Ca12 Mg60</td>
<td>Ca11 Mg61</td>
<td>Ca13 Mg60</td>
</tr>
<tr>
<td>Cr2O3</td>
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<td>0.32</td>
<td>0.39</td>
<td>0.91</td>
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<td>Wt.%</td>
<td>0.69</td>
<td>0.47</td>
<td>0.20</td>
<td>0.20</td>
<td>0.63</td>
<td>0.47</td>
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<td>0.06</td>
<td>0.04</td>
<td>0.03</td>
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<td>-</td>
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<tr>
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<td>0.15</td>
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<td>TiO2</td>
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<td>0.15</td>
<td>0.17</td>
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<tr>
<td>Wt.%</td>
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<td>0.16</td>
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<tr>
<td>OPX</td>
<td>0.09</td>
<td>0.07</td>
<td>0.08</td>
<td>0.11</td>
<td>-</td>
<td>0.08</td>
</tr>
<tr>
<td>ol</td>
<td>0.08</td>
<td>0.06</td>
<td>-</td>
<td>-</td>
<td>0.14</td>
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mineral compositions of the rock samples over very small thicknesses. Particularly obvious are variations in Fe$^{2+}$:Mg$^{2+}$ ratio, and in the Cr$_2$O$_3$ and TiO$_2$ contents of the minerals.

In Table No. 5 data showing the extent of the mineral composition variations within the solid solution series and in Cr$_2$O$_3$ and TiO$_2$ content for certain Kalskaret and Almklovdalen mineral assemblages are tabulated. Specimens N69, N70 and N71 from Lien, Almklovdalen occur within a few feet of each other in the field (data from O'Hara and Mercy 1963). Similarly N21, N23, N26, T96 and T153 are associated assemblages from the Kalskaret garnet peridotite lens, collected within 50 feet of each other (data for N21, N23 and N26 from O'Hara and Mercy 1963).

From Table No. 5 the surprisingly large extent of the variations in mineral compositions over very small thicknesses is apparent (see also Table No. 6). The question then arises as to whether such variations are likely to result solely from cryptic layering effects during igneous crystallization. It is therefore important to consider the extent of the cryptic layering variations found in large layered basic complexes.

In the Stillwater Complex, Montana (Hess 1960) the variations in mineral compositions noted over some 17,000 feet were orthopyroxene (enstatite 87 → 66), clinopyroxene (Ca$_{40}$Mg$_{52}$Fe$_8$ → Ca$_{39}$Mg$_{42}$Fe$_{19}$), plagioclase (An 86 → 63) and olivine (Fo 86 → 77 - but does not occur over much of the layered series).

In the Skaergaard Intrusion of East Greenland (Wager and Deer 1939) the mineral composition variations noted over some 3000 metres of layered basic and ultrabasic accumulates were olivine (Fo 70 → 2.5) and plagioclase (An 69 → 34).

In/
Outline of garnet peridotite stability field

a. after Ringwood, MacGregor and Boyd (1964)
b. after MacGregor (1965)

Full descriptions are given in the text.
In both the Stillwater and Skaergaard complexes, the most extensive mineral composition variations occur in the late-stage differentiates, while cryptic layering variations are minimal in the initial ultrabasic accumulates. It therefore appears likely that the variations within the solid solution series of the garnet peridotite minerals are greater than those expected to result through cryptic layering over such small thicknesses.

As a result it would appear unlikely that the Kalskaret and Almklovdalen garnet peridotites represent simply local segregations of garnet and pyroxenes within peridotites precipitated by normal igneous crystallization processes, whether in situ or at depth.

Another feature brought out by the mineral composition data which is worth noting is that although all the coexisting minerals of the garnet peridotites from both Kalskaret and Almklovdalen become increasingly more ferriferous in composition with increasing modal contents of garnet and clinopyroxene in the rock type, the minerals of the Kalskaret garnet peridotites as a group have a higher $\text{Fe}^{2+}:\text{Mg}^{2+}$ ratio than those of the Almklovdalen garnet peridotites.

Both these features are illustrated by the linear bulk rock composition trends on Fig. No. 20.

In view of the experimental studies of Boyd and England (1960), which indicated that orthopyroxenes from high-pressure alumina-bearing environments are likely to have high $\text{Al}_2\text{O}_3$ contents, the low $\text{Al}_2\text{O}_3$ contents (0.90-1.60 wt.%) of the orthopyroxenes from these garnet peridotites is surprising. The $\text{Al}_2\text{O}_3$ contents of the clinopyroxenes are also correspondingly low (1.66-3.08 wt.%).

The clinopyroxenes have generally low jadeite contents (4.4-8.6 %), being significantly lower than the jadeite contents of clinopyroxenes from eclogite/
FIG. 20

GARNET PERIDOTITES

○ KALSKARET
● LIEN
■ RODHAUGEN
X UGELVIK
X CZECH. FIALA 1965

$\text{Mg}^{2+} \rightarrow \text{Fe}^{2+} + \text{Fe}^{3+}$

$\text{Ca}^{2+}$
eclogite lenses in the gneisses (Lappin 1962).

The low CaO contents of the orthopyroxenes (0.26–0.53 wt.%) reflect the low degree of mutual solubility between coexisting pyroxenes in these rocks.

The significance of this feature, and also of the low Al₂O₃ contents of the pyroxenes is discussed in a subsequent section.

Included among the garnet analyses Table No. 3 is the analysis of garnet (T271) from the large eclogite lens within the Kaldhussaeter peridotite. The significantly higher CaO content and Fe²⁺/Mg²⁺ ratio, and the lower Cr₂O₃ content of this garnet compared with garnets from the Kalskaret and Almklovdalen garnet peridotites is apparent.

2. UGELVIK PERIDOTITES.

Unlike the garnet peridotites of the Kalskaret and Almklovdalen districts, the Ugelvik garnet peridotites are of much more widespread distribution interlayered with garnet-free peridotites, and are far more uniform mineralogically.

Here the general garnetiferous rock type is a garnet lherzolite with relatively low modal contents of garnet and pyroxenes. The garnet content varies from a few% to perhaps 25%, averaging about 10%. The clinopyroxene content averages about 5%, or perhaps even slightly less.

The mineral assemblage U₆ is exceptional in containing an abnormally high concentration of clinopyroxene (~47%). Its restricted occurrence was noted in the last chapter.

Optical determinations of the 2V of olivines from seven different Ugelvik peridotite samples ranging from garnet-free to garnet-rich assemblages, gave values of 85–87° +ve thought to be accurate with ± 1.5°. The composition of these olivines was therefore estimated as ranging between 6–9% Fa + about/
about 2% Fa - from the graph given by Deer, Howie and Zussman 1962 Vol. 1 Fig. 11.

Measurements of the 2V of orthopyroxenes in three of the rock samples gave values of 74-75° ± ve, estimated to be accurate within ± 1°.

Estimated from the graph given by Deer, Howie and Zussman 1963 Vol. 2, Fig. 10, these 2V values correspond to compositions in the range 6-7% Fs ± about 1% Fs.

All 2V determinations were made on the universal stage by direct measurement across the acute bisectrix.

The compositions of chemically analysed olivines and orthopyroxenes (Table No. 1) were olivine U 16 7.2% Fa, and orthopyroxene U 8 7.3% Fs. The optical determinations compare favourably with these values.

In Table No. 6 the range of mineral compositions from the Ugelvik peridotites is compared with the range from the Kalskaret and Almklovudalen peridotites, and that from analysed kimberlite garnet peridotite inclusions (O'Hara and Mercy 1963).

It is apparent that while the minerals of the Kalskaret and Almklovudalen garnet peridotites show notably increased Fe²⁺:Mg²⁺ ratios with increasing garnet and clinopyroxene contents of the mineral assemblage (a feature noted in the last section), the minerals of the Ugelvik peridotites have relatively constant Fe²⁺:Mg²⁺ ratios irrespective of the mineral assemblage (garnetiferous or garnet free) of the particular rock type, or the modal proportions of the constituent minerals.

All the principal primary minerals (Ol, OPX, CPX and GNT) of the Ugelvik peridotites are far more restricted in composition range, are more magnesian and also notably more Cr₂O₃ rich than the minerals of the Kalskaret and Almklovudalen garnet peridotites. On the other hand, the minerals of the garnet peridotite nodules in South African kimberlites are far more comparable in/
TABLE NO. 6.

ANALYSED MINERAL DATA + DATA FROM O'HARA AND MERCY (1963).

<table>
<thead>
<tr>
<th>KALSKARET PERIDOTITES:</th>
<th>TOTAL RANGE OF COMPOSITION</th>
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</thead>
<tbody>
<tr>
<td>OLIVINES</td>
<td>7.0 - 18.0% Fa</td>
</tr>
<tr>
<td>ORTHOPYROXENES</td>
<td>7.8 - 16.6% Fs</td>
</tr>
<tr>
<td>GARNETS</td>
<td></td>
</tr>
<tr>
<td>Ca 11.4</td>
<td>12.9</td>
</tr>
<tr>
<td>Mg 61.2</td>
<td>55.6</td>
</tr>
<tr>
<td>Ca 46.4</td>
<td>47.4</td>
</tr>
<tr>
<td>Mg 49.3</td>
<td>48.0</td>
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<table>
<thead>
<tr>
<th>AIMKLOVDALEN PERIDOTITES:</th>
<th>TOTAL RANGE OF COMPOSITION</th>
</tr>
</thead>
<tbody>
<tr>
<td>OLIVINES</td>
<td>7.0 - 11.2% Fa</td>
</tr>
<tr>
<td>ORTHOPYROXENES</td>
<td>7.5 - 11.5% Fs</td>
</tr>
<tr>
<td>GARNETS</td>
<td></td>
</tr>
<tr>
<td>Ca 12.0</td>
<td>12.2</td>
</tr>
<tr>
<td>Mg 65.7</td>
<td>63.6</td>
</tr>
<tr>
<td>Ca 45.9</td>
<td>49.7</td>
</tr>
<tr>
<td>Mg 51.4</td>
<td>47.7</td>
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<table>
<thead>
<tr>
<th>UGELVIK PERIDOTITES:</th>
<th>CHEMICAL</th>
</tr>
</thead>
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<tr>
<td>OLIVINES</td>
<td>7.2% Fa</td>
</tr>
<tr>
<td>ORTHOPYROXENES</td>
<td>7.3% Fs</td>
</tr>
<tr>
<td>*GARNETS</td>
<td>6-9% Fa ± 2% Fa</td>
</tr>
<tr>
<td>Ca 10.8</td>
<td>12.4</td>
</tr>
<tr>
<td>Mg 74.6</td>
<td>72.6</td>
</tr>
<tr>
<td>*CLINOPYROXENES</td>
<td>6-8% Fs ± 1% Fs</td>
</tr>
<tr>
<td>Ca 46.4</td>
<td>46.8</td>
</tr>
<tr>
<td>Mg 51.9</td>
<td>51.5</td>
</tr>
</tbody>
</table>

* Excluding analysed assemblage U6

<table>
<thead>
<tr>
<th>GARNET PERIDOTITE NODULES IN KIMBERLITES:</th>
<th>CHEMICAL</th>
</tr>
</thead>
<tbody>
<tr>
<td>OLIVINES</td>
<td>6.4 - 8.2% Fa</td>
</tr>
<tr>
<td>ORTHOPYROXENES</td>
<td>5.7 - 8.6% Fs</td>
</tr>
<tr>
<td>GARNETS</td>
<td></td>
</tr>
<tr>
<td>Ca 10.2</td>
<td>12.0</td>
</tr>
<tr>
<td>Mg 78.2</td>
<td>72.7</td>
</tr>
<tr>
<td>Ca 44.8</td>
<td>46.4</td>
</tr>
<tr>
<td>Mg 52.2</td>
<td>51.3</td>
</tr>
</tbody>
</table>

* Excluding analysed assemblage U6
in composition with the minerals of the Ugelvik garnet peridotites.

The compositions of the minerals of rock specimen U₆ are rather exceptional in many respects when compared with the compositions of the minerals from the other analysed Ugelvik peridotite assemblages. The exceptionally high modal clinopyroxene content and the restricted occurrence of this rock type have already been noted.

The garnet of U₆ is particularly more calcic and ferriferous than all the other analysed Ugelvik garnets. Its high CaO content (6.02 wt.%) is unusual, being appreciably higher than the CaO content of all other garnets from garnet peridotites analysed in the course of this study or by O'Hara and Mercy (1963).

The clinopyroxene of U₆ has appreciably lower Al₂O₃, Cr₂O₃ and Na₂O contents than all the other analysed Ugelvik clinopyroxenes, and appreciably lower Na₂O and Al₂O₃ contents than all other clinopyroxenes from garnet peridotites analysed in the course of this study or by O'Hara and Mercy (1963).

The low Al₂O₃ content (0.50 wt.% and Na₂O content (0.56 wt.%) of this clinopyroxene indicate the low degree of jadeite substitution (1.0%) in this clinopyroxene. However, the degree of substitution of Ca-Tschermak's molecule calculated after the formation of NaR₃Si₂O₆ is not significantly lower than in certain other Norwegian garnet peridotite clinopyroxenes.

An obvious explanation of the exceptional mineral chemistry of the assemblage U₆ is lacking. This rock sample is the most 'eclogitic' of all the analysed Ugelvik garnet peridotites in the sense that it has by far the lowest modal olivine content (~30% by volume) yet it contains the clinopyroxene with/
with the lowest jadeite content. This is the converse of the situation in the Kalskaret and Almklovdalen garnet peridotites where the most 'eclogitic' assemblages, T153 and N75 respectively, contain the clinopyroxenes with the highest jadeite contents.

This rock type, $U_6$, has clearly not resulted solely from the mechanical concentration of the minerals present throughout the Ugelvik peridotite mass in which it occurs, because of the mineral composition anomalies involved.

It is considered here that the most feasible explanation of the development of this rock type is that either hydrothermal introduction or some mechanism of metasomatic concentration (wet or dry diffusion) has somehow produced the high CaO content of this thin band resulting in the high modal content of clinopyroxene, the abnormally high CaO content of the garnet, and the apparent general 'disequilibrium' distribution of elements ($Al_2O_3$, $CaO$, $Cr_2O_3$ etc) among the mineral phases of this assemblage compared with the relatively constant distribution ratios of these elements noted among the minerals of the other garnet peridotite assemblages. (c.f. Table Nos. 7-9).

However, with the above reservations about the mineral compositions of the assemblage $U_6$, it is apparent that in general the minerals of the Ugelvik peridotites are relatively uniform in composition, particularly with respect to $Fe^{2+}/Mg^{2+}$ ratio.

Minor inhomogeneities occur, however, in certain minor elements, $Cr_2O_3$ and $TiO_2$ in particular. The differing $Cr_2O_3$ contents of the two garnet fractions of $U_{16}$ separated suggest minor variations in the $Cr_2O_3$ contents of the garnets even in large hand specimens. Initially only a small amount of garnet was separated from this rock sample and analysed $U_{16}$. However, this/
However, this garnet was so significantly rich in chromium that as plenty of this rock sample was available a second garnet fraction $U_{16}2$ was separated and analysed. The $Cr_2O_3$ content of the garnets appears to be closely controlled by the $Al_2O_3$ content, which in turn controls the modal proportions of garnet in the assemblage. Hence the $Cr_2O_3$ content of the garnets decreases with increasing modal garnet content.

For the most part, however, it would appear that the mineral layering observed in the Ugelvik peridotite masses could be adequately explained by the sorting (either igneous or later mechanical sorting) of an eclogite facies assemblage ($Ol + GNT + CPX + OPX +$ chromite) precipitated by igneous crystallization processes. As pointed out previously a similar explanation cannot explain the Kalskaret and Almklovdalen garnetiferous peridotite layering.

The limited extent of the mutual solubility between the coexisting pyroxenes of the assemblage $U_8$, and the low $Al_2O_3$ contents of both OPX and CPX from the Ugelvik garnet peridotites, are comparable with the situation in the Kalskaret and the Almklovdalen garnet peridotites and also in the kimberlite garnet peridotite nodules (O'Hara and Mercy 1963).

3. EQUILIBRATION OF THE GARNET PERIDOTITE ASSEMBLAGES.

While the mineral compositions vary quite strikingly between adjacent layers of the garnet peridotite masses at Kalskaret and Almklovdalen (see Table 5), these variations are related to changes in the bulk rock chemistry of the adjacent layers while the distribution ratios of the elements between the various mineral phases appear to remain relatively constant. Element partitions/
partitions among coexisting minerals in the Ugelvik garnet peridotites are also quite constant.

The data tabulated in Table No. 7 indicates that while the bulk rock Fe\(^{2+}\)Fe\(^{3+}\)Mn\(^{2+}\)/Mg\(^{2+}\) ratio varies considerably (approximately three fold) among the Norwegian garnet peridotites, the distribution ratio of Fe\(^{2+}\)Fe\(^{3+}\)Mn\(^{2+}\)/Mg\(^{2+}\) among the coexisting minerals from almost all the analysed mineral assemblages is comparatively constant even for assemblages from the three different localities - Kalskaret, Almklovdalen and Ugelvik.

However, the Ugelvik assemblage U\(_6\) has an exceptionally iron rich garnet with respect to its coexisting clinopyroxene, and the Almklovdalen assemblage N71 appears to have a rather abnormally iron rich garnet and orthopyroxene with respect to its coexisting clinopyroxene. As, however the Fe\(^{2+}\)Fe\(^{3+}\)Mn\(^{2+}\) contents of the clinopyroxenes are low, hence increasing the effect of any possible analytical error on the distribution ratio of Fe\(^{2+}\)Fe\(^{3+}\)Mn\(^{2+}\)/Mg\(^{2+}\) among the mineral phases when standardized against clinopyroxene, the distribution ratio was recalculated standardized against orthopyroxene (Table No. 8) to see if the anomalies still persist. N71 does, in fact, still appear exceptional. The orthopyroxene of U\(_6\) unfortunately has not been analysed.

In all the analysed Norwegian garnet peridotite assemblages the order of decreasing Fe\(^{2+}\)Fe\(^{3+}\)Mn\(^{2+}\)/Mg\(^{2+}\) ratio is GRT > OL > OPX > CPX.

Experimental work in the system MgO-FeO-SiO\(_2\) (Bowen and Schairer 1935) showed that at equilibrium the ratio Fe\(^{2+}\)/Mg\(^{2+}\) in olivine is greater than in the coexisting Ca-poor pyroxene, the equilibrium Fe/Of ratio being about 1.6. It must be borne in mind, however, that the experimental work involved coexisting clinohyperthene and olivines, not orthopyroxenes (although/
**TABLE NO. 7**

<table>
<thead>
<tr>
<th>SAMPLE NO.</th>
<th>OL.</th>
<th>OPX</th>
<th>CPX</th>
<th>GNT</th>
<th>SP</th>
<th>AMP</th>
<th>BULK ROCK COMPOSITION</th>
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</thead>
<tbody>
<tr>
<td>Kalskaret (N23)</td>
<td>very scarce</td>
<td>1.603</td>
<td>1.440</td>
<td>-</td>
<td>1.784</td>
<td>0.265</td>
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<td>Garnet (N26)</td>
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<td>very scarce</td>
<td>1.394</td>
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<td>-</td>
<td>0.204</td>
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<tr>
<td>Peridotites (T96)</td>
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<td>1.426</td>
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<td>-</td>
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</tr>
<tr>
<td>Peridotites (T153)</td>
<td>-</td>
<td>-</td>
<td>1.672</td>
<td>-</td>
<td>-</td>
<td>0.260</td>
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<td>Almklovdalen (N69)</td>
<td>1.443</td>
<td>1.397</td>
<td>1.435</td>
<td>-</td>
<td>1.513</td>
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<td>1.582</td>
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<tr>
<td>Peridotites (N71)</td>
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<td>1.872</td>
<td>1.621</td>
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<tr>
<td>Garnet (U6)</td>
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<td>Bellinzona</td>
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<td></td>
</tr>
</tbody>
</table>

**PARTITION OF $^{2+}$Fe $^{3+}$Mn $^{2+}$ Mg $^{2+}$ AMONG THE MINERAL PHASES IN RATIO AGAINST CPX**
<table>
<thead>
<tr>
<th>SAMPLE NO.</th>
<th>OL.</th>
<th>OPX.</th>
<th>CPX.</th>
<th>GNT.</th>
<th>SP.</th>
<th>AMP.</th>
<th>BULK ROCK COMPOSITION</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kalskaret</td>
<td>N23</td>
<td>very scarce</td>
<td>1</td>
<td>0.624</td>
<td>2.770</td>
<td>-</td>
<td>1.113</td>
</tr>
<tr>
<td>Garnet</td>
<td>N26</td>
<td>very scarce</td>
<td>1</td>
<td>0.691</td>
<td>2.849</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Peridotites</td>
<td>T96</td>
<td>1.152</td>
<td>1</td>
<td>0.716</td>
<td>3.121</td>
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</tr>
<tr>
<td>Peridotites</td>
<td>T153</td>
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<td>1</td>
<td>0.740</td>
<td>3.390</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Almklovdalen</td>
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<td>1</td>
<td>0.716</td>
<td>3.121</td>
<td>-</td>
<td>1.083</td>
</tr>
<tr>
<td>Garnet</td>
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<td>1</td>
<td>0.716</td>
<td>3.390</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Peridotites</td>
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<td>1</td>
<td>0.534</td>
<td>3.435</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Ugelvik</td>
<td>U3</td>
<td>not analysed</td>
<td>1</td>
<td>0.690</td>
<td>3.166</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Garnet</td>
<td>U6</td>
<td>not analysed</td>
<td>1</td>
<td>0.740</td>
<td>3.390</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Peridotites</td>
<td>U8</td>
<td>1.033</td>
<td>1</td>
<td>0.534</td>
<td>3.435</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Bellinzona</td>
<td>A2</td>
<td>1.033</td>
<td>1</td>
<td>0.801</td>
<td>2.747</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>GNT. Peridotite</td>
<td>1</td>
<td>0.969</td>
<td>1</td>
<td>1.320</td>
<td>3.564</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Peridotite</td>
<td>2</td>
<td>0.878</td>
<td>1</td>
<td>1.172</td>
<td>3.809</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Nodules in</td>
<td>3</td>
<td>0.982</td>
<td>1</td>
<td>1.168</td>
<td>3.809</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Basalts</td>
<td>4</td>
<td>0.986</td>
<td>1</td>
<td>1.119</td>
<td>3.499</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Ross et al. (1954)</td>
<td>5</td>
<td>1.050</td>
<td>1</td>
<td>1.239</td>
<td>3.499</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>6</td>
<td>1.014</td>
<td>1</td>
<td>0.984</td>
<td>2.899</td>
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<td>-</td>
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<tr>
<td></td>
<td>7</td>
<td>1.054</td>
<td>1</td>
<td>1.101</td>
<td>2.899</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>8</td>
<td>1.012</td>
<td>1</td>
<td>1.598</td>
<td>2.899</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>9</td>
<td>1.012</td>
<td>1</td>
<td>1.598</td>
<td>14.121</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>
(although the Ca-poor pyroxene phase may have been a protohypersthene on crystallization).

Bowen and Schairer (1935) and Ramberg and De Vore (1951) have argued that the inversion from monoclinic to orthorhombic Ca-poor pyroxene will not, however, significantly alter the distribution of Fe\(^{2+}\) and Mg\(^{2+}\).

Ramberg and De Vore's (1951) contention that the equilibrium distribution is changed by different temperatures and pressures of crystallization corresponding to differing environments of formation, was discounted by O'Hara (1963d).

O'Hara (1963d) has demonstrated that the peridotites and garnet peridotites which occur as lenses in gneiss or schist in Norway or Switzerland show a distribution of iron between olivine and coexisting Ca-poor pyroxene that resembles that found in garnet peridotites and peridotites occurring as nodules in kimberlite pipes, and approaches the experimental equilibrium relationship. It was also suggested that all igneous assemblages depart at least slightly from equilibrium (as a result principally of compositional zoning, corona formation and crystal fractionation processes), the departure being greatest in those rocks that have crystallized most rapidly.

Data on the composition of coexisting Ca-poor and Ca-rich pyroxenes in basic and intermediate plutonic rocks have gradually been accumulating since Hess (1941) drew attention to the relationship of such pairs in his study on the pyroxenes of common mafic magmas. In this connection he noted that a tie-line in the system CaSiO\(_3\)-MgSiO\(_3\)-FeSiO\(_3\) joining the compositions of coexisting augites and hypersthenes if extended,
extended, intersected the \( \text{MgSiO}_3 - \text{CaSiO}_3 \) side line in a position approximately \( \text{En}_{25} \text{Wo}_{75}\). His data were largely drawn from the pyroxene pairs of the Skaergaard and Stillwater layered intrusions.

In terms of the Mg-Fe partition of the pyroxene pair, this tie-line relationship denoted a higher relative concentration of iron in the orthopyroxene member of the pair. It was noted, however, that pyroxene pairs in phenocryst bearing volcanic rocks do not normally follow this relationship, as in these assemblages there is a sequence of crystallization of the two pyroxenes with the phenocryst phase relatively enriched in Mg, a relationship probably corresponding to non-equilibrium conditions.

Since Hess published his account further data have become available, notably from the Skaergaard intrusion (Brown 1957), confirming this relationship and indicating that it also applies where the Ca-poor pyroxene is a pigeonite.

O'Hara (1960b) considered that slight differences in the distribution of \( \text{Fe}^{2+} \) and \( \text{Mg}^{2+} \) between coexisting pyroxenes can probably result from variations in the conditions of formation, but concluded that such differences are of little significance in defining definite thermal histories as proposed by Wilson (1960).

Bartholomé (1962) considered the distribution of \( \text{Fe}^{2+} \) and \( \text{Mg}^{2+} \) between coexisting pyroxenes from the thermodynamic viewpoint and suggested that there is a genuine correlation with temperature, and hence that accurately determined \( \text{Fe}^{2+}:\text{Mg}^{2+} \) ratios can yield useful temperature data. However, he warned that anomalies are likely to arise in some assemblages due to lack of equilibrium, analytical errors and possibly errors due to other causes. He considered that minor constituents \( \text{Al}_2\text{O}_3, \text{TiO}_2, \text{Fe}_2\text{O}_3, \text{MnO}, \text{Na}_2\text{O} \) etc. do not exert appreciable influence on the equilibrium relations.
relations, a conclusion directly opposed to that of De Vore (1957).

Kretz (1961) has also attempted to correlate differences in the distribution of $\text{Mg}^{2+}$ and $\text{Fe}^{2+}$ between coexisting pyroxenes with differences in the temperature of crystallization. However, O'Hara and Mercy (1963) have pointed out several factors (disequilibrium, effect of pressure, effect of other constituents - $\text{Al}_2\text{O}_3$, CaO etc.) which may affect the validity of such a correlation.

However, these uncertainties as to whether or not the distribution of $\text{Fe}^{2+}$ and $\text{Mg}^{2+}$ among coexisting pyroxenes can yield useful temperature data do not invalidate the general conclusion that the equilibrium distribution of $\text{Fe}^{2+}$ and $\text{Mg}^{2+}$ between coexisting olivines and pyroxenes is expected to be that olivine has a greater relative concentration of iron than its coexisting orthopyroxene, which in turn has a higher relative concentration of iron than its coexisting clinopyroxene.

This is in fact the $\text{Fe}^{2+}\text{:Mg}^{2+}$ distribution found in the Norwegian garnet peridotite assemblages.

Considering again Tables No. 7 and 8, the Bellinzona garnet peridotite assemblage A2 (O'Hara and Mercy 1963) shows the same order of distribution of $\text{Fe}^{2+}\text{Fe}^{3+}\text{Mn}^{2+}/\text{Mg}^{2+}$ ratios as the Norwegian garnet peridotite assemblages. Also given for comparison are the distribution ratios for analysed assemblages ($\text{Ol} + \text{OPX} + \text{CPX} + \text{Spinel}$) from peridotite nodules in basalts (Ross et al. 1954). In these assemblages the general distribution order is $\text{Sp} > \text{CPX} > \text{OPX} > \text{Ol}$, probably representing a disequilibrium distribution resulting from an igneous origin of these assemblages. Furthermore the actual ratio values are considerably more variable than in the Norwegian garnet peridotite assemblages.

Table/
Table No. 9 shows that Cr$_2$O$_3$ is distributed among the mineral phases of most Norwegian garnet peridotites in a remarkably constant ratio considering the variations in the total Cr$_2$O$_3$ content of the rocks and minerals. Again, however, U$_6$ is exceptional in having a GNT/CPX Cr$_2$O$_3$ ratio of 4.60 considerably greater than that of the other mineral assemblages. Also N71 is unusual in having a rather high OPX/CPX Cr$_2$O$_3$ ratio of 0.22. However, with these two exceptions the distribution ratio of Cr$_2$O$_3$ among the mineral phases of these garnet peridotites (including the Bellinzona garnet peridotite A2) is remarkably constant, particularly when compared with the Cr$_2$O$_3$ partition among the mineral phases of peridotite nodules in basalts (Ross et al. 1954).

If one considers the partition of other elements (NiO, TiO$_2$, CaO, Al$_2$O$_3$ etc.) among the mineral phases of the Norwegian garnet peridotite assemblages similarly consistent ratios are found with further anomalies in the mineral assemblages of N71 and U$_6$, particularly in the latter.

With these two possible exceptions, it is considered that the important evidence that all the Norwegian garnet peridotite assemblages have the expected equilibrium order of distribution of Fe$^{2+}$/Fe$^{3+}$/Mn$^{2+}$/Mg$^{2+}$ combined with possible supporting evidence from the consistent partition ratios of other elements among the mineral phases indicates the likely general attainment of, or close approach to, equilibrium during the production of the present mineral assemblages of these rocks, whatever their origin might have been. Furthermore, as failure to attain equilibrium appears to be a common feature of igneous crystallization (viz. crystal fractionation, mineral/
<table>
<thead>
<tr>
<th>SAMPLE NO.</th>
<th>OL</th>
<th>Opx</th>
<th>Cpx</th>
<th>Gnt</th>
<th>Sp.</th>
<th>Bulk Rock Composition</th>
</tr>
</thead>
<tbody>
<tr>
<td>N23</td>
<td>0.14</td>
<td>0.20</td>
<td>0.39</td>
<td>-</td>
<td>-</td>
<td>0.21</td>
</tr>
<tr>
<td>Kalskaret</td>
<td>0.03</td>
<td>0.63</td>
<td>0.91</td>
<td>-</td>
<td>-</td>
<td>0.34</td>
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<tr>
<td>Garnet</td>
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<td>1.79</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Peridotites</td>
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<td>0.47</td>
<td>0.84</td>
<td>-</td>
<td>-</td>
<td>0.29</td>
</tr>
<tr>
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<td>0.08</td>
<td>0.16</td>
<td>0.25</td>
<td>-</td>
<td>-</td>
<td>0.12</td>
</tr>
<tr>
<td>N69</td>
<td>0.12</td>
<td>0.69</td>
<td>0.80</td>
<td>-</td>
<td>-</td>
<td>0.43</td>
</tr>
<tr>
<td>Almklovdaen</td>
<td>0.06</td>
<td>0.47</td>
<td>0.95</td>
<td>-</td>
<td>-</td>
<td>0.49</td>
</tr>
<tr>
<td>Garnet</td>
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<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Peridotites</td>
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<td>0.20</td>
<td>0.32</td>
<td>-</td>
<td>-</td>
<td>0.32</td>
</tr>
<tr>
<td>U2</td>
<td>0.13</td>
<td>1.81</td>
<td>2.77</td>
<td>4.60</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>U6</td>
<td>0.15</td>
<td>1.30</td>
<td>2.32</td>
<td>1.00</td>
<td>1.00</td>
<td>-</td>
</tr>
<tr>
<td>Bellinzona</td>
<td>0.04</td>
<td>0.36</td>
<td>1</td>
<td>20.25</td>
<td>20.25</td>
<td>-</td>
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<td>1</td>
<td>20.25</td>
<td>20.25</td>
<td>-</td>
</tr>
<tr>
<td>A3</td>
<td>2</td>
<td>0.65</td>
<td>1</td>
<td>20.26</td>
<td>20.26</td>
<td>-</td>
</tr>
<tr>
<td>N 4</td>
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<td>1</td>
<td>11.86</td>
<td>11.86</td>
<td>-</td>
</tr>
<tr>
<td>Basalts</td>
<td>7</td>
<td>0.53</td>
<td>1</td>
<td>11.87</td>
<td>11.87</td>
<td>-</td>
</tr>
<tr>
<td>Ross et al.</td>
<td>9</td>
<td>0.75</td>
<td>1</td>
<td>32.59</td>
<td>32.59</td>
<td>-</td>
</tr>
</tbody>
</table>

**Table No. 9.**

**Cr$_2$O$_3$ Contents (wt. %) and Distribution of Cr$_2$O$_3$ Among the Mineral Phases in Ratio Against Cpx Cr$_2$O$_3$ Content.**
mineral zoning, corona formation etc.) this suggests that a metamorphic origin is more likely.

The slight anomalies noted for the mineral assemblages of N71 and U6 may indicate some degree of failure to attain the ideal equilibrium distribution of elements among the mineral phases of these two assemblages, the reason for which is not immediately apparent.

The unusual mineral content of U6 has already been noted and contrasts with all the other analysed Ugælvik garnet peridotite assemblages. The mineral assemblage of N71 is quite rich in garnet and clinopyroxene and lacks olivine, but is not significantly different from certain other analysed Almklovdalen and Kalskaret garnet peridotite assemblages. However, it is just conceivable that in both cases these rock types represented somewhat extreme original compositional rock types and have hence failed to attain complete equilibrium during subsequent metamorphic recrystallization.

It is to be stressed, however, that equilibrium has been generally attained or closely approached in the formation of the present mineral assemblages of the Norwegian garnet peridotites.

4. STABLE CONDITIONS OF FORMATION OF THESE MINERAL ASSEMBLAGES.

At this juncture it is relevant to consider the likely stable physical conditions of formation of garnet peridotite mineral assemblages. It is therefore intended in this section to give a short review of recent experimental petrology studies which have thrown some light on this problem, and are hence critical in any discussion on the petrogenesis of garnet peridotites.
The pressure-temperature stability fields of the different mineral assemblages in the system MgO-Al₂O₃-SiO₂ have been discussed by Ringwood, MacGregor and Boyd (1964) and are illustrated in Fig. 19a. In this diagram the approximate stability fields of spinel peridotite (olivine + 2 pyroxenes + spinel), aluminous pyroxene peridotite (olivine + aluminous pyroxenes) and garnet peridotite (olivine + alumina poor pyroxenes + garnet) are outlined and their relationship shown to possible oceanic and Pre-Cambrian shield geotherms.

Ringwood (1962a,b), Green and Ringwood (1963), and Clark and Ringwood (1964) have discussed likely models of the upper mantle mineralogy in terms of a hypothetical rock type called 'pyrolite'. This rock type in fact represents a peridotite with low but essential amounts of Al₂O₃, CaO and Na₂O. It was chosen as a likely primitive upper mantle material on the assumption that such material must be capable of yielding a basaltic composition partial melt, leaving behind a peridotitic or dunitic residual.

It was therefore considered by Ringwood, MacGregor and Boyd (1964) that material of pyrolite composition may occur in any one of the three mineral assemblages outlined in Fig. 19a (and also in a plagioclase bearing assemblage - Olivine + plagioclase + pyroxenes + accessory chromium spinel), depending on the physical conditions of formation of the assemblage.

The line AC in Fig. 19a is the boundary for the reaction:

\[ 4\text{MgSiO}_3 + \text{MgAl}_2\text{O}_4 \rightarrow \text{Mg}_2\text{Al}_2\text{Si}_3\text{O}_{12} + \text{Mg}_2\text{SiO}_4 \]  

enstatite spinel  pyrope  forsterite

In more complex natural systems, solid solution effects would be expected.
expected to broaden this linear boundary into a transition zone.

Ringwood, MacGregor and Boyd (1964) stress that the concentration and distribution of the trivalent elements, particularly \(\text{Al}^{3+}\), in aluminous pyroxene pyrolite strongly influences its stability field with respect to garnet pyrolite.

The line CD in Fig. 19a is the boundary for the reaction:

\[
3\text{MgSiO}_3 + x\text{Al}_2\text{O}_3 \rightleftharpoons x\text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12} + 3(1-x)\text{MgSiO}_3
\]

aluminous enstatite pyrope enstatite

where \(x \leq 0.045\) (Boyd and England 1964)

It was assumed in the construction of Fig. 19a that the solubility of pyrope garnet in clinopyroxene is similar to that of pyrope in orthopyroxene under similar temperature and pressure conditions. It was considered that the error likely to be introduced by this assumption was quite small.

Preliminary runs on the reaction Al-rich diopside + garnet (MacGregor and Ringwood 1964) with natural mineral pairs from a kimberlite garnet peridotite nodule indicated that at \(1500^\circ\text{C}\) and 30-35 kb. the diopside on the equilibrium curve has approximately 1 mole% more total \(\text{R}_2\text{O}_3\) (less \(\text{R}_2\text{O}_3\) associated with \(\text{Na}_2\text{O}\)) than the enstatite under the same conditions. This experimental evidence is not substantiated by a study of coexisting pyroxenes from the Norwegian garnet peridotites where in most cases the OPX has slightly higher total mole \% \(\text{R}_2\text{O}_3\) (less \(\text{R}_2\text{O}_3\) associated with \(\text{Na}_2\text{O}\)) than the coexisting CPX (see Table No.13). In one assemblage the opposite relationship was found, however. Furthermore the pyroxenes from the Norwegian garnet peridotites were probably recrystallized at appreciably lower temperatures (\(\sim 600-700^\circ\text{C}\) - see Section 5), than those of the experimental conditions studied by MacGregor and Ringwood (1964). However, in general it appears that the degree/
degree of solubility of fictive garnet molecule is likely to be reasonably similar in both coexisting clinopyroxenes and orthopyroxenes formed under the same physical conditions.

The slopes of the $R_2O_3$ contours in Fig. 19a were taken from the results of studies of the synthetic system enstatite-pyrope (Boyd and England 1964).

It is to be expected that the formation of garnet from aluminous pyroxenes would extend continuously over a wide pressure interval.

The maximum average $R_2O_3$ content of the pyroxenes was taken as 4.5 mole% by Ringwood, MacGregor and Boyd (1964), corresponding to the situation where all the $R_2O_3$ in the rock occurs in solid solution in pyroxene and spinel is not present as a stable phase.

The maximum $R_2O_3$ contents of pyroxenes in naturally occurring pyroxene peridotites quoted in the literature are those from the Lizard primary peridotite assemblages (90683 and 90681 - Green 1964), although some peridotite nodules in basalts (Ross et al. 1954) contain pyroxenes with $R_2O_3$ contents of much the same order of magnitude.

Relevant data for these Lizard assemblages are tabulated below,

Table No.10.

<table>
<thead>
<tr>
<th>Lizard Sample No.</th>
<th>Wt.% $Al_2O_3$</th>
<th>Mole% $R_2O_3$ in pyroxenes</th>
</tr>
</thead>
<tbody>
<tr>
<td>90683</td>
<td>OPX 6.59</td>
<td>3.96</td>
</tr>
<tr>
<td></td>
<td>CPX 7.09</td>
<td>4.65</td>
</tr>
<tr>
<td></td>
<td>Spinel 47.58</td>
<td></td>
</tr>
<tr>
<td>90681</td>
<td>OPX 6.30</td>
<td>3.81</td>
</tr>
<tr>
<td></td>
<td>CPX 6.59</td>
<td>4.68</td>
</tr>
<tr>
<td></td>
<td>Spinel 46.88</td>
<td></td>
</tr>
</tbody>
</table>
The $R_2O_3$ contents (comprising mostly $Al_2O_3$) of these pyroxenes agree roughly with the value of 4.5 mole % chosen by Ringwood, MacGregor and Boyd (1964). However, the experimental work of Boyd and England (1960, 1964) and others (see Section 5B) indicates that higher $R_2O_3$ contents in pyroxenes are theoretically possible under the physical conditions under consideration here (Fig. 19a).

The $R_2O_3$ contents of the pyroxenes will, however, be restricted by the total $R_2O_3$ content of the rock. The bulk rock $R_2O_3$ content of the Lizard specimen 90683 is 4.55 wt. %, being of the same order as envisaged $R_2O_3$ contents of likely mantle compositions based on the pyrolite model (4.43 wt. %) and the chondrite model (3.90 wt. %) – see Table No. 25.

However, the Lizard pyroxenes coexist with a highly aluminous spinel phase which may therefore restrict their $R_2O_3$ contents. Also an assemblage with only one pyroxene phase and without aluminous spinel might well be expected to contain a pyroxene with an $R_2O_3$ content > 4.5 mole %.

In conclusion it appears that peridotites of likely mantle compositions may be expected to crystallize, at least in theory, to assemblages containing a pyroxene (or perhaps even two pyroxenes) of $R_2O_3$ content > 4.5 mole %.

The boundary limits between the spinel pyrolite and the aluminous pyroxene pyrolite field are as yet ill-defined. However, Ringwood, MacGregor and Boyd (1964) consider that spinel is unstable in rocks of pyrolite composition above about 1000°C, owing to the stability of $Al^{3+}$, $Cr^{3+}$, and $Fe^{3+}$ in the pyroxenes.

Ringwood, MacGregor and Boyd (1964) say that the position of the pyrolite solidus in Fig. 19a is little more than a plausible guess based on/
on miscellaneous results including those of O'Hara (1963a).

According to Fig. 19a, olivine and aluminous pyroxenes are expected to be the phases stable at the solidus down to a depth of about 150 km., and hence that equilibria involving these minerals and not garnet assemblages would control the formation of magmas in the mantle by direct fractional melting of rocks of pyrolite composition.

However, MacGregor (1965) has demonstrated that when CaO is introduced into the experimental system MgO-Al$_2$O$_3$-SiO$_2$, it depresses the stability field of garnet peridotite to appreciably lower pressures for likely mantle peridotite compositions (Fig. 19b). Hence garnet peridotite may be the stable assemblage at the solidus at depths perhaps slightly less than 100 kms.

In Fig. 19b, curve 1 (equivalent to AC in Fig. 19a) is the reaction curve $4\text{MgSiO}_3 + \text{MgAl}_2\text{O}_3 \rightleftharpoons 3\text{MgAl}_2\text{Si}_3\text{O}_{12} + 2\text{Mg}_2\text{Si}_4\text{O}_{10}$ (MacGregor 1964). Curves 2, 3 and 4 are coincident. Curves 2 and 3 are equilibrium curves for the reaction: CPX ± OPX + Spinel + Forsterite $\rightleftharpoons$ CPX + Forsterite + CNT + Spinel - for compositions which lie on the joins Garnet - Forsterite and Pyroxene - Spinel with enstatite/diopside ratios of 25:75 and 75:25 respectively. No experiments have been carried out at sufficiently high pressures to give products that lie in the 2 phase field Forsterite - Garnet.

Curve 4 is the boundary curve for the reaction:

$\text{CaMgSi}_2\text{O}_6 + 2\text{MgSiO}_3 + \text{MgAl}_2\text{O}_4 \rightleftharpoons 2\text{Mg}_2\text{CaAl}_2\text{Si}_3\text{O}_{12} + 2\text{Mg}_2\text{Si}_4\text{O}_{10}$ (Kushiro and Yoder 1965).

MacGregor (1965) considered that the true equilibrium boundaries between natural spinel and garnet peridotites lie between curves (1) and (2,3,4) in Fig. 19b.

Thus the equilibrium curves outlined in Figs. 19a and b. broadly delineate/
delineate the likely pressure and temperature stability fields of expected naturally occurring mantle peridotite compositions.

In the light of this experimental data it is to be expected that in continental areas at depths in excess of probably about 10 kms, most peridotite compositions will (under dry conditions) lie within the garnet peridotite stability field. This conclusion has an important bearing on the origin of the large spinel peridotite masses found along the axes of old mountain belts. With normal geothermal gradients, it is unlikely that such peridotites have formed by inversion from a high pressure garnet bearing assemblage but they must have formed directly at high temperatures, and pressures on the low pressure side of curve \((2,3,4)\) in Fig. 19b.

This conclusion suggests that these spinel peridotites have probably formed by fractional crystallization of basaltic magmas in or near the base of the crust, and that they, therefore, do not represent a primary upper mantle composition. A similar origin of these spinel peridotites has been suggested by O'Hara and Mercy (1963).

It is also expected from the experimental data that the pressures and temperatures attained in the high grade regionally metamorphosed cores of mountain belts will probably lie within the garnet peridotite stability field and hence that their possible occurrence under these conditions is consistent with the experimental data. However, the absence of garnet peridotites among the metamorphic rocks of every orogenic belt outside of Europe and their very sparse occurrence in European orogenic belts, indicates that if these garnet peridotites are, in fact, of metamorphic origin then the physical conditions for their formation have been only rarely attained.

In oceanic areas, it is expected that there will be a delicate balance/
balance between the intersection of the geothermal gradient and the spinel peridotite→ garnet peridotite equilibrium curve, depending on bulk compositions and local physical conditions. Also in tectonically active continental areas with exceptionally high geothermal gradients it is just conceivable that spinel peridotites may be stable over a significant depth range, probably sandwiched within the garnet peridotite stability field.

It is therefore apparent from the experimental data that garnet peridotites, such as those which occur in the Norwegian gneiss complex, may represent tectonically emplaced fragments of the upper mantle or they may have been formed by metamorphism in the lower part of the crust of peridotite which originally crystallized to another mineral facies assemblage. The problem then is to decide between these two alternatives.

5. ESTIMATION OF TEMPERATURE OF FORMATION OF MINERAL ASSEMBLAGES.

In this section a discussion is given of mineralogical features which throw some light on the likely temperatures of crystallization of the primary mineral assemblages of the Norwegian garnet peridotites.

A. SOLID SOLUTION LIMITS IN COEXISTING PYROXENES.

Davis (1963) has shown that the solubility of enstatite in diopside in the synthetic system is essentially the same at 30 kilobars pressure as at atmospheric pressure (Boyd and Schairer 1964). Therefore in principal it should be possible to use this solvus curve as a geothermometer that is virtually independent of pressure.

Given below (Table No. II) are estimates of the limits of composition (as/
(as atomic proportions) of the two pyroxene field in the system MgSiO$_3$-CaMgSi$_2$O$_6$ for various temperatures. These estimates are from a diagram based on experimental studies at atmospheric pressure in this system (Boyd and Schairer 1964). It should be noted that the results of Boyd and Schairer (1964) show more restricted solid solution between the coexisting pyroxenes in this system than the results of Atlas (1952) for all temperatures except for the leg of the solvus defining the limit of solubility of diopside in rhombic enstatite.

**TABLE NO. II**

<table>
<thead>
<tr>
<th>℃</th>
<th>Ca</th>
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<tbody>
<tr>
<td>700</td>
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<tr>
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<tr>
<td>900</td>
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<td>1.6</td>
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</tr>
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</tr>
<tr>
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</tr>
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</tr>
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</tr>
<tr>
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<td>34.0</td>
<td>66.0</td>
<td>8.0</td>
<td>92.0</td>
</tr>
</tbody>
</table>

However, O'Hara and Mercy (1963) point out that these results from the synthetic MgSiO$_3$-CaMgSi$_2$O$_6$ system are not directly comparable with natural pyroxene assemblages because most natural assemblages contain FeO and Al$_2$O$_3$ as important constituents. From their plots of the compositions of naturally occurring coexisting pyroxenes from pyroxene hornfels facies igneous assemblages and granulite facies metamorphic assemblages in the system CaSiO$_3$-MgSi$_2$-FeSiO$_3$, it is apparent that for the estimated/
estimated temperatures of crystallization of these assemblages, the apparent solvus limits extrapolated to the MgSiO₃-CaMgSi₂O₆ join are somewhat wider than those observed in the experimental Al₂O₃ free system (Schairer and Boyd 1957) at the same temperature. However, compared with the revised experimental solvus limits (Boyd and Schairer 1964) the extrapolated solvus limits for the granulite facies assemblages correspond closely to those expected for estimated temperatures of crystallization (700°C ± 100°C), although the extrapolated solvus limits for the higher temperature igneous assemblages are still somewhat wider than those expected at about 1100°C, their estimated temperature of crystallization.

Melting experiments at 30 kbar (O'Hara 1963a) on a coexisting pyroxene pair from a garnet Iherzolite nodule in kimberlite (the pyroxenes of which are for the present purposes essentially analogous in composition with the Norwegian pyroxenes under consideration) has indicated considerable mutual solubility between the pyroxenes at high temperatures near the beginning of melting of such garnet peridotite assemblages.

O'Hara and Mercy (1963) considered that the very limited mutual solubility between analysed coexisting pyroxenes in the Norwegian garnet peridotites indicated crystallization at relatively low temperatures, ~600°C.

Fig. 21 shows the very limited extent of the mutual solubility between analysed coexisting pyroxene pairs from the Norwegian garnet peridotites. To compare the solvus limits of these natural pyroxene pairs with the solvus limits of pyroxene pairs in the synthetic MgSiO₃-CaMgSi₂O₆ system (Boyd and Schairer 1964) two extrapolations are involved, namely an extrapolation to zero Fe/Fe+Mg ratio and one to zero R₂O₃ content.

The extrapolated solvus limits on the MgSiO₃-CaMgSi₂O₆ join (Fig. 21)
Variation in Ca/Ca+Mg+Fe vs Fe/Mg+Fe for coexisting pyroxenes from the Norwegian garnet peridotites.
\[
\frac{CA \times 100}{MG + FE + CA}
\]

\[
\frac{FE \times 100}{MG + FE}
\]
(Fig. 21) are approximately $\text{Ca}_{48.5}\text{Mg}_{51.5}$ and $\text{Ca}_{0.5}\text{Mg}_{99.5}$. This extrapolation, particularly for the Ca-rich pyroxene phase, is very small.

The effect of the $\text{R}_2\text{O}_3$ contents of the pyroxenes on the solvus limits is uncertain (see Bartholomé 1962, De Vore 1957) but the $\text{R}_2\text{O}_3$ content of these pyroxenes is low (0.90 - 3.08 wt.%) so that again the extrapolations involved for comparison with the synthetic $\text{MgSiO}_3$-$\text{CaMgSi}_2\text{O}_6$ system are small.

The extrapolated solvus limits for these Norwegian garnet peridotite pyroxene pairs are similar to those of pyroxene pairs from granulite facies assemblages with an estimated temperature of crystallization of $700^\circ\text{C} \pm 100^\circ\text{C}$ (O'Hara and Mercy 1963, Fig. 138).

This similarity and comparison with temperature estimates of the solvus limits of pyroxene pairs in the synthetic $\text{MgSiO}_3$-$\text{CaMgSi}_2\text{O}_6$ system (Boyd and Schairer 1964) indicates that temperatures probably as low as $600 - 700^\circ\text{C}$ were likely during the crystallization of the present primary mineral assemblages of the Norwegian garnet peridotites.

B. ALUMINA SUBSTITUTION IN THE PYROXENES

In the course of a study on the stability field of pyrope, Boyd and England (1960) found that 14-19 wt.% $\text{Al}_2\text{O}_3$ would dissolve in enstatite at 20 kbs. and $1400^\circ\text{C}$. This high solubility of $\text{Al}_2\text{O}_3$ in enstatite was interpreted as an effect of pressure. A maximum $\text{Al}_2\text{O}_3$ content of about 6% in enstatite at atmospheric pressure was noted in an orthopyroxene taken to be stable on the join diopside-pyrope (O'Hara and Schairer 1963).

The maximum $\text{Al}_2\text{O}_3$ content found in natural orthopyroxene is about 9 wt.% from hypersthenes in garnet granulites (Eskola 1952).

It/
It was thought that still higher Al$_2$O$_3$ contents might be found in orthopyroxenes from garnet peridotites, in particular in the garnet peridotite nodules in kimberlite pipes, as these are believed to have formed at depths of 100 kms, or more in the mantle (O’Hara and Mercy 1963).

However, analysed orthopyroxenes from garnet peridotite nodules in kimberlite contain uniformly low amounts of Al$_2$O$_3$ - of the order of 1-2% (O’Hara and Mercy 1963, Banno et al. 1963, Nixon et al. 1963).

Similarly, orthopyroxene from the Norwegian garnet peridotites contain surprisingly low amounts of Al$_2$O$_3$ (0.90 - 1.60 wt.%).

It must be borne in mind, however, that these orthopyroxenes come from assemblages which include pyrope-rich garnets and the solubility of Al$_2$O$_3$ in enstatite in such assemblages is expected to be governed by the reaction:

$$\text{Al-rich enstatite} \rightarrow \text{Al-poor enstatite} + \text{pyrope}.$$  

Boyd and England (1963) have studied this reaction in the synthetic system MgO-Al$_2$O$_3$-SiO$_2$, and have shown that when pyrope is stable the solubility of Al$_2$O$_3$ in enstatite does in fact drop according to the above reaction (see Table No.12).

<table>
<thead>
<tr>
<th>P kbs.</th>
<th>T°C</th>
<th>Wt. % Al$_2$O$_3$</th>
<th>Saturated assemblage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Data from Atmospheric</td>
<td>1300</td>
<td>$\leq 2\frac{1}{2}$</td>
<td>Pr enst$_{ss} +$ Fo + Co</td>
</tr>
<tr>
<td>Boyd and England</td>
<td>19.7</td>
<td>1400</td>
<td>14-19</td>
</tr>
<tr>
<td>(1963)</td>
<td>28.7</td>
<td>1400</td>
<td>9+1</td>
</tr>
<tr>
<td></td>
<td>40.0</td>
<td>1400</td>
<td>11+1</td>
</tr>
</tbody>
</table>

Comparison of the results obtained at 29-40 kb, with those at 20 kb, show that the saturation limit of Al$_2$O$_3$ in enstatite does in fact drop markedly when/
when pyrope appears as a stable phase. Nevertheless, the solubilities obtained at 29-40 kb, are still much higher than those found in natural enstatites from garnet peridotites. Moreover, continued increase in pressure produces a slight but significant increase in the solubility of Al$_2$O$_3$ in enstatite according to their data.

However, Boyd and England (1964) have conducted further experimental studies in the synthetic MgSiO$_3$-Mg$_2$Al$_2$Si$_3$O$_{12}$ system and clearly demonstrate that the Al$_2$O$_3$ content of enstatites coexisting with pyrope decreases with increasing pressure at constant temperature.

MacGregor and Ringwood (1965) have obtained similar results with mixtures of natural enstatite and pyrope-rich garnet, although the natural orthopyroxene in equilibrium with pyrope-rich garnet has rather higher total R$_2$O$_3$ content (less the R$_2$O$_3$ associated with Na$_2$O and K$_2$O) than a synthetic enstatite at equilibrium with pyrope under the same conditions. However, with increasing pressure the differences decrease so that at pressures > 50 kb, it is negligible, and both synthetic and natural enstatites can be directly compared.

Boyd and England (1964) have also demonstrated that the Al$_2$O$_3$ content of an enstatite in equilibrium with pyrope is not only a function of pressure, but is also a function of temperature. At 30 kb, pressure the Al$_2$O$_3$ content of enstatite from the synthetic CaO-free system enstatite-pyrope increases with temperature from ~5 wt.% at 1100°C to about 16 wt.% at 1650°C.

Melting experiments on garnet peridotites (O'Hara 1963a) and bimodal eclogite (O'Hara 1963b) at 30 kbs, also indicated that at high temperatures there is very extensive solid solution of potential garnet molecule in both CPX and OPX resulting in high Al$_2$O$_3$ contents of these pyroxenes.
pyroxenes. With decreasing temperature, however, the solubility diminishes appreciably. O'Hara (1965b) noted up to 11 wt.% \( \text{Al}_2\text{O}_3 \) in both CPX and OPX of GNT + CPX + OPX assemblages at the beginning of melting (1515 - 1565°C) at 30 kbar pressure.

Extensive solid solution of \( \text{Al}_2\text{O}_3 \) in diopside and along the \( \text{CaMgSi}_2\text{O}_6-\text{CaAl}_2\text{Si}_2\text{O}_6 \) join is known from experimental work (Segnit 1953, Hytonen and Schairer 1961), although these clinopyroxenes have been subsequently shown to be at least partly metastable (O'Hara and Schairer 1963).

Preliminary runs on the reaction Al-rich diopside \( \rightarrow \) Al-poor diopside + garnet with natural analysed minerals (MacGregor and Ringwood 1964) has shown that at 1500°C and 30-35 kbar., the diopside on the equilibrium curve has approximately 1 mole% more total \( \text{R}_2\text{O}_3 \) (less \( \text{R}_2\text{O}_3 \) associated with \( \text{Na}_2\text{O} \) and \( \text{K}_2\text{O} \)) than the coexisting enstatite.

The \( \text{R}_2\text{O}_3 \) contents of coexisting pyroxenes in the Norwegian garnet peridotite assemblages are tabulated in Table No. 13 which shows:

A: the excess \( \text{R}^{3+} \) (recalculated to \( \text{Al}_2\text{O}_3 \) wt.%) in clinopyroxenes after subtraction of \( \text{R}^{3+} \) equivalent to \( \text{Na}^{1+} \) to form \( [\text{Na}^{1+} \text{[R}^{3+}] \text{Si}_2\text{O}_6] \), compared with \( \text{Al}_2\text{O}_3 \) wt.% for coexisting orthopyroxenes from the Norwegian garnet peridotites.

B: the excess mole% \( \text{R}_2\text{O}_3 \) after subtraction of \( \text{R}_2\text{O}_3 \) associated with \( \text{Na}_2\text{O} \) for the coexisting CPX and OPX.

<table>
<thead>
<tr>
<th></th>
<th>T96</th>
<th>N23</th>
<th>N69</th>
<th>N70</th>
<th>N71</th>
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<td>OPX</td>
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<td>1.08</td>
<td>1.28</td>
<td>1.09</td>
<td>1.29</td>
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<tr>
<td>CPX</td>
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<td>1.39</td>
<td>0.98</td>
<td>1.61</td>
<td>0.50</td>
<td>0.85</td>
</tr>
<tr>
<td>OPX</td>
<td>0.50</td>
<td>1.46</td>
<td>1.13</td>
<td>1.06</td>
<td>1.18</td>
<td>0.70</td>
</tr>
<tr>
<td>CPX</td>
<td>0.69</td>
<td>1.03</td>
<td>0.59</td>
<td>0.88</td>
<td>0.80</td>
<td>0.47</td>
</tr>
</tbody>
</table>
It is debatable whether total $R_2O_3$ ($Al_2O_3 + Fe_2O_3 + Cr_2O_3$) or just $Al_2O_3$ is the best basis for comparison. Total $R_2O_3$ is theoretically more suitable, but errors can result, however, from errors in $Fe_2O_3$ attributable to inaccuracies in the analytical methods of determining $FeO$ ($Fe_2O_3$ being determined by difference from the total iron value).

The data for the chemically analysed pairs from the Norwegian garnet peridotites fails to substantiate the experimental evidence (MacGregor and Ringwood 1964) that diopside clinopyroxenes contain slightly more mole$%$ $R_2O_3$ than the coexisting enstatites (the basis for comparison being total mole$%$ $R_2O_3$, after subtraction of $R_2O_3$ to match the $Na_2O$ in the analysis as pyroxene $NaRSi_2O_6$). In fact, in all but one case the opposite seems to be the case. It must be borne in mind, however, that the pyroxenes from the Norwegian garnet peridotites were probably recrystallized at appreciably lower temperatures ($\sim 600-700^0C$) than those of the experimental conditions studied by MacGregor and Ringwood (1964).

In both clin- and orthopyroxenes the amounts of excess total $R_2O_3$ are small.

The experimental studies reviewed here suggest that, even when coexisting with a pyrope-rich garnet, comparatively high $R_2O_3$ contents are to be expected in pyroxenes formed at high temperatures near the beginning of melting in the eclogite mineral facies except at very high pressures ($>50$ kbs.).

The remarkably low $R_2O_3$ contents of both clin- and orthopyroxenes from the Norwegian garnet peridotites probably indicate that the present mineral assemblages were equilibrated at temperatures considerably below the temperatures of the beginning of melting.

O'Hara/
O’Hara and Mercy (1963) have noted that granulite facies orthopyroxenes of highest $\text{Al}_2\text{O}_3$ contents are found in assemblages which are CPX free and the OPX coexists with a CaO poor garnet. When CPX is present or even in the presence of a relatively CaO rich garnet, the $\text{Al}_2\text{O}_3$ content of the OPX is much lower than the observed maximum value. Extending the explanation to the case of the low $\text{Al}_2\text{O}_3$ contents of OPX in the garnet peridotites, they suggested that such orthopyroxenes were either formed in an $\text{Al}_2\text{O}_3$ poor environment, or in the presence of CPX, or a garnet which is, in this context, relatively CaO rich. Such an explanation is in agreement with results obtained from experiments on natural mineral assemblages from garnet peridotites (O’Hara 1963c). However, O’Hara and Mercy 1963) also stressed that the relatively low temperatures of equilibration of most natural orthopyroxene and garnet bearing granulites and eclogites are also very important factors in limiting the extent of $\text{Al}_2\text{O}_3$ substitution in the OPX.

The present author considers that the apparent relatively low temperatures of equilibration of the present primary mineral assemblages of the Norwegian garnet peridotites is probably the most important single factor in limiting the extent of the $\text{Al}_2\text{O}_3$ substitution in both clinopyroxene and orthopyroxenes.

In Fig. 22, adapted from Boyd and MacGregor (1964), the stability field of garnet peridotite in the synthetic system $\text{MgO-Al}_2\text{O}_3-SiO_2$ is subdivided by isopleths giving the expected solubility of $\text{Al}_2\text{O}_3$ in enstatite in equilibrium with pyrope under differing temperature and pressure conditions (Boyd and England 1964). Natural enstatites and pyrope-rich garnets contain small amounts of $\text{Fe}^{3+}$ and $\text{Cr}^{3+}$ which substitute for $\text{Al}^{3+}/$
Stability field of garnet peridotite subdivided by isopleths giving the expected solubility of $\text{Al}_2\text{O}_3$ in enstatite in equilibrium with pyrope under differing physical conditions - after Boyd and MacGregor (1964).
FIG. 22.

WT.% AL₂O₃ IN ENSTATITE

12/10/8/6/4/

OCEANIC GEOTHERM

PRECAMBRIAN SHIELD GEOTHERM

GRAPHITE ⇄ DIAMOND

EN + SP ⇄ PY. + FO

T°C

DEPTH KMS.

10 20 30 40 50 60 70 80 90 100

PRESSURE KBS.
Al$^{3+}$ in these structures. However, the experimental results obtained with mixtures of natural enstatite and pyrope-rich garnets (MacGregor and Ringwood 1964) are in relatively good agreement with those obtained for the synthetic system. The effect of FeO on the phase relations of Fig. 22 will be to enlarge the stability field of garnet to lower pressures, but it is not yet certain how FeO will affect the solubility of Al$_2$O$_3$ in enstatite, although it will probably reduce it.

In Fig. 22, the pyrope stability boundary En + Sa + Si = Py is from Boyd and England (1962), the curve for the reaction 4En + Sp = Py + Fo from MacGregor (1964), the diamond = graphite curve from Bundy et al. (1961) A - the oceanic geotherm and B - the Pre-Cambrian shield geotherm from Ringwood, MacGregor and Boyd (1964), and C is equivalent to curves (2,3,4) in Fig. 19b. - see MacGregor (1965).

The exact location of the low pressure boundary of the garnet peridotite stability field is uncertain (see MacGregor 1965) but in terms of Fig. 22 it is conceivable to interpret the Norwegian garnet peridotites with orthopyroxenes containing 0.90-1.60 wt.% Al$_2$O$_3$ as being due to relatively low temperatures of formation ($\sim$400-500°C) under metamorphic conditions in the crust, or alternatively they may represent solid intrusions from the mantle, possibly coming from depths less than those required to form diamond. However, the independent evidence, from the limited mutual solubility of coexisting pyroxene pairs from the Norwegian garnet peridotites, for low temperatures ($\sim$600-700°C) of formation suggests a metamorphic origin at depth within the crust of the present primary mineral assemblage of these garnet peridotites.

Jadeite (NaAlSi$_2$O$_6$) and Ca Tschermak's molecule (CaAl$_2$Si$_4$O$_{10}$) are important,
important indicators of the physical conditions of formation of clinopyroxenes.

Comparison of isothermal sections on the join diopside-albite and diopside-
anorthite (Kushiro 1965c) indicates that the content of Ca Tschermak's
component in diopside coexisting with plagioclase is higher than that of
jadeite under the pressure-temperature conditions where garnet of the
pyrope-grossularite series is not stable. In the stability field of garnet,
however, the content of jadeite becomes higher than that of Ca Tschermak's
molecule - the content of the latter in fact decreasing with increasing
pressure in the garnet stability field according to the reaction:

\[ \text{Al-rich clinopyroxene} \rightarrow \text{Al-poor clinopyroxene} + \text{garnet}. \]

It is to be expected, therefore, that the content of Ca Tschermak's
component relative to the jadeite component will be small in clinopyroxene
from eclogite mineral facies assemblages.

White (1964) has demonstrated excellent separation in the
compositions of clinopyroxenes from true eclogite and granulite facies
assemblages, attributable to the higher pressures of formation of the eclogite
assemblage. The basis of this distinction was the ratio of jadeite molecule:
Ca Tschermak's molecule in the clinopyroxene. He showed that all clinopyroxenes
with a molecular ratio of jadeite: Ca Tschermak's molecule of \( \frac{1}{5} \) were from
granulites, while those with a ratio of \( \frac{1}{5} \) were all from eclogites.

In table 14 data for analysed clinopyroxenes from the Norwegian
garnet peridotites are given as follows:

I. Percentage \( R_2O_3 \) (as mole% of total \( R_2O_3 \)) not involved in substitution
   of the form \( Na^{1+}R^{3+}\text{Si}_2O_6 \), hence involved in Ca Tschermak's substitution.

II. Percentage \( R_2O_3 \) (as mole% of total \( R_2O_3 \)) involved in jadeite \( (Na^{1+}Al^{3+}
   \text{Si}_2O_6) \) substitution, calculated after formation of \( NaFe^{3+}\text{Si}_2O_6 \) and
   \( NaCr^{3+}\text{Si}_2O_6 \).
TABLE NO. 14.

<table>
<thead>
<tr>
<th></th>
<th>T96</th>
<th>N23</th>
<th>N26</th>
<th>N69</th>
<th>N70</th>
<th>N71</th>
<th>T153</th>
<th>U₂</th>
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<tbody>
<tr>
<td>I</td>
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<td>25.5</td>
<td>32.4</td>
<td>44.4</td>
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<td>20.9</td>
<td>14.8</td>
<td>43.5</td>
<td>19.0</td>
</tr>
<tr>
<td>II</td>
<td>42.9</td>
<td>32.8</td>
<td>56.0</td>
<td>33.8</td>
<td>30.7</td>
<td>52.7</td>
<td>45.0</td>
<td>53.1</td>
<td>14.5</td>
<td>51.8</td>
</tr>
</tbody>
</table>

These clinopyroxene compositions do not endorse such a clear separation of eclogite and granulite facies clinopyroxenes on the basis of jadeite: Ca Tschermak's molecule ratio as suggested by White (1964). Nevertheless, the majority of the clinopyroxenes from these Norwegian garnet peridotites contain a higher content of jadeite molecule than of Ca Tschermak's molecule. However, in three cases (N23, N70 and U₆) the content of Ca Tschermak's molecule is higher, being significantly so in the clinopyroxene from the Ugelvik assemblage U₆.

Boyd and England (1965) have followed Solar et al. (1964) in demonstrating the stability of clinoenstatite as the low temperature polymorph of enstatite below about 630-700°C over the pressure range 0-30 kbs. This raises a problem in the case of the Norwegian garnet peridotite assemblages, which it has just been suggested have been equilibrated at temperatures of about 600-700°C. Rhombic enstatite is the form which occurs in these assemblages and no evidence of inversion to or from clinoenstatite has been observed, as might be expected from the experimental work.

Clinoenstatite has, in fact, only exceptionally been reported in any natural mineral assemblages, namely in a Papuan tholeiite (see Tilley, Yoder and Schairer 1964) and in some chondritic meteorites (Mason 1962).
The absence of clinoenstatite from the garnet peridotite assemblages may indicate that the $R_2O_3$ contents of these enstatites enlarge the stability field of rhombic enstatite relative to clinoenstatite as suggested by Boyd and England (1965), hence displacing the inversion to even lower temperatures.

C. TEXTURAL EVIDENCES

Although it has just been demonstrated that the solvus limits between coexisting pyroxene pairs and the alumina contents of these pyroxenes suggest low temperatures of equilibration of the present primary mineral assemblages of the Norwegian garnet peridotites, there is some textural evidence of an original high temperature history of these rocks.

The principal evidence consists of rare orientated lamellae of garnet in clinopyroxene (Plate 6) similar to textures described in certain eclogite nodules from Kimberlite (Williams 1932 Plates 82 and 83). The best example of this texture occurs in a rather amphibolitized but only mildly cataclasized garnet lherzolite from the Grubse peridotite mass at Almklovldalen, which contains large clinopyroxene grains with in certain cases regular garnet lamellae $//$(010). Elsewhere, however, more irregular blebs of garnet have been occasionally noted in clinopyroxenes in garnet peridotites from Almklovldalen and also from Kalskaret.

The clinopyroxenes of the Grubse garnet lherzolite also illustrate another significant textural feature. The clinopyroxene of this rock has not been analysed but is probably a chrome-diopside being strikingly green in thin section (significantly more so than the clinopyroxenes from most of/
of the garnet peridotites). It can be seen recrystallizing marginally to colourless diopside thronged with minute inclusions of a red brown mineral considered probably to be chrome spinel.

Clinopyroxenes from the Ugelvik garnet peridotites similarly appear to have mostly recrystallized to a colourless diopside and to be expelling chrome spinel inclusions, while the orthopyroxene in these rocks also appears to have recrystallized and to be expelling exsolved inclusions. Similar textures have been noted among pyroxenes from certain Kalskaret and other Almklovdalen garnet peridotites.

It is considered that these textural exsolution features indicate the likely higher original $R_2O_3$ contents, than their general present $R_2O_3$ contents, of pyroxenes from these garnet peridotites, in turn indicating higher temperatures of formation than those estimated ($\sim 600-700^0C$) to explain certain mineralogical features in sections 5A and B above.

It appears that the textural features described above in clinopyroxenes are rare relics of an original high temperature history of the Norwegian garnet peridotites. In most cases, however, any previous high temperature history has been totally obscured by subsequent recrystallization at lower temperatures within the eclogite mineral facies. It is considered that the low $R_2O_3$ contents of pyroxenes from all the analysed garnet peridotite assemblages reflect the predominant low temperature equilibration of these assemblages while rare relict pyroxenes with higher $R_2O_3$ contents have been overlooked by bulk mineral separation and analysis, just the average $R_2O_3$ content being obtained.

The application of electron microprobe analytical techniques on single/
single pyroxene grains is called for to demonstrate the extent of the variations in $R_2O_3$ content between apparent 'original' and recrystallized pyroxene grains. It is hoped that an opportunity to perform such analyses may arise in the future.

Abundant textural evidence of recrystallization of the eclogite mineral facies assemblages of these garnet peridotites is apparent being consequent upon the deformation of these rocks resulting from their tectonic emplacement origin. However, as the percentage of unrecrystallized strained porphyroclasts is quite high it seems unlikely that this recrystallization in evidence was the one responsible for the apparent generally complete equilibration of the eclogite mineral facies assemblages at comparatively low temperature ($\sim 600-700^\circ C$). Rather it may represent yet a second phase of recrystallization of the mineral assemblages of these rocks.

In fact it may be that during their emplacement into their present position, and perhaps catalysed by the tectonic forces of deformation, the mineral assemblages of these garnet peridotites have suffered essentially continuous recrystallization in attempts to re-establish equilibrium with the physical conditions of the environments through which they have passed.

Whether or not recrystallization was so repetitive, it is clear from the mineralogical and textural features described above that despite the evidence of a general equilibration of these mineral assemblages at relatively low temperatures ($\sim 600-700^\circ C$) within the eclogite mineral facies, relics of an original higher temperature history exist.

**B. SECONDARY MINERAL ASSEMBLAGES:**

1./
1. TAFJORD AND ALMKLOVDALEN PERIDOTITES.

Chlorite and amphibole occur as sparse secondary grains in the garnet-free peridotites. Both are free of any major mechanical deformation effects and are therefore later in growth than the extensive deformation associated with the tectonic emplacement of these peridotite masses. They are generally of random orientation commonly askew to and growing over the principal foliation in these rocks. Chlorite flakes are sometimes orientated in the shear slip planes of microfolded dunites, however, and also in the foliation planes.

The chlorite, blue grey in hand specimen but colourless in thin section with prominent lamellar twinning \(\|(001)\) and rather anomalous interference colours, has been X-rayed by Lappin (1962) and identified as a kammerite, a chlorite containing \(-2\% \text{Cr}_2\text{O}_3\).

X-ray fluorescence analysis on a chlorite from a Kaldhussaeter peridotite (T237) indicated about 2.7\% \text{Cr}_2\text{O}_3.

Chlorite forms rims round chromite in some garnet-free peridotites but does not appear to be related to the opaque ore fraction in the garnet peridotites. However, in some garnet peridotites chlorite forms a broad outer rim round garnets with an inner amphibole alteration rim.

O'Hara and Mercy (1963) have analysed the amphibole from a garnet-free Kalskaret peridotite (N6) and have shown it to be essentially a tremolite, with low \(\text{Al}_2\text{O}_3\), \(\text{Na}_2\text{O}\) and \(\text{Fe}^{2+}/\text{Mg}^{2+}\) ratio. It is suggested that its CaO content (11.10 wt.\%) indicates the original presence of some clinopyroxene in these rocks. Such amphibole is colourless in thin section.

O'Hara and Mercy (1963) have also analysed amphiboles from a garnet/
garnet peridotite (N69) and a garnet websterite (N23) and shown them to be magnesian hornblende. These amphiboles are pleochroic from pale yellow to pale green in thin section, and contain much higher Al₂O₃, Na₂O and Fe²⁺/Mg²⁺ ratio than the tremolitic amphibole of the garnet-free peridotites. This amphibole occurs marginally to garnets and clinopyroxenes and appears to have formed by reaction between these minerals and the olivine of the matrix. The more highly altered garnet peridotites contain garnets with pronounced reaction rims of this amphibole sometimes with a concentration of spinel inclusions. Occasionally there is a less well defined outer reaction rim of chlorite flakes.

The degree of serpentinization of these peridotites is generally very slight and as serpentine can be observed pseudomorphing the dimensionally orientated olivine matrix and cutting through deformed porphyroclasts, it is considered likely to have occurred subsequent to the tectonic emplacement of these masses.

Textural evidence outlined in Chapter II indicates secondary growth of enstatite in some of the peridotites, in particular in the Ranukdalen harzburgites. Experimental studies on the system MgO-SiO₂-H₂O led Bowen and Tuttle (1949) to suggest that hydrothermal transfer of SiO₂ may convert olivine into enstatite at low temperatures if there is insufficient water available for complete conversion into more hydrous phases (talc, serpentine, brucite etc.). Such an origin could explain the occurrence of secondary enstatite in these rocks. Talc is commonly found as an alteration mineral associated with this secondary enstatite.

2. UGELVIK PERIDOTITES

Garnets/
Garnets in the Ugelvik garnet peridotites often possess very extensive kelyphitic reaction rims (Plate 7). These reaction rims consist of the most part of a pale brown fibrous semi-opaque material which is very difficult to resolve optically. However, detailed optical examination of numerous examples indicates that this fibrous reaction rim consists primarily of orthopyroxene with minute inclusions of chrome-spinel. Less prominent streaks of a higher birefringence material with oblique extinction and similar relief are probably of clinopyroxene but may be of amphibole.

Occasionally around the external margins of the reaction rims, the fibrous material appears to be recrystallized into clear orthopyroxene with separate granules of chrome-spinel. External to that concentrations of colourless amphibole grains often occur.

These reaction rims may be interpreted in terms of a retrogression of the eclogite facies assemblage (olivine + garnet) to a granulite facies assemblage (pyroxenes + spinel). This reaction involved may be generalized as follows, with an assumed garnet composition:

\[
\text{olivine} + \text{garnet} \rightarrow \text{orthopyroxene} + \text{chrome-spinel}
\]

\[
(Mg,Fe)_2SiO_4 + (Mg,Fe)_2Ca(Al,Cr)_2Si_2O_6 \rightarrow 2(Mg,Fe)SiO_3 + Ca(Mg,Fe)Si_2O_6 + (Mg,Fe)(Al,Cr)_2O_4
\]

The proportion of clinopyroxene formed according to this reaction will depend upon the calcium content of the garnet. The orthopyroxenes from this environment contain only very limited amounts of CaO (Table 1), hence a restricted amount of clinopyroxene would be expected to form to absorb the excess of CaO released by the breakdown of the calcium poor garnet.
garnet (Table 3) found in these rocks.

The concentration of amphibole grains often observed as a partial external rim around the garnets may indicate the later onset of amphibolite facies metamorphic conditions and introduction of water, perhaps associated with the extensive serpentinization of the peridotites.

In these garnet peridotites serpentine veins can be observed in several cases to cut across the inner fibrous garnet reaction rims, suggesting that the formation of these reaction rims preceded the onset of serpentinization. Furthermore, serpentine veins can be observed passing straight through highly deformed pyroxene and olivine grains, and also pseudomorphing highly distorted grains.

On the other hand, the amphibole grains which are generally clear, unstrained and interstitial in habit to the deformed grains are not shot through with serpentine veins as are all the other minerals but in fact often appear to be growing over and obscuring serpentine veins.

These relationships therefore suggest the following conclusions:

1. Serpentinization occurred after the formation of the fibrous garnet reaction rims.
2. Serpentinization was subsequent to the intense deformation of the minerals and hence probably subsequent to the tectonic emplacement of these peridotite masses.
3. Growth of amphibole was a late feature being post deformation and probably also mostly subsequent to or contemporaneous with the serpentinization.

The reaction rims around the garnets in the Ugelvik garnet peridotites are quite distinct from those around the garnets in the Kalskaret and/
and Almklovdalen garnet peridotites. At these other localities the reaction rims are clearly composed of pale green amphibole, sometimes with a broad outer rim of chlorite.

Chlorite has not been observed, however, as an alteration mineral in the Ugelvik garnet peridotites.

The Ugelvik garnet kelyphitic rims seem most closely paralleled by those described by Nixon et al. (1963) around garnets in garnet peridotite nodules from Basutoland kimberlites. They describe the kelyphitic rims as consisting of ...'an inner zone of brown, semi-opaque, optically unresolvable fibrous material clearing outwardly into a border of spinel granules, phlogopite, chlorite and amphibole.' From an examination of electron micrographs prepared from diluted macerated fibrous kelyphite they observed three forms which they considered represented chrome-bearing spinel, enstatite and/or hornblende, mica and/or chlorite.

They determined the presence of orthopyroxene and spinel in one of their specimens but other than that no actual determinations have been made.

Wagner (1914) also detected optically orthopyroxene and spinel in kelyphitic garnet reaction rims from garnet peridotite nodules in kimberlites.

Nixon et al (1963) considered that the kelyphite formed rapidly from chrome pyrope by the addition of water, the reaction being promoted by a decrease of pressure probably due to kimberlite intrusion.

However, in the case of the Ugelvik garnet peridotites it seems likely that the addition of water associated with serpentinization occurred after the formation of the kelyphitic reaction rims. It is considered that the inner fibrous reaction rims are possibly just the direct result of decreasing pressure conditions connected with the emplacement of these peridotites/
peridotites initiating from greater depths.

It is interesting to note how the apparent retrogression of the Ugelvik gneisses from granulite to amphibolite facies grade appears to be paralleled in the garnet reaction rims with an inner rim of orthopyroxene and spinel, and an outer rim of amphibole.

This does not necessitate, however, that these garnet peridotite masses were emplaced in their present position during the apparent partial retrogression of the gneiss complex from granulite to amphibolite facies grade. Rather the garnet peridotites may have passed from an environment of granulite facies conditions to one of amphibolite facies conditions during their intrusion from depths within the crust or upper mantle. The pressure and temperature conditions of these facies probably, in fact, overlap, availability of water being the most important distinction between them.

C. SUMMARY OF CONCLUSIONS:

The following points raised in the discussion of the mineralogy of the Norwegian peridotites in this chapter, are particularly important in further discussions on the petrogenesis of these rock masses to follow (Chapter V).

1. The large extent of the mineral composition variations, both within the solid solution series and in minor element content (Cr$_2$O$_3$, TiO$_2$ etc.), renders it unlikely that the Kalskaret and Almklovdalen garnet peridotites simply represent local segregations of garnet and pyroxenes within the peridotites precipitated by normal igneous crystallization processes, whether in situ or at depth. The minerals of these peridotites/
peridotites show notably increased Fe\(^{2+}\):Mg\(^{2+}\) ratios with increasing garnet and clinopyroxene contents of the mineral assemblage.

2. The Ugelvik garnet peridotites are far more uniform mineralogically, both in mineral compositions and in modal contents of constituent minerals, than the Kalskaret and Almklovdalen garnet peridotites. Furthermore their constituent minerals are particularly more magnesian and chromium rich than those generally occurring in the other garnet peridotites.

It would appear that the mineral layering observed in the Ugelvik peridotite masses may be adequately explained by the sorting of an eclogite facies assemblage (OL + GNT + CPX + OPX + chromite) precipitated by igneous crystallization processes. The assemblage \(U_6\) is exceptional, however, and requires another explanation.

3. All the Norwegian garnet peridotite assemblages have the expected equilibrium distribution of Fe\(^{2+}\)+Fe\(^{3+}\)+Mn\(^{2+}\)/Mg\(^{2+}\) among constituent minerals indicating that equilibrium has been generally attained or closely approached in the formation of the present primary mineral assemblages of these rocks, in turn suggesting the likelihood of a metamorphic origin of these mineral assemblages.

4. The solvus limits of coexisting pyroxene pairs and the low R\(_2\)O\(_3\) contents of the pyroxenes indicate equilibration of the present primary mineral assemblages of the Norwegian peridotites at relatively low temperatures (\(\sim\)600-700\(^0\)C) within the eclogite mineral facies.

5. A review of relevant experimental petrology studies, in particular with reference to the R\(_2\)O\(_3\) contents of the pyroxenes, indicates that the Norwegian garnet peridotites may represent tectonically emplaced fragments of the upper mantle or they may have been formed by metamorphism at greater/
greater depths in the crust of peridotite which originally crystallized to another mineral facies assemblage.

The independent evidence for a low temperature equilibration of the mineral assemblages of these garnet peridotites (conclusions 3 and 4) suggests that a metamorphic origin at depth within the crust was more likely.

On the other hand it seems improbable in the light of the expected pressure and temperature stability field of garnet peridotite that they were of primary igneous crystallization in their present environment as envisaged by Eskola (1921). Furthermore evidence from the textural and contact relations of these masses (Chapter II) is also contrary to such an origin.

Textural evidence of rare exsolution textures indicates, however, higher original $\text{R}_2\text{O}_3$ contents in the pyroxenes which suggests an original higher temperature history of these garnet peridotites. It appears that only occasional relics of this original higher temperature history have survived a subsequent period of extensive recrystallization and general equilibration of the mineral assemblages of these rocks at appreciably lower temperatures within the eclogite mineral facies.
A. PRESENTATION OF ANALYSES AND DISCUSSION OF NORMS.

13 rock samples have been collected from a section across the Kalskaret garnet peridotite lens (see Fig. 8) and analysed to show the extent of the chemical variation which has resulted in the production of these garnetiferous rock types.

The exact location of collection of each of these specimens and details of their mineral assemblages are noted in the Appendix III. Estimated volume proportions of minerals in these specimens are given in Table No. 4.

The chemical compositions of these 13 samples are given in Table No. 15. Several of these specimens have been analysed by both rapid silicate 'wet' and X-ray fluorescence techniques. The choice of the preferred values of such specimens, as given in Table No. 15, is discussed in the Appendix IV, in a section dealing with the methods of analysis used.

The C.I.P.W. norms calculated from the analyses of these rock samples are given in Table No. 16. As a few of the samples show a moderate degree of serpentinization or amphibolitization and hence have significant $\text{H}_2\text{O}^+$ contents, the norms of these particular samples have been based on recalculated water-free analyses (Table No. 17). Recalculation of the relevant analyses assuming that serpentinization was a constant composition reaction with addition of water and without leaching of other components yielded quite acceptable original compositions when compared with bulk rock compositions calculated using the modal proportions of the minerals and likely mineral compositions (see Appendix I).
### Table No. 15

**Bulk Rock Analyses of Kalskaret Peridotites**

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<td>0.24</td>
<td>0.36</td>
<td>0.30</td>
<td>0.25</td>
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<td>0.33</td>
<td>0.78</td>
<td>0.35</td>
<td></td>
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</tr>
</tbody>
</table>

\(^{*}\)\(\text{Fe+Mn+Ni} \div \text{Fe+Mn+Ni+Mg}\) Molecular ratio in normative ferromagnesian silicates.
<table>
<thead>
<tr>
<th></th>
<th>833</th>
<th>834</th>
<th>835</th>
<th>842</th>
<th>843</th>
<th>845</th>
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<tr>
<td>SiO₂</td>
<td>42.51</td>
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<td>45.06</td>
<td>47.49</td>
<td>43.29</td>
<td>44.76</td>
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<tr>
<td>TiO₂</td>
<td>0.04</td>
<td>0.01</td>
<td>0.48</td>
<td>0.23</td>
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<td>0.04</td>
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<tr>
<td>Al₂O₃</td>
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<td>0.51</td>
<td>7.24</td>
<td>8.62</td>
<td>0.29</td>
<td>3.72</td>
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<tr>
<td>Cr₂O₃</td>
<td>0.38</td>
<td>0.30</td>
<td>0.19</td>
<td>0.32</td>
<td>0.37</td>
<td>0.58</td>
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<tr>
<td>*FeO</td>
<td>7.49</td>
<td>6.41</td>
<td>12.03</td>
<td>8.58</td>
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<td>6.56</td>
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<td>0.15</td>
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<tr>
<td>MgO</td>
<td>45.89</td>
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<td>CaO</td>
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<td>8.38</td>
<td>0.10</td>
<td>4.26</td>
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<tr>
<td>Na₂O</td>
<td>0.10</td>
<td>0.04</td>
<td>0.56</td>
<td>0.91</td>
<td>0.03</td>
<td>0.49</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.03</td>
<td>0.03</td>
<td>0.07</td>
<td>0.12</td>
<td>0.01</td>
<td>0.07</td>
</tr>
<tr>
<td>TOTAL</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
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</tr>
</tbody>
</table>

*FeO = Total iron calculated as FeO
Bulk rock analyses of 5 Ugelvik peridotite specimens have been performed - the values being given in Table No. 18. The high degree of serpentinization of these rocks, one specimen contains just over 14% $H_2O^+$, necessitates an attempt to recalculate the original water free compositions.

It was found that recalculation assuming that serpentinization was essentially a constant composition reaction with addition of water, increase in volume, and without significant leaching of other components, when compared with the bulk rock compositions calculated using the mineral analyses and modal proportions of the minerals gave reasonably acceptable original compositions in only two of the five specimens - namely $U_3$ and $U_6$.

Serpentinization of $U_2$, $U_8$ and $U_{16}$, however, appears to have involved considerable leaching of $FeO$, $NiO$ and $MgO$ in particular so that the recalculated water free compositions assuming simply addition of water without leaching are significantly low in $MgO$, $FeO$, and $NiO$ with respect to $SiO_2$.

An attempt has been made to correct for the amount of leaching of these elements in the recalculation of the analyses of $U_2$, $U_8$ and $U_{16}$. The form of this recalculation and the calculations suggesting that significant leaching has in fact occurred in these cases are discussed in detail in the section on 'Serpentinization' in the Appendix.

However, despite these corrections; uncertainties as to the exact extent of the leaching in these two cases and the possibility that minor leaching may also have occurred in the serpentinization of the other rock samples, leave slight reservations in one's mind as to just how precisely the $MgO+FeO+NiO/SiO_2$ ratios in the assumed original compositions represent those of the original pre-serpentinization compositions of these Ugelvik peridotites.
<table>
<thead>
<tr>
<th>Analysis No.</th>
<th>Field Sample No.</th>
<th>WHOLE ROCK ANALYSES OF UGELVIK PERIDOTITES</th>
<th>WHOLE ROCK ANALYSES RECALCULATED ON WATER FREE BASIS</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>U_2</td>
<td>U_3</td>
<td>U_6</td>
</tr>
<tr>
<td>SiO_2</td>
<td>39.67</td>
<td>36.72</td>
<td>49.49</td>
</tr>
<tr>
<td>TiO_2</td>
<td>0.01</td>
<td>&lt;0.005</td>
<td>0.07</td>
</tr>
<tr>
<td>Al_2O_3</td>
<td>2.00</td>
<td>0.13</td>
<td>3.70</td>
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<tr>
<td>Cr_2O_3</td>
<td>0.33</td>
<td>0.17</td>
<td>0.71</td>
</tr>
<tr>
<td>Fe_2O_3</td>
<td>4.32</td>
<td>5.22</td>
<td>2.45</td>
</tr>
<tr>
<td>FeO</td>
<td>2.21</td>
<td>1.25</td>
<td>4.12</td>
</tr>
<tr>
<td>MnO</td>
<td>0.10</td>
<td>0.10</td>
<td>0.16</td>
</tr>
<tr>
<td>NiO</td>
<td>0.29</td>
<td>0.31</td>
<td>0.12</td>
</tr>
<tr>
<td>MgO</td>
<td>39.65</td>
<td>42.17</td>
<td>25.86</td>
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<tr>
<td>CaO</td>
<td>1.06</td>
<td>0.05</td>
<td>10.40</td>
</tr>
<tr>
<td>Na_2O</td>
<td>0.14</td>
<td>0.06</td>
<td>0.28</td>
</tr>
<tr>
<td>K_2O</td>
<td>0.06</td>
<td>0.01</td>
<td>0.02</td>
</tr>
<tr>
<td>H_2O^+</td>
<td>10.71</td>
<td>14.02</td>
<td>3.23</td>
</tr>
<tr>
<td>TOTAL</td>
<td>100.55</td>
<td>100.21</td>
<td>100.61</td>
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</table>
The preferred recalculated water free analyses are also given in Table No. 18.

The estimated modal proportions (by volume) of the minerals in these analysed Ugelvik peridotite assemblages are given in Table No. 4.

The C.I.P.W. norms based on the recalculated water free analyses of these Ugelvik peridotite samples are given in Table No. 19. It is considered that the small amount of nepheline in the norm of U_2 and the small amounts of nepheline, kaliophilite and silica deficiency in the norm of U_16, are insignificant because of uncertainties regarding the original MgO+FeO+NiO/SiO_2 ratios in these rocks resulting from their high degree of serpentinization and the evidence of leaching of MgO etc. It should also be borne in mind that in the recalculated analyses all the iron is assumed to be in the ferrous state which is unlikely. Only relatively small amounts of Fe_2O_3 calculated as magnetite in the norm would be sufficient to eliminate all the nepheline, kaliophilite and SiO_2 deficiency from the norms of U_2 and U_16.

For comparison the C.I.P.W. norms of the analyses of U_2 and U_16 recalculated water free assuming that no leaching occurred during serpentinization are also given in Table No. 19. It can be seen that these norms contain considerable amounts of orthopyroxene correlating with the lower MgO+FeO/SiO_2 ratio of the assumed original composition.

The most significant feature demonstrated by the normative mineral calculations of the Ugelvik peridotite compositions is the consistency of the Fe+Mn+Ni/Fe+Mn+Ni+Mg ratios in the normative ferromagnesian silicates of the garnet-free assemblage U_3, the garnet-poor assemblages U_2 and U_16, and the garnet-rich assemblage U_8.

This feature is notably different from the situation in the Kalskaret and Almklovtdalen peridotites where with increasing GNT and CPX contents the Fe/
**TABLE NO. 19.**

C.I.P.W. NORMS OF ANALYSES OF UGELVIK PERIDOTITES RECALCULATED ON A WATER FREE BASIS

<table>
<thead>
<tr>
<th>SAMPLE NO.</th>
<th>U_2</th>
<th>U_3</th>
<th>U_6</th>
<th>U_8</th>
<th>U_16</th>
<th>C.I.P.W. NORMS OF ALTERNATIVE WATER FREE COMPOSITIONS</th>
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<tr>
<td>Cr</td>
<td>0.36</td>
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<td>0.12</td>
<td>-</td>
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<tr>
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<td>0.59</td>
<td>0.71</td>
<td>-</td>
<td>1.35</td>
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<tr>
<td>An</td>
<td>4.85</td>
<td>0.06</td>
<td>11.61</td>
<td>19.14</td>
<td>1.99</td>
<td>5.19</td>
</tr>
<tr>
<td>Ne</td>
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<td>-</td>
<td>-</td>
<td>-</td>
<td>0.50</td>
<td>-</td>
</tr>
<tr>
<td>Kp</td>
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<td>-</td>
<td>-</td>
<td>-</td>
<td>0.10</td>
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</tr>
<tr>
<td>Co</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Di</td>
<td>0.46</td>
<td>0.51</td>
<td>0.17</td>
<td>0.19</td>
<td>-</td>
<td>0.47</td>
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<td>Hd</td>
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<td>0.02</td>
<td>5.24</td>
<td>4.32</td>
<td>-</td>
<td>0.05</td>
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<tr>
<td>En</td>
<td>-</td>
<td>7.94</td>
<td>25.01</td>
<td>14.34</td>
<td>-</td>
<td>11.21</td>
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<tr>
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<td>-</td>
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<td>86.00</td>
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<tr>
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<td>100.01</td>
<td>100.00</td>
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<td>100.02</td>
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*Fe+Mn+Ni/Fe+Mn+Ni+Mg*  Molecular ratio in normative ferromagnesian silicates

Ab/Ab+An  Ratio of plagioclase contents.
Fe+Mn+Ni/Fe+Mn+Ni+Mg ratio of the normative ferromagnesian silicates increases markedly from its value in the garnet free peridotites (Table No. 16; see also Mercy and O'Hara 1965a, Table 2).

Comparing the bulk composition of a garnet lherzolite (N69) and a garnet websterite (N71) collected only a few feet apart at Lien, Almklovdalen - O'Hara and Mercy (1963) noted that these rocks cannot be regarded as the eclogite facies metamorphosed equivalents of basic igneous rocks formed by igneous differentiation under pyroxene hornfels facies conditions because, added to the chemical abnormality of the samples there is an anomalous variation of Fe/Fe+Mg ratio of the normative ferromagnesian silicates in the C.I.P.W. norm, when compared with the composition of the bulk normative felspars. The assemblage with the more calcic normative felspar (N71) also has the more iron-rich normative ferromagnesian silicates, a behaviour which is contrary to that of normal igneous differentiation. Mercy and O'Hara (1965a) have elaborated on this point with respect to further analyses of the Almklovdalen and Kalskaret garnet peridotites.

The C.I.P.W. norms of the analysed Kalskaret garnet peridotites (Table No. 16) and the Ugelvik peridotites (Table No. 19) also illustrate within each group certain anomalous variations in the Fe/Fe+Mg ratios of the normative ferromagnesian silicates when compared with the compositions of the bulk normative felspars.

Fig. 20 is a Ca₂⁺: Mg²⁺: Fe²⁺+Fe³⁺ plot of the bulk rock compositions of analysed Kalskaret peridotites (Table No. 15), Ugelvik peridotites (Table No. 18), and Kalskaret and Almklovdalen garnet peridotites [Lappin (1962), and Mercy and O'Hara (1965a)].

The composition points on this diagram define apparent linear correlations/
correlation between samples from the different localities, a feature first noted by Mercy and O'Hara (1965a). The linear trends of compositions from the Lien and Rodhaugen localities are well defined, that of the Kalskaret samples less well so. Nevertheless it is apparent that the samples from Kalskaret have a systematically higher \( \frac{Fe^{2+} + Fe^{3+}}{Mg^{2+}} \) ratio for any given \( Ca^{2+} \) content than the samples from Lien, and those in turn than the samples from Rodhaugen.

The linear composition trend of the Ugelvik peridotites is notably transverse to those defined by the samples from the other three localities, showing only a very slight increase in \( \frac{Fe^{2+} + Fe^{3+}}{Mg^{2+}} \) ratio with increasing \( Ca^{2+} \) content, a feature which can be attributed solely to the higher \( \frac{Fe^{2+} + Fe^{3+}}{Mg^{2+}} \) ratio of the garnet compared with its coexisting minerals (see Table Nos. 7 and 8). By comparison the linear composition trends of the Kalskaret, Lien and Rodhaugen peridotite compositions show markedly increased \( \frac{Fe^{2+} + Fe^{3+}}{Mg^{2+}} \) ratios with increasing \( Ca^{2+} \) contents (i.e. increasing modal proportions of GNT and CPX).

This distinction agrees with that noted above in the discussion of the norms, where it was noted that while in the Kalskaret and Almklovdalen peridotites the \( \frac{Fe+Mn+Ni}{Fe+Mn+Ni+Mg} \) ratio of the normative ferromagnesian silicates increases markedly with increasing modal contents of GNT and CPX, the ratio in the normative ferromagnesian silicates of the Ugelvik peridotites remains quite constant.

B. STATISTICAL ANALYSIS OF THE COMPOSITION VARIATIONS.

A covariance matrix analysis has been performed on the variation in bulk rock chemistry shown by the 13 closely related rock samples collected across the Kalskaret garnet peridotite lens (Table No. 15). For convenience MnO/
MnO and K$_2$O were eliminated from the analyses it being considered that any variation trends shown by these oxides would be expressed by the other oxide components present.

The variation about the mean of the oxides for all the analyses was calculated, then standardized to unit variance (one standard deviation about the mean) and the covariance matrix calculated.

Table No. 20 shows the covariance matrix together with the standard deviation and the standard deviation as a percentage of the mean, for all the elements. This table is useful in that it demonstrates any major correlations between the elements. For example NiO is demonstrated as showing a very strong positive correlation with MgO and a negative correlation with Al$_2$O$_3$, CaO and Na$_2$O, which in turn show a strong positive correlation amongst themselves. The mineralogical significance of this correlation obviously lies in the fact that the NiO in these rocks is concentrated in the olivines rather than in the coexisting garnets or clinopyroxenes.

There are, however, clearly too many variables here to be able to generalize about the problem in hand, an explanation of the origin of the compositional variation amongst these rock samples. The problem is that one is dealing with a number of different variates (element compositions) which cannot be isolated but must be considered together as they are interdependent. One therefore must economize in the number of variates and hence reduce the dimensions of the problem. This can be done by transferring to new uncorrelated variates which account for as much of the total variation as possible in descending order of magnitude.

The covariance matrix was therefore fed into an Atlas Computer to calculate the chief variates responsible for the total observed variation in/
### TABLE NO. 20

**COVARIANCE MATRIX**

<table>
<thead>
<tr>
<th></th>
<th>SiO$_2$</th>
<th>TiO$_2$</th>
<th>Al$_2$O$_3$</th>
<th>Cr$_2$O$_3$</th>
<th>Fe$_2$O$_3$</th>
<th>FeO</th>
<th>NiO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na$_2$O</th>
<th>H$_2$O$^+$</th>
</tr>
</thead>
<tbody>
<tr>
<td>SD</td>
<td>1.7716</td>
<td>0.2192</td>
<td>3.3517</td>
<td>0.1088</td>
<td>0.4652</td>
<td>2.5776</td>
<td>0.0977</td>
<td>10.0386</td>
<td>3.4036</td>
<td>0.3296</td>
<td>1.8217</td>
</tr>
<tr>
<td>SD/M</td>
<td>4</td>
<td>81</td>
<td>66</td>
<td>36</td>
<td>17</td>
<td>35</td>
<td>43</td>
<td>31</td>
<td>65</td>
<td>60</td>
<td>91</td>
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<td>SiO$_2$</td>
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<td>+0.7394</td>
<td>+0.8564</td>
<td>-0.4564</td>
<td>-0.1616</td>
<td>+0.6825</td>
<td>-0.8795</td>
<td>-0.8589</td>
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</tr>
<tr>
<td>TiO$_2$</td>
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<td>+1.0000</td>
<td>+0.8795</td>
<td>-0.7576</td>
<td>+0.3820</td>
<td>+0.7558</td>
<td>-0.8896</td>
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<td>+0.8795</td>
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<td>+0.6094</td>
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<td>-0.9765</td>
<td>+0.9541</td>
<td>+0.9354</td>
<td>-0.3151</td>
</tr>
<tr>
<td>Cr$_2$O$_3$</td>
<td>-0.4564</td>
<td>-0.7576</td>
<td>-0.6416</td>
<td>+1.0002</td>
<td>-0.5205</td>
<td>-0.4540</td>
<td>+0.6774</td>
<td>+0.6297</td>
<td>-0.5692</td>
<td>-0.4867</td>
<td>+0.1980</td>
</tr>
<tr>
<td>Fe$_2$O$_3$</td>
<td>-0.1616</td>
<td>+0.3820</td>
<td>+0.2386</td>
<td>-0.5205</td>
<td>+0.9994</td>
<td>+0.0065</td>
<td>+0.2133</td>
<td>+0.2472</td>
<td>+0.1937</td>
<td>+0.1148</td>
<td>+0.3768</td>
</tr>
<tr>
<td>FeO</td>
<td>+0.6825</td>
<td>+0.7558</td>
<td>+0.6094</td>
<td>-0.4540</td>
<td>+0.0065</td>
<td>+1.0000</td>
<td>-0.6729</td>
<td>-0.7172</td>
<td>+0.6675</td>
<td>+0.7029</td>
<td>-0.7560</td>
</tr>
<tr>
<td>NiO</td>
<td>-0.8795</td>
<td>-0.8896</td>
<td>-0.9872</td>
<td>+0.6774</td>
<td>-0.2133</td>
<td>-0.6729</td>
<td>+0.9997</td>
<td>+0.9861</td>
<td>-0.9530</td>
<td>-0.9443</td>
<td>+0.3614</td>
</tr>
<tr>
<td>MgO</td>
<td>-0.8589</td>
<td>-0.9178</td>
<td>-0.9765</td>
<td>+0.6297</td>
<td>-0.2472</td>
<td>-0.7172</td>
<td>+0.9861</td>
<td>+1.0000</td>
<td>-0.9789</td>
<td>-0.9723</td>
<td>+0.3927</td>
</tr>
<tr>
<td>CaO</td>
<td>+0.8517</td>
<td>+0.9119</td>
<td>+0.9541</td>
<td>-0.5692</td>
<td>+0.1937</td>
<td>+0.6675</td>
<td>-0.9530</td>
<td>-0.9789</td>
<td>+1.0000</td>
<td>+0.9830</td>
<td>-0.4157</td>
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<td>Na$_2$O</td>
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<td>+0.8604</td>
<td>+0.9354</td>
<td>-0.4867</td>
<td>+0.7029</td>
<td>-0.9443</td>
<td>-0.9723</td>
<td>+0.9830</td>
<td>+1.0000</td>
<td>+0.4366</td>
<td>-0.1000</td>
</tr>
<tr>
<td>H$_2$O$^+$</td>
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<td>-0.4893</td>
<td>-0.3151</td>
<td>+0.1981</td>
<td>+0.3768</td>
<td>-0.7560</td>
<td>+0.3614</td>
<td>+0.3827</td>
<td>-0.4157</td>
<td>-0.4366</td>
<td>+1.0000</td>
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### TABLE NO. 21

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<th>( L_1 )</th>
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<th>( L_3 )</th>
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<th>( L_5 )</th>
<th>( L_6 )</th>
<th>( L_7 )</th>
<th>( L_8 )</th>
<th>( L_9 )</th>
<th>( L_{10} )</th>
<th>( L_{11} )</th>
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<tbody>
<tr>
<td>( \lambda )</td>
<td>0.7053</td>
<td>0.1593</td>
<td>0.0738</td>
<td>0.0335</td>
<td>0.0142</td>
<td>0.0083</td>
<td>0.0034</td>
<td>0.0015</td>
<td>0.0003</td>
<td>0.0002</td>
<td>0.0000</td>
</tr>
<tr>
<td>( \text{SiO}_2 )</td>
<td>-0.32012</td>
<td>+0.24298</td>
<td>+0.19195</td>
<td>+0.27060</td>
<td>-0.15597</td>
<td>+0.62109</td>
<td>+0.22037</td>
<td>-0.46935</td>
<td>+0.04585</td>
<td>+0.18191</td>
<td>-0.11955</td>
</tr>
<tr>
<td>( \text{TiO}_2 )</td>
<td>-0.34058</td>
<td>-0.11437</td>
<td>-0.21539</td>
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<td>-0.25124</td>
<td>-0.42222</td>
<td>-0.28015</td>
<td>-0.65516</td>
<td>+0.10161</td>
<td>+0.23804</td>
<td>-0.00685</td>
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<td>( \text{Al}_2\text{O}_3 )</td>
<td>-0.34326</td>
<td>-0.09614</td>
<td>+0.24310</td>
<td>+0.08728</td>
<td>+0.01562</td>
<td>+0.10608</td>
<td>-0.64246</td>
<td>+0.29879</td>
<td>+0.49280</td>
<td>-0.22285</td>
<td>+0.01309</td>
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<td>( \text{Cr}_2\text{O}_3 )</td>
<td>+0.24418</td>
<td>+0.33971</td>
<td>+0.41713</td>
<td>-0.70899</td>
<td>-0.15853</td>
<td>+0.14773</td>
<td>-0.25856</td>
<td>-0.13090</td>
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<tr>
<td>( \text{Fe}_2\text{O}_3 )</td>
<td>-0.07208</td>
<td>-0.67663</td>
<td>-0.28813</td>
<td>-0.39861</td>
<td>-0.28048</td>
<td>+0.44783</td>
<td>+0.06874</td>
<td>+0.08637</td>
<td>-0.02556</td>
<td>-0.01596</td>
<td>-0.05250</td>
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<tr>
<td>( \text{FeO} )</td>
<td>-0.28100</td>
<td>+0.21134</td>
<td>-0.44511</td>
<td>-0.38547</td>
<td>+0.66206</td>
<td>+0.09232</td>
<td>+0.01749</td>
<td>-0.06064</td>
<td>+0.08849</td>
<td>-0.16378</td>
<td>-0.17492</td>
</tr>
<tr>
<td>( \text{NiO} )</td>
<td>+0.34975</td>
<td>+0.06804</td>
<td>-0.17651</td>
<td>-0.12574</td>
<td>-0.19680</td>
<td>-0.13951</td>
<td>+0.19021</td>
<td>-0.18142</td>
<td>+0.47996</td>
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<td>+0.32729</td>
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<tr>
<td>( \text{MgO} )</td>
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<td>+0.06313</td>
<td>-0.15048</td>
<td>+0.10947</td>
<td>-0.12920</td>
<td>+0.00493</td>
<td>+0.00795</td>
<td>+0.01910</td>
<td>+0.41521</td>
<td>+0.17580</td>
<td>+0.78507</td>
</tr>
<tr>
<td>( \text{CaO} )</td>
<td>-0.34699</td>
<td>-0.02507</td>
<td>+0.19805</td>
<td>-0.14433</td>
<td>-0.23190</td>
<td>-0.38466</td>
<td>+0.20377</td>
<td>+0.16452</td>
<td>+0.28460</td>
<td>-0.51896</td>
<td>-0.44107</td>
</tr>
<tr>
<td>( \text{Na}_2\text{O} )</td>
<td>-0.34294</td>
<td>+0.04243</td>
<td>+0.24958</td>
<td>-0.22790</td>
<td>+0.03813</td>
<td>-0.12225</td>
<td>+0.54944</td>
<td>+0.19659</td>
<td>+0.48629</td>
<td>+0.39468</td>
<td>+0.13103</td>
</tr>
<tr>
<td>( \text{H}_2\text{O}^+ )</td>
<td>+0.18705</td>
<td>-0.52546</td>
<td>+0.48721</td>
<td>-0.00769</td>
<td>+0.50944</td>
<td>-0.08785</td>
<td>+0.09306</td>
<td>-0.36597</td>
<td>+0.08982</td>
<td>-0.09559</td>
<td>-0.12078</td>
</tr>
</tbody>
</table>
in the chemistry of these rock samples. The results are shown in Table No. 21, in which these new variates are listed in descending order of magnitude. This shows that 86.46% of the total variation can be expressed in terms of two new variates $L_1$ and $L_2$, while the contribution of the nine other variates to the total variation is comparatively small.

It is therefore apparent that all the compositions fall in a rather narrow ribbon in multidimensional space, the co-ordinates of which are the oxide components.

The new variates $L_1$ and $L_2$, with their direction cosines along the various oxide co-ordinates expressed in the first two vertical columns of Table No. 21, may be calculated back into the unit variance matrix of all the oxides in the various rock samples to show which rock samples represent the most extreme members of the variation trends in either case. The results of this calculation have been plotted in Fig. 23.

This diagram shows that along the $L_1$ variate, which accounts for a considerable proportion (70.53%) of the total variation, rocks which are essentially dunites with a little accessory chromite (T268 and T156) plot at one extreme while the rock (T153) which represents the closest approach to a pure eclogite (GNT and CPX only) plots at the other extreme. The statistical analysis indicates that there is a basic relationship between all these rock types with considerable gradations in composition between these extreme composition rock types, although there is something of a gap apparent in the intermediate composition range along the $L_1$ co-ordinate of Fig. 23, corresponding to a scarcity of garnet peridotites from Kalskaret with only low modal proportions of garnet and pyroxenes.

It is quite clear from the distribution of the rock types along the/
the L₂ co-ordinate of Fig. 23 that this second principal variate, responsible for some 15.93% of the total observed composition variation, is connected with the partial secondary alteration, serpentinization and amphibolitization, of some of these rock types, with introduction of H₂O⁺ and oxidation of FeO to Fe₂O₃⁺.

No geological significance of the L₃ variate is immediately apparent. However, this variate together with the other eight minor variates only constitute a relatively small proportion of the total variation.

A discussion of the various possible geological interpretations of the major L₁ variate will be delayed until the next chapter, as such interpretations are key in discussions on the petrogenesis of the Kalskaret garnet peridotites. Let it suffice to point out here that the magnitude of this variate indicates a strong linear composition trend amongst the analysed Kalskaret peridotite samples thereby endorsing the linear composition correlation revealed in the CHF plot (Fig. 20) and demonstrating that this linear correlation extends to all the components. Furthermore the logical interpretation of a linear trend is the random mixing of two end members. The mixing of three or more phases would not be expected to produce such compositional linearity.

Subsequently the Kalskaret peridotite composition data were also treated by a totally computerized multivariate statistical technique, namely Q-mode factor analysis as outlined by Mason and Imbrie (1964). The data were computed by Imbrie and Darland on an IBM computer at Columbia University, New York.

The object of this mathematical treatment was to determine the minimum number of factors required to describe the total composition variations of all the observed samples, and further to determine the most compositionally extreme/
extreme end member samples along each factor axis, hence the significant end members in the spectrum of diversity under study. An excellent account of this treatment is given by Imbrie and Van Andel (1964). Only a brief account of the various computer operations involved is attempted here.

The compositional data computed were the chemical compositions of 16 bulk rock samples from the Kalskaret garnet peridotite locality, the 13 listed in Table No. 15 together with the compositions of N21, N23, and N26 - Mercy and O'Hara (1965a).

Firstly theoretical orthogonal vectors were calculated these representing the principal axes of the multidimensional swarm of composition points. These are the eigenvectors, the corresponding eigenvalues representing the proportions of the total information explained by each vector. Mathematically the problem is to calculate in order the positive eigenvalues of the \cos \theta matrix, hence noting how many axes are required to describe most of the composition variation.

The positive eigenvalues are given in Table No. 28 and indicate that the first two eigenvalues alone account for 99.8% of the total variation pointing to only two factors being responsible for virtually all the variation.

The positions of the sample vectors in relation to these two factor axes are determined by projecting the vectors onto the axes, the projections being called factor loadings the size and the number indicating the extent to which each factor axis controls the position of each sample vector. The sum of the squared factor loadings for a specific sample vector is referred to as its communality and reflects the degree to which that sample vector has been explained by the set of factor axes. According to the technique/
## Table No. 28

### Table of Positive Eigenvalues

<table>
<thead>
<tr>
<th>Eigenvalue</th>
<th>% Communality Over All 16 Factors (Compositions)</th>
<th>Cumulative % Communality Over All 16 Factors</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>15.446</td>
<td>96.5</td>
</tr>
<tr>
<td>2.</td>
<td>0.520</td>
<td>3.2</td>
</tr>
<tr>
<td>3.</td>
<td>0.024</td>
<td>0.2</td>
</tr>
<tr>
<td>4.</td>
<td>0.006</td>
<td>0.0</td>
</tr>
<tr>
<td>5.</td>
<td>0.003</td>
<td>0.0</td>
</tr>
<tr>
<td>6.</td>
<td>0.002</td>
<td>0.0</td>
</tr>
<tr>
<td>7.</td>
<td>0.000</td>
<td>0.0</td>
</tr>
</tbody>
</table>
technique used here a communality of 1.0 indicates a perfect explanation.

As, however, the eigenvectors located by this procedure are rarely in the most meaningful positions and are hence generally difficult to interpret, owing to the fact that the theoretical end-members derived are only sophisticated average samples representing the average orthogonal directions of the composition variation, they have been rotated according to the varimax technique to positions close to the actual extremes of compositional variation. The theoretical reference vectors in this new orthogonal scheme were selected to fit the data best in a least-squares sense. The resulting rotated factor matrix is given as Table No. 29. This gives the communality (the degree of explanation) of all the 16 samples in terms of the two rotated factor axes, and in terms of each individual factor axis, together with the percentage of each factor required to account for each sample.

It is apparent that with only two factors involved the communalities of all the samples are very high, never less than 0.992 and for only two samples less than 0.997. This indicates that the composition variations involved in this set of data can be almost entirely explained in terms of two factors (end members). From Table No. 29 it is apparent that the serpentinized dunite composition 834/T268 and the most eclogitic composition 838/T153 are compositionally the most extreme of the 16 samples.

This is in close agreement with the interpretation of the previous statistical treatment (cf. Fig. No.23). However, while the previous treatment distinguished between one major variation trend with dunite and 'eclogite' as its extremes and a much more minor one due predominately to partial serpentinization of some of the samples, this factor analysis treatment incorporates the two together giving serpentinized dunite and the most/
### TABLE NO. 29.

**ROTATED FACTOR MATRIX**

<table>
<thead>
<tr>
<th>SAMPLE NO.</th>
<th>COMMUNALITY</th>
<th>FACTOR NUMBER</th>
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</thead>
<tbody>
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<td></td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>833/T200</td>
<td>1.000</td>
<td>8.293</td>
</tr>
<tr>
<td>834/T268</td>
<td>0.997</td>
<td>0.536</td>
</tr>
<tr>
<td>835/T150</td>
<td>0.996</td>
<td>0.494</td>
</tr>
<tr>
<td>836/T151</td>
<td>0.999</td>
<td>0.766</td>
</tr>
<tr>
<td>837/T152</td>
<td>0.997</td>
<td>0.793</td>
</tr>
<tr>
<td>838/T153</td>
<td>0.998</td>
<td>0.801</td>
</tr>
<tr>
<td>839/T154</td>
<td>0.998</td>
<td>0.901</td>
</tr>
<tr>
<td>840/T97</td>
<td>0.999</td>
<td>0.766</td>
</tr>
<tr>
<td>841/T96</td>
<td>0.997</td>
<td>0.753</td>
</tr>
<tr>
<td>842/T155</td>
<td>0.992</td>
<td>0.804</td>
</tr>
<tr>
<td>843/T156</td>
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</tr>
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<td>844/T162</td>
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<td>0.578</td>
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<td>845/T100</td>
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<td>0.628</td>
</tr>
<tr>
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<td>0.588</td>
</tr>
<tr>
<td>N23</td>
<td>0.998</td>
<td>0.846</td>
</tr>
<tr>
<td>N26</td>
<td>1.000</td>
<td>0.745</td>
</tr>
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</table>
most 'eclogitic' composition as the end members of the linear variation trend accounting for almost all the composition variation involved. This is fair enough as obviously serpentinization is related to the more dunitic rather than eclogitic compositions. However, the partial amphibolitization of a few of the more eclogitic compositions adds a minor complicating factor obscured by the factor analysis treatment. This probably accounts for the two most amphibolitized samples 842/T155 and 835/T150 having lowest communalities (Table No. 29).

However, it seems clear that only two end members are sufficient to account for the composition variations involved any minor deviations being accountable by minor complications added through partial serpentinization and amphibolitization of certain of the samples. To be certain about this the complete factor analysis treatment was in fact run through assuming that four factors were involved but the last two proved entirely negligible.

The procedure was also run through with the 16 samples above plus the dunite composition 843/T156 recalculated anhydrous and the extrapolated 'eclogitic' composition III - see Table No. 22 and Fig. 27. In this case the procedure was carried somewhat further as the computer was programmed to search out and identify the two most extreme vectors in the varimax configuration and an oblique projection matrix derived (Table No. 30), in which the proportional composition of each sample in terms of these two most compositionally extreme vectors (now representing actual samples) are listed. The serpentinized dunite sample 834/T268 and the extrapolated 'eclogitic' composition (18) were picked out as the compositional extremes. It was found that the recalculated anhydrous composition 843/T156 had a slightly lower communality on factor axis 1, than the actual hydrous composition which represents a mildly serpentinized dunite and a more notably lower communality/
**TABLE NO. 30.**

**OBLIQUE PROJECTION MATRIX**

<table>
<thead>
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<th>SAMPLE NO.</th>
<th>SAMPLE INDEX</th>
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<td>833/T200</td>
<td>0.949</td>
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<tr>
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<td>835/T150</td>
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</tr>
<tr>
<td>836/T151</td>
<td>0.424</td>
</tr>
<tr>
<td>837/T152</td>
<td>0.475</td>
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<tr>
<td>838/T153</td>
<td>0.193</td>
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<tr>
<td>839/T154</td>
<td>0.556</td>
</tr>
<tr>
<td>840/T97</td>
<td>0.403</td>
</tr>
<tr>
<td>841/T62</td>
<td>0.582</td>
</tr>
<tr>
<td>842/T155</td>
<td>0.458</td>
</tr>
<tr>
<td>843/T156</td>
<td>0.989</td>
</tr>
<tr>
<td>844/T162</td>
<td>0.890</td>
</tr>
<tr>
<td>845/T100</td>
<td>0.810</td>
</tr>
<tr>
<td>N21</td>
<td>0.875</td>
</tr>
<tr>
<td>N23</td>
<td>0.363</td>
</tr>
<tr>
<td>N26</td>
<td>0.600</td>
</tr>
<tr>
<td>17</td>
<td>0.987</td>
</tr>
<tr>
<td>18</td>
<td>0.000</td>
</tr>
</tbody>
</table>

|   | 834/T268 | 18 |

Sample No. 17 is dunite composition 843/T156 recalculated anhydrous.

Sample No. 18 is extrapolated 'eclogitic' end member composition. Table No. 22 III - see also Fig. 27.
communality than 834/T268 which represents a more highly serpentinized
dunite. Thus again serpentinized dunite and the most 'eclogitic' composition
are indicated as the end members.

The factor analysis treatment described above therefore gives
strong support to the conclusion that a very strong linear composition
trend, which can be interpreted in terms of the mixing of two end members
only, exists among the rock samples collected from the Kalskaret garnet
peridotite locality. Whether or not the most extreme rock samples picked
out by the mathematical treatment represent the actual end members of the
trend or whether such end members have more extreme compositions along the
trend is another matter, and one which is left for discussion in the next
chapter.

C. MODELS OF BULK ROCK COMPOSITION VARIATION.

In this section the hypothesis that the Norwegian garnet
peridotites have resulted from the eclogite facies metamorphism in deep
levels of the crust of peridotitic rock masses which originally crystallized
to another mineral facies assemblage shall be considered. Such an origin
has been postulated by O'Hara and Mercy (1963), and also inferred by Boyd
and MacGregor (1964) and MacGregor (1965) on the basis of the experimental
evidence reviewed in the last chapter.

The varying mineral assemblages which can result from the
crystallization of peridotitic rocks, of broadly comparable bulk chemical
composition under differing physical conditions (pressure, temperature and
water pressure) have recently been stressed and outlined by several workers,
O'Hara and Mercy (1963), Green and Ringwood (1963), Ringwood, MacGregor and
Boyd (1964), Boyd and MacGregor (1964), Kushiro and Yoder (1965) and
MacGregor (1965).
It is to be expected, however, that minor yet significant variations in bulk chemistry may result from the differing affinities for the various minor (Ti, Cr, Mn, Ni etc.) and trace (Rb, Sr, Co, V, Ba, etc.) elements of the various minerals crystallizing under the varying physical conditions, on the assumption that there was an extra phase present (namely liquid) with which to exchange and remove the 'residual' elements. Also, the ease of substitution of certain cations in particular minerals (for example Al$^{3+}$, Cr$^{3+}$ and Fe$^{3+}$ substitution in pyroxenes) and the degree of solid solution between the end members in the different mineral series varies with the physical conditions under which crystallization occurs. Such differences will result in minor compositional differences which may indicate from which original mineral assemblage particular metamorphosed peridotites may have crystallized.

In the case under consideration here, the bulk rock chemistry of the Norwegian garnet peridotites is compared directly with that of all other known peridotite mineral assemblages from which it is remotely conceivable that the former may have originated.

It was considered that the method of comparison best suited to this particular problem was to outline the range of compositions encompassed by the mineral compositions of each particular assemblage in a series of two dimensional variation diagrams. The range of bulk compositions shown by the Norwegian garnet peridotites can then be plotted in these diagrams to see if they comply with the range of bulk compositions to be expected from the sorting of minerals from any other particular mineral assemblage.

This method of comparison can first be discussed in terms of a hypothetical case (Fig. 24a). Plotted in this diagram are the compositions of three minerals A, B and C in terms of the weight percentages of two components/
components X and Y. Mineral A has a ratio X:Y of 4:1, the point A representing the absolute amounts of X and Y in this mineral, namely 20 wt.% of X and 5 wt.% of Y. Similarly mineral B contains 25 wt.% of X and Y and has a X:Y ratio of 1:1, and mineral C contains 5 wt.% X and 10 wt.% of Y and has a X:Y ratio of 1:2.

The lines OA, OB and OC are only significant in that they represent the locus of points having the X:Y ratio of each particular mineral up to the actual concentration of X and Y in that mineral, but they do not have any significance in forming boundary conditions to the range of bulk compositions encompassed by all proportions of these three minerals, which is represented rather by the triangle ABC. Any composition outside ABC could not crystallize to these three minerals only.

In the case of only three minerals as here the modal proportions of the minerals to represent any composition point within the triangle ABC can be calculated. For example, composition K would comprise PC of A, MA of B, and NA of C, namely 50% A, 25% B, and 25% C.

However, the majority of the peridotite assemblages contain at least four mineral phases. The composition of the fourth mineral may lie outside the composition triangle ABC (as in Fig. 24b) or inside (as in Fig. 24c), or may even coincide with one of the sides of the triangle ABC.

In the first case (Fig. 24b) the range of bulk compositions which may be represented by this mineral assemblage has now been extended into a quadrilateral ABCD. However, it is now no longer possible to specify the modal proportions of the minerals for any particular composition point. For example, the composition point K can be represented by a mixture of all four minerals, the proportions of which cannot now be calculated, or by a mixture of A, B and C, or A, B and D in both cases the relative proportions of the minerals/
minerals of which can be calculated.

In the case where the composition point of the fourth mineral lies either on one side or within the composition triangle ABC (Fig. 24c) the range of bulk compositions has not been extended but again problems arise in trying to give a quantitative estimate of the mineral assemblage of any particular composition point within ABC. For example, the composition K can now be represented by mixtures of A, B and C; A, B and D; or all four minerals A, B, C and D.

With a fifth mineral the number of possible mineral mixtures which can represent a particular bulk composition is even greater.

However, despite the fact that when dealing with more than three minerals one cannot discuss any particular composition in terms of a unique mixture of the minerals, the compositions of the minerals still outline the range of bulk compositions which can result from the sorting of the minerals of that particular mineral assemblage.

The differing peridotite mineral assemblages have therefore been plotted on a series of such diagrams and the range of bulk compositions demonstrated by the Norwegian garnet peridotite localities at Kalskaret, Almklovdaelen and Ugelsvik compared with the range expected from the respective mineral assemblages. The diagrams (Fig. Nos. 37-48) are presented and discussed in the Appendix II.

The peridotite mineral assemblages used for comparison were:

1. Lizard primary high temperature peridotite 90681 (Green (1964)).
   Assemblage: O1 + CPX + OPX + Spinel

2. Lizard recrystallized low temperature peridotite Green (1964) 90686.
   Assemblage: O1 + CPX + OPX + Plagioclase + Chromite.

3. Peridotite nodule from basalt No. 4, Ross et al. (1954).
   Assemblage: O1 + CPX + OPX + Spinel.
4. Stillwater layered basic intrusion. EB.43 Hess (1960)
   Assemblage Plagioclase + CPX + OPX

   Assemblage Ol + CPX + Spinel.

6. Hypersthene eclogite nodule in tuff - Salt Lake Crater, Oahu 66118
   Yoder and Tilley (1962)
   Assemblage GNT + OPX + CPX.

   Assemblage GNT + CPX + OPX + Ol.

8. Spinel herzolite - Pyrenees 104,58 O'Hara and Mercy unpublished data.
   Assemblage Ol + CPX + OPX + Spinel.

    Assemblages Ol + CPX + OPX + GNT. N23 - Kalskaret
    T96 - Kalskaret
    U8 - Ugelvik

The most illustrative variation diagrams used for this purpose were:
   a. CaO wt.% : Al₂O₃ wt.%
   b. Fe²⁺ + Fe³⁺ + Mn²⁺ : Mg²⁺
   c. Na₂O wt.% : Al₂O₃ wt.%
   d. Cr₂O₃ wt.% : Al₂O₃ wt.%
   e. Ni²⁺ : Mg²⁺
   f. TiO₂ wt.% : SiO₂ wt.%

The following conclusions have been drawn from a study of these diagrams.

1. It is impossible to explain the mineralogy and range of composition of the Norwegian garnet peridotites in terms of the isochemical eclogite facies metamorphism of other possible peridotite mineral assemblages, in particular/
particular in terms of the sorting of such mineral assemblages as Ol + CPX + OPX + Plagioclase + chromite (pyroxene hornfels mineral facies) or Ol + CPX + OPX + Spinel (granulite mineral facies), the most obvious possibilities. Invocation of cryptic layering as a mechanism producing sufficiently large compositional variations in the original mineral assemblages appears unsatisfactory considering the small thicknesses over which these compositional variations occur (see Chapt.III, section 1A).

2. The composition range of the Norwegian garnet peridotites is best covered by the eclogite facies peridotite mineral assemblage (Ol + OPX + CPX + GNT) itself. However, in detail the chemical variations among these garnet peridotites as a group are such that they extend beyond the range of compositions to be expected from the sorting of any analysed garnet peridotite assemblage (either from the Norwegian gneisses or from kimberlite pipes) even when allowance is made for possible cryptic layering compositional effects.

Several Kalskaret bulk compositions are markedly incompatible with the sorting of the Kalskaret mineral assemblages (T96 and N23) for most composition indices.

In most respects the Almklovdalen bulk compositions, with the notable exception of the Rodhaugen eclogite N75, are compatible with the sorting of the Lien mineral assemblage N69. However, only a limited range of rock types has been analysed from here, the analyses of garnet poor peridotites being lacking.

The Ugelvik bulk compositions are quite compatible with the sorting of the Ugelvik mineral assemblage U9, unlike several of the Kalskaret and Almklovdalen compositions. There is, however, a minor anomaly in TiO₂ content, this assemblage being notably TiO₂ rich with respect to the other analysed Ugelvik assemblages. Furthermore the Ugelvik compositions are highly/
highly incompatible with the Kalskaret (T96 and N23) and Almklovdalen (N69) mineral assemblages, their being a particularly important anomaly in $\text{Fe}^{2+}\text{Fe}^{3+}\text{Mn}^{2+}/\text{Mg}^{2+}$ ratio. Also they show a much closer correlation to the range of bulk compositions expected from the sorting of the kimberlitic garnet peridotite mineral assemblage A3 than do most of the Kalskaret and Almklovdalen peridotite compositions.

The possibility that, under the extreme conditions of metamorphism to be expected in the eclogite facies, metamorphism involved the setting up of composition gradients resulting from the varying chemical potentials or activities of the different elements, must be considered. Under these conditions chemical diffusion processes may be important but a quantitative discussion of the compositional gradients involved is at this stage impossible, particularly as one can only guess as to the compositions of the original rock types which perhaps reacted together during metamorphism to produce the present composition distributions.

If metamorphism in fact involved reaction between rock types of high variance, the composition diffusion paths are unlikely to have been linear as that would imply the unlikely restriction of equality among the diffusion rates of all the components (Clark and Rhines 1959). As a result such a strong linear composition trend as demonstrated in the last section for the Kalskaret garnet peridotites would not be expected.

Furthermore, the sorting of three or more mineral phases would not be expected to produce such compositional linearity, suggesting the unlikeliness of the eclogite facies metamorphism of a sorted igneous mineral assemblage as the mechanism producing the present mineral assemblages of these garnet peridotites.

The effects of interstitial liquid as a mechanism producing added chemical/
chemical variation in the igneous mineral assemblages which were considered here perhaps to have been metamorphosed to the eclogite facies assemblages may be significant and should therefore have been considered. However, the invocation of interstitial liquid effects adds yet another phase to the system and renders the existence of such a strong linear composition trend even more unlikely.
CHAPTER V. THE PETROGENESIS OF THE PERIDOTITES

A. THE PETROGENESIS OF THE KALSKARET GARNET PERIDOTITES.

In the last chapter strong linear trends in the compositions of the garnet peridotites were demonstrated, with particular reference to the Kalskaret garnet peridotite occurrence. It is relevant here to discuss the likely origin of this compositional linearity.

To date only the Kalskaret garnet peridotites, for which most bulk rock chemical data is now available, have been dealt with statistically. However, as the garnet peridotite compositions from both the Lien and Rodhaugen localities form roughly parallel trends on the $\text{Ca}^{2+} : \text{Mg}^{2+} : \text{Fe}^{2+} : \text{Fe}^{3+}$ plot (Fig. 20) to that of the Kalskaret compositions, it is considered that they possess a similar compositional linear trend of the same origin, whatever that origin might have been. Furthermore if all the peridotite compositions from each of the three localities are plotted on variation diagrams with their content of the various oxides against MgO content, a similar and generally pronounced linear trend is shown by each oxide for each group (see Figs. 25 and 26).

The linear trend of the Ugelvik garnet peridotites on Fig. 20 is notably transverse to the other trends and is considered to be of different origin. This is further substantiated by differences in the trends of various oxides (notably $\text{Cr}_2\text{O}_3$ and FeO) when the Ugelvik compositions are plotted on a variation diagram against MgO content (Fig. 26). The origin of the Ugelvik garnet peridotites is discussed in a subsequent section.

It is considered that the development of such a strong linear composition trend as demonstrated for the Kalskaret garnet peridotite compositions is only possible with the random mixing of two end member phases.
FIG. 25 KALSKARET GARNET PERIDOTITES COMPOSITION VARIATION DIAGRAM
phases (minerals, rock types, or one mineral or rock type together with a trapped liquid phase) a conclusion which has been substantiated by the results of the factor analysis. The repetition of such a linear trend for three independent localities would be an exceptional coincidence with the mixing of any more than two phases.

It has been demonstrated in the case of the Kalskaret composition trend that the extreme rock types on the trend are essentially dunites (T268 and T156) at one end and the nearest approach to a bimineralic eclogite (T153) at the other. The question then arises as to whether or not these rock types represent the real end member phases of the composition trend.

In Fig. 27 the contents of the various oxides are plotted against MgO content for the extreme rock types T156 and T153, together with two intermediate composition rock types T100 and T151, and the calculated mean composition of the thirteen analysed Kalskaret garnet peridotites (Table No.15). The linear trends of the oxide contents in these five compositions are drawn and in most cases are very well defined.

It is notable how the Al$_2$O$_3$, CaO, Na$_2$O and TiO$_2$ contents all tend to zero in the composition of the dunite T156 suggesting that this composition in fact closely represents the most extreme rock type end member feasible in that direction.

The other end member is, however, less well defined. Although the NiO and Cr$_2$O$_3$ contents are low in the eclogitic rock type T153 they have not quite fallen to zero. The true end member rock type in that direction may therefore have a rather more extreme composition than T153. Extrapolating the oxide linear trends the NiO content falls to zero at a MgO/
Mgo content of 9.80 wt.%. Four hypothetical rock compositions slightly more extreme than T153 have been calculated, their compositions and C.I.P.W. norms being given in Table No. 22.

The features of these increasingly extreme compositions are a notable decrease in the normative olivine contents (the most extreme composition calculated in fact containing normative quartz), increase in the Fe+Ni+Mn/Fe+Ni+Mn+Mg ratio in the normative ferromagnesian silicates, and slight increases in normative pyroxene and plagioclase contents. This composition trend appears to approach the major element composition of some tholeiitic basalts. However, the low Na2O contents and likely negligible K2O and P2O5 contents contrast with all extrusive magma types.

There are only a limited number of conceivable mechanisms apparent to the author which may have produced the combination of the observed linear composition trend and the eclogite facies mineralogy of the Kalskaret garnet peridotites. These mechanisms are as follows:

1. Mechanical mixing of tholeiitic basalt and dunite, subsequently metamorphosed to an eclogite facies assemblage.

In fact two possibilities exist within this hypothesis depending on the depth at which the mixing is assumed to have occurred, whether at high levels in the crust or at greater depths within the crust or perhaps even the top of the mantle. The first possibility entails the transportation of the mass down to greater depths to produce the eclogite facies mineralogy then subsequently back up again to its present environment, a rather complex procedure. Furthermore the Na2O, K2O and P2O5 contents of the likely basaltic end member of the composition trend (see Table No. 22) are too low in comparison with likely high level basaltic magma types. The first/
<table>
<thead>
<tr>
<th>TABLE NO. 22</th>
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**POSTULATED 'ECLOGITIC' END MEMBERS OF KALSKARET COMPOSITION TREND**

<table>
<thead>
<tr>
<th>ANALYSES WT.%</th>
<th>C.I.P.W. NORMS</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>TiO₂</strong></td>
<td><strong>I</strong></td>
</tr>
<tr>
<td>10.83</td>
<td>4.00</td>
</tr>
<tr>
<td>18.87</td>
<td></td>
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<td>0.47</td>
<td>0.0</td>
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<tr>
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<td>3.30</td>
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<td>8.65</td>
<td>9.06</td>
</tr>
<tr>
<td>0.22</td>
<td>0.23</td>
</tr>
<tr>
<td>0.06</td>
<td>0.04</td>
</tr>
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<td>14.26</td>
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</tr>
<tr>
<td>0.004</td>
<td>(-0.004)</td>
</tr>
<tr>
<td>TOTAL</td>
<td>99.57</td>
</tr>
</tbody>
</table>

*Fe+Ni+Mn*  
*Fe+Ni+Mn+Mg*  

In Normative Ferromagnesian Silicates
first possibility would therefore seem unlikely, while both possibilities raise a problem regarding the intimate mixing required to produce the present interlayering of the rock types. Mechanical mixing during tectonic transport into their present environment may supply a feasible explanation of the layering noted in the case of the Kalskaret garnet peridotites which occur in a very restricted lens in which garnet and clinopyroxene rich rock types predominate, but seems unlikely in the case of the Lien locality at Almklovdalen where garnetiferous peridotites occur intimately interlayered with garnet free peridotites over about 200 metres and garnet peridotites with only low modal proportions of garnet and clinopyroxene are prevalent.

2. Mechanical mixing of eclogite and dunite.

It is again considered unlikely that simply the mechanical mixing of eclogite and dunite could produce the compositional variations resulting in the present distribution of the rock types, especially the preponderance of rock types of intermediate composition as occur at the Lien locality.

3. Igneous crystallization with the sorting of olivine and aluminous clinopyroxene, the sequence being later metamorphosed to an eclogite facies assemblage.

Close inspection of the composition variations among the Kalskaret garnet peridotites renders it unlikely that they could have been derived solely from the sorting of olivine and aluminous clinopyroxene for the following reasons:

a. An additional chrome rich phase is necessary to give the high Cr₂O₃ content of the dunitic end members of the composition trend.

b./
b. The $\text{Al}_2\text{O}_3/\text{CaO}$ ratio in the most eclogitic compositions is too high to have been produced only by the sorting of stable aluminous clinopyroxene of any known composition, even those realised experimentally.

c. It is also unlikely that there would be such a striking increase in FeO/MgO ratio towards the eclogitic compositions solely as a result of increase in clinopyroxene content with respect to coexisting olivine. The equilibrium distribution expected is that olivine would have a higher FeO/MgO ratio than its coexisting clinopyroxene (see Chapt.III, Section 3). Even if disequilibrium conditions prevailed and the relationship was reversed, the reversal is unlikely to have been so extreme as to result in such a notable increase in FeO/MgO ratio of the bulk rock compositions.

4. Igneous crystallization of olivine + orthopyroxene cumulates with the residual liquid imperfectly filter pressed out and crystallized in pockets within the peridotite.

Variations on this hypothesis result depending on the depth at which, and the mineral facies within which, the crystallization is thought to have occurred. If crystallization is assumed to have occurred at high crustal levels then the mass must have had a complex tectonic history, namely transport to considerably greater depths in order to produce the eclogite facies mineralogy followed by transport back up again into its present environment of disequilibrium with the enclosing gneiss complex. Furthermore a low temperature metamorphic origin of the eclogite facies assemblage does not explain the relict exsolution textures which are considered indicative of an original higher temperature mineral assemblage, unless the original crystallization is assumed to have involved the crystallization of aluminous pyroxenes (granulite or eclogite facies assemblages). In that case original crystallization must probably be relegated to the upper levels of/
of the mantle and this hypothesis becomes closely akin to the mantle partial melting hypothesis (5) outlined below, except that in this case the composition series would not be expected to include the original upper mantle source composition.

The hypothesis that the composition variations and the present mineralogy of the Kalskaret peridotites has resulted from igneous crystallization at high levels in the mantle to give olivine + orthopyroxene crystal cumulates with the residual liquid imperfectly filter pressed out and crystallized in pockets within the peridotite and the whole mass subsequently recrystallized to a low temperature eclogite facies assemblage during tectonic emplacement into higher levels within the crust, therefore seems to be a valid one or at least one that is very difficult to eliminate conclusively on the evidence available at present. According to this hypothesis the intricate interlayering of the garnetiferous rock types could be accounted for through imperfect filter pressing with the restricted lenses of 'eclogitic' rocks representing trapped pockets of residual liquid. 5. Partial melting of an intermediate peridotite composition resulting in segregation into 'basaltic' partial melt and dunitic residual compositional fractions.

In this case the garnet free peridotites are considered as representing the residual left after the extraction of a partial melt of 'basaltic' composition, and the garnetiferous peridotites the result of imperfect filter pressing out of the partial melt fraction. This hypothesis poses three important questions.

a. Is the most eclogitic rock composition (T153) or any of the extrapolated rock compositions (see Table No. 22) likely to represent that of either a trapped partial melt or a crystal cumulate from a partial melt?
b. Did the partial melting process take place in the mantle or the crust, and consequently to what mineral facies did the initial and resultant rock types belong?

c. Do any rock types of intermediate composition occur, which are possibly representative of the unmodified source peridotite material?

Experimental melting of garnet peridotite and bimineralic eclogite at pressures of 30 kbs. (O'Hara 1963a,b) yielded liquids which were far removed from silica saturated tholeiitic basalt in composition and were considered to resemble picritic basalts. O'Hara and Yoder (1963) considered that the liquid formed by partial fusion of garnet peridotite should lie slightly to the SiO₂ poor side of the GNT-CPX-OPX plane (see Fig. 28). This plane has been demonstrated to be a thermal divide at this pressure (O'Hara 1963a,b) so that liquid derived from garnet peridotite cannot fractionate through this plane to give a quartz eclogite (or tholeiitic basalt) composition at this pressure.

O'Hara and Mercy (1963) concluded that the garnet peridotite nodules in kimberlite pipes may be samples of the upper mantle and the accompanying eclogite nodules their partial fusion products, melted and crystallized at great depth. However, for the partial melt composition to lie in the GNT-CPX-OPX plane, demonstrated to be a thermal divide at 30 kbs, would seem an extraordinary coincidence. Therefore bimineralic eclogites now seem best interpreted as crystal accumulates of the partial melt liquids, rather than the liquid compositions themselves (O'Hara and Yoder 1963).

Yoder and Tilley (1962) and O'Hara and Yoder (1963) also demonstrated that sorting of GNT-CPX assemblages could produce considerable compositional variations, illustrated by variations in the weight percentages of the C.I.P.W. normative constituents. Tholeiitic characteristics are associated with/
Figure 28. The tetrahedron CaO-MgO-Al₂O₃-SiO₂ showing the composition plane MgSiO₃ (enstatite, En) – CaSiO₃ (wollastonite, WOL) – Al₂O₃. The diopside (Di) – pyrope (Py) join lies in this composition plane, as do the joins Di – Ca Tschermak's molecule (Ca Tsch.) and pyrope – grossularite (GROSS.). The range of clinopyroxene solid solutions (CFX ss) that coexist with garnet solid solutions (GNT ss) at high pressures is shown diagrammatically. This diagram is after O'Hara and Yoder (1963).
with GNT rich compositions whereas alkaline characteristics are associated with CPX rich compositions.

Davis and Schairer (1965) have studied the melting relations in the join Diopside-Forsterite-Pyrope at 40 kbars pressure and considered the quaternary isobaric invariant point OI + CPX + OPX + GNT + Liquid to lie just to the SiO₂ poor side of the GNT - CPX - OPX plane (in fact at Di₄⁴,₇Fo₃Py₄⁷En₃ in the CaO-Al₂O₃-MgO-SiO₂ system). The primary liquid at 40 kbars is an enstatite rich picrite (similar to that envisaged by O'Hara and Yoder, 1963 at 30 kbars) with a C.I.P.W. norm of An₃₂.₄Di₁₈.₈En₂₆.₀ Fo₁₉.₉.

It therefore appears that the primary partial fusion extract of garnet peridotite, at least in the pressure range 30-40 kbars, is likely to be picritic with its composition lying slightly to the SiO₂ poor side of the GNT - CPX - OPX plane.

The most extreme Kalskarek eclogite rock type, Tl53, would in the light of the above experimental evidence have a suitable composition for it to represent a garnet peridotite partial melt extract formed at high pressures (30-40 kbars). Its composition is, in fact, that of a pyroxenitic picrite having a C.I.P.W. norm of Plag. 35.7 CPX 22.1 OPX 18.3 OI. 17.0. This composition is closely similar in the amounts of respective C.I.P.W. normative constituents to that of the primary liquid composition given by Davis and Schairer (1965). This Kalskarek garnet websterite rock type consists predominately of GNT and CPX but also contains small amounts of OPX and to a lesser extent OI. Its composition would therefore have lain just to the SiO₂ poor side of the GNT-OPX-CPX plane at the pressure of its formation.

The localized occurrence of the garnet websterite rock type (Tl53) in/
in two lenses within the garnet peridotite outcrop (see Fig. 8) is also in keeping with it representing trapped pockets of partial melt liquid (or residual liquid according to hypothesis-4). It should be noted that the Kalskaret garnet websterite specimen N23 analysed by Mercy and O'Hara (1965a) is similar mineralogically and chemically to Tl53 (c.f. C.I,P,W. norms). Any compositional difference between them may be attributed to slight sorting of GNT and CPX as crystal accumulate or perhaps subsequent mechanical sorting.

The extrapolated end member 'eclogitic' compositions (Table No.22) trend towards a SiO$_2$ saturated tholeite composition. It is unlikely that such a composition represents that of a partial melt extract of garnet peridotite formed at high pressures (30-40 kbs.) but it may represent that of a crystal accumulate of such a partial melt (O'Hara and Yoder 1963). This then suggests a possible variation on the partial melting hypothesis for the origin of the Kalskaret composition trend namely that it has resulted from the mechanical mixing of an eclogite accumulate of a partial melt and a dunitic residual. Such an origin does not, however, provide such an elegant explanation of the intimate interlayering of the various garnetiferous and garnet free rock types and envisages a hypothetical end member which seems unnecessary in view of the demonstration that the most eclogitic rock type present, Tl53, possesses a suitable composition for it to represent a likely partial melt extract of garnet peridotite formed at high pressures.

However, correlation of the composition of Tl53 with the partial melting product of garnet peridotite produced at pressures of 30-40 kbs. implies depths of origin in the order of 95-130 kms., perhaps rather excessive for these peridotite masses. It is therefore worth considering at this stage the likely depths to which orogenic tectogenes might extend thus/
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thus indicating possible depths from which material may have been brought up during orogenesis.

Since Veining Meinesz (1934) first discovered the negative gravity anomalies associated with the island arcs off S.E. Asia, and suggested that they were the result of down-buckling of the sialic crust, tectogene formation has become an integral part of most modern theories on orogenesis. However, direct evidence of just how far down the crustal disturbance during orogenesis is likely to extend is hard to come by, although figures of 60-70 kms., sometimes as high as 100 kms., are often mentioned.

Seismic studies (e.g. Gutenberg 1951, Benioff 1954, 1962) of earthquakes in the Japanese area have suggested the existence of broad inclined shear zones extending beneath the continental margins. The most frequent earthquakes associated with these movement zones have foci extending down to depths of 60-70 kms., but deeper earthquake foci indicate that movement extends down to depths of about 700 kms. Although the shallow zone appears to be the most active tectonically from the point of view of earthquake frequency, it may only be that while in the shallow zone movement takes place principally through brittle fracture and is hence abrupt, at greater depths movement occurs mostly through plastic flow with the stress being slowly dissipated. If such shear zones are commonly directly related to tectogene formation they provide an ideal mechanism for the emplacement of mantle material into the orogenic rock sequence.

The demonstration by Gorshkov (1958) of the probable existence of localized magma reservoirs between 50 to 70 kms. below the active orogenic zone of the Kamchatka and Kurila Islands Volcanic arc is also significant.
It therefore seems possible that during a profound orogenic episode, such as the European Caledonian one undoubtedly was, the tectogene may extend down to depths of 60-70 kms., at about which level local partial melting and magma generation may occur. At such depths tectonic overpressure produced through the downbuckling of the orogenic tectogene may bring the pressure under which partial melting of mantle garnet peridotite material might occur to about 30 kbs. Mantle peridotite material modified by partial melting processes under such conditions may then become involved in the orogenic movements and so be subsequently tectonically emplaced into higher crustal levels. It is also conceivable that the garnet peridotite melting relations observed at 30-40 kbs. persist to somewhat lower pressures, say down to 20 kbs. In that case it would not be necessary to invoke tectonic overpressure effects.

The apparent restriction of garnet peridotite occurrences within metamorphic gneiss and schist complexes to Europe, and to rocks of undoubted Caledonian age or rocks of possible Caledonian age exposed as 'basement' during the Hercogian and Alpine orogenies may indicate that only during the European Caledonian orogeny did the tectogene extend to sufficient depths or were the tectonic conditions such that mantle garnet peridotites were involved, incorporated and tectonically emplaced into higher crustal levels.

The alternative suggestion that the composition variation has resulted from the partial melting of granulite facies peridotite at considerably lower pressures (say 10-15 kbs. or less) at or near the base of the crust is open to the following objections.

1. Lack of evidence of the wholesale production of basalt magmas at such shallow depths. Seismic evidence for the depth of origin of basalt magmas (e.g.)
(e.g. Gorshkov 1958, Eaton and Murata 1960) suggests depths of the order of 50–70 km, clearly well within the upper mantle. At such depths mantle peridotites are expected to be well within the garnet peridotite stability field (c.f. Fig. 19b) although at the likely temperatures of melting (Fig. 19a) aluminous pyroxene peridotite is likely to be the stable assemblage.

2. The partial melt of peridotite under these conditions is likely to be nepheline normative in composition (O’Hara 1965, Green and Ringwood 1965, and Kushiro 1965 a,b) which does not fit with the inferred Kalskaret partial melt composition.

It is therefore considered that if the partial melting hypothesis for the explanation of the composition trend among the Kalskaret garnet peridotites is accepted, then the partial melting involved rocks of eclogite facies within the upper mantle, a conclusion which is enhanced by the occurrence of other garnet peridotite masses (namely those of Ugelvike) of possible mantle composition (see later) close by in the same orogenic belt.

Although it has been found impossible to eliminate conclusively the olivine cumulate-residual liquid hypothesis (4) as an explanation of the composition trend among the Kalskaret garnet peridotites, this author considers that preference for the partial melting hypothesis can be justified on several grounds.

1. Demonstration that the most eclogitic Kalskaret rock type represents a close approach to the expected composition of the partial melt of mantle garnet peridotite formed at 30–40 kbs. pressure.

2. Figure 29 is a \(\text{Ca}^{2+} : \text{Mg}^{2+} : \text{Fe}^{2+} + \text{Fe}^{3+}\) plot of peridotite and eclogite nodules from kimberlites — data from Williams (1932), Dawson (1962), Nixon et al (1963), and analyses (Holmes 1936, and O’Hara and Mercy personal communication) of peridotite nodules A1, A3, A6, A10, A15 and A17/
FIG. 29

NODULES IN KIMBERLITE PIPES

X GARNET PERIDOTITES
○ KYANITE ECLOGITES
○ ECLOGITES
examined by O'Hara and Mercy (1963).

With the exception of two pyroxenic nodules (370/2- Williams 1932, and A17 - O'Hara and Mercy 1963), there is an appreciable geochemical hiatus between the garnet peridotite and the eclogite nodules as was noted by O'Hara and Mercy (1963) who considered that... 'The higher Fe/Fe+Mg and Na/Na+Ca ratios, the increased fMg and Mn/Mg ratios observed in the eclogites relative to the peridotites; the absence of intermediate types; the relative abundances and the discrepancy between the eclogite composition and the inferred composition of the upper mantle from the chondrite model all indicate that the eclogites are partial fusion products of the garnet peridotites formed and consolidated at great depth.' They also noted that... 'the majority of the analyses of such eclogites exhibit a SiO₂ poor tholeiitic character rather than an alkali basalt or melilitite basalt character.' Although such eclogites are now interpreted as crystal accumulates of the partial melt liquids rather than the liquid compositions themselves (O'Hara and Yoder 1963, O'Hara 1965) it is considered significant here that the Kalskaret and Almklovdalen garnet peridotite compositions (Fig. 20) span the gap between the 'mantle' garnet peridotite and eclogite compositions (Fig. 29) suggesting that they have been involved in the process of production of partial melts within the mantle.

3. It is also considered significant that the Kalskaret and Almklovdalen garnet peridotite compositions trend towards those of the eclogites which occur as lenses in the Norwegian gneisses (Fig. 34). A discussion as to whether these eclogites represent primary partial melts of garnet peridotites, crystal accumulates of such partial melts, partial melts substantially fractionated at lower pressures, or whether they are of any other origin is given in the next chapter.
FIG. 30

GARNETS

- KALSKARET AND ALMKLOVDALEN GARNET PERIDOTITES
- UGELVIK GARNET PERIDOTITES
- CZECH GARNET PERIDOTITES
- KIMBERLITE GARNET PERIDOTITES
- KIMBERLITE ECLOGITES

\[ \text{Fe}^{2+} + \text{Fe}^{3+} \]
4. Comparison of published compositional data for garnets from kimberlitic garnet peridotite and eclogite nodules indicates notably increased Fe/Fe+Mg ratios and decreased Cr$_2$O$_3$ contents in the garnets from the eclogites.

Figure 30 is a Cr$^{3+}$:Mg$^{2+}$:Fe$^{2+}$+Fe$^{3+}$ plot of garnet compositions from kimberlitic garnet peridotite and eclogite nodules (Wagner 1914, Williams 1932, Nixon et al 1963, O'Hara 1963b, O'Hara and Mercy 1963, and MacGregor and Ringwood 1964), from Norwegian garnet peridotites (Eskola 1921, O'Hara and Mercy 1963, and analyses presented in this thesis), and from garnet peridotites from the Czech Massif (Fiala 1965).

The garnets from garnet peridotites which are considered here as possible representatives of primary mantle material - namely the garnet peridotite nodules in kimberlites, the Ugelvik and the Czech garnet peridotites (this hypothesis is further elaborated shortly) outline a compositional field with minimum Fe$^{2+}$+Fe$^{3+}$/Mg$^{2+}$ ratios of comparatively restricted range and high through variable Cr$_2$O$_3$ contents. The compositions of the garnets from the Kalskaret and Almklovdalen garnet peridotites trend towards those of garnets from the eclogite nodules in kimberlites showing increasing Fe$^{2+}$+Fe$^{3+}$/Mg$^{2+}$ ratios and decreasing Cr$_2$O$_3$ contents, indicating the possibility that they are from mantle garnet peridotites which have been modified by partial melting processes.

Lappin (1962) has analysed garnets from both the Almklovdalen garnet peridotites and the eclogite lenses in the gneisses of the Almklovdalen and Selje districts. He does not give Cr$_2$O$_3$ values, however, which is unfortunate as Cr$_2$O$_3$ content seems to provide an excellent compositional index for distinction between the two groups of garnet parageneses. Figure 31 is a Ca$^{2+}$:Mg$^{2+}$:Fe$^{2+}$+Fe$^{3+}$ plot of the garnets analysed by Lappin. Similar to Fig. 30 it shows the increased Fe$^{2+}$+Fe$^{3+}$/Mg$^{2+}$ ratios of the garnets from the/
FIG. 31

GARNETS
○ GARNET PERIDOTITES
● ECLOGITES IN GNEISS
the eclogites as compared with those from the garnet peridotites. It also illustrates the variable CaO contents of garnets from several of the eclogites which contrasts with the very restricted range of CaO contents among garnets from garnet peridotite assemblages (O'Hara and Mercy 1963 P.288 and Fig. no.21).

5. Although there are significant compositional differences between the compositions of clinopyroxenes from garnet peridotites and eclogites notably in $\text{Cr}_2\text{O}_3$ content, $\text{Fe}^{2+}/\text{Fe}^{3+}+\text{Mg}^{2+}$ ratio and jadeite content the differences are not so obvious as with the garnets and it is difficult to devise a single diagram to differentiate clearly between the two groups of garnet parageneses. An attempt has been made in Figure 32, however, where the clinopyroxene compositions (using data from Eskola 1921, Nixon et al. 1963, O'Hara 1963b, O'Hara and Mercy 1963, MacGregor and Ringwood 1964 and analyses presented in this thesis) are plotted on the basis of $\text{Mg}^{2+}:\text{Cr}^{3+}:10\text{Na}^{2+}+\text{Al}^{3+}+\text{Fe}^{3+}$.

The compositions of three clinopyroxenes from eclogite nodules in kimberlites are quite clearly separated from those from garnet peridotites considered as possible representatives of primary mantle material (nodules in kimberlites and those from Ugelvik) the distinction resulting from the lower $\text{Cr}_2\text{O}_3$ contents, higher $\text{Fe}^{2+}/\text{Fe}^{2+}+\text{Mg}^{2+}$ ratios and the greater extent of $\text{NaR}^{3+}$ substitution in the clinopyroxenes of the former group compared with the latter. The clinopyroxenes from the Kalskaret and Almklovålen garnet peridotites again tend to fall between the two groups, in harmony with the explanation of these rocks as mantle garnet peridotite material modified by partial melting processes. However, with the mineralogical evidence in these rocks of a low temperature equilibration of most of the mineral assemblages it appears that the $\text{NaR}^{3+}$ substitution index for distinguishing/
FIG. 32

CLINOPYROXENES

ORNAMENTS AS FIG. 30

$\text{Mg}^{2+}$ $\text{Na}^{+} + \text{Al}^{3+} + \text{Fe}^{3+}$ $\text{Cr}^{3+} \times 10$
distinguishing between the different clinopyroxene parageneses has been somewhat obscured by the recrystallization.

Lappin (1962) has also analysed several clinopyroxenes from both the Almklovdalen garnet peridotites and the eclogite lenses in the gneisses, but unfortunately again does not give the Cr$_2$O$_3$ contents. Figure 33 is a Mg$^{2+}$:Fe$^{2+}$:Na$^{2+}$+Al$^{3+}$+Fe$^{3+}$ plot of the clinopyroxenes analysed by him and indicates that while several of the clinopyroxenes from eclogites have appreciably higher Na$_2$O and R$_2$O$_3$ contents than the clinopyroxenes from the garnet peridotites others have quite comparable contents.

The composition distinction among the clinopyroxenes from the eclogites may be the result of differing parageneses (3 out of 4 of the higher NaH$^{3+}$ substitution clinopyroxenes are from kyanite bearing eclogite assemblages) or it may just reflect low temperature recrystallization and equilibration of the clinopyroxenes from certain of the eclogites. Whichever is the case it is apparent that at least some of the eclogites in the gneisses contain clinopyroxenes with appreciably higher Na$_2$O and R$_2$O$_3$ contents than those in the garnet peridotites.

If the composition trend among the Kalskaret garnet peridotites is the result of partial melting modification of mantle garnet peridotite material, then the mantle peridotite composition must lie within the composition series, obviously being intermediate between the residual and partial melt composition fractions. If the partial melting process has not been carried to completion the possibility exists that some mantle peridotite material still persists. It is therefore important to consider whether any peridotites of intermediate composition such that they might represent suitable primary mantle material do occur within the series.

If, on the other hand, the composition series has resulted through/
through the alternative olivine cumulate-residual liquid mechanism, then intermediate mantle peridotite compositions would not be expected.

In terms of the likely mantle composition suggested by the chondrite meteorite model (see Table No.25) it is expected that the primary upper mantle rock type would be a garnet peridotite with low modal contents of GNT and CPX. Such rock types are scarce among the Kalskaret garnet peridotites but do occur (e.g. N21 - O'Hara and Mercy 1963, Mercy and O'Hara 1965a) although on the evidence on hand at present it cannot be proved irrefutably that they represent primary mantle material while the more garnetiferous peridotites and the garnet free peridotites represent modified mantle peridotites. However, it may be possible to test some of the intermediate composition peridotites as possible primary mantle material in terms of their Rb, Sr, K, U and Th contents, and Rb/Sr and Sr$^{87}$/Sr$^{86}$ ratios as suggested by the work of Faure and Hurley (1963), Roe (1964), Aldrich et al (1964), Hart et al (1965) and Engel et al (1965). Mantle peridotite material must have suitable contents of these trace elements such that the contents of these elements in basalts can be derived from them and their radioactive element contents are sufficient to produce rates of heat flow from the mantle, similar to those observed. It is hoped in the future to have an opportunity to be able to follow up these possible tests.

In the meantime preliminary determinations of Sr and Rb in the Kalskaret and Ugelvik peridotites have been performed by X-ray fluorescent techniques. These indicated Rb values of at least <10 ppm, probably <5 ppm, in all these rock samples. Sr, values were as follows:

**TABLE NO.31/**
TABLE NO. 31

<table>
<thead>
<tr>
<th>SAMPLE NO.</th>
<th>Sr. ppm.</th>
<th>SAMPLE NO.</th>
<th>Sr. ppm.</th>
</tr>
</thead>
<tbody>
<tr>
<td>833/T200</td>
<td>40</td>
<td>846/U2</td>
<td>17</td>
</tr>
<tr>
<td>834/T268</td>
<td>14</td>
<td>847/U3</td>
<td>13</td>
</tr>
<tr>
<td>835/T150</td>
<td>65</td>
<td>848/U6</td>
<td>19</td>
</tr>
<tr>
<td>836/T151</td>
<td>51</td>
<td>849/U8</td>
<td>21</td>
</tr>
<tr>
<td>837/T152</td>
<td>47</td>
<td>850/U16</td>
<td>12</td>
</tr>
<tr>
<td>838/T153</td>
<td>74</td>
<td>84.6/U2</td>
<td>17</td>
</tr>
<tr>
<td>839/T154</td>
<td>42</td>
<td>84.7/U3</td>
<td>13</td>
</tr>
<tr>
<td>840/T97</td>
<td>66</td>
<td>84.8/U6</td>
<td>19</td>
</tr>
<tr>
<td>841/T96</td>
<td>59</td>
<td>84.9/U8</td>
<td>21</td>
</tr>
<tr>
<td>842/T155</td>
<td>96</td>
<td>84.6/U2</td>
<td>17</td>
</tr>
<tr>
<td>843/T156</td>
<td>14</td>
<td>84.7/U3</td>
<td>13</td>
</tr>
<tr>
<td>844/T162</td>
<td>30</td>
<td>84.8/U6</td>
<td>19</td>
</tr>
<tr>
<td>845/T100</td>
<td>89</td>
<td>84.9/U8</td>
<td>21</td>
</tr>
</tbody>
</table>

The low Sr contents of the Kalskaret garnet peridotites further substantiate the unlikelihood that a basaltic liquid which has been extensively fractionated at low pressures has been involved in the production of the present rock types. Faure and Hurley (1963) have given average Rb and Sr concentrations in high level crustal basalts as Rb $17^{±6}$ ppm., Sr $390^{±55}$ ppm. for 5 olivine basalts, and Rb $30^{±7}$ ppm., Sr $504^{±167}$ ppm. for 6 tholeiitic basalts. By comparison even the most eclogitic Kalskaret rock types have appreciably lower contents of Rb and Sr, and also of $K_2O$ and $P_2O_5$. 

Of/
Of all the basaltic magma types exposed at the earth's surface for which data is available, the oceanic tholeiites dredged from the ocean floors have lowest contents of all these elements and are therefore closest to the likely liquid composition postulated as having been involved in the production of the Kalskaret garnet peridotites. Engel et al (1965) has given the average contents of these elements in 10 such oceanic tholeiites as Sr 130 ppm, Rb <10 ppm, K₂O 0.16%, and P₂O₅ 0.16%. Hart et al (1965) gave the following contents for a similar Mid Atlantic Ridge basalt: K 1400 ppm, Rb 0.98 ppm, and Sr 102 ppm.

Engel et al (1965) considered such oceanic tholeiitic basalts as a chemically distinctive group representing primary magma erupted from the mantle. This author would agree with them to the extent of considering such basalts as being more primitive in composition than other groups of crustal basaltic magma types and probably parental to certain more highly fractionated basalts, but would not consider that they necessarily represent a primary magma which infers that they represent the actual composition of the partial melt extract formed in the mantle. As pointed out by O'Hara (1965) it seems in fact unlikely that any liquid produced in the mantle is extruded at the earth's surface in a truly primary nature, that is without any composition fractionation whatsoever having occurred during transit to the surface. Thus while it appears that the oceanic tholeiites represent probably the most primitive basaltic magma type erupted at the surface this only indicates that they have undergone least fractionation in transit to the surface.

In view of the likely contents of Sr, Rb, K, P etc. of the liquid considered to have been involved in the production of the Kalskaret garnet peridotites, together with its picritic nature and major element compositional/
compositional similarity with the partial melt of garnet peridotite formed experimentally at 30-40 kbars pressure, it is considered that this liquid was of a more primitive nature than that of the oceanic tholeiitic basalt: and was hence closer to the composition of a *trans* primary partial melt liquid formed at high pressure in the mantle.

The hypothesis that the composition trend among the Kalskaret garnet peridotites is the result of a crystal-liquid mixing process which occurred in a high pressure environment is therefore supported by the composition and primitive nature of the liquid phase involved. As already stated this process is considered to have been either a partial melting process involving mantle garnet peridotite or perhaps less likely a crystal cumulate-residual liquid crystallization process.

However, either crystal-liquid mixing process still raises a problem in trying to explain all the intermediate compositions on the linear composition trend (see Fig. 25).

With the partial melt -residual crystals hypothesis it is apparent that if the most eclogitic composition T153 is taken as representing that of the partial melt liquid itself then compositions occur in the range 0-30% liquid + 70-100% residual crystals and 60-85% liquid + 15-40% residual crystals. Thus assuming that MgO content provides a reliable compositional index of the composition trend (in effect it probably constitutes the best such index) there is a gap in the range 30-60% liquid + 40-70% crystals, which is probably significant but may just be fortuitous. A gap in approximately the same composition range is demonstrated by several of the other elements, namely Al$_2$O$_3$, Na$_2$O, NiO and TiO$_2$ but is absent with Cr$_2$O$_3$, SiO$_2$ and FeO$^+$ (see Fig. 25).

Compositions in the range 0-30% liquid can be readily explained/
explained as representing predominantly residual crystals with trapped interstitial liquid. Alternatively some of these compositions may represent unmodified primary mantle garnet peridotite material. However, compositions which represent 60-85% liquid + 15-40% crystals are more difficult to explain. In normal circumstances such proportions of crystals would be expected to sink out of the liquid phase and a much larger gap in the intermediate crystals + liquid range would be expected.

The intermediate compositions may, however, have been developed according to one or a combination of the following explanations.

1. The liquid composition may not in fact be represented by T153 but may be more intermediate in the composition series. If N23 or T97 represents the liquid composition, rather than T153 which might then represent a crystal cumulate from the liquid, then the composition gap widens somewhat to about the 35-70% liquid range. However, these compositions with high modal OPX and O1 contents, respectively, are not so satisfactory as T153 as possible partial melt liquids of garnet peridotites at high pressures, in view of the fact that O1 and OPX are likely to bear a reaction relationship to such a partial melt liquid (O'Hara and Yoder (1963)).

2. Under natural conditions the liquid phase at pressures of the order envisaged (perhaps up to 30 kbs.) may be more viscous than expected from experimental studies at high pressures, hence restricting crystal settling especially if the partial melting process was relatively quickly frozen.

3. It is possible that some of the more eclogitic compositions may represent the product of the mixing of GNT and CPX crystal cumulates of the partial melt with O1 and CPX crystals of the residuum. Such a process need not necessarily lead to much deviation from the linear composition trend./
trend. It might in fact explain the minor deviations from ideal linearity (c.f. Fig. 25) which suggests perhaps a certain amount of segregation of GNT from CFX or vice versa in several of the more eclogitic compositions. However, any mechanical segregation process cannot have proceeded all that far otherwise the compositional linearity would not be expected to be so strong.

4. A further explanation may lie in the fact that with the envisaged partial melting process, the liquid is considered to have developed essentially in situ from an intermediate composition in the composition series. Hence the liquid may have been trapped as minor pockets or stringers within the peridotite with which following crystallization it was mechanically mixed, remembering the evidence for the subsequent low temperature recrystallization of the sequence.

However, both these last two explanations bring back the possibility of mechanical mixing which was previously rejected as a mechanism for the origin of the linear composition sequence through the mixing of basalt or eclogite with crystalline dunite into which it was intruded. It is still considered here that an entirely mechanical mixing process is unlikely to have produced the intricate layering and the range of bulk compositions which occur, in particular the predominance of intermediate types in the Lien locality which probably represent something like 0-30% 'basalt' together with 70-80% dunite.

The possibility exists, however, that a limited degree of mechanical mixing during tectonic emplacement superimposed upon the crystal-liquid mixing process has been a factor in producing the intermediate rock compositions which are difficult to explain by the crystal-liquid process alone.

The/
The crystal cumulate-residual liquid hypothesis alone provides a similar problem in explaining the intermediate compositions which represent the apparent mixing of a large proportion of liquid with a smaller but significant proportion of crystals.

Therefore although neither crystal-liquid mixing process postulated provides by itself an entirely satisfactory explanation of all the intermediate compositions on the linear composition trend, this is not considered to be a sufficient reason for rejection of such a mechanism for the origin of the composition variations noted. Certainly all other hypotheses apparent to this author seem far less satisfactory.

Returning to the mineral and bulk rock composition trends outlined earlier in this section it is considered that the olivine cumulate-residual liquid mechanism may have resulted in a differentiation trend which might have been similar to that described among the Kalskaret and Almklovdalen garnet peridotites, but only by coincidence would the compositions of these peridotites then have lain precisely in the intermediate range between suggested mantle garnet peridotite and eclogite compositions.

The hypothesis favoured by this author to explain the composition trend among the Kalskaret garnet peridotites is then that it represents the fossilized process of segregation of eclogitic and dunitic composition fractions produced by the partial melting of original mantle peridotite material. The partial melting process in this instance is considered not to have been carried to completion, that is the partial melt phase has not been completely filter pressed out and removed but has been trapped and crystallized within the residual peridotite. The possibility of subsequent limited mechanical mixing during tectonic emplacement of these masses as a factor/
factor leading to the production of certain intermediate rock compositions should be borne in mind, however.

From the apparent parallelism of the composition trends among the Lien and Rodhaugen garnet peridotites with that of the Kalskaret garnet peridotites (see Figs. 20, 25 and 26) it is probable that a similar origin was also responsible for the composition variation trends noted among them. More data are needed to confirm this, however, and in particular to show the complete range of compositions present at these localities.

In the meantime it is considered that all these peridotite masses with their enclosed garnet peridotite layers have originated in the mantle and have been involved in partial melting processes there before their emplacement into their present environment. The garnet free peridotites with very low CaO and Al_2O_3 and negligible alkali contents (see Table Nos. 35 and 36) are assumed to represent residuals after the removal of a partial melt fraction, while the Kalskaret and Almklovdan garnet peridotite layers are considered to represent fossilized attempts at the production of such partial melts. Thus these garnet peridotite outcrops as a whole are considered to represent neither original mantle garnet peridotite material, nor trapped ecolitic partial melts of the original garnet peridotite, nor just the residual from such a process, but probably a combination of all three types of material.

The garnet peridotites with low modal contents of GNT and CPX which outcrop extensively within the Lien peridotite mass and to a much more limited extent in the Kalskaret mass, may represent original mantle material which has escaped the partial melting process and has only been modified by the subsequent low temperature recrystallization.
The displacement of the linear composition trends for the
Kalskaret, Lien and Rodhaugen localities (see Fig. 20) indicates
systematic variations in the \( \text{Fe}^{2+}/\text{Fe}^{3+}/\text{Mg}^{2+} \) ratios of the rock types
from these three localities, which may stem from inherent variations
in \( \text{Fe}^{2+}/\text{Fe}^{3+}/\text{Mg}^{2+} \) ratio of the original mantle peridotite compositions
which have been involved in the partial melting process.

B. THE LOW TEMPERATURE EQUILIBRATION OF THE GARNET PERIDOTITES

Recognition of the apparent low temperature (~600-700°C)
equilibration of the present garnet peridotite assemblages as demonstrated
in Chapter III might be thought at first to raise a major objection to their
suggested high temperature origin in the mantle.

The only irrefutable evidence of any previous origin prior to
the production of the low temperature assemblages is the occurrence in certain
mineral assemblages of pyroxenes (with significantly higher \( \text{R}_{2} \text{O}_3 \) contents
than those of the present mineral assemblages) which show evidence of
exsolution of garnet and chrome spinel.

On these grounds alone the present low temperature garnet peridotite
assemblages might, however, be considered to have resulted from the eclogite
facies recrystallization within the crust of spinel bearing peridotites
containing high alumina pyroxenes.

However, most estimates of the continental thermal gradient are
such (c.f. Clark and Ringwood 1964; Ringwood, MacGregor and Boyd 1964) that
the conditions of equilibration deduced from the garnet peridotite mineral
assemblages and from the likely garnet peridotite stability field outlined
by/
by Ringwood, MacGregor and Boyd (1964) and MacGregor (1965) — namely about 600-700°C, pressure greater than about 10 kbs., and water free conditions — are unlikely to have been attained under normal lithostatic conditions within the crust, in fact at depths less than about 50 kms. (c.f. Figs. 19a and b).

Unfortunately the boundary conditions between the garnet peridotite and the aluminous pyroxene spinel peridotite stability fields are as yet only roughly defined (see MacGregor 1965) especially as the effect of FeO on the stability relations in the system MgO-Al₂O₃-CaO-SiO₂ has not yet been experimentally determined. The exact position of the boundary is exceedingly critical in deciding whether or not crystallization of eclogite facies mineral assemblages is possible in the crust. However, in the light of the data at present available it seems possible that with a somewhat steepened geothermal gradient as to be expected under regional metamorphism (although not too steep otherwise crystallization will occur in the aluminous pyroxene spinel peridotite field) and/or tectonic overpressure effects produced through the downbuckling of the orogenic tectogene, the necessary temperature and pressure conditions for the production of the observed garnet peridotite mineral assemblages, as outlined above, may be attained in the lower levels of the crust.

O’Hara (in press) has, in fact, postulated the formation of ‘crustal’ garnet peridotites through the metamorphism of spinel bearing aluminous pyroxene peridotite masses under orogenic conditions with tectonic overpressure near the base of the crust.

However, it is considered that at least in the case of the Kalskaret and Almkløvdalen garnet peridotites such an hypothesis must be rejected on the following grounds.

1/
1. Attempts to explain the bulk rock chemical variations among these rocks in terms of the metamorphism of spinel bearing aluminous pyroxene peridotite assemblages \(e.g.,\) primary Lizard assemblage - Green (1964) have proved unsuccessful (see Chapter IV section C and Appendix II).

2. Statistical analysis of the composition variations noted among these rocks indicates that they have resulted essentially from a process involving the mixing of only two end member phases. After consideration of the various conceivable possibilities it was concluded in the last section that the most feasible origin of the observed compositional and mineralogical variations was through the mixing of partial melt and residual fractions produced by a partial melting process operating on an intermediate composition garnet peridotite.

As regards the present low temperature equilibrated garnet peridotite assemblages, it is therefore considered that they have resulted through recrystallization of previous higher temperature eclogite facies assemblages. This recrystallization is thought to have occurred during the tectonic emplacement of these peridotite masses into their present environment from their origin in the mantle, and may represent an essentially continuous process of the peridotite assemblages attempting to re-establish equilibrium with the physical conditions of the environments through which they have passed. The 600-700°C equilibration may therefore only represent the final phase of extensive recrystallization within the eclogite facies of these peridotite assemblages, perhaps 600-700°C representing the temperature below which cataclasis finally outpaced and ceased to be accompanied by recrystallization.

C.
C. DISCUSSION OF OTHER TAFJORD PERIDOTITES

Although the main line of research in this project has been concerned with the garnetiferous peridotite occurrences, it should not be forgotten that they form only a very minor fraction within the preponderant garnet free peridotites. A short discussion of the latter group of peridotites is therefore given here in relation to the bulk rock analyses presented in Table Nos. 35 and 36 (see Appendix III).

To fit with the preferred explanation of the Kalskaret garnet peridotites it has been stated in the first section of this chapter that the garnet free peridotites represent residuals after the removal of a partial melt fraction. This conclusion was based on the fact that they are clearly for the most part impoverished in many of the elements enriched in the postulated partial melt phase (e.g. TiO₂, Al₂O₃, MnO, CaO and Na₂O) and enriched in others (Cr₂O₃, NiO and MgO). It should be borne in mind, however, that these peridotites would also have been consistent with being crystal accumulates if the hypothesis that the garnetiferous peridotites represent pockets of trapped residual liquid within such peridotites had been preferred.

In Table No. 35, the bulk rock analyses of 11 specimens from the Kaldhussaeter peridotite mass are listed. None of these shows any evidence of either garnet or clinopyroxene in hand specimen, while the degree of serpentinization varies considerably.

Specimens T80 and T81 (Table No. 27) are amphibolites from the restricted occurrence of mafic layering within this peridotite mass (see Fig. 11 and Plate 2B) with T75, T76 and T79 from the immediately adjacent peridotite layers. The compositions of T80 and T81 appear to show similar impoverishment and/
and enrichment in particular elements compared to the dunite specimens as do the Kalskaret garnet peridotites. This is particularly demonstrated by their notably higher Fe/Fe+Mg ratios than all the other analysed Kaldhussaeter peridotites. These compositional features suggest that they represent amphibolitized garnet peridotites (see also discussion in Chapter VI), probably of the same origin as the Kalskaret garnet peridotites. Unfortunately the fact that the primary minerals are for the most part unobtainable, owing to the high degree of amphibolitization, makes it impossible to test this rigorously.

Specimens T75 and T79 with their low Al₂O₃, and very low CaO and Na₂O contents can have contained originally only very minor amounts of garnet and negligible clinopyroxene. They do, however, appear to have rather high Fe/Fe+Mg ratios compared with the other CaO and Al₂O₃ impoverished peridotites. The reason for this is not immediately apparent but would appear to be related to their close field association with the ferriferous rock types T30 and T61.

T76 contains appreciable Al₂O₃, CaO and Na₂O which must be primarily contained in the secondary chlorite and amphibole, but probably indicates the original presence of garnet and clinopyroxene in the primary assemblage. It does not, however, have a higher Fe/Fe+Mg ratio than the more truly monomineralic dunites.

T37 likewise shows a correlation between Al₂O₃, CaO and Na₂O contents which again suggests the original presence of a certain amount of primary clinopyroxene and possibly also garnet. As with T76 this rock type is no more ferriferous than the true dunites.

T207, with its notably high SiO₂ content, represents a small isolated harzburgite nodule, about 4 inches across. Its occurrence within sheared/
sheared platy dunites and the lack of cataclasis of the enstatite suggest the late stage secondary growth of the enstatite probably through hydrothermal transfer of SiO₂ resulting in the conversion of the primary olivine to enstatite (c.f. Bowen and Tuttle 1949). This rock specimen also has an appreciable content of chromite. The secondary enstatitization of the peridotites is far more restricted, however, in the Kaldhussaeter mass than in the Runukdalen peridotite masses, of which T133 represents a typical orthopyroxenite and T146 a more intermediate harzburgite (see Table No.36).

The Kaldhussaeter rock types so far described, with the exception of T207 - the secondary harzburgite nodule, represent the exceptionally chlorite and amphibole rich varieties while the other 6 peridotite listed in Table No.35, namely T214, T215, T237, T242, T246 and T247, are representative of the most typical rock type within the Kaldhussaeter mass, a more or less true dunite. These dunites contain very low to negligible contents of Al₂O₃, CaO and Na₂O which could be accounted for by only very minor amounts of primary garnet and/or clinopyroxene. There is not always a perfect correlation between CaO and Na₂O contents as would be expected if the contents of these elements are related solely to primary clinopyroxene, but the CaO content is also expected to be related to primary garnet content while the possibility also exists of a certain degree of migration of alkalies during partial serpentinization.

In any group of partially serpentinized peridotites one invariably seems to find slight irregularities in the distribution of alkalies (c.f. Challis 1965a).

The Fe/Fe+Mg ratio of most of these dunites is quite constant indicating at the most only slight mineral composition variations, in particular of olivine the predominate mineral. This would seem to be confirmed by a number of chemical and X-ray (O'Hara and Mercy 1963) and optical determinations of olivine and orthopyroxene compositions from these peridotites (see Chapter III), which demonstrates/
demonstrates the very restricted composition range of these minerals in these rock types.

Therefore as regards the Kaldhussaeter peridotites, and this also seems to be the case for other peridotite masses for which some analyses are presented in Table No. 36, the preponderant dunites with low $\text{Al}_2\text{O}_3$ and CaO contents seem to be in harmony with the interpretation that they represent mantle peridotite residuals left after the subtraction of a 'basaltic' partial melt fraction. Their $\text{Al}_2\text{O}_3$ and CaO contents are clearly too low for likely primary mantle material, in particular in the light of the chondrite meteorite model (c.f. Table No. 25). It appears that what composition variations there are amongst these peridotites can be attributed to sorting of minor proportions of orthopyroxene, and to an even lesser extent garnet and clinopyroxene in primary assemblages consisting for the most part of olivine.

It is apparent, however, that certain peridotite rock types do occur (e.g. T80, T61, T177 and T178) within these peridotite masses which probably originally contained appreciable amounts of garnet and clinopyroxene, in view of their present $\text{Al}_2\text{O}_3$ and CaO contents, which together with their considerably higher Fe/Fe+Mg ratios than the dunites suggests that they are more akin to the Kalskaret garnet peridotites and may thus also represent stringers of trapped partial melt liquid. It is unfortunate that in these instances the high degree of secondary amphibolitization and chloritization has more or less completely obliterated the original mineral assemblages.

Rock types such as T76 and T37 with small but significant $\text{Al}_2\text{O}_3$ and CaO contents but which do not show any considerable enrichment in Fe with respect to Mg may represent more primary mantle peridotite compositions. They are clearly not baren enough in $\text{Al}_2\text{O}_3$, CaO and alkalies to have had much if any basaltic fraction extracted from them.

Other composition variations amongst these peridotites as a group would seem to be accounted for by secondary serpentinization and enstatitization effects.
Considerable confusion has resulted regarding the specimen from Almklovldalen referred to as the Rodhaugen eclogite.

Eskola (1921) gave the mineral analyses and calculated a bulk rock analysis of an eclogite specimen marked 'Almklovldalen' from the Oslo Museum. Through comparison with specimens later collected by him, Eskola was certain that this specimen came from 'the hill called Rodhaugen.'

There are, however, two occurrences of 'eclogitic' rocks within the peridotite in the Rodhaugen area.

1. Just behind the Rodhaugen scarp feature there is interlayering of heterogeneous garnetiferous rock types within the peridotite including some thin bands of garnet websterite (GNT + CPX + minor OPX) but no true bimineralic eclogites.

2. A little to the north of the above locality there is a lens of what appeared in the field to this author to be essentially homogeneous bimineralic eclogite exposed within peridotite in a small stream section near Ekraemsaeter farm. The rock type at this locality consists almost entirely of garnet and clinopyroxene along with minor pale green amphibole, and rutile and an opaque ore mineral as accessories. It contains neither olivine nor orthopyroxene and therefore represents the only eclogite (sensus stricto) found among the Almklovldalen peridotites.

This latter rock type was the one analysed by Mercy and O'Hara (1965a) as specimen no. N75 and also coincides with the petrographic description given by Eskola of his Rodhaugen eclogite specimen.

Comparison of the bulk rock analysis of N75 and Eskola's calculated bulk/
bulk analysis of the Rodhaugen eclogite (see Mercy and O'Hara 1965a Table 1) reveals, however, several disturbing discrepancies, notably in Fe/Fe-Mg ratio and Na₂O content. As noted by Mercy and O'Hara (1965a) the clinopyroxene from N75 must have an appreciably higher Na₂O content, probably as jadeite molecule, than the clinopyroxene analysed by Eskola.

Analyses are now available of the garnet and clinopyroxene from N75 (Mercy - personal communication). The garnet has the composition Ca₃₀.₆Mg₄.₃Fe₂₅.₄Mn₀.₆, with a very low Cr₂O₃ content (0.034 wt.%) and an extremely high CaO content (11.62 wt.%), which contrasts strongly with the garnet composition given by Eskola as from the Rodhaugen eclogite - namely Ca₁₂Mg₆₁Fe₂₆Mn₁ with Cr₂O₃ of 0.22 wt.% and CaO of 4.71 wt.%.

The clinopyroxene as previously suspected has an appreciably higher Na₂O content (2.67 wt.%) than that analysed by Eskola (0.83 wt.%).

In the light of the composition features of garnets and clinopyroxenes from garnet peridotites and eclogites discussed in the preceding section (see Figs. 30-33 inclusive), it is apparent that while the mineral composition of the garnet and clinopyroxene from N75 indicate a likely true eclogite paragenesis the mineral compositions of the rock type referred to by Eskola as the Rodhaugen eclogite are far more comparable with the compositions of garnet and clinopyroxene from certain of the Almklovdaalen garnet peridotite assemblages.

Particularly impressive in this respect is the CaO content of 11.62 wt.% of the garnet from N75 compared with 4.71 wt.% for the garnet analysed by Eskola. It can be demonstrated that the garnets from the garnet peridotite assemblages (GNT + CPX ± OL + OPX) show very limited variation in CaO content, being restricted almost entirely to the range 4-5 wt.% (see data in O'Hara and Mercy 1963, Nixon et al 1963, Fiala 1965 and presented in/
in this thesis). However, in true eclogite assemblages, lacking olivine and orthopyroxene, the CaO content in the garnet is no longer closely restricted but shows considerable variations (see data in Wagner 1914, Williams 1932, Lappin 1962, Nixon et al 1963, and O'Hara 1963b). O'Hara (1960a) has given a discussion of just this point.

This therefore suggests that the garnet from N75 was from a true eclogite assemblage, lacking olivine and orthopyroxene, while the garnet analysed by Eskola was from a mineral assemblage which might have contained at least a trace of olivine or orthopyroxene, or both.

Thus although the petrographic description of the Rodhaugen eclogite given by Eskola coincides with the N75 specimen analysed by Mercy and O'Hara (1965a), the mineral and bulk rock compositions of the two specimens appear too diverse for them to represent the same rock type from one locality.

Alternative explanations are either that ....

1. the specimen analysed by Eskola did not come from the same Rodhaugen locality where N75 was collected but rather from the garnet peridotite locality directly behind the Rodhaugen hill, or less likely from another locality altogether.

or

2. the Rodhaugen eclogite of which N75 definitely represents a true specimen is compositionally more variable than is suspected by this author.

It appears that Eskola had the misfortune to have performed much of his analytical work on poorly labelled specimens from the Oslo Museum collected by earlier workers. As with the Rodhaugen eclogite specimen he later tried to match them in the field, an exceedingly tricky task. Eskola himself/
himself expresses doubts about the exact location of certain of his
analysed specimens, in particular the Duen type eclogite and the Lyngenes
eclogite.

Regarding the actual paragenesis of the true Rodhaugen eclogite,
because of its homogeneous bimineralic nature and its nepheline normative
composition it is considered to possibly represent an eclogite accumulate
of a garnet peridotite partial melt trapped and crystallized within the
peridotite. As well as the mineral composition features (low Cr$_2$O$_3$,
high Fe/Fe+Mg, high Na$_2$O in CPX, high CaO in GNT) its content of accessory
rutile also suggests that it is more related both chemically and
mineralogically to certain of the eclogite lenses in the adjacent gneisses
(c.f. Lappin 1962) than to the garnet peridotites. Rutile is a conspicuous
accessory mineral in almost all the eclogites examined from the gneisses,
while it is absent in the garnet peridotite rock types.

The hypothesis that the eclogite lenses in the gneisses also
represent accumulates of primary or fractionated garnet peridotite partial
melts is considered in the next chapter.

E. THE PETROGENESIS OF THE UGELVIK PERIDOTITES

In the earlier discussion of the linear composition trends
(Figs. 20, 25 and 26) it was noted that the trend of the Ugelvik garnet
peridotite compositions is notably different from the Kalskaret, Lien and
Rodhaugen trends and was therefore considered to be of different origin.
It was pointed out that while the Kalskaret, Lien and Rodhaugen peridotite
compositions show markedly increased Fe$^{2+}$+Fe$^{3+}$/Mg$^{2+}$ ratios with increasing
Ca$^{2+}$ content correlating with the increasing 'eclogitic' nature of the
assemblage/
assemblage (a feature subsequently attributed to their involvement in partial melting processes), the Ugelvik peridotite compositions show only a slight increase in Fe\textsuperscript{2+} + Fe\textsuperscript{3+} / Mg\textsuperscript{2+} with increasing Ca\textsuperscript{2+} content.

The exceptional mineral chemistry of the Ugelvik assemblage U\textsubscript{6} was stressed in Chapter III, where reasons for possibly considering this rock type as a modified 'disequilibrium' assemblage were pointed out. Its coincidence on the Ca\textsuperscript{2+}:Mg\textsuperscript{2+}:Fe\textsuperscript{2+} + Fe\textsuperscript{3+} plot with the compositions of two pyroxenic nodules from kimberlite pipes (c.f. Figs. 20 and 29) is interesting and may be more than just coincidence. The fact that in Figure 20 the bulk rock composition of U\textsubscript{6} lies on the extension of an apparent linear trend through the other Ugelvik peridotite compositions may, however, be coincidental. In fact, the best fit line through the other 4 Ugelvik peridotite compositions passes rather to the right of the U\textsubscript{6} composition point. More analyses are obviously required to define this linear composition trend if in fact it genuinely exists.

This author considers from the data available at present that, with the exception of the rock composition U\textsubscript{6}, the apparent linear trend through the other 4 compositions principally reflects the sorting of garnet among these rock types. The most notable mineral variation among these 4 rock types is undoubtedly in garnet content (0-20 vol.%), although the modal proportions of the respective minerals among these 4 rock types as a group show much less variation than among the Kalskaret and Almklovdalen peridotites (see Table No.4).

Furthermore it is apparent from the mineral composition data discussed in Chapter III, that, with the exception of the assemblage U\textsubscript{6}, the minerals of the Ugelvik peridotites are relatively uniform in composition, particularly/
particularly with respect to Fe/Fe+Mg ratio. It would therefore appear that the mineral layering in the Ugelvik peridotite masses could be adequately explained by the sorting of an eclogite facies assemblage precipitated by igneous crystallization processes (see also discussion of composition variation diagram Fig. 45).

As pointed out earlier a similar explanation cannot, however, hold for the Kalskaret and Almklovdalen garnetiferous peridotite layering because of the extreme mineral composition variations involved.

While in the Kalskaret and Almklovdalen peridotites the Fe+Mn+Ni/Fe+Mn+Ni+Mg ratio in the C.I.P.W. normative ferromagnesian silicates increases markedly with increasing modal garnet and clinopyroxene contents of the mineral assemblage, this ratio is quite constant in the C.I.P.W. norms of the Ugelvik garnet free assemblage U₃, the garnet poor assemblages U₂ and U₁₆, and the garnet rich assemblage U₈.

In the discussion of the mineral data in Chapter III (see in particular Table No. 6) it was demonstrated that the range of mineral compositions from the Ugelvik peridotites is much more comparable with that of the minerals from the garnet peridotite nodules in kimberlites analysed by O'Hara and Mercy (1963) than with that of the minerals from the Kalskaret and Almklovdalen peridotites. The average garnet composition data given in Table No. 23 further exemplifies the statement that while the Ugelvik garnets are closely comparable in composition to those from South African and Siberian garnet peridotite nodules in kimberlites, they are quite distinct from garnets from the Kalskaret and Almklovdalen garnet peridotite masses which have much higher $\frac{Fe^{2+}}{Fe^{2+}+Mg^{2+}}$ ratios and lower Cr₂O₃ contents. The correlation between the mineral compositions from the/
<table>
<thead>
<tr>
<th></th>
<th>I</th>
<th>II</th>
<th>III</th>
<th>IV</th>
<th>V</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ca</td>
<td>11.6</td>
<td>11.8</td>
<td>12.2</td>
<td>11.8</td>
<td>13.4</td>
</tr>
<tr>
<td>Mg</td>
<td>73.7</td>
<td>75.2</td>
<td>61.5</td>
<td>72.3</td>
<td>72.0</td>
</tr>
<tr>
<td>Fe^{2+}</td>
<td>13.8</td>
<td>12.2</td>
<td>25.3</td>
<td>15.2</td>
<td>14.1</td>
</tr>
<tr>
<td>Mn</td>
<td>0.9</td>
<td>0.7</td>
<td>0.9</td>
<td>0.7</td>
<td>0.5</td>
</tr>
<tr>
<td>Cr_{2}O_{3} wt.%</td>
<td>3.16</td>
<td>3.21</td>
<td>0.64</td>
<td>2.18</td>
<td>?</td>
</tr>
</tbody>
</table>

**COLUMN I**  Average of 4 garnet compositions from the Ugelvik garnet peridotites, excepting assemblage U6 (Table No.3).

**II**  Average of 3 garnet compositions from the garnet peridotite nodules in South African kimberlites (O'Hara and Mercy 1963).

**III**  Average of 7 garnet compositions from the Kalskaret and Almklovdalen garnet peridotites (Table No.3 and O'Hara and Mercy 1963).

**IV**  Average of 7 garnet compositions from the Czech garnet peridotites (Fiala 1965 Table 2).

**V**  Average of 20 analyses of garnet from garnet peridotite inclusions in Siberian kimberlites (Sobolev 1963, quoted by Fiala 1965).
the Ugelvik peridotites and those from the kimberlitic garnet peridotites is further demonstrated on the garnet and clinopyroxene composition plots (Figs. 30 and 32).

Consideration of the mineralogy and field relations of the garnet peridotites which occur in the Norwegian gneisses and those which occur as nodules in Kimberlite pipes, led O'Hara and Mercy (1963) to the conclusion that ... 'both types of garnet peridotite are derived from the mantle, but that only the nodules in Kimberlite can be representative of the unmodified composition of the upper mantle'.

This author endorses their conclusion regarding the Kalskaret and Almklovdalen garnet peridotites from the Norwegian gneisses as unlikely representatives of unmodified upper mantle material, but suggests that the Ugelvik garnet peridotite masses can be correlated both chemically and mineralogically with the garnet peridotite nodules in Kimberlites and that the restricted range of bulk rock and mineral chemistry among the Ugelvik and kimberlitic garnet peridotites as a group indicates that they are all possibly representative of primary upper mantle material.

The Ugelvik garnet peridotites are mineralogically similar to the kimberlitic garnet peridotites in containing generally small modal proportions of garnet (0-20 vol.%) and clinopyroxene (0-15 vol.%) - see O'Hara and Mercy 1963 Table IV, and the Ugelvik modes Table No.4. Most of the analysed Kalskaret and Almklovdalen garnet peridotites (O'Hara and Mercy 1963, and Kalskaret modes Table No.4.) contained appreciably higher modal contents of garnet and clinopyroxene.

However, although the Ugelvik peridotites are to a large extent chemically, mineralogically and texturally similar to the peridotite nodules in Kimberlite pipes described by Wagner (1914), Williams (1932), Holmes (1936) Dawson (1962)/
Dawson (1962), Nixon et al (1963) and O'Hara and Mercy (1963), an apparent distinction appears to exist in the lower OPX/OL ratio of the Ugelvik peridotites.

O'Hara and Mercy (1963) have noted average modal proportions of olivine as about 50% by volume, and orthopyroxene about 40% by volume in the peridotite nodules in kimberlites analysed by them.

Nixon et al (1963) gave the following modal proportions of minerals in the garnet peridotite nodules from Basutoland Kimberlites:

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Range</th>
<th>Mean</th>
</tr>
</thead>
<tbody>
<tr>
<td>Olivine</td>
<td>20-69</td>
<td>55</td>
</tr>
<tr>
<td>Enstatite</td>
<td>7-41</td>
<td>26</td>
</tr>
<tr>
<td>Chrome Diopside</td>
<td>0-35</td>
<td>11</td>
</tr>
<tr>
<td>Chrome Pyrope</td>
<td>2-20</td>
<td>8</td>
</tr>
</tbody>
</table>

Dawson (1962) noted that olivine invariably constitutes more than 60% by volume of the peridotite nodules in Basutoland Kimberlites studied by him.

Excepting the rather anomalous rock type U6, the average OPX content of the analysed Ugelvik peridotites is about 8% by volume, and that of olivine + serpentine about 80% by volume.

A possible explanation of the apparent low OPX content of most Ugelvik peridotites as compared with the kimberlite peridotite inclusions, may lie in the more extensive cataclasis and serpentinization of the former. Textural evidence from the Norwegian garnet peridotites as a group, indicates that OPX is the mineral most rapidly cataclased during deformation, the/
the exceedingly fine grain size of the resultant cataclasized OPX being therefore likely to result in an underestimation of the OPX content of the rock.

Furthermore with the ensuing serpentinization of the tectonically emplaced and deformed Ugelvik peridotites, and with textural evidence that OPX as well as olivine has been extensively serpentinized, it is apparent that the original OPX content of these rocks may have been somewhat higher than is now obvious. The high OPX content (about 35% by volume) of one relatively uncataclasized and only mildly serpentinized Ugelvik garnet lherzolite (U₁₀), also suggests that some of the more highly cataclasized and serpentinized peridotites may originally have had reasonably high OPX contents, more comparable with the OPX contents of the peridotite nodules in kimberlites.

A higher OPX content could also explain the low MgO+FeO+NiO/SiO₂ ratios of the recalculated water free analyses U₂, U₈ and U₁₆, when compared with the analyses calculated from mineral compositions assuming that all the serpentine was derived from olivine. In that case leaching of MgO, FeO and NiO may not in fact have occurred during serpentinization, or at least have been more restricted in influence than would appear.

As the NiO ratio of OL:OPX in these rocks is about 4:1, accurate NiO determinations of all the constituent minerals including the serpentine and ore fractions, should indicate whether the OPX content has been underestimated or whether leaching has in fact occurred during serpentinization — see Appendix I on 'Serpentinization'.

Fiala (1965) has described garnet peridotite masses from the Czech Massif which appear more akin to the Ugelvik garnet peridotites than any/
any other crustal garnet peridotite masses described in the literature. These garnet peridotites are likewise highly serpentinized, are interlayered with peridotites devoid of garnet and occur predominately in a granulite facies complex. The general mineral assemblage 01 (Serp.) + GNT + OPX + CPX + accessory amphibole, is similar to that of the Ugelvik peridotites as are the modal proportions of garnet in the peridotites, ranging from 2.5-14% in the 7 rock samples from which garnet has been analysed.

The two bulk rock analyses of Czech garnet peridotites given by Fiala (1965) are likewise very similar to the bulk analyses of the Ugelvik garnet peridotites (see Fig. 20).

The average of the 7 garnet samples analysed from the Czech garnet peridotite masses is given in Table No. 23, and the garnet compositions are also plotted on Fig. 30. It is apparent that these garnets have Ca:Mg:Fe²⁺:Mn ratios of very restricted composition range, and one which is closely similar to that of garnets from the Ugelvik and kimberlitic garnet peridotites. However, the Cr₂O₃ contents of garnets from the Czech garnet peridotites are very variable (0.85-6.85 wt.%) and on the average somewhat lower than those of garnets from the Ugelvik and kimberlitic garnet peridotites. 5 of the 7 garnets have Cr₂O₃ contents below the range reported from garnet peridotite inclusions in kimberlite 1.9-7.5% (Nixon et al 1963, O'Hara and Mercy 1963) and that for the Ugelvik garnet peridotites (2.30-3.96 wt.%).

Following the demonstration by Kopesky and Sattran (1962) that in the Jihlava diatreme in Czechoslovakia the source of garnet peridotite xenoliths is a layer or lens of 'crustal' garnet peridotite within the gneiss complex penetrated at shallow depth by the diatreme, Davidson/
Davidson (1964) has claimed that all the xenoliths found in kimberlite pipes (including all the garnet peridotite and eclogite nodules) have been derived from crustal metamorphic formations.

However, O'Hara and Mercy (1965 & 1966) have refuted this claim pointing out that the lower Cr$_2$O$_3$ contents of the garnets and the lower OPX/Ol ratios of the Czech, Kalskaret and Almåkløvdalen 'crustal' garnet peridotites and the higher GNT and CPX contents, Fe/Fe$^2+$Mg ratios etc. of the Kalskaret and Almåkløvdalen garnet peridotites rule these Czech and Norwegian crustal occurrences out as potential sources of the garnet peridotite nodules such as are found in the South African kimberlite pipes (O'Hara and Mercy 1963).

O'Hara and Mercy (1966) report Cr$_2$O$_3$ contents in garnets of 0.62 wt.% from the Linkorka 'crustal' garnet peridotite mass and 0.56 and 1.57 wt.% from nodules in the diatreme itself, all these values being outside the range reported in nodules from the South African kimberlites. 4 of the garnets analysed by Fiala (1965) are from the same 'crustal' garnet peridotite mass and have Cr$_2$O$_3$ contents of 0.85, 0.95, 1.79 and 6.85 wt.%.

The extremely high Cr$_2$O$_3$ garnet is from an assemblage containing only 0.2% CPX and 0.3% OPX together with 2.5% GNT.

Consideration of the garnet composition data for the Norwegian, Czech and Kimberlite garnet peridotite occurrences indicates that the Cr$_2$O$_3$ contents of the garnets are considerably more variable than the Ca:Mg:Fe$^2+$Mn ratios. The Cr$_2$O$_3$ content of the garnets from all these localities is apparently closely related to the total Al$_2$O$_3$ content, which in turn controls the modal proportions of garnet in the assemblage. Hence garnets from peridotites with lowest garnet contents have highest Cr$_2$O$_3$ contents.
It therefore appears that one must beware basing conclusions on the \( \text{Cr}_2\text{O}_3 \) contents of garnets until the whole range of \( \text{Cr}_2\text{O}_3 \) contents in garnets from any particular locality has been determined. The Ca:Mg:Fe\(^{2+}\):Mn ratios of the garnets are far less variable and provide a much more reliable composition index for distinction between different groups of garnet peridotites.

Nevertheless, despite the variability in \( \text{Cr}_2\text{O}_3 \) contents, it is apparent that the garnets from the Ugelvik and kimberlite peridotites all have higher \( \text{Cr}_2\text{O}_3 \) contents than those from the Kalskaret and Almkløvdalen peridotites and than most of the garnets from the Czech peridotites.

The garnet composition data presented in Table No.23 is considered to support the grouping of the Ugelvik and kimberlite garnet peridotites together on mineralogical and chemical grounds (see also Table No.6) as possible representatives of primary upper mantle material, while the Kalskaret and Almkløvdalen garnet peridotites are mineralogically and chemically distinct and are considered on other evidence to represent upper mantle material modified by partial melting processes. The apparent lower OPX/OI ratios and absence of garnet harzburgites characteristic of the Southern Africa kimberlite peridotite nodules (O'Hara - personal communication) from the Ugelvik garnet peridotites may, however, represent a valid distinction between the two occurrences. In the case of the Ugelvik garnet peridotites the OPX/OI ratio of the primary assemblages has, however, been obscured in most instances by the high degree of serpentinization.

On the basis of the similarity in Ca:Mg:Fe\(^{2+}\):Mn ratios of the garnets from the Czech garnet peridotites (see Table No. 23 and Fig.30) with/
with those from the Ugelvik and Kimberlite garnet peridotites and the
general mineralogical and chemical similarity between the Czech and the
Ugelvik garnet peridotites, the former may also be representative of
upper mantle material although it is perhaps presumptive to make such
a conclusion on the basis of the sparse data at present available for
the Czech occurrences.

The distinction drawn by O'Hara and Mercy (1963, 1966) between
garnet peridotites which occur as Kimberlite xenoliths, and are considered
possibly to be representative of primary mantle material, and 'crustal'
garnet peridotites has become obscured by the discovery of the Ugelvik
garnet peridotites. The true distinction to be made between the various
types of garnet peridotites observed at the earth's surface may rather be
between primary upper mantle material and modified upper mantle material,
irrespective of the environment in which they now occur.

Garnets from the Kalskaret and Almklovdalen garnet peridotites
with low modal garnet contents have not been analysed by O'Hara and Mercy
(1963) nor in the course of this study. This has been the result of the
initial concentration of interest on the more 'eclogitic' rock types exposed
at these localities. However, significant amounts of garnet poor peridotites
do occur among the Almklovdalen garnet peridotites, especially in the Lien mass
where their volume is considerably greater than those of the more eclogitic
rock types, which have been extensively analysed and sampled. Minor amounts
of garnet poor peridotite also occur in the more restricted Kalskaret garnet
peridotite outcrop. Mercy and O'Hara (1965a) have given the analysis of a
Kalskaret garnet peridotite (N21) with only 10% by volume of garnet. This
rock type has a comparable composition to the Ugelvik garnet peridotites,
but unfortunately the garnet composition from this rock is not yet available.

It/
It is therefore conceivable that the garnet composition of this rock and the bulk rock and garnet compositions of other Kalskaret and Almklovdalen garnet poor peridotites are much more akin to the inferred primary mantle peridotite material as discussed above than to the more 'eclogitic' Kalskaret and Almklovdalen garnet peridotites.

Although it has been suggested above that the Ugelvik garnet peridotites represent primary mantle peridotite material it is not denied that they at least have been modified by subsequent low temperature recrystallization still under eclogite facies conditions, by partial metamorphic retrogression under first granulite facies and then amphibolite facies conditions, and by serpentinization. However, it is considered that they have not suffered the appreciable compositional modifications which appear to have occurred in the Kalskaret and Almklovdalen garnet peridotites considered as most likely the result of involvement in partial melting processes at their site of origin in the upper mantle.

F. THE CHONDRITE MODEL OF LIKELY MANTLE COMPOSITIONS

As it has been suggested in the last section that the Ugelvik garnet peridotites, together with the kimberlite garnet peridotites, represent little modified mantle peridotites it is important to consider whether their bulk compositions are compatible with likely mantle compositions suggested by the chondrite meteorite model.

In a series of papers, Ringwood has discussed the chemical evolution of the earth in terms of the chondrite meteorite model and has strengthened the case for the acceptance of such a model (see Ringwood 1962c, 1966 in particular).
Ringwood (1966) has compared the average abundances of elements (relative to silicon) in the different groups of chondrite meteorites with their average abundances in Type I carbonaceous chondrites (Orgueil and Ivuna meteorites). He showed that there is a relative depletion of many elements in all groups of chondrites compared with Type I carbonaceous chondrites, and that it would be possible to obtain the compositions of enstatite chondrites, ordinary chondrites and Types II and III carbonaceous chondrites solely from the removal (by suitable chemical fractionation processes) of the appropriate amounts of trace and minor elements from Type I carbonaceous chondrites.

It is also considered extremely significant that the abundance in Type I carbonaceous chondrites of all but the most volatile elements are in reasonable agreement with the estimated primordial or cosmic abundance (Anders 1964).

This fact, supported by their highly oxidised state, high content of water, carbonaceous compounds and other volatiles, suggest that these particular chondrites have had at the most only a mild thermal and metamorphic history, and hence that they represent the closest approach to the primordial dust of the solar nebula from which the terrestrial planets accreted, that we possess. However, Birch (1965) has stressed limitations in the chondrite model for the earth claiming that such meteorites are samples of an unsuccessful planet, with its own complex and different history and as such only provide a kind of fossilised chemistry suggesting early stages of chemical and mineralogical differentiation of the earth. However, he still appears to accept that carbonaceous chondrites, such as Orgueil, represent the best available composition guide to the undifferentiated primitive earth material.

Considering/
Considering the possibility of the primordial abundances of 'non volatile' elements as obtained from the composition of Type 1 carbonaceous chondrites yielding an earth model with the correct core/mantle ratio (according to the Bullen Model A, the mantle + crust contains 69% of the earth's mass, and the core about 31%) and acceptable compositions for both these major phases, Macdonald (1959) showed that such a model necessitated the presence of 10-20% of elemental silicon in the earth's core. In his model the mantle was considered as largely made up of \((Mg, Fe)_2 SiO_4\) either as olivine or as some high pressure polymorph with \(Mg/Fe = 9\). It was found that according to the chondrite model after the reduction of iron oxide to provide sufficient metal phase to form the core, the remaining oxide phases gave rise to an approximate 1:1 ratio of \(Si:Mg, Fe\) and \(Ca\). This would imply a silicate mantle composed almost entirely of pyroxene or polymorphs of the pyroxene composition, which is incompatible with the physical properties of the upper mantle (Birch 1952). The most likely explanation of this anomaly is that substantial amounts of silicon have been reduced and have entered the core of the earth.

Furthermore studies of the physical properties of the earth's core by Birch (1952) have indicated that it is about 10-20% less dense than Ni-Fe and that its seismic velocity is substantially greater than that of Ni-Fe under comparable pressure-temperature conditions.

The presence of substantial quantities of silicon in the metal phase of enstatite chondrites shows that chemical conditions during the formation of the solar system were favourable for the reduction of silicates, at least in certain regions (Ringwood 1961).

Macdonald (1959) has strengthened support for the chondrite model of/
of the earth's composition by showing that the abundance of the non-volatile elements in the solar atmosphere is similar (in all cases within a factor of two) to their abundance in chondrites and hence that the latter provide a good sample of the non-volatile element composition of the solar system.

In Table No.25 below, column 1 gives the average composition of the analysed Ugelvik peridotites (excepting U6), column 2 the mantle composition derived by Ringwood (1966) from the chondrite meteorite model, and column 3 the hypothetical mantle pyrolite composition derived by Ringwood (1965) on the assumption that the primary mantle peridotite composition must be capable of yielding a basaltic composition partial melt and a dunitic or peridotitic residual.

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From the likely mantle compositions (columns 2 and 3) derived/
derived from the chondrite and pyrolite models respectively, it is apparent that the mantle is expected to be composed of peridotite with low but significant amounts of CaO, Al₂O₃ and alkalies. In the light of the experimental petrology studies reviewed in Chapter III it is to be expected that the upper mantle would be composed for the most part of eclogite facies mineral assemblages. It would therefore appear that the likely mantle rock type would be a garnet peridotite with low modal contents of garnet and clinopyroxene.

The Ugelvik garnet peridotites (Table No. 25 Column 1) represent just such a rock type as do the garnet peridotites which occur as xenoliths in kimberlite pipes and those which occur in the Czech Massif (Fiala 1965).

It is not suggested that there is perfect chemical similarity between the Ugelvik, and the chondrite and pyrolite model compositions. There are in fact quite significant discrepancies in Fe/Fe+Mg ratio, CaO, Al₂O₃ and alkali contents between the compositions. However, the compositions derived from the chondrite and pyrolite models are hypothetical and can only be expected to give an indication of the approximate mantle composition. The significance of these compositions lies in the fact that combined with petrological evidence they indicate that the likely primary mantle material is expected to be a garnet peridotite with only low modal proportions of GNT and CFX.

Perhaps the most disturbing discrepancy from the chondrite model mantle composition lies in the low alkali contents of the Ugelvik peridotites, which appear to raise a major problem concerning the generation of basaltic liquids by partial fusion of such peridotite material.

Tholeiitic basalts, the common basaltic type lowest in alkalies contain
contain about 2% Na$_2$O and 0.7% K$_2$O, while alkali basalts contain considerably greater concentrations.

It would require 10 to 20 times concentration at the very least to generate the alkali content of even the low alkali tholeiitic basalt magmas from the partial melting of peridotite such as the Ugelvik peridotites. In other words if every trace of Na$_2$O and K$_2$O in the peridotite were somehow concentrated in the partial melt magma, that magma would comprise only 5-10% of the source rock. This may not, however, represent such an impossibility as it perhaps first appears. It does not imply that all the alkali content of the mantle peridotite was somehow taken up in a partial melt fraction which comprised only 5-10% of the source rock at the site of generation. The initial partial melt may, however, have comprised a much greater proportion of the source rock but have been fractionated during transit to the surface leaving behind peridotite accumulates in the particular pressure and temperature regimes through which it has passed (O'Hara 1965). In such a fractionation process the alkali content would be expected to be concentrated in the extruded 'residual' - namely the basalt magma extruded at the surface, which may by then in fact only represent a limited proportion (5-10%) of the original mantle source rock.

In the search for peridotites with likely mantle compositions one continually runs up against this problem of low alkali contents apparently precluding the material as a suitable source rock for the generation of a basaltic composition partial melt.

Hamilton and Mournjoy (1965) have discussed the low alkali contents of alpine-type peridotites, which they consider preclude such material as being representative of upper mantle material, from which basalt magma might be/
be derived. However, the alkali contents of these peridotites are appreciably lower than those of the Ugelvik peridotites. Hamilton and Mountjoy note median values of 0.004% Na$_2$O and 0.003% K$_2$O for 40 analysed alpine-type peridotite specimens. They suggest that such peridotites have originated either as magmatic crystal precipitates or as mantle residues left after fusion and removal of basalt magmas, a conclusion which is fully endorsed by this author.

Reliable analyses of alkali contents in fresh peridotite nodules from kimberlites are exceedingly scarce. Probably the best available are those quoted by Nixon et al. (1963) for two peridotite nodules from Basutoland kimberlites. They note Na$_2$O contents of 0.24% and 0.18% and K$_2$O contents of 0.00% and 0.06%.

The low Na$_2$O and K$_2$O contents of the constituent minerals from garnet peridotite nodules in kimberlites (O'Hara and Mercy 1963) suggest similarly low bulk rock Na$_2$O and K$_2$O contents. Recent analyses of such garnet peridotites show Na$_2$O contents in the range 0.10-0.34% (O'Hara and Mercy, personal communication).

It therefore appears that the Na$_2$O and K$_2$O contents of garnet peridotite nodules from kimberlites are of the same order as those found in the Ugelvik peridotites.

This author knows of no analyses of peridotites of suitable primary upper mantle mineralogy and composition with significantly higher alkali contents than the Ugelvik and kimberlite garnet peridotites.

It therefore appears likely that mantle peridotites have in fact low alkali contents probably averaging about 0.20-0.25% Na$_2$O and about 0.02-0.06% K$_2$O, the alkali content having been drastically concentrated in the basaltic partial melt fraction by the time such magmas are/
are extruded at the surface. It is thought that basically two processes may have been involved in producing the remarkable concentration of alkalies in basaltic magmas compared with the likely alkali contents of the source rocks. Firstly, partial melting in the mantle may have involved a 'sweating out' of the low melting fraction of the source rocks, in particular of alkalies. Secondly, the occurrence during transit to the surface of such partial melts of such fractionation processes as outlined by O'Hara (1965) would be expected to result in further appreciable concentration of alkalies in the extruded basalt magmas.

It is therefore concluded that the relatively low alkali contents of the Ugelvik peridotites do not preclude such material as possible mantle peridotite material. It is considered that these peridotites together with the garnet peridotite nodules in kimberlites and may well represent relatively unmodified mantle peridotites or at least the nearest approach to such material exposed at the earth's surface.

G. SUMMARY OF CONCLUSIONS

In Chapter I, section D the various possible origins which might explain any random body of garnet peridotite exposed at the earth's surface were outlined. It is worth reconsidering these basic possibilities here in terms of the field, mineralogical and chemical evidence outlined in the subsequent chapters for the various garnet peridotite occurrences studied.

1. Crystallization in situ to an eclogite facies assemblage.

Such an origin can be rejected for all the Norwegian garnet peridotite masses studied on the following grounds.
a. Recognition of the fact that the garnet peridotites belong to
the eclogite mineral facies, while the regional gneiss complex in
which they occur is mostly of amphibolite but locally of granulite
facies grade.

b. The lack of any recognizable high temperature thermal aureole
around these peridotite masses.

c. Cataclastic textures, preferred orientation fabrics and the sheared
textures of their contacts all point to the tectonic emplacement of these
masses in at least a semi-crystalline state.

d. The pressure and temperature stability field of garnet peridotites as
outlined by experimental petrology studies is such that it is restricted
for the most part to depths within the mantle but such mineral assemblages
may just have been formed under conditions of intense regional metamorphism
in the lower regions of the crust.

As the peridotites with enclosed garnet peridotite layers are
for the most part remarkably fresh with only minor secondary alteration and
serpentinitization, at least in the Kalskaret and Almklovaldalen areas, it
is apparent that they must have been emplaced late on in the metamorphic
and injection history of the gneiss complex. It is therefore exceedingly
unlikely that they could have crystallized to an eclogite facies assemblage
in such an environment where the gneiss complex had suffered considerable
injection and was predominately of amphibolite facies grade.

e. In the case of the Kalskaret and Almklovaldalen garnet peridotites it
is considered that the large extent of the mineral composition variations
renders it unlikely that the layering observed at these localities
represents simply local segregation of garnet and clinopyroxene within
the peridotites precipitated by normal igneous crystallization processes.
2. Crystallization in situ to another mineral facies assemblage, but subsequently metamorphosed to an eclogite facies assemblage.

Such an origin can be similarly rejected on the first four grounds as noted for the first hypothesis.

Furthermore although this hypothesis eliminates the objection (e) to the first hypothesis, for the case of the Kalskaret and Almklovdalen garnet peridotites, attempts to explain the mineralogy and range of compositions of all the garnet peridotites in terms of the isochemical eclogite facies metamorphism of any other possible peridotite mineral assemblages have proved unsuccessful. It might be invoked that metamorphism was not isochemical but involved chemical diffusion processes, but such an explanation is contrary to the evidence of strong linear composition trends demonstrated among the Kalskaret and Almklovdalen garnet peridotites.

3. Crystallization at depth within the crust or mantle to another mineral facies assemblage, then subsequently both metamorphosed to an eclogite facies assemblage and tectonically emplaced into their present position.

Such an origin is in agreement with the field and textural evidence but is rejected because it fails to explain firstly the composition variations which have produced the mineral layering and also the linear composition trends as noted under hypothesis 2.

4. Crystallization at depth within the crust or mantle to an eclogite facies assemblage, with subsequent tectonic emplacement into their present position.

With/
With the rejection of the other three basic hypotheses, this origin is the one favoured to explain the Norwegian garnet peridotite occurrences studied. The garnet peridotite stability field outlined by experimental petrology studies, indicates the likelihood that original eclogite facies crystallization would have occurred within the mantle rather than in the lower regions of the crust. Such an origin as envisaged here is in agreement with the field occurrence and textural features of these peridotite masses.

It has been demonstrated, however, in the course of this thesis that two variations on this basic mode of origin are necessary to explain the mineral and bulk rock chemistry of the Kalaskaret and Almklovdalen garnet peridotites on one hand, and the Ugelvik garnet peridotites on the other.

a. The Kalaskaret and Almklovdalen garnet peridotites.

The mineral layering at these localities cannot be explained by the simple sorting of an eclogite facies mineral assemblage precipitated by igneous crystallization processes, whether in situ or at depth, because of the extreme mineral composition variations involved. Furthermore strong linear composition trends have been demonstrated among the rock types from these garnet peridotite localities.

The explanation favoured by this author to account for the composition variations involved is that these peridotites have been involved in partial melting processes at their site of origin in the upper mantle. The garnet free peridotites which predominate at these localities are assumed to represent residuals after the removal of a partial melt fraction, while the/
the garnet and clinopyroxene rich garnet peridotites are considered to represent fossilized attempts at the production of such partial melts which have not been carried to completion. The partial melt phase in these instances has not been completely filter pressed out and removed but instead has been trapped and crystallized within the residual peridotite. There are, however, some Kalskaret and Almklovdalen garnet peridotites with only low modal contents of garnet and clinopyroxene, particularly from the Lien locality, which may represent original mantle peridotite material which has escaped involvement in such partial melting processes.

The alternative hypothesis that the linear composition variations among these garnet peridotites have resulted from the igneous crystallization at high levels in the mantle of magma (generated in the mantle) to give olivine cumulates with the residual liquid imperfectly filter pressed out and crystallized in pockets within the peridotite, cannot be conclusively eliminated on the evidence available at present but is considered on the whole less likely. With either of these two crystal-liquid processes it may, however, be necessary to invoke subsequent limited mechanical mixing during tectonic emplacement to account for certain of the intermediate compositions on the linear composition trend.

b. The Ugelvik garnet peridotites.

These peridotites do not show the extensive mineral and bulk rock chemical variations present among the rocks from the above two localities. The mineral layering within these peridotites may therefore be adequately explained by minor sorting of an eclogite facies assemblage precipitated by igneous crystallization processes. There are, in fact, two possibilities regarding/
regarding this igneous crystallization. It may have occurred following the mantle melting differentiation process considered by many (e.g. Ringwood 1966) as likely to have followed the terrestrial accretion process, or alternatively it may have occurred through the crystallization of isolated bodies of magma (produced from the mantle) in the upper levels of the mantle. In either case these peridotites are considered to have been derived from the mantle, but in the former case they would represent more primitive mantle material.

It is considered that these garnet peridotites, together with the garnet peridotite nodules in kimberlites, are mineralogically and chemically suitable as primary upper mantle peridotite representatives in the light of their broad mineralogical and chemical uniformity, and their compatibility with likely mantle compositions suggested by the chondrite meteorite model and with likely upper mantle mineralogy inferred from experimental high pressure and temperature petrology studies.

Although it is considered that the Ugelvik garnet peridotites may fundamentally represent primary upper mantle material and the Kalskaret and Almklovdalen garnet peridotites similar material modified by involvement in partial melting processes, it is also recognized that all these rock types have been subsequently modified by recrystallization and equilibration, still under eclogite facies conditions, at relatively low temperature (600-700°C) during their tectonic emplacement into their present environment, and also by partial metamorphic retrogression and in certain cases serpentinization.
Although an investigation of the eclogites which occur as inclusions within the gneiss complex of the Sunnmøre district of Norway was not the primary aim of this research project, a certain amount of field and chemical evidence has nevertheless been brought to light that is relevant to speculations on the origin of these rock masses. This evidence centres round the discovery of 'country rock' type eclogites within the peridotite masses at Kaldhussaeter, Tafjord and Hormindal, Nordfjord. These occurrences are apparently contrary to the statement by O'Hara and Mercy (1963) that ....'Field and petrological evidence show that in the Norwegian province there is no relationship in space or time between the peridotite masses and the small eclogite bodies in the layered gneiss series.'

Experimental evidence (Yoder and Tilley 1961, 1962) indicates that the characteristic minerals of eclogites are not likely to be stable at the liquidus of basalt at likely crustal pressures. Acceptance of this evidence, together with the lack of evidence of contact metamorphic or chill features, rules out a direct igneous crystallization of these eclogites in their present environment as envisaged by Eskola (1921). Recognition of the high pressure nature of the eclogite facies assemblages, the fact that it is at variance with the mineral facies of the enclosing gneisses, the boudin like form of the eclogite masses, and the partial retrogression of the mineral assemblages under amphibolite facies conditions are considered to be contrary to the hypotheses of Gjelsvik (1952), Hernes (1953), Kolderup (1960) and Schmitt (1964) that the eclogites were formed by a metamorphic origin in their present position during the Caledonian orogeny. On the other hand it appears likely
likely that the eclogite masses have been tectonically emplaced into their present environment in the gneiss complex as crystalline rocks of either metamorphic or igneous crystallization at greater depths (c.f. Lappin 1962, O'Hara and Mercy 1963).

The bulk rock analyses and C.I.P.W. norms of 7 eclogite specimens are given in Table No. 26. T205, T238 and T271 are eclogite specimens from inclusions within the Kaldhussaeter peridotite mass; T223 represents a quartzo-felspathic segregation within the main eclogite horizon in this peridotite mass; H4 is a sample of the eclogite mass within the Hornindal peridotite; and T212 and H2 are specimens from eclogite boudins directly enclosed within the gneisses at Kaldhussaeter and Hornindal, respectively. The field occurrences of these rock samples has already been described in detail in Chapter II.

These eclogites range from quartz normative to nepheline normative compositions, but all have appreciably lower Cr$_2$O$_3$ and NiO contents and higher Na$_2$O, Al$_2$O$_3$, P$_2$O$_5$ and TiO$_2$ contents and Fe/Fe+Mg ratios than any of the analysed garnet peridotites. The analysis of garnet from T271 is given in Table No. 3. It can be seen to be appreciably more calciferous and to have a higher Fe/Fe+Mg ratio than any of the garnets from the garnet peridotites. The composition of this garnet and the bulk rock are quite comparable to those of the Glenelg eclogite (A.L36 Alderman 1936, 35090 Yoder and Tilley 1962).

All the analysed eclogites contain garnet, pyroxene symplektite - amphibolitized to varying degrees, amphibole and minor amounts of rutile and an opaque ore mineral. Specimens T212, T223, T271 and H4 all contain modal quartz. Mica and clinzoisite occur as accessory minerals in certain of/
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<td>0.67</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Fe</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>7.11</td>
<td>-</td>
<td>-</td>
<td>6.01</td>
</tr>
<tr>
<td>Fa</td>
<td>5.35</td>
<td>2.70</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>3.28</td>
<td>4.54</td>
<td></td>
</tr>
<tr>
<td>Mg</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>4.49</td>
<td>5.22</td>
<td>6.50</td>
<td>6.38</td>
<td>3.89</td>
</tr>
<tr>
<td>Tm</td>
<td>1.54</td>
<td>2.01</td>
<td>1.35</td>
<td>1.79</td>
<td>1.33</td>
<td>1.58</td>
<td>1.62</td>
<td></td>
</tr>
<tr>
<td>Cr</td>
<td>0.07</td>
<td>0.03</td>
<td>0.07</td>
<td>0.05</td>
<td>-</td>
<td>0.06</td>
<td>0.02</td>
<td></td>
</tr>
<tr>
<td>Ap</td>
<td>0.40</td>
<td>0.37</td>
<td>0.14</td>
<td>0.21</td>
<td>0.28</td>
<td>0.07</td>
<td>0.11</td>
<td></td>
</tr>
<tr>
<td>Rest</td>
<td>0.70</td>
<td>1.01</td>
<td>0.74</td>
<td>0.54</td>
<td>0.72</td>
<td>0.85</td>
<td>0.53</td>
<td></td>
</tr>
<tr>
<td>TOTAL</td>
<td>100.56</td>
<td>100.42</td>
<td>99.89</td>
<td>100.38</td>
<td>99.06</td>
<td>99.43</td>
<td>99.42</td>
<td></td>
</tr>
<tr>
<td>Fe$_T$ / Fe$_T +$ Mg</td>
<td>0.26</td>
<td>0.34</td>
<td>0.21</td>
<td>0.24</td>
<td>0.29</td>
<td>0.27</td>
<td>0.29</td>
<td></td>
</tr>
<tr>
<td>Ab/ Ab+An</td>
<td>0.52</td>
<td>0.31</td>
<td>0.40</td>
<td>0.50</td>
<td>0.31</td>
<td>0.48</td>
<td>0.52</td>
<td></td>
</tr>
</tbody>
</table>

* Wet Methods Analyses All others X-ray Fluorescence Determinations.

Fe$_T$ / Fe$_T +$ Mg Ratio in Normative Ferromagnesian Silicates
of the specimens, while kyanite has not been noted in any of them.

The textures of all these specimens with the omphacitic clinopyroxene having been altered firstly to a diopsidic clinopyroxene + plagioclase symplektite and secondly to an amphibole + plagioclase intergrowth, and kelyphitic reaction rims of amphibole and plagioclase around the garnets are typical of partially retrogressed eclogites (c.f. Alderman 1936). However, the more extensive the amphibolitization the more obscure this texture becomes.

All these eclogites are far more extensively amphibolitized than those of the Almklovdalen and Selje districts studied by Lappin (1962). Lappin noted the following apparent abundances of the different eclogite types in the gneisses.

<table>
<thead>
<tr>
<th>Type</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>GNT + CPX + Quartz</td>
<td>very common</td>
</tr>
<tr>
<td>GNT + CPX</td>
<td>common</td>
</tr>
<tr>
<td>GNT + CPX + Quartz + Kyanite</td>
<td>fairly common</td>
</tr>
<tr>
<td>GNT + CPX + Opx</td>
<td>rather rare</td>
</tr>
<tr>
<td>GNT + CPX + Amphibole</td>
<td>rare</td>
</tr>
<tr>
<td>GNT + CPX + Quartz + Zoisite</td>
<td>rare</td>
</tr>
<tr>
<td>GNT + CPX + Kyanite + Amphibole</td>
<td>rare</td>
</tr>
</tbody>
</table>

Amphibole is considered to occur as an accessory in many of these.

The specimens analysed by Lappin also show a considerable range of compositions but most are olivine tholeiites in composition, while only two are quartz normative and one nepheline normative. However, it should be borne in mind that many of these bulk rock compositions were calculated from modes and mineral analyses.

Mineral analyses by Lappin indicate that while the garnets from the/
the garnet peridotites are pyrope rich and have very restricted grossular contents, those from the eclogites in the gneisses are more almandine rich and have much more variable and often higher grossular contents (see Fig. 31). Furthermore the clinopyroxenes from the eclogites commonly have higher jadeite contents than those in the garnet peridotites (see Fig. 33).

It can be stated that broadly speaking the same bulk rock and mineral composition variations exist between the eclogites in the gneisses and the postulated 'primary' mantle peridotites among the Norwegian gneisses (namely those at Ugelvik and possibly also the garnet poor peridotites from Kalskaret and Almklovdaalen) as exists between the eclogites and the garnet peridotites which occur as xenoliths in kimberlite pipes (compare Figs. 20 and 29-34). Furthermore as pointed out in the last chapter the bulk rock and mineral compositions of the garnet rich Kalskaret and Almklovdaalen peridotites define a composition trend between the eclogites and the 'primary' mantle garnet peridotites which it was postulated has resulted from their involvement in partial melting processes at depth in the mantle prior to their emplacement.

It is therefore interesting to speculate as to whether it is possible that the eclogites in the gneisses represent partial melt products of mantle garnet peridotites. Proof of such an origin would further substantiate the origin suggested to explain the composition variations among the Kalskaret and Almklovdaalen garnet peridotites.

O'Hara and Mercy (1963) in fact stated that the most satisfactory hypothesis is that these eclogites represent partial fusion products of mantle peridotites or differentiates of such liquids. However, O'Hara and Yoder (1963) considered that ... these eclogites have probably formed by the/
Figure 34. \( \text{Ca}^{2+}:\text{Mg}^{2+}:\text{Fe}^{2+}+\text{Fe}^{3+} \) plot of eclogites from the Norwegian gneisses.
FIG. 34

- SELJE AND ALMKLOV.
- TAFJORD AND HORNINDAL

CA$^{2+}$

MG$^{2+}$

$\rightarrow$ Fe$^{2+}$ + Fe$^{3+}$
metamorphic recrystallization of deep-seated gabbros, or crustal basic igneous rocks carried down in orogenic movements... and no longer considered them as ...'possible representatives of the partial fusion product at great depth because of the probability that the liquids from which they are derived have undergone differentiation under conditions in which these eclogite facies minerals are not stable in the presence of the liquid.'

Experiments on the melting of garnet peridotite and eclogite compositions at 30 kbs. pressure (O'Hara 1963a,b) have raised a major problem in the way of the derivation of quartz eclogites as derivatives of partial melts of garnet peridotites at such pressures. It was found that partial melting of garnet peridotite or bimineralic eclogite at pressures of 30 kbs, yields liquids which are far removed from SiO₂ saturated tholeiitic basalt (equivalent to quartz eclogite) in composition, and can only resemble picritic basalts.

Temperatures of beginning of melting of GNT + CPX and GNT + CPX + OPX assemblages (1515-1565°C) are higher than the temperatures at which these same assemblages begin to melt when olivine or quartz is added to the assemblage (1425-1450°C and 1350-1375°C respectively). It therefore appears that the GNT - CPX - OPX plane (Fig. 28) is a thermal barrier at 30 kbs. pressure and therefore that quartz-bearing eclogite, or a liquid of that composition, cannot be derived as a differentiate of a partial melt of garnet peridotite at this pressure. O'Hara and Yoder (1963) considered that the relationship at the beginning of melting of garnet peridotite at this pressure is CPX + GNT + Spinel ⇒ Ol + OPX + Liquid, the liquid composition lying within or to the MgO poor side of the subsolidus plane Spinel + CPX + GNT.
GNTss and to the SiO₂ poor side of the CPXss - GNTss plane (Fig. 28).

A metamorphic origin of eclogites, such as those which occur in the Norwegian basal gneiss complex, through the recrystallization of crustal basic igneous rocks carried down in orogenic movements poses the following problems.

1. If the eclogite assemblages are the result of the metamorphism of cool crustal basic igneous rocks, what then was the motive force for their remobilization and intrusion into rocks of lower mineral facies at higher crustal levels? They might have been expected to move upwards under tectonic stresses if they had possessed either a lower specific gravity or a higher temperature than their enclosing country rocks. The former is clearly unlikely as these rocks are among the densest known, while the alternative infers that they had an initial magmatic stage so that a temperature gradient existed between them and their enclosing country rocks and is hence contrary to a metamorphic origin.

2. Although the compositions of these eclogites are broadly speaking basaltic they cannot be matched exactly with high level basaltic types. Their K₂O contents are decidedly low, and their SiO₂ and P₂O₅ contents also rather low in most cases. Their low K₂O and P₂O₅ contents perhaps suggest that they have been derived from more 'primitive' liquids which have not been so extensively fractionated as the high level basalts. Lappin (1962) also notes that most of the eclogites in the gneisses show considerable relative Mg enrichment when compared with basalts of equivalent Ca content, while SiO₂ is rather erratic in relation to MgO which may suggest that fractional crystallization of olivine was not a prime cause of the original compositional variation among these rock types.
3. Both the eclogites in the gneisses and the garnet peridotites have the following features in common.

a. The mineral facies (eclogite) to which they both belong is clearly distinct from the mineral facies (amphibolite) of the gneiss complex in which they occur and to which they show partial retrogression. They are thus both foreign and metastable in their present environment and therefore cannot have been metamorphosed to their present mineralogy there.

b. Neither show any contact metamorphic or chill features to suggest that they represent igneous intrusions into their present environment. Furthermore igneous crystallization to eclogite facies assemblages in such an environment is at variance with the experimental evidence for the likely stable conditions of crystallization of such assemblages.

c. The boudin-like form of the outcrops of these rock types suggests the influence of tectonic forces, as do the textures of the garnet peridotites.

d. For both rock types the only satisfactory explanation of their field relations and mineralogy is that they have been tectonically emplaced into their present environment in a crystalline or at least quasi-crystalline state.

e. A certain spatial relationship does appear to exist between the eclogites and the peridotites as both occur together abundantly in a broad zone in the gneiss complex of the Sunnmøre district, and in at least three cases (Kaldfussaeter, Hornindal and Rodhaugen) eclogites in fact occur as inclusions directly within the peridotites.

If the peridotites with their enclosed garnetiferous layers are accepted as having been derived from the mantle (as concluded in Chapter V) it would appear to represent a rather remarkable coincidence that they are now/
now intimately associated in the field with eclogite bodies of similar mineral facies but which originated, unassociated with the peridotite masses, as metamorphically recrystallized crustal basic igneous rocks. Such an origin for the eclogite lenses in the gneisses therefore appears less satisfactory than one which envisages that they represent either the primary or differentiated partial fusion products of mantle garnet peridotites or crystal cumulates from such liquids.

There is no major barrier in the way of considering the olivine tholeiite composition or even the nepheline normative composition eclogites as either the partial melt liquids of mantle garnet peridotites or the crystal accumulates of such liquids, derived and crystallized at high pressures. However, the experimental evidence which appears to preclude the derivation of quartz eclogites as differentiates of partial melts of garnet peridotites, at least at 30 kbs. pressure, has already been described. They may, however, represent lower pressure differentiates from such liquids but in that case it is very unlikely that they would have crystallized directly to an eclogite facies assemblage, rather have been subsequently recrystallized to such an assemblage.

Chemical characteristics, such as low $K_2O$ and $P_2O_5$ contents, suggest that the eclogites as a group cannot be closely matched with high level crustal basic rock types but rather have been derived from less highly fractionated and hence more 'primitive' liquids than high level basaltic rocks.

This author therefore favours the hypothesis that the eclogites represent a series of differentiates or crystal cumulates (perhaps formed over a considerable pressure range) from liquids which were produced as partial/
partial melts of mantle garnet peridotite. Although in many respects a primary crystallization of the eclogites at high pressures would seem more satisfactory, in view of the depth factor involved and the problem of the genesis of the quartz eclogites at high pressures, it seems necessary to postulate metamorphic recrystallization for the origin of at least some and possibly all the eclogites.

Field evidence from the eclogite lenses enclosed in the gneisses indicates that they were intruded and boudinaged before the main injection and metamorphism of the gneiss complex, while the fresh and relatively unaltered appearance of the peridotites suggest emplacement after the main metamorphic episode. Furthermore, the extensive amphibolitization of the eclogite masses within the Kaldhussaeter and the Hornindal peridotites contrasts with the fresh appearance of the enclosing peridotites, and suggests that the eclogite was exposed to amphibolitization before inclusion within the peridotite.

However, the existence of a time gap between the intrusion of the eclogites and that of the peridotites is not considered to be contrary to the hypothesis of a genetic relationship between them.

A consideration of the field relations, the likely stable conditions of formation, and the chemistry of the eclogites in the gneisses indicates that the metamorphosed Sunnmøre dolerites, with CNT + CPX + Plagioclase assemblages, cannot be considered as rocks transitional into eclogites, as thought by Gjelsvik (1952), but represent stable SiO₂ deficient granulite facies assemblages.

The following significant chemical distinctions exist between the eclogites and the metamorphosed dolerites described by Gjelsvik (1952).
1. The eclogites have considerably lower $K_2O$ contents.

<table>
<thead>
<tr>
<th></th>
<th>Range</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fresh eclogites in the gneisses</td>
<td>0.02 - 0.39 wt.%</td>
<td>0.14 wt.%</td>
</tr>
<tr>
<td>(Table No. 26, also Lappin 1962)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sunnmøre dolerites</td>
<td>0.33 - 1.00 wt.%</td>
<td>0.70 wt.%</td>
</tr>
<tr>
<td>(Gjelsvik 1952)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

2. The eclogites also have somewhat lower $P_2O_5$ contents.

<table>
<thead>
<tr>
<th></th>
<th>Range</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fresh eclogites in the gneisses</td>
<td>0.03 - 0.16 wt.%</td>
<td>0.07 wt.%</td>
</tr>
<tr>
<td>Sunnmøre dolerites</td>
<td>0.04 - 0.35 wt.%</td>
<td>0.25 wt.%</td>
</tr>
</tbody>
</table>

3. They also have rather lower $TiO_2$ contents.

<table>
<thead>
<tr>
<th></th>
<th>Range</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fresh eclogites in gneisses</td>
<td>0.25 - 1.06 wt.%</td>
<td>0.72 wt.%</td>
</tr>
<tr>
<td>Sunnmøre dolerites</td>
<td>0.95 - 2.23 wt.%</td>
<td>1.74 wt.%</td>
</tr>
</tbody>
</table>

4. All the eclogites analysed here have higher Cr contents (138-345 ppm) than the Sunnmøre dolerites (all <100 ppm, but not detected in most instances).

5. The dolerites are generally less magnesian than the eclogites (see Fig. 36) particularly in their picritic varieties.

6. Among the eclogites, Lappin (1962) notes that $SiO_2$ is rather erratic in relation to $MgO$ which he suggest may indicate that fractional crystallization of olivine was not the prime cause of the original composition variation.

7. There is also a different distribution of $Ca^{2+}:Na^{2+}:Al^{3+}$ in the eclogites compared with the dolerites (see Fig. 35). This probably results from the fact that the $Ca^{2+}:Na^{2+}/Al^{3+}$ ratio was controlled by plagioclase crystallization in the dolerites but not in the eclogites.

8. None of the dolerites analysed by Gjelsvik have quartz in the C.I.P.W. norm, as do several of the eclogites, and almost all have lower normative CPX contents than the eclogites.

It is therefore concluded that the hypothesis that the eclogites were/
### Table No. 27

**MISCELLANEOUS AMPHIBOLITES**

<table>
<thead>
<tr>
<th>SAMPLE NOS.</th>
<th>T60</th>
<th>T61</th>
<th>T127</th>
<th>T195</th>
<th>T256</th>
<th>T183</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>SiO₂</strong></td>
<td>47.33</td>
<td>44.27</td>
<td>45.50</td>
<td>46.24</td>
<td>42.01</td>
<td>45.26</td>
</tr>
<tr>
<td><strong>TiO₂</strong></td>
<td>0.30</td>
<td>0.58</td>
<td>1.06</td>
<td>0.60</td>
<td>0.53</td>
<td>0.50</td>
</tr>
<tr>
<td><strong>Al₂O₃</strong></td>
<td>9.21</td>
<td>11.77</td>
<td>16.93</td>
<td>11.40</td>
<td>17.06</td>
<td>17.23</td>
</tr>
<tr>
<td><strong>Cr₂O₃</strong></td>
<td>0.09</td>
<td>0.04</td>
<td>0.01</td>
<td>0.04</td>
<td>0.04</td>
<td>0.03</td>
</tr>
<tr>
<td><strong>Fe₂O₃</strong></td>
<td>3.11</td>
<td>5.15</td>
<td>4.42</td>
<td>3.25</td>
<td>4.10</td>
<td>2.35</td>
</tr>
<tr>
<td><strong>FeO</strong></td>
<td>6.72</td>
<td>9.24</td>
<td>8.71</td>
<td>8.23</td>
<td>9.93</td>
<td>6.18</td>
</tr>
<tr>
<td><strong>MnO</strong></td>
<td>0.23</td>
<td>0.50</td>
<td>0.22</td>
<td>0.19</td>
<td>0.19</td>
<td>0.14</td>
</tr>
<tr>
<td><strong>NiO</strong></td>
<td>0.04</td>
<td>0.04</td>
<td>&lt;0.01</td>
<td>0.03</td>
<td>0.03</td>
<td>0.05</td>
</tr>
<tr>
<td><strong>MgO</strong></td>
<td>17.23</td>
<td>12.67</td>
<td>9.25</td>
<td>12.00</td>
<td>9.71</td>
<td>13.09</td>
</tr>
<tr>
<td><strong>CaO</strong></td>
<td>12.43</td>
<td>14.02</td>
<td>8.40</td>
<td>14.46</td>
<td>12.54</td>
<td>11.41</td>
</tr>
<tr>
<td><strong>Na₂O</strong></td>
<td>1.73</td>
<td>0.97</td>
<td>1.68</td>
<td>2.28</td>
<td>1.60</td>
<td>1.42</td>
</tr>
<tr>
<td><strong>K₂O</strong></td>
<td>0.30</td>
<td>0.10</td>
<td>1.53</td>
<td>0.10</td>
<td>0.22</td>
<td>0.16</td>
</tr>
<tr>
<td><strong>P₂O₅</strong></td>
<td>0.004</td>
<td>0.07</td>
<td>0.03</td>
<td>0.008</td>
<td>0.007</td>
<td>n.d.</td>
</tr>
<tr>
<td><strong>H₂O</strong></td>
<td>1.78</td>
<td>0.86</td>
<td>1.17</td>
<td>1.46</td>
<td>1.56</td>
<td>1.72</td>
</tr>
<tr>
<td><strong>TOTAL</strong></td>
<td>100.48</td>
<td>100.28</td>
<td>99.31</td>
<td>100.28</td>
<td>99.52</td>
<td>99.54</td>
</tr>
</tbody>
</table>

* Wet Methods Analyses

All Others X-ray Fluorescence Determinations
were formed in situ during the Caledonian orogeny or specifically from the Sunnmøre dolerites cannot be upheld.

In Table No. 27 the analyses of six amphibolites from the Tafjord district are listed. These amphibolites are also plotted on the Ca$^{2+}$:Mg$^{2+}$:Fe$^{2+}$Fe$^{3+}$ plot (Fig. 36) and the Ca$^{2+}$:Na$^{2+}$:Al$^{3+}$ plot (Fig. 35) where their compositions can be compared with those of the Sunnmøre dolerites, the eclogites in the gneisses and the garnet peridotites.

Specimens T30 and T31 are from the restricted sequence of amphibolite interlayering within the Kaldhussaeter peridotite (Fig. 11). T30 is an OPX bearing amphibolite: T31 a garnet amphibolite with relict CPX. On the basis of their field occurrence it was considered (Chapter II) that they probably represent amphibolitized garnet peridotite layers indigenous to the enclosing peridotites. Such a conclusion seems to be borne out by their association with 'eclogitic' garnet peridotite compositions on the Ca$^{2+}$:Na$^{2+}$:Al$^{3+}$ plot, and their rather higher Cr$\text{O}_3$ and NiO contents and lower P$_2$O$_5$ and TiO$_2$ contents compared with the country rock type eclogites, including those which occur as inclusions within the Kaldhussaeter peridotite (Table No. 26). However, the chemical distinction is only slight, as might be expected according to the explanation put forward in this thesis to account for the garnet peridotites and the eclogites, and the composition trend between them.

Specimen T127 was collected from a large, coarse-grained garnet amphibolite mass remote from any of the peridotite masses. It shows no textural evidence (kelyphitic garnet reaction rims or pyroxene symplektite intergrowth alterations) to suggest that it represents a highly retrogressed eclogite. Its composition is typical of the metamorphosed dolerites described by Gjelsvik (see Figs. 35 and 36), having higher TiO$_2$ and K$_2$O contents and lower/
FIG. 35

MOST ECLOGICTIC GARNET PERIDOTITES
FROM KALSKARET

OTHER ORNAMENTS AS FIG. 36
lower Cr$_2$O$_3$ and NiO contents than the eclogites.

Specimen T195 was from a narrow amphibolite lens within one of the Kalskaret peridotite masses. It contains relict CPX much of which has a symplektite alteration texture indicating breakdown into a hornblende + plagioclase intergrowth. However, although it is likely that it represents a highly retrogressed eclogite it is difficult on the basis of composition alone to decide whether it represents an extensively amphibolitized 'eclogitic' garnet peridotite layer indigenous to the peridotite environment in which it occurs, or an amphibolitized tectonically emplaced country rock type eclogite. Its field occurrence results in a preference for the latter alternative.

Specimen T183 was of garnet amphibolite which occurs below the small altered peridotite mass at the western end of the Kalskaret col. Its low K$_2$O and TiO$_2$ contents and significant Cr$_2$O$_3$ and NiO contents together with plagioclase-amphibole symplektite textural evidence indicates the likelihood that it represents a highly retrogressed eclogite.

Specimen T256 was of garnet amphibolite which occurs as an isolated mass adjacent to one of the Kalskaret peridotite masses. It does not contain any CPX and although there is suggestion of hornblende-plagioclase reaction rims round the garnets, its texture does not really resemble that of a partially altered eclogite. However, its composition with low P$_2$O$_5$, K$_2$O and TiO$_2$ contents, and notable Cr$_2$O$_3$ and NiO contents suggests a much closer correlation with the eclogites than with the metamorphosed dolerites. Therefore on compositional grounds (see also Figs. 35 and 36) it appears more likely that this rock also represents a highly retrogressed eclogite rather than a metamorphosed basic igneous rock (e.g. the Sunnmøre dolerites).

This/
This emphasizes the difficulty involved in distinguishing between such rock types on textural grounds alone, a difficulty which has probably contributed largely to confusion in the identification of amphibolitized eclogites as distinct from garnet amphibolites and hence their false correlation by certain previous field workers.
<table>
<thead>
<tr>
<th>CONCLUSION</th>
<th>IN AGREEMENT WITH</th>
<th>IN LIMITED AGREEMENT WITH</th>
<th>IN DISAGREEMENT WITH</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Tectonic emplacement of the peridotite masses: not crystallized in situ</td>
<td>O'Hara and Mercy</td>
<td>Lappin</td>
<td>Eskola</td>
</tr>
<tr>
<td>2. Peridotites affected by E-W axis and N-S axis folding</td>
<td></td>
<td></td>
<td>Lappin</td>
</tr>
<tr>
<td>3. Emplacement of the peridotites late in the metamorphic and injection history of the gneiss complex</td>
<td>Lappin</td>
<td>O'Hara and Mercy</td>
<td>Eskola</td>
</tr>
<tr>
<td>4. Peridotites derived from the mantle (specific origin undefined)</td>
<td>Lappin</td>
<td>O'Hara &amp; Mercy</td>
<td>Eskola</td>
</tr>
<tr>
<td>5. Low temperature equilibration of the present garnet peridotite assemblages.</td>
<td>Lappin</td>
<td>O'Hara &amp; Mercy</td>
<td>Eskola</td>
</tr>
<tr>
<td>6. Certain of the garnet peridotites, in particular those at Ugelvik may represent primary upper mantle material.</td>
<td></td>
<td></td>
<td>Lappin</td>
</tr>
<tr>
<td>8. Chemical relationship between peridotites and eclogites, even if rather indirect.</td>
<td>Eskola</td>
<td>O'Hara and Mercy</td>
<td>Mercy</td>
</tr>
<tr>
<td>9. Close relationship in space between 'country-rock' type eclogites and peridotites; the former occurs as inclusions in latter.</td>
<td></td>
<td></td>
<td>Mercy</td>
</tr>
<tr>
<td>10. Time gap between the intrusion of the eclogites and the peridotites.</td>
<td>Lappin</td>
<td>O'Hara and Mercy</td>
<td>Eskola, Gjelsvik, Hernes</td>
</tr>
<tr>
<td>11. Eclogites tectonically emplaced: not crystallized in present environment.</td>
<td>Lappin</td>
<td>O'Hara and Mercy</td>
<td>Gjelsvik, Schmitt</td>
</tr>
<tr>
<td>12. Eclogites chemically distinct from the Sunnmøre dolerites.</td>
<td></td>
<td></td>
<td>Gjelsvik</td>
</tr>
<tr>
<td>13. Probable metamorphic re-crystallization origin of the eclogite assemblages.</td>
<td>Lappin</td>
<td>O'Hara and Mercy</td>
<td>Eskola</td>
</tr>
</tbody>
</table>
APPENDIX I. SERPENTINIZATION RECALCULATIONS

The serpentinization of peridotites has for long been a controversial subject both as regards the origin of the water producing the reaction and the nature of the reaction involved (see Turner and Verhoogen 1960, p. 316-321). In particular it is uncertain whether the reaction involved is principally a constant volume reaction, involving addition of $\text{H}_2\text{O}$ and leaching of $\text{MgO}, \text{SiO}_2$ etc., or a constant composition reaction, involving the addition of $\text{H}_2\text{O}$ without leaching and hence with a large increase in volume.

In general the degree of serpentinization in the large peridotite masses observed in the course of this study is low, but the high degree of serpentinization of the Ugelvik peridotites necessitates an attempt to recalculate the original water free compositions.

If the serpentinization was principally a constant composition reaction with addition of water and without significant leaching of other components, then the straightforward water free recalculation of the bulk rock analysis should be in reasonably good agreement with the bulk rock composition calculated using the mineral analyses and the estimated modal proportions of the minerals.

This has been tested for the various analysed Ugelvik samples. $\text{U}_{850}/\text{U}_{16}$. The 'residual' $\text{GNT}$ and $\text{Ol}$ from this rock, assumed to represent original mineral compositions of the unserpentinized peridotite, have been analysed. Mineral analyses and optical determinations on minerals from the suite of Ugelvik rock types indicates only minor compositional differences especially in major element content, e.g. $\text{Fe}/\text{Fe+Mg}$ ratio. The composition of the analysed 'residual' $\text{OPX}$ from $\text{U}_8$ was therefore used in the calculation of/
of the bulk analysis of $U_{16}$. Only minor amounts of an opaque ore mineral, considered to be a chrome spinel, occur. The composition of a chrome spinel from serpentinite quoted by Deer, Howie and Zussman Vol.5, Table 11, No.8, was therefore considered to give a reasonable approximation and was hence used in the calculation.

A further problem arises regarding the estimation of the proportions by weight of the constituent minerals. Volume proportions estimated by point counting are given in Table No.4. However, the cataclastic textures, variable grain sizes and partial serpentinization are not conducive to accurate estimations of the modes of these rocks by point counting.

Consideration of the mineral analyses indicates, however, that a reasonably valid mode can be set up solely from a consideration of the bulk chemical composition. Modal contents of GNT and CPX are closely restricted by the present CaO, Al$_2$O$_3$ and Cr$_2$O$_3$ contents (if no leaching of these elements has occurred), with chrome spinel to bring up the Cr$_2$O$_3$ content after first taking into consideration enough GNT to give the Al$_2$O$_3$ content. Furthermore it is apparent that the Ol/OPX ratio is closely restricted by the present NiO content if this is assumed to be that of the original peridotite. This may at first be considered to be suspect when it is realised that the NiO is largely contained in olivine which has been extensively serpentinized. However, it is unlikely that NiO has been introduced, rather it may have been diminished by leaching, so that the present NiO content is considered to at least set a minimum value on the Ol/OPX ratio. The Ol/OPX ratio of NiO consistently averages about 4/1 (see Table No.1; also O'Hara and Mercy 1963 Tables V and VII).

With these considerations in mind the modal proportions estimated by point counting (Table No.4) have where necessary been adjusted for the present/
present purposes.

For the calculation of the bulk composition of $U_16$, the modal proportions by weight of the minerals were considered to be ol 87.5%, GNT 5%, CPX nil, OPX 7%, and chrome spinel 0.5%.

<table>
<thead>
<tr>
<th></th>
<th>1.</th>
<th>2.</th>
<th>3.</th>
<th>4.</th>
<th>5.</th>
</tr>
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<tbody>
<tr>
<td>SiO$_2$</td>
<td>43.26</td>
<td>40.31</td>
<td>42.41</td>
<td>40.51</td>
<td>40.62</td>
</tr>
<tr>
<td>TiO$_2$</td>
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<td>0.04</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
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<td>1.64</td>
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<td>1.67</td>
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<tr>
<td>Cr$_2$O$_3$</td>
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<td>0.32</td>
<td>0.33</td>
<td>0.34</td>
<td>0.34</td>
</tr>
<tr>
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<td>7.37</td>
<td>7.28</td>
<td>7.51</td>
<td>7.25</td>
</tr>
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<td>0.11</td>
<td>0.11</td>
<td>0.11</td>
<td>0.11</td>
</tr>
<tr>
<td>NiO</td>
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<td>0.36</td>
<td>0.36</td>
<td>0.37</td>
<td>0.36</td>
</tr>
<tr>
<td>MgO</td>
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<td>49.61</td>
<td>47.57</td>
<td>49.15</td>
<td>49.27</td>
</tr>
<tr>
<td>CaO</td>
<td>0.42</td>
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<td>0.38</td>
<td>0.40</td>
<td>0.40</td>
</tr>
<tr>
<td>Na$_2$O</td>
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<td>-</td>
<td>0.11</td>
<td>0.11</td>
<td>0.11</td>
</tr>
<tr>
<td>K$_2$O</td>
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<td>-</td>
<td>0.03</td>
<td>0.03</td>
<td>0.03</td>
</tr>
<tr>
<td>TOTAL</td>
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<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
</tr>
</tbody>
</table>

Column 1 gives the recalculated water free composition from the bulk rock analysis.

Column 2 gives the bulk rock composition calculated from the mineral analysis.

From a comparison of these two analyses it is obvious that the MgO/SiO$_2$ ratio of the recalculated water free composition (col.1) is considerably lower than that expected from the composition calculated from the mineral analyses (col.2).

There/

possible origins of this discrepancy.

Column 3 gives the bulk rock composition calculated from the mineral analysis by the following equations:

\[ 2H_2O + Mg_3Si_2O_9 \rightarrow 2H_4Mg_2Si_2O_9 + 4MgO + SiO_2 \]

55 gm. 161 gm. 60 gm.

\[ 2H_2O + Fe_3Si_2O_9 \rightarrow 2H_4Fe_2Si_2O_9 + 4FeO + SiO_2 \]

391 gm. 60 gm.

\[ 2H_2O + Ni_3Si_2O_9 \rightarrow 2H_4Ni_2Si_2O_9 + 4NiO + SiO_2 \]

299 gm. 60 gm.
present purposes.

For the calculation of the bulk composition of $U_{16}$, the modal proportions by weight of the minerals were considered to be 01 37.5%, GNT 5%, CPX nil, OPX 7%, and chrome spinel 0.5%.

<table>
<thead>
<tr>
<th></th>
<th>Column 1</th>
<th>Column 2</th>
<th>Column 3</th>
<th>Column 4</th>
<th>Column 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO$_2$</td>
<td>43.26</td>
<td>40.31</td>
<td>42.41</td>
<td>40.51</td>
<td>40.62</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>0.01</td>
<td>0.04</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>1.56</td>
<td>1.64</td>
<td>1.41</td>
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<td>1.47</td>
</tr>
<tr>
<td>Cr$_2$O$_3$</td>
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<td>0.32</td>
<td>0.33</td>
<td>0.34</td>
<td>0.34</td>
</tr>
<tr>
<td>FeO</td>
<td>6.96</td>
<td>7.37</td>
<td>7.28</td>
<td>7.51</td>
<td>7.28</td>
</tr>
<tr>
<td>MnO</td>
<td>0.12</td>
<td>0.11</td>
<td>0.11</td>
<td>0.11</td>
<td>0.11</td>
</tr>
<tr>
<td>NiO</td>
<td>0.34</td>
<td>0.36</td>
<td>0.36</td>
<td>0.37</td>
<td>0.36</td>
</tr>
<tr>
<td>MgO</td>
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<td>49.61</td>
<td>47.57</td>
<td>49.15</td>
<td>49.27</td>
</tr>
<tr>
<td>CaO</td>
<td>0.42</td>
<td>0.34</td>
<td>0.38</td>
<td>0.40</td>
<td>0.40</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>0.12</td>
<td>-</td>
<td>0.11</td>
<td>0.11</td>
<td>0.11</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>0.03</td>
<td>-</td>
<td>0.03</td>
<td>0.03</td>
<td>0.03</td>
</tr>
<tr>
<td>TOTAL</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
</tr>
</tbody>
</table>

Column 1 gives the recalculated water free composition from the bulk rock analysis.

Column 2 gives the bulk rock composition calculated from the mineral analysis.

From a comparison of these two analyses it is obvious that the MgO/SiO$_2$ ratio of the recalculated water free composition (col.1) is considerably lower than that expected from the composition calculated from the mineral analyses (col.2).

There/
There are three possible origins of this discrepancy.

1. Considerable underestimation of CPX and/or GNT in the mode. This, however, can be ruled out as the CPX and GNT contents are closely restricted by the bulk rock CaO, Al$_2$O$_3$ and Cr$_2$O$_3$ contents.

2. Considerable underestimation of the OPX/01 ratio in the mode. This is more of a possibility but the OPX/01 ratio is quite closely restricted by the NiO content. Furthermore to account for the total discrepancy in MgO/SiO$_2$ ratio the OPX/01 ratio in the mode would have to be adjusted from about 1/12 to 1/3, and there is no evidence of quite such large amounts of OPX in the rock.

3. Serpentinization involved the introduction of SiO$_2$ as well as H$_2$O, without any leaching. This explanation raises problems regarding the origin of the SiO$_2$ and the large volume increases which would have been involved as a result of serpeninization in such a case, and therefore seems less likely than the final one.

4. Serpentinization involved leaching of MgO with respect to SiO$_2$. This last possibility seems the most likely and suggests that in this case, at least, the serpeninization may well have involved a constant volume reaction with addition of H$_2$O and leaching of MgO, SiO$_2$ etc.

An attempt has been made to recalculate the bulk rock analysis assuming loss of MgO, FeO, NiO and SiO$_2$ and addition of H$_2$O according to the following equations.

**EQUATION 1.** \[5\text{Mg}_2\text{Si}_0\text{.}_4 + 4\text{H}_2\text{O} \rightarrow 2\text{H}_4\text{Mg}_3\text{Si}_2\text{O}_9 + 4\text{MgO} + \text{SiO}_2\]

\[700 \text{ gm.} \quad 72 \text{ gm.} \quad 552 \text{ gm.} \quad 161 \text{ gm.} \quad 60 \text{ gm.}\]

**EQUATION 2.** \[5\text{Fe}_2\text{Si}_0\text{.}_4 + 4\text{H}_2\text{O} \rightarrow 2\text{H}_4\text{Fe}_3\text{Si}_2\text{O}_9 + 4\text{FeO} + \text{SiO}_2\]

\[391 \text{ gm.} \quad 60 \text{ gm.}\]

**EQUATION 3.** \[5\text{Ni}_2\text{Si}_0\text{.}_4 + 4\text{H}_2\text{O} \rightarrow 2\text{H}_4\text{Ni}_3\text{Si}_2\text{O}_9 + 4\text{NiO} + \text{SiO}_2\]

\[299 \text{ gm.} \quad 60 \text{ gm.}\]
In the recalculated analysis (col.3, Table No.32) the amount of $H_2O^+$ in the bulk rock analysis was redistributed according to the composition of the olivine (52.51 wt.% MgO, 7.27 wt.% FeO, 0.40 wt.% NiO), and the estimated amounts of MgO, FeO, NiO and SiO$_2$ lost according to the above equations replaced in the analysis in place of $H_2O^+$. 

In the recalculated analysis (col.4, Table No.32) the SiO$_2$ released through the serpentinization of the olivines according to the equations 1, 2 and 3 above, was considered not to have been leached out but to have reacted according to the following equation to form more serpentine from olivine.

EQUATION 4. $3Mg_2SiO_4 + 4H_2O + SiO_2 \rightarrow 2H_4Mg_2Si_2O_9$

This reaction results in an increase in volume with addition of $H_2O$ but no loss. Thus the %$H_2O^+$ attributed to %SiO$_2$ lost in the recalculation (col.3) above is recalculated here as a straightforward constant composition anhydrous calculation. It can be clearly seen that this recalculated analysis gives a much closer approximation to the MgO/SiO$_2$ ratio of the bulk rock composition calculated from the mineral compositions (col.2) than does (col.3).

Recalculation of the analysis as given in Col.5 (Table No.32) was essentially similar to that in col.4, except that it was assumed that only 75% of the FeO and NiO released according to equations 2 and 3 was in fact lost during serpentinization. The other 25% is considered probably to have been retained in secondary magnetite and Ni bearing chlorite. Col.5 gives a somewhat better fit for FeO and NiO than col.4 when compared with the analysis calculated from the mineral compositions, and is therefore considered to give quite a good estimate of the original pre-serpentinization composition. The figure of 75% loss of FeO and NiO was chosen purely arbitrarily because it appears to give the best FeO and NiO fit, for the recalculations of $U_{16}$ and/
and \( U_2^* \).

From the foregoing calculations it is apparent that it is possible to derive a suitable pre-serpentinization bulk rock composition in terms of appropriate serpentinization reactions.

It is therefore considered that in this case, at least, significant leaching of MgO in particular has occurred during serpentinization. It seems unlikely that OPX has been sufficiently underestimated in the mode to account for such a large discrepancy in MgO/SiO\(_2\) ratio. However, irrefutable proof of whether leaching has in fact occurred or whether the original rock type contained an appreciably higher OPX content than is now apparent, rests on accurate NiO determinations of all the constituent minerals, including the serpentine and ore fractions. In the meantime on the available evidence the case for leaching of MgO etc. having occurred during serpentinization of this rock sample is quite strong.

\( \text{SiO}_2/\text{U}_2^* \). Given below are the recalculated compositions of this rock sample, the columns corresponding directly to those described above for \( U_16^* \).

<table>
<thead>
<tr>
<th></th>
<th>1.</th>
<th>2.</th>
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<tbody>
<tr>
<td>( \text{SiO}_2 )</td>
<td>44.35</td>
<td>40.47</td>
<td>43.34</td>
<td>41.30</td>
<td>41.42</td>
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<tr>
<td>( \text{TiO}_2 )</td>
<td>0.01</td>
<td>0.05</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>( \text{Al}_2\text{O}_3 )</td>
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<td>2.13</td>
<td>2.00</td>
<td>2.08</td>
<td>2.09</td>
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<tr>
<td>( \text{Cr}_2\text{O}_3 )</td>
<td>0.37</td>
<td>0.46</td>
<td>0.33</td>
<td>0.35</td>
<td>0.35</td>
</tr>
<tr>
<td>( \text{FeO} )</td>
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<td>7.15</td>
<td>7.18</td>
<td>7.43</td>
<td>7.18</td>
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<tr>
<td>( \text{MnO} )</td>
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<td>0.11</td>
<td>0.10</td>
<td>0.10</td>
<td>0.10</td>
</tr>
<tr>
<td>( \text{NiO} )</td>
<td>0.33</td>
<td>0.34</td>
<td>0.35</td>
<td>0.36</td>
<td>0.35</td>
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<tr>
<td>( \text{MgO} )</td>
<td>44.34</td>
<td>43.16</td>
<td>45.43</td>
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<tr>
<td>( \text{CaO} )</td>
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<td>1.07</td>
<td>1.06</td>
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<td>1.11</td>
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<tr>
<td>( \text{Na}_2\text{O} )</td>
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<td>0.06</td>
<td>0.14</td>
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<td>0.15</td>
</tr>
<tr>
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<td>-</td>
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<td>0.06</td>
<td>0.06</td>
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<tr>
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<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
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</tr>
</tbody>
</table>
The modal proportions (weight %) of the minerals used in the calculations were OL 84.0%, GNT 6%, CPX 3%, OPX 6%, and chrome spinel 1%. Mineral compositions used were GNT U₁₂, CPX U₁₀, OL U₁₆, OPX U₈ and chrome spinel (Deer, Howie and Zussman, Vol. 5, Table 11, No. 8).

To account for the total discrepancy in MgO /SiO₂ ratio between cols. 1 and 2, the OPX/OL ratio in the mode would have to be adjusted from 1/14 to almost 1/2, an inconceivably large underestimation of OPX content.

From a consideration of the compositions in Table No. 33, it is apparent that col. 5 again gives the best fit with the bulk rock composition calculated from the mineral compositions (col. 2). It is therefore considered that once again a strong case can be made out for leaching of MgO etc., with respect to SiO₂ during serpentinization.

Given below are the recalculated compositions of this rock sample, the numbering of the columns corresponding directly to those for U₁₆, with column k eliminated and an additional column (6) substituted. Column 5 represents a recalculation essentially similar to that of column 5, except that 50% rather than 75% Fe₀ and Ni₀ was assumed lost during serpentinization.

<table>
<thead>
<tr>
<th>TABLE NO. 34</th>
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<td>1.00</td>
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<td>6.65</td>
<td>6.87</td>
<td>6.86</td>
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<td>0.17</td>
<td>0.17</td>
<td>0.17</td>
</tr>
<tr>
<td>NiO</td>
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<td>0.21</td>
<td>0.23</td>
<td>0.23</td>
<td>0.22</td>
</tr>
<tr>
<td>MgO</td>
<td>33.82</td>
<td>36.58</td>
<td>35.14</td>
<td>35.94</td>
<td>35.99</td>
</tr>
<tr>
<td>CaO</td>
<td>4.46</td>
<td>4.25</td>
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<tr>
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<tr>
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<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
<td>100.00</td>
</tr>
</tbody>
</table>
The modal proportions (weight %) of the minerals used in the calculations were Ol 45%, GNT 25%, CPX 14%, OPX 15%, and chrome spinel 1%. Mineral compositions used in calculating the bulk rock composition were GNT U₈, CPX U₈, OPX U₈, Ol U₁₆ and chrome spinel (Deer, Howie and Zussman Vol. 5, Table II, No.8).

Once again there is a large discrepancy in MgO/SiO₂ ratio between columns 1 and 2. The OPX/Ol ratio would have to be brought down from 1/3 to 1/1 to account for it which seems unlikely as it would also bring the FeO and NiO values down too low.

On the other hand, the recalculated composition (col.6) assuming loss of some MgO, FeO and NiO during serpentinization gives a satisfactory fit with the bulk rock composition calculated from the mineral compositions (col.2). It is therefore considered that leaching of some MgO etc. during serpentinization has again occurred in this case.

For the remaining two Ugelvik specimens, U₃ and U₆, the straightforward water free recalculation of the analysis compared favourably with the bulk rock composition calculated using the mineral analyses and the estimated modal proportions of the minerals. It is therefore considered that in these cases serpentinization involved essentially a constant composition reaction without significant leaching of MgO etc.

It will be noticed that in the bulk rock compositions calculated from the mineral compositions (Col.2, Tables 31, 32 and 33) the Na₂O content is low compared with that of the recalculated bulk rock analysis compositions. However, this is probably just the result of the alkali contents of the minerals having been determined only for the clinopyroxenes, rather than Na₂O having been introduced during serpentinization.

Similar tests on certain of the more serpentinized Kalskaret peridotites, namely 833/T200, 834/T268, and 845/T100, also indicated that the straightforward/
straightforward water free recalculation of the bulk rock analysis gave a suitable pre-serpentinization composition. Therefore in all cases, other than the three above - U₂, U₈ and U₁₆ - where it is considered that there is a strong case for the leaching of MgO etc. having occurred during serpentinization, the straightforward water free recalculation of the bulk rock analysis is given as closely representing the pre-serpentinization composition.
In Chapter IV Section C the possibility that the Norwegian garnet peridotite mineral assemblages studied have resulted from the eclogite facies metamorphism of peridotitic rock masses which originally crystallized to another mineral facies assemblage was considered and discussed in terms of a series of bulk rock composition variation diagrams.

The differing peridotite mineral assemblages have been plotted on a series of diagrams (Fig. Nos. 37-48) and the range of bulk compositions shown by the Norwegian garnet peridotite localities at Kalskaret, Almklovdalen, and Ugelvik compared with the range expected from the sorting of the respective mineral assemblages. These diagrams are discussed in more detail here and the reasons outlined as to why the above hypothesis to explain the origin of the composition variation and mineralogy of these masses has been rejected.

**FIGURE NO. 37.** Lizard primary high temperature peridotite - 90631 Green (1964.)

Assemblage 01 + Opx + Cpx + Spinel.

A. CaO wt.% plotted against Al$_2$O$_3$ wt.%.

The range of Norwegian garnet peridotite compositions with respect to CaO and Al$_2$O$_3$ is encompassed by the Lizard assemblage but to derive most of the compositions in terms of this assemblage the original mineral assemblage would have to be considerably rich in Cpx and spinel (Lizard peridotites contain only 1-2 vol.% spinel in the mode).

B. Fe$^{2+}$Fe$^{3+}$Mn$^{2+}$ plotted against Mg$^{2+}$.

The majority of compositions can be derived from the Lizard assemblage/

- Kalskaret Garnet Peridotites
- Almklovdalen Garnet Peridotites
  (R Rødhaugen Eclogite)
- Ugelvik Garnet Peridotites
FIG. 42

Diagram A: CAO vs. \( \text{Al}_2\text{O}_3 \)

Diagram B: \( \text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mn}^{2+} \) vs. \( \text{Mg}^{2+} \)

Diagram C: \( \text{Na}_2\text{O} \) vs. \( \text{Al}_2\text{O}_3 \)

Diagram D: \( \text{Cr}_2\text{O}_3 \) vs. \( \text{Al}_2\text{O}_3 \)

Diagram E: \( \text{Ti}_2\text{O}_3 \) vs. \( \text{Si}_2\text{O}_3 \)
assemblage, although in many cases the original assemblage would again have to be very rich in spinel, which would appear unlikely. However, the Rodhaugen eclogite composition and one Kalskaret composition are too ferriferous, while several Kalskaret compositions, one Almklovdalen composition and the majority of the Ugelvik compositions are too magnesian.

C. $\text{Na}_2\text{O}$ wt.$\%$ plotted against $\text{Al}_2\text{O}_3$ wt.$\%$.

All the garnet peridotite compositions lie outside the composition range outlined by this Lizard assemblage, being too rich in $\text{Na}_2\text{O}$ with respect to $\text{Al}_2\text{O}_3$ to have been derived by metamorphism of this assemblage.

D. $\text{Cr}_2\text{O}_3$ wt.$\%$ plotted against $\text{Al}_2\text{O}_3$ wt.$\%$.

Only a few of the garnet peridotite compositions, mostly from Ugelvik could be derived from the Lizard assemblage with respect to $\text{Cr}_2\text{O}_3$ and $\text{Al}_2\text{O}_3$. Most have too low a $\text{Cr}_2\text{O}_3/\text{Al}_2\text{O}_3$ ratio but a few have too high a ratio.

E. $\text{Ni}^{2+}$ plotted against $\text{Mg}^{2+}$.

Most of the Norwegian garnet peridotite compositions are encompassed by this Lizard assemblage, but a few Kalskaret and most of the Ugelvik compositions have too low a $\text{Ni}^{2+}/\text{Mg}^{2+}$ ratio.

F. $\text{TiO}_2$ wt.$\%$ plotted against $\text{SiO}_2$ wt.$\%$.

Only a very few of the garnet peridotite compositions fall within the composition field outlined by the Lizard assemblage.

Clearly therefore in the light of these diagrams and the compositional differences involved neither the Kalskaret, Almklovdalen nor the Ugelvik garnet peridotites/
peridotites can have been derived from the isochemical eclogite facies metamorphism of peridotite such as that produced by the sorting of a mineral assemblage similar to that of the primary high temperature Lizard peridotite. Cryptic layering (changes in mineral compositions within the solid solution series) would be expected to produce only very minor bulk compositional changes in such small peridotite bodies as those in the Norwegian gneiss complex, when compared with the extent of cryptic layering variations found in large basic layered complexes, (see Chapt.III, Section 1A).

The principal features demonstrated by the comparison of the Norwegian garnet peridotite compositions with the other mineral assemblages were as follows.

FIGURE NO. 38. Lizard recrystallized low temperature peridotite, 90686 Green (1964)

Assemblage Ol + OPX + CPX + Plagioclase + chromite

Despite uncertainties as to the exact composition of the plagioclase and chromite of this assemblage (appropriate analyses from the literature have been plotted) this assemblage can likewise be ruled out as a possible one for explaining the bulk composition range of the Norwegian garnet peridotites on the following grounds:

1. All except the Rodhaugen eclogite composition have too high a Ni$^{2+}$/Mg$^{2+}$ ratio.

2. Several of the Kalskaret compositions and one Ugelvik composition have slightly too high Na$_2$O/Al$_2$O$_3$ ratios, while many of the Almklovdalen compositions would require rather unlikely concentrations of chromite in the original assemblage to give their Na$_2$O/Al$_2$O$_3$ ratio.

3. The majority of the Ugelvik and a few of the Kalskaret compositions are too magnesian, while many of the Almklovdalen and Kalskaret compositions would/
would again require unlikely original concentrations of chromite to
give their $\text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mn}^{2+} / \text{Mg}^{2+}$ ratios.

4. Several of the Ugelvik and Kalskaret compositions are incompatible
   in terms of TiO$_2$ content.

FIGURE NO.39. Peridotite nodule from basalt. Dreiser Weiher, Germany
Ross et al. (1954) No. 4.

Assemblage Ol + CPX + OPX + Spinel.

This assemblage likewise fails to explain the composition range
of the garnet peridotites from any of the three localities on the following
grounds:

1. Only 3 Ugelvik and 2 Kalskaret compositions are compatible in terms
   of Cr$_2$O$_3$/Al$_2$O$_3$ ratio.

2. The Rodhagen eclogite composition is incompatible in virtually every
   respect.

3. The majority of the Ugelvik compositions, 4 Kalskaret compositions and
   one Almklovdalen composition are too magnesian, while many of the Kalskaret
   and Almklovdalen compositions would require unlikely original concentrations
   of spinel to give their $\text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mn}^{2+} / \text{Mg}^{2+}$ ratios.

4. Many of the compositions from Kalskaret, Almklovdalen and Ugelvik are
   incompatible in terms of Ni$^{2+}$/Mg$^{2+}$ ratio.

FIGURE NO.40. Stillwater layered basic intrusion EB 43 Hess (1960)

Assemblage Plagioclase + OPX + CPX.

This assemblage is clearly incompatible with all the garnet peridotite
compositions with respect to Cr$_2$O$_3$/Al$_2$O$_3$ and $\text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mn}^{2+} / \text{Mg}^{2+}$ ratios, and
also for most compositions with respect to Na$_2$O/Al$_2$O$_3$ ratio and SiO$_2$ content.

This/
This assemblage together with a magnesian olivine would ease the $\text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mn}^{2+} / \text{Mg}^{2+}$ and SiO$_2$ but not the Cr$_2$O$_3$/Al$_2$O$_3$ and Na$_2$O/Al$_2$O$_3$ anomalies.

**FIGURE No. 41.** Large dunite pluton - Twin Sisters, Washington No. 12

Assemblage OL + CFX + Spinel

Once again the garnet peridotite compositions show a haphazard distribution with respect to the range of compositions to be expected from the sorting of this mineral assemblage. All the compositions are incompatible in terms of Cr$_2$O$_3$/Al$_2$O$_3$ ratio, and many of them from all three localities incompatible with respect to Na$_2$O/Al$_2$O$_3$, CaO/Al$_2$O$_3$ and Ni$^{2+}$/Mg$^{2+}$ ratios and TiO$_2$ content.

**FIGURE No. 42.** Hypersthene eclogite nodule in tuff - Salt Lake Crater, Oahu 66118 Yoder and Tilley (1962).

Assemblage GNT + CFX + CFX.

Sorting of this mineral assemblage again fails to explain the range of Norwegian garnet peridotite compositions on the following grounds.

1. Almost all the compositions lie outside the range of $\text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mn}^{2+} / \text{Mg}^{2+}$ ratios expected.

2. Similarly only two Kalskaret and the Rodhaugen eclogite composition lie within the expected range of Cr$_2$O$_3$/Al$_2$O$_3$ ratios.

3. Several of the compositions from all three localities are also incompatible with respect to CaO/Al$_2$O$_3$ and Na$_2$O/Al$_2$O$_3$ ratios and TiO$_2$ content.

The additional sorting of a magnesian olivine in this assemblage would obviously ease the $\text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mn}^{2+} / \text{Mg}^{2+}$, CaO/Al$_2$O$_3$ and Na$_2$O/Al$_2$O$_3$ ratio anomalies for several of the compositions, but clearly not the Cr$_2$O$_3$/Al$_2$O$_3$ anomaly.
Sorting of this mineral assemblage again fails to explain the range of composition exhibited by the Norwegian garnet peridotites in the following respects:

1. All the compositions lie outside the range of Cr$_2$O$_3$/Al$_2$O$_3$ ratios expected.

2. Several Kalskaret compositions and the Redhaugen eclogite composition are too ferriferous, while other Kalskaret compositions and all but one Ugelvik composition are too magnesian.

3. The absolute amounts of Al$_2$O$_3$ in many compositions would require unlikely concentrations of spina in the original assemblage.

The assemblage, identical to that of the majority of the Norwegian garnet peridotites, covers the range of compositions of most of the latter with respect to CaO/Al$_2$O$_3$ and Na$_2$O/Al$_2$O$_3$ ratios, with the notable exception of the Redhaugen eclogite in both cases.

However, the following compositional anomalies do arise.

1. Several of the Kalskaret and two of the Almklovndalen compositions are too ferriferous, although three Ugelvik and two Kalskaret compositions are slightly too magnesian.

2. Most of the garnet rich rocks from both Kalskaret and Almklovndalen have too low a Cr$_2$O$_3$/Al$_2$O$_3$ ratio.
3. Minor anomalies also arise concerning the Ni\(^{2+}/\text{Mg}^{2+}\) ratio and TiO\(_2\) content of certain compositions from all three localities.

**FIGURE NO. 45.** Garnet peridotite - Ugelvik, Otteroy 349/8

Assemblage O1 + GNT + CPX + OPX

This mineral assemblage covers the range of compositions of the Norwegian garnet peridotites with respect to CaO/Al\(_2\)O\(_3\) and Na\(_2\)O/Al\(_2\)O\(_3\) ratios, with the notable exception of the Rodhaugen eclogite in both cases. All the Ugelvik compositions can in fact be derived from varying proportions of the mineral assemblage, apart from a minor anomaly in TiO\(_2\) content (this assemblage is notably TiO\(_2\) rich with respect to all the other analysed Ugelvik assemblages).

However, most of the other garnet peridotites, namely those from Kalskaret and Almklovdalen, cannot be interpreted in terms of the sorting of this mineral assemblage, with the minerals having the compositions represented here, on several counts. Most have too high a Ni\(^{2+}/\text{Mg}^{2+}\) ratio while the most garnetiferous peridotites have too high a Fe\(^{2+}\)+Fe\(^{3+}\)+Mn\(^{2+}/\text{Mg}^{2+}\) ratio and too low a Cr\(_2\)O\(_3/\text{Al}_2\)O\(_3\) ratio.

**FIGURE NO.46.** Garnet websterite - Kalskaret, Tafjord N 23 O'Hara & Mercy (1963)

Assemblage GNT + CPX + OPX + secondary amphibole (10%) 

This assemblage likewise covers the range of compositions of most of the garnet peridotites with respect to CaO/Al\(_2\)O\(_3\) and Na\(_2\)O/Al\(_2\)O\(_3\) ratios with the notable exception of the Rodhaugen eclogite. However, it again fails to cover anything like the complete range of Cr\(_2\)O\(_3/\text{Al}_2\)O\(_3\), Ni\(^{2+}/\text{Mg}^{2+}\), and Fe\(^{2+}\)+Fe\(^{3+}\)+Mn\(^{2+}/\text{Mg}^{2+}\) ratios or TiO\(_2\) values even for the Kalskaret compositions alone.

**FIGURE NO.47.**
FIGURE NO. 47. Garnet peridotite - Kalskaret, Tafjord 841/T96

Assemblage 01 + OPX + CPX + GNT

Despite broad agreement, notably in CaO/Al₂O₃ and Na₂O/Al₂O₃ ratios, this assemblage again fails to encompass the range of compositions of all the garnet peridotites, even from Kalskaret alone, on much the same grounds as with assemblage N23. However, the appearance of olivine in this assemblage results in a greater range of Ni²⁺/Mg²⁺ values compared with N23.

FIGURE NO. 48. Garnet lherzolite Lien Farm, Almklovdelen N69 O'Hara and Mercy (1963)

Assemblage 01 + CPX + GNT + OPX + secondary amphibole (5%)

This assemblage covers the range of compositions with respect to CaO/Al₂O₃ and Na₂O/Al₂O₃ ratios, again with the exception of the Rodhaugen eclogite. Although the broad range of compositions encompassed by this assemblage covers most of the garnet peridotite compositions, in detail several incompatibilities arise as follows:

1. Certain Almklovdelen and Kalskaret compositions have too low a Cr₂O₃/Al₂O₃ ratio, while some Kalskaret and Ugelvik compositions have too high a ratio.

2. Most Ugelvik compositions, several Kalskaret compositions, the Rodhaugen eclogite and one other Almklovdelen composition are incompatible in terms of Fe²⁺+Fe³⁺+Mn²⁺/Mg²⁺ ratio.

3. Minor anomalies arise mostly concerning Kalskaret and Ugelvik compositions with respect to Ni²⁺/Mg²⁺ ratio and TiO₂ content.

The conclusion which can be drawn from the foregoing discussion of these/
these variation diagrams is that it is impossible to explain the mineralogy and range of bulk composition of the Norwegian garnet peridotites solely in terms of the isochemical eclogite facies metamorphism of other possible peridotite mineral assemblages.

It is further apparent from the composition variation diagrams that although the composition range of the Norwegian garnet peridotites is most closely encompassed by the eclogite facies peridotite assemblage - Ol + GNT + CPX + OPX - chemical variations among the Kalskaret and Almkløvdalen garnet peridotites are such that they extend beyond the range of compositions to be expected from the sorting of any analysed garnet peridotite assemblage (either from the Norwegian gneisses or from kimberlite pipes) even when allowance is made for possible cryptic layering compositional effects. Over such small thicknesses as the Kalskaret garnet peridotites occur (about 50 feet) cryptic layering effects are expected to be insignificant.

On the other hand, only small mineral composition differences occur in the Ugelvik peridotites so that apart from a minor TiO₂ anomaly all the Ugelvik bulk rock compositions are quite compatible with the sorting of the Ugelvik mineral assemblage U₈.
APPENDIX III. LOCATIONS AND DESCRIPTIONS OF THE ROCK TYPES ANALYSED
BY THE AUTHOR IN THIS THESIS.

KALSKARET GARNET PERIDOTITE LENS

Specimens collected along Section AB Fig.8. Distances measured planimetrically on ground surface. See also Map I for exact location of garnet peridotite outcrop.

833/T200 PERIDOTITE 150' above top of main garnet peridotite lens.
Assemblage: OL + Amphibole + chlorite.

834/T268 DUNITE 80' above top of main garnet peridotite lens.
Assemblage: OL + OPX. Traces of Amphibole, Chlorite and Opaques.

835/T150 AMPHIBOLITIZED GARNET PERIDOTITE 30' above base of main garnet peridotite lens.
Assemblage: OL + Amphibole + Chlorite. Traces of OPX, CPX and Opaques.

836/T151 GARNET LHERZOLITE 25' above base of main garnet peridotite lens.
Assemblage: OL + CPX + GNT + OPX + Amphibole. Trace of Opaques.

837/T152 GARNET WEHRLITE 21' above base of main garnet peridotite lens.
Assemblage: OL + CPX + GNT + Amphibole + Opaques.

838/T153 GARNET WEBSTERITE 19' above base of main garnet peridotite lens.
Assemblage: CPX + GNT + OPX + Amphibole. Trace of OL and Opaques.

839/T154 GARNET WEHRLITE 14' above base of main garnet peridotite lens.
Assemblage: OL + CPX + GNT + Amphibole. Trace of OPX and Opaques.

840/T155 GARNET WEHRLITE 9' above base of main garnet peridotite lens.
Assemblage: OL + CPX + GNT + Amphibole + Opaques.

841/T196 GARNET LHERZOLITE 6' above base of main garnet peridotite lens.
Assemblage: OL + CPX + GNT + OPX + Amphibole. Trace of Opaques.

842/T155/
AMPHIBOLITIZED GARNET PERIDOTITE

Assemblage: OL + OPX + Amphibole + Chlorite + Opaques.

DUNITE

Assemblage: OL + Chlorite. Traces of amphibole and opaques.

AMPHIBOLITIZED GARNET PERIDOTITE

Assemblage: OL + Amphibole + Opaques.

ALTERED GARNET PERIDOTITE

Assemblage: OL + Amphibole + Chlorite + Opaques.

UGELVIK PERIDOTITES

Specimen 84.9/U_8 was collected from the north side of the peridotite mass to the north-east of Ugelvik village (see Fig. 17). All other specimens were collected from the Ugelvik village peridotite mass, the localities being shown on Fig. 18.

GARNET LHERZOLITE

Assemblage: OL (Serp.) + CPX + GNT + OPX + Opaques.

Serpentinite

Assemblage: Serpentine after OL + Serpentine after OPX + OL + Opaques.

GARNET LHERZOLITE

Assemblage: Serpentine + CPX + GNT + OPX

GARNET LHERZOLITE

Assemblage: OL (Serp.) + CPX + GNT + OPX + Opaques.

GARNET DUNITE

Assemblage: OL (Serp.) + GNT + OPX + Opaques.
KALDHUSSAETER PERIDOTITES

The various peridotite outcrops in the Kaldhussaeter mass are shown on Map II. The analyses of these specimens are listed in Table No.35.

Specimens T76, T79 and T75 were collected adjacent to the mafic layering (Fig. 11) in the small peridotite outcrop at a height of 1080 m. in the central portion of the mass.

888/T76  CHLORITE AMPHIBOLITE

Assemblage: Chlorite + Amphibole + OL (Serp.) + Chromite.

889/T79  ALTERED HARZBURGITE

Assemblage: OL + OPX + Chlorite + Chromite. Traces of amphibole.

898/T75  ALTERED HARZBURGITE

Assemblage: OL (Serp.) + OPX + Amphibole + Chlorite + Chromite.

897/T37  ALTERED PERIDOTITE Collected at about 1040 m. in the southernmost peridotite outcrop.

Assemblage: OL (Serp.) + OPX + Chlorite + Amphibole + Talc + Chromite.

899/T207  HARZBURGITE NODULE IN PERIDOTITE Collected about 4 m. above the mafic layering in the small central peridotite outcrop at height of 1080 m.

Assemblage: OPX + OL + Chromite.

900/T214  DUNITE Collected at height of about 980 m. at the northern end of the mass.

Assemblage: OL + minor OPX and chromite with rims of chlorite.

901/T215  DUNITE Adjacent to 900/T214.

Assemblage: OL + OPX + Chlorite + Chromite. Trace of amphibole.

902/T237  DUNITE Collected at height of about 1075 m. just above the eclogite lens in the northernmost group of peridotite outcrops.

Assemblage:/
Assemblage: OL + minor OPX, Chlorite, Amphibole and Chromite.

903/T242 PERIDOTITE Collected at about 1055 m. in central portion of the mass; the tourist hut bearing 30°N. of W.

Assemblage: OL + OPX + Chlorite + Amphibole + trace of Chromite.

904/T246 DUNITE Collected at about 985 m. in central portion of the mass; the tourist hut bearing 28°N. of W.

Assemblage: OL + Amphibole + trace of Chlorite and Chromite.

905/T247 HARBURGITE Adjacent to 904/T246.

Assemblage: OL + OPX + Chlorite + Amphibole + Chromite.

OTHER PERIDOTITES Analyses listed in Table No. 36.

906/T133 ORTHOPYROXENITE Southern margin of main Ranukdaleen peridotite mass.

Assemblage: OPX + Chlorite + Chromite.

907/T259 SERPENTINIZED DUNITE Western margin of the largest Upper Ranukdaleen peridotite outcrop.

Assemblage: OL (Serp.) + minor OPX, Chlorite and Chromite.

908/T260 SERPENTINIZED HARBURGITE Adjacent to 907/T259.

Assemblage: OL (Serp.) + OPX + minor Chlorite and Chromite.

909/T262 DUNITE Centre of largest Upper Ranukdaleen peridotite outcrop.

Assemblage: OL + minor OPX, Chlorite, Chromite and Amphibole.

910/T146 HARBURGITE Southern margin of main Ranukdaleen peridotite mass.

Assemblage: OL + OPX + Talc + Chromite + Chlorite.

911/T83 ALTERED PERIDOTITE Femre Roddal mass, Tafjord (Large mass about 3 kms. N.E. of Kaldhussaeter mass).

Assemblage: OL+ Amphibole + Phlogopite + Chromite + Chlorite.

912/T132 DUNITE Collected at about 750 m. from the largest peridotite outcrop of the group overlooking Onilsa Watn on the N.E. slope of Jordhom (see Map I and Fig. 5).

Assemblage:/
Assemblage: CL + minor OPX + trace of Chlorite and Chromite.

918/T177  CHLORITIZED PERIDOTITE Small peridotite outcrop overlooking Herdalen at the W. end of Kalskaret coll.

Assemblage: Chlorite + Amphibole + CPX + Chromite.

919/T178  CHLORITIZED PERIDOTITE Adjacent to 918/T177.

Assemblage: Chlorite + Amphibole + CPX + Chromite.

921/T258  HARZBURGITE Eastern lip of Kalskaret coll, overlooking Tafjord.

Assemblage: OL + OPX + Chlorite + minor Amphibole and Chromite.

924/H9   ALTERED HARZBURGITE Hornindal, Nordfjord.

Assemblage: OL + OPX + Chlorite + Talc + Chromite.

925/H10   DUNITE Hornindal, Nordfjord.

Assemblage: OL + OPX + Chlorite + Chromite.

ANORTHOSITES (Table No. 37).

916/T270  ANORTHOSITE S.E. side of Kalskaret coll.

Assemblage: Labradorite + minor Epidote and Chlorite.

917/V10   ANORTHOSITE N.E. end of Viksvatn.

Assemblage: Labradorite + minor Epidote, Amphibole and Chlorite.

ECLOGITES

892/T205  ECLOGITE 10' lens within small peridotite outcrop at 1085 m. in the central portion of the Kaldhussaeter mass (see Map II).

Assemblage: Garnet + altered Pyroxene Symplektite + Amphibole + Rutile + Opaques.

893/T212  QUARTZ ECLOGITE Lens in gneiss - Kaldhussaeter; height 1100 m. in gneiss crags just above location of 892/T205.

Assemblage: Garnet + altered Pyroxene Symplektite + Amphibole + Phlogopite + Quartz + Rutile + Opaques.
894/T238  **ECLOGITE** Lens at 970 m. within peridotite at northern end of Kaldhussaeter mass.

Assemblage: Garnet + altered Pyroxene Symplektite + Amphibole + Clinozoisite + Rutile + Opaques.

895/T271 **QUARTZ ECLOGITE** Lens at about 1050 m. within peridotite at northern end of Kaldhussaeter mass.

Assemblage: Garnet + altered Pyroxene Symplektite + Amphibole + Quartz + Phlogopite + Clinozoisite + Rutile + Opaques.

896/T223 **QUARTZ RICH SEGREGATION WITHIN ECLOGITE** Easternward extension of lens within Kaldhussaeter peridotite where 895/T271 collected.

Assemblage: Garnet + Amphibole + Quartz + Plagioclase + Mica + Opaques.

922/H2 **ECLOGITE** Lens within gneiss just above Hornindal peridotite mass.

Assemblage: Garnet + Pyroxene Symplektite + Amphibole + Rutile + Opaques.

923/H4 **QUARTZ ECLOGITE** Lens within Hornindal peridotite mass.

Assemblage: Garnet + altered Pyroxene Symplektite + Amphibole + Quartz + Clinozoisite + Rutile + Opaques.

**AMPHIBOLITES**

890/T80 **ENSTATITE AMPHIBOLITE** Mafic layer within small peridotite outcrop in central portion of Kaldhussaeter mass. Height about 1080 m. (Fig. 11).

Assemblage: Hornblende + Enstatite + Opaques.

891/T81 **GARNET AMPHIBOLITE** Adjacent layer to 890/T80.

Assemblage: Garnet + Clinopyroxene + Hornblende + Sphene + Opaques.

913/T127/
913/T127  **GARNET AMPHIBOLITE**  Mass within gneiss-Muldal, Tafjord.
   Assemblage: Garnet + Hornblende + Biotite + Plagioclase + Opaques.

914/T195  **AMPHIBOLITE**  Mafic lens within Kalskaret peridotite
   Assemblage: Altered Pyroxene Symplektite + Amphibole + Plagioclase + Sphene + Opaques.

915/T256  **GARNET AMPHIBOLITE**  Adjacent to Kalskaret peridotite mass.
   Assemblage: Garnet + Hornblende + Plagioclase + Opaques.

920/T183  **GARNET AMPHIBOLITE**  Just below the chloritized peridotite mass at the western end of Kalskaret coll.
   Assemblage: Garnet + Hornblende + Plagioclase + Phlogopite + Rutile + Opaques.
<table>
<thead>
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<th>ANAL. NO.</th>
<th>FIELD NO.</th>
<th>T76</th>
<th>T79</th>
<th>T37</th>
<th>T75</th>
<th>T207</th>
<th>T214</th>
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* Wet Methods Analyses. All Others X-Ray Fluorescence Determinations
### TABLE NO. 36

**BULK ROCK ANALYSES OF MISCELLANEOUS PERIDOTTITES.**

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* Wet Methods Analyses. All Others X-ray Fluorescence Determinations.

| Fe/Fe+Mg | 0.08 | 0.08 | 0.07 | 0.07 | 0.08 | 0.09 | 0.08 | 0.16 | 0.17 | 0.10 | 0.10 | 0.08 |
# TABLE NO. 37

**BULK ROCK ANALYSES OF ANORTHOSITES**

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Whenever possible rock samples for bulk rock analysis were of 1-2 kgm. weight, although some hand specimens of the layered garnet peridotite sequences were somewhat smaller. However, these latter specimens were collected to show the composition variations between adjacent layers and not individually to represent the bulk composition of any substantial volume of rock.

These rock samples after scrubbing, rinsing with deionised water and drying were then split into small pieces of about 2 cm. size with a hydraulic rock splitter. Any weathered material was then discarded and only fresh material retained. After mixing by hand a substantial grab sample (500-700 gms.) was reduced to 10 mesh size by passing it through thoroughly cleaned 'Manchester' rollers. A sixth of this material (amounting to somewhere in the order of 100 gm. in most cases) was then taken, by means of an automatic sample divider, for further grinding. This sample was then reduced to 100 mesh powder by grinding in an agate mortar. Throughout the grinding process care was taken to avoid contamination and to create as little dust as possible so to keep sample loss to a minimum.

The rock powder was then placed in a glass bottle, thoroughly mixed, and dried at 110° C in an oven for at least 2 hours, in most cases overnight.

**WET METHODS OF ANALYSIS**

Eleven Kalskaret peridotites were analysed for Si, Al, Ti, Fe (total), Fe^{2+}, Mn, Mg, Ca, Na, K, Cr and H by wet chemical methods. The results of these analyses are given in Table No. 38, together with Ni values for/
for all the samples and Cr values for 4 of them determined by X-ray fluorescence spectroscopy.

Si was determined gravimetrically. After fusion with Na$_2$CO$_3$ and evaporation of the solution with HCl, any remaining soluble SiO$_2$ was precipitated by addition of gelatin at 70°C.

Al was determined according to the spectrophotometric method given by Mullen and Riley. After adding $\alpha$-dipyridyl hydroxylamine, BeSO$_4$ and sodium acetate and adjusting the pH to 4.9-5.1, the aluminium was extracted with 8-hydroxyquinoline dissolved in chloroform. Interference was limited to that of titanium, and a correction factor was applied to adjust for it.

Total iron (as Fe$_2$O$_3$) was determined spectrophotometrically by means of the red colour produced by complexing with sodium citrate and ammoniacal thioglycollic acid.

Oxidation of an acid solution of the rock with periodate produces a purple permanganate colour from the Mn, which can then be conveniently determined spectrophotometrically.

P was only determined as a check in a limited number of cases. The P contents of the peridotites are negligible ($< 0.01\%$) even in the garnetiferous varieties. Of the analysed rock types only the quartz eclogites contain significant amounts ($\sim 0.05 - 0.17\%$). The method used for determining P was a spectrophotometric one involving the production of a molybdenum blue complex through the addition of a reducing solution of ammonium molybdate and ascorbic acid to a solution of the rock of carefully controlled acidity.

Ti was determined spectrophotometrically according to the Tiron method. 4% aqueous tiron (1,2-dihydroxy benzene 3,5-disulphonate disodium) was/
was added to a solution of the rock which was then neutralized with
$\text{NH}_4\text{OH}$ to give a blood red colour due to $\text{Fe}$, and a buffer solution (of pH 4.7)
consisting of equal volumes of molar acetic acid and molar sodium acetate
added. Subsequently sodium dithionite was added carefully removing the
red colour by reducing $\text{Fe}^{3+}$ to $\text{Fe}^{2+}$ and leaving a pale yellow colour due
to $\text{Ti}^{4+}$ which was then determined.

Following fusion of rock powder with a 10:1 mixture of $\text{Na}_2\text{CO}_3 :$
$\text{KNO}_3$ and oxidation of chromic chromium to chromate by heating in acid
solution with ammonium persulphate in the presence of a trace of $\text{Ag NO}_3$
which acts as a catalyst in accelerating the oxidation, Cr in the rock samples
was determined spectrophotometrically by means of the violet colour produced
through complexing with diphenylcarbazide.

Spectrophotometric determination of Cr as a test of the X-ray
fluorescence values for several chrome rich garnets and clinopyroxenes was
done directly on the yellow chromate colour in strongly alkaline solution
following fusion with 10 : 1 mixture of $\text{Na}_2\text{CO}_3 : \text{KNO}_3$.

Mg and Ca were determined by titration with ethylene-diamine-
tetra-acetic acid (E.D.T.A.). Prior to titration, Fe, Al, and Ti were
removed by double precipitation with $\text{NH}_4\text{OH}$ from a solution containing $\text{NH}_4\text{Cl}$,
and sulphides of Ni, Pb, Co, Pt etc. removed by precipitation with $\text{H}_2\text{S}$
under first alkaline and then acidic conditions. Ca and Mg together were
titrated at a pH of 9 after the addition of KCN to complex heavy metals
and triethanolamine to complex any Al remaining in solution, using eriochrome
black T as indicator. As the $\text{Ca}^{2+}/\text{Mg}^{2+}$ ratio is very low in many of these
rocks, $\text{Ca}^{2+}$ was concentrated by precipitation as calcium oxalate and a
restricted amount of $\text{Mg}^{2+}$ ions subsequently added so that in the titrate
the ratio of $\text{Ca}^{2+}/\text{Mg}^{2+}$ ions was approximately unity. Ca alone was then
titrated/
**TABLE NO. 38.**

**WET METHODS ANALYSES OF KALSKARET PERIDOTTES**

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* X-ray Fluorescence Determinations of Cr and Ni
titrated at a pH of 12, again after addition of KCN and triethanolamine, using Patton and Reeder's reagent as indicator this time. The amount of Mg was obtained by difference.

An EEL Model A flame photometer was used to determine Na and K. Suitably diluted sample solutions with acid concentrations of 0.1-1% H₂SO₄ were compared with standard solutions containing 5-25 ppm. Na₂O or K₂O (as sulphate) in 0.1% H₂SO₄ read at the same time.

Ferrous iron in all the analysed rocks and minerals was determined by titration with dichromate using sodium diphenylamine sulphonate as indicator. Considerable difficulty was experienced in achieving complete solution when simmering with H₂SO₄ and HF, particularly with rocks containing garnet and spinel. The time allowed for simmering was varied from 3-20 minutes in an attempt to reach a balance between the time required for complete solution and the possible oxidation effects introduced through prolonged simmering. In general it was found that about 10 minutes represented the optimum time. If solution was obviously incomplete, following titration the sample was recovered, washed and the process repeated again. In the case of the garnets, the process was repeated at least twice, in many cases three times. The precision obtained by this method for these rock types and minerals was not very good, and the method may also be rather inaccurate. It is therefore considered that the FeO/Fe₂O₃ ratio of many of the analysed rocks and minerals should be treated with reserve as in many cases, the garnet analyses in particular, the FeO content indicated by this method may well be somewhat lower than the true value.

The determination of H (as H₂O) was done gravimetrically. A mixture of 1 gm. of rock powder and 3 gm. of a flux (2 parts PbO + 1 part PbCrO₄) was heated for 5 minutes in a Pyrex test-tube and the water driven off/
off collected in the cooled upperpart of the test-tube. A correction factor was added, 10% for \( <0.02 \text{ gm. H}_2\text{O} \) and 0.002 gm. for \( >0.02 \text{ gm. H}_2\text{O} \), to allow for the escape of any uncondensed steam. All determinations were done in duplicate or until concordant results were obtained. Considering the rather crude method of determination it was considered that the precision obtained even at the 14% level was surprisingly good. At such high concentrations the amounts of rock sample and flux were cut down, however.

X-RAY FLUORESCENCE SPECTROSCOPY

All the analysed rocks and minerals, including those analysed by wet methods, have had Ti, Al, Cr, Fe, Mn, Ni, Mg and Ca determined by X-ray fluorescence spectroscopic techniques.

The operating conditions for the analysis of each element are summarized in Table No. 40.

A chromium target was used for the light elements Mg, Ca and Al, for all the other elements a tungsten target was used for generation of the X-rays.

In all cases, other than the determination of trace amounts of Cr \((<0.4 \text{ wt.\% Cr}_2\text{O}_3)\), a fusion disc preparation was used in an attempt to minimize the differing matrix absorption effects of the rock types analysed and of the standard rock powders. A 1 : 1 : 4 ratio of rock powder (100 mesh) : \( \text{La}_2\text{O}_3 : \text{LiB}_2\text{O}_7 \) was mixed and fused together in an oven at 1000°C for 10 minutes. After cooling the fused beads were reweighed and brought up to standard weight through the addition of boric acid. The beads were then crushed and ground together with the boric acid in a tungsten carbide ball-mill until the material would pass a 200 mesh sieve. Some of this material was then made into a compact smooth surfaced disc by backing it with boric acid and/
and pressing under 15 tons for 1 minute. Through this fusion process the rock powder matrix absorption effects are minimized both by dilution and by the addition of \( \text{La}_2\text{O}_3 \) which has a large coefficient of absorption hence becoming the limiting absorption factor.

As La interferes with the Cr \( \text{K}_{\alpha_1} \) peak and Mn with the Cr \( \text{K}_{\beta_1} \) peak, determination of Cr raised a problem. This was overcome by determining Cr at low concentrations (\(<\) about 0.40 wt.\% \( \text{Cr}_2\text{O}_3 \)) on 100 mesh rock powder covering a film of Mylar. For this determination the Cr \( \text{K}_{\alpha_1} \) peak at a 2\( \theta \) of 69.31 for LiF was used. It is considered that at such low concentrations the error resulting from differing matrix absorption effects is very small. However, as at greater concentrations this error is likely to become significant, Cr in higher concentrations (\(>\) about 0.40 wt.\%) was determined on fusion discs using the \( \text{K}_{\beta_1} \) peak at a 2\( \theta \) of 100.48 for Topaz. It was found possible to determine Cr down to concentrations of about 0.40 wt.\% \( \text{Cr}_2\text{O}_3 \) before interference by Mn, at the low MnO concentrations (\(<0.60 \text{ wt.}\%\)) found in these rocks and minerals.

In the setting up of the X-ray spectrograph for the determination of a given element it was necessary to determine the location of the peak using a pure compound of the element and to determine the optimum discriminator settings (lower level, channel width and attenuation) to cut out any spectral overlap produced by reflections of different orders and to maximize the peak/background ratio and stability of counting.

Determinative curves were drawn up for each element using a number of recognized standard materials prepared similar to the unknown rock powders. Peak minus background counts or, in certain cases of high peak/background ratio; peak only counts were plotted against concentrations of the particular element in question.
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*SiO₂, FeO, Na₂O, K₂O, H₂O WET METHODS DETERMINATIONS*
The standards used were as follows:

U.S.G.S. standards G-1 and W-1 - the values given by Fleischer and Stevens (1962) being used.

U.S. Bureau of standards - Milford Granite 4983
Silica Brick 102
Opal Glass 91
Argillaceous Limestone 1A
Burnt Refractories 76 and 77.

C.A.A.S. standard - Syenite Sy-1 - using the values given by Webber (1965)

Rock samples N23, N75, N76 analysed by Mercy and O'Hara (1965a) and the compositions of the peridotites analysed by wet methods (Table No.38) were also used as approximate standards in defining the working curves.

Special standards for the Ni and Cr determinations were prepared by adding appropriate amounts of both these elements to one of the rock samples.

In all cases samples were irradiated in batches of 4, one of which was an internal standard against which the unknowns were standardized to counteract the effects of any drift of the generator E.H.T.

Table No.39 gives the results of the X-ray fluorescence analysis of the Kalskaret peridotites which were also analysed by wet methods (c.f. Table No.38). In most instances the results from the two methods of analyses are quite comparable so that the analyses listed in Table No.15 are mostly averages of the two values obtained. However, the X-ray fluorescence determinations of low concentrations of CaO are considered more reliable, the wet method used being very complex. Therefore the X-ray fluorescence values of CaO for 834/T268 and 843/T156 were preferred. Also the wet method MgO value for 843/T156 gives rather too high a total, therefore the X-ray fluorescence value was again preferred.
### TABLE NO. 40

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**OPERATING CONDITIONS OF X-RAY FLUORESCENCE SPECTROGRAPH**


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HART, S.R./


KITAHARA, S./


LEECH, G.B./


MURET, G.,/


Thayer, T.P.,/


Transvaal Leader, Johannesburg.


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Plate 1. Microfolding in the Kaldhussaeter peridotite mass.


Note the tendency for the secondary chlorite flakes (grey-green) to be orientated within the incipient strain-slip cleavage planes.

B. Microfolding in the field.
Plate 2. A and B. Mafic layering in the small peridotite outcrop at a height of about 1080 metres in the central portion of the Kaldhussaeter mass.
Plate 3. 

A. Eclogite boudin (just above and to the right of the hammer head) in gneiss at Kaldhussaeter, Tafjord.

B. Ranukdalen peridotite diapir (large intrusion at the northern end of Ranukdalen - see Fig. 5).
Plate 4.  A. The intricately sheared contact between eclogite (the darker rock) and anorthosite at the southern end of Viksvatn.

B. Small eclogite lens (pale coloured area just above the hammer head) in injected amphibolite gneiss - at Ugelvik, Otterøy.
Plate 5.  


B. Dimensional orientation and cataclasis of olivine grains in Ranukdalen chromite dunite. In the lower portion with the finely cataclased olivine grains note also the cataclasis of the large chromites (isotropic) leaving trains of smaller grains. Photomicrograph. Magnification x 8. Crossed nicols.
Plate 6. Photomicrographs of garnet exsolution in clinopyroxene within a garnet peridotite from the Grubse mass, Almkløvdalen. Two garnet lamellae can be seen in the large clinopyroxene grain on the left side of the photograph.

A. Magnification x 8. Plane polarized light.
   The high relief of the garnet lamellae is obvious.

B. Magnification x 8. Crossed nicols.
   The garnet lamellae are isotropic.
Plate 7. A and B. Pronounced reaction rims around garnets in Ugelvik garnet serpentinites.


Plate 9.  

A. Western extremity of garnetiferous peridotite lens (the darker rock) within dunite at Kalskaret, Tafjord.

B. Eastern extremity of the main garnet peridotite lens (the darker rock in left centre of photograph) within dunite at Kalskaret. Note also the peridotite with sparse pseudomorphs after garnet in the foreground.
Plate 10. A. Photomicrograph of folding and shearing in the Kaldhussaeter peridotite mass. Note the large unclastic amphibole grains along the right margin of the photograph. These are considered to be of secondary growth after the intense microfolding and shearing connected with the tectonic emplacement of the peridotite masses.
Magnification x 5. Plane polarized light.

B. Folding in peridotite in the Upper Ranukdalen mass, outlined by thin enstatite rich layers. Note the strain slip cleavage particularly obvious on the left hand side of the photograph.