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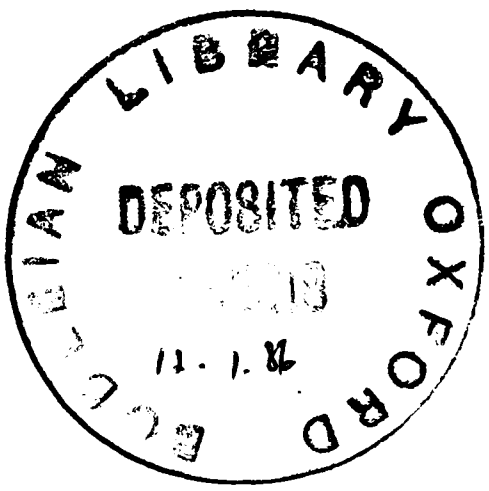

RADIOGENIC ISOTOPE STUDIES OF CRUST-FORMING PROCESSES IN THE LOFOTEN-VESTERÅLEN PROVINCE OF NORTH NORWAY

(Volume One: Text)

by

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Short Abstract

The Lofoten-Vesterålen province of North Norway consists almost exclusively of Precambrian granulite-facies rocks. The oldest rocks in the province are monzonitic and dioritic migmatitic gneisses, the protoliths of which were formed at 2.7 Ga.

The migmatites are overlain by a series of supracrustal gneisses, from 2.1 Ga, largely volcanogenic in origin, but with interbedded marbles and banded ironstones. The first occurrence of marble in western Lofoten is reported. Deposition in a subsiding back-arc basin or in an Andean-type environment on a thin continental margin is inferred.

Both gneiss sequences were intruded by basic rocks at 1.8 Ga. The basic rocks could not have been formed simply by extraction from the mantle at 1.8 Ga. The required contribution from 2.7 Ga migmatites could be as much as 37%, but less if contamination took place via anatectic melts. The first report of eclogitic rocks from the Lofoten-Vesterålen province is made in this study; their formation is associated with shear deformation.

Both gneiss sequences and the basic rocks were intruded by mangeritic rocks at 1.8 - 1.7 Ga. Their chemical compositions can be explained by fractional crystallization from magmas formed from 2.7 Ga and 2.1 Ga gneisses and 1.8 Ga mantle-derived magmas. Parameters derived from Rb-Sr, Sm-Nd and U-Pb systems to express the relative proportions of crust and mantle contributions to the mangerites mutually correlate, supporting the crust-mantle source model for the mangerites. Mixing calculations suggest that the late Archaean contributes in excess of 50% by mass for almost all mangerites.

Anatectic veins present, especially in the Moskenesøy supracrustal gneisses, are inferred to represent partial melts which coalesced to form the mangerites at higher structural levels. Anatexis was caused by basaltic underplating associated with limited crustal extension.

Later rock-forming events were the emplacement of dolerite dykes; the 1.65 Ga Lødingen Granite; the Leknes Group metasediments and the Caledonian granite pegmatites.

Long Abstract

Crust-forming processes in the Lofoten-Vesterålen province of North Norway have been investigated using detailed fieldwork, petrography, geochemical and isotopic methods.

The Lofoten and Vesterålen islands form an archipelago off the west coast of Norway opposite Narvik.

The oldest rocks in the province are a series of migmatitic gneisses which are most extensive on Moskenesøy and Langøy, at the deepest structural level in the area. These migmatites are monzonitic or dioritic in composition, they are foliated in some places but not in others and have granulite or amphibolite-facies mineral assemblages. The development of amphibolite-facies assemblages is a retrogressive stage. Sm-Nd model ages have been used to date the age of formation of the protoliths of the migmatites. A late Archaean age, 2.7 Ga has been obtained for migmatites from Langøy and the existence of Archaean gneisses on Austvågøy has been demonstrated for the first time. The Moskenesøy migmatites are distinctly younger. The Rb-Sr isotopic systems of the late Archaean rocks at Festvåg, Austvågøy and Sandset, Langøy have been disturbed subsequent to the formation of the gneiss protoliths. The present scattered distribution of data-points for these gneisses can be explained as due to Rb-depletion during granulite - facies metamorphism at 1.8 Ga.

The migmatitic gneisses are overlain by a series of supracrustal gneisses. The bulk of these gneisses are volcanogenic in origin, ranging from basaltic trachyandesites to rhyolites. Marble layers are present in this gneiss sequence. The present study has revealed the first reported marble layer in western Lofoten. The mineral assemblages in this zoned

marble layer have been used to obtain a minimum temperature for the granulite-facies metamorphism at Ølkona, Moskenesøy of 750°C. Other work (Krogh, 1975) has shown that the temperature reached 950°C in the structurally deeper parts of the province, so a relatively small temperature increase could have produced granitic melts. The migmatitic and supracrustal gneisses of Moskenesøy and Flakstadøy are more basic than those at higher structural levels which could be the result of extraction of relatively acid melts.

The mineral assemblages and the reactions in the Ølkona Marble show that P_{CO_2} decreased in the layer with time. The presence of marbles and banded ironstones in the thick volcanogenic pile suggests that the supracrustal gneisses were deposited in a back-arc basin in which the rate of basin subsidence was similar to the rate of infill from the andesitic lavas, limestones and haematitic cherts formed when the marginal sea was at the appropriate depths. The formation of the protoliths of the rhyolitic gneisses at Vik, western Vestvågøy has been dated at 2.1 Ga using Sm-Nd model ages. A supracrustal gneiss sample from Flakstadøy yields a Sm-Nd T_{DM} age of 2.3 Ga suggesting that a component of its Nd may have been derived from the migmatitic gneisses. The Rb-Sr isotopic systems of the Vik gneisses have been disturbed. The scattered distribution of data-points can by explained, in a similar way to that for the late Archaean migmatites, by disturbance during the retrograde metamorphism at 1.1 Ga.

Both the migmatitic and supracrustal gneisses are intruded by basic rocks, either small gabbroic bodies or larger intrusions containing anorthosites. The basic intrusion on Flakstadøy has well-developed cumulate textures and magmatic sedimentation features. Rb-Sr and Sm-Nd isochron ages for the Flakstadøy Basic Complex are both within error of 1.8 Ga. The precision of the Rb-Sr age is very poor because of the very low Rb concentrations

in the basic rocks. Evolution diagrams for the Sm-Nd system show that the basic rocks could not have been derived entirely from the depleted (or undepleted) mantle at 1.8 Ga. Estimates of the proportion of late Archaean migmatite-derived material in the mantle-derived magmas to produce the observed isotopic characteristics in the basic rocks have been made using typical values for Nd concentrations and isotopic compositions of the components involved. Similar calculations are made using Pb isotopes. An origin from the mantle at 1.8 Ga can be ruled out for any finite present day μ value in the basic rocks, since the samples with the least radiogenic Pb plot very close to the 1.8 Ga first-stage isochron on the $^{207}Pb/^{204}Pb$ v. $^{206}Pb/^{204}Pb$ diagram. For typical Nd and Pb concentrations in an ancient lower crustal rock the crustal contribution by mass may be as much as 37%. If contamination took place via anatectic melts, the necessary crustal contribution by mass would be reduced. The high crustal contribution to the Napp Gabbro implied by Pb isotopes is interpreted as due to marginal assimilation of the adjacent South-West Lofoten Mangerite, because of the clear relationship between the marginal basic samples and a mangerite component in chemical variation diagrams for the Flakstadøy Basic Complex.

The first report of eclogitic rocks from the Lofoten-Vesterålen province is made in this study. Eclogite mineral assemblages were developed in shear zones in the gabbroic rocks. Grain size reduction, surface area increase and an increase in pathways causes reactions to eclogite parageneses to take place in parts of the rock that have undergone shear deformation while the undeformed parts remain in granulite facies. The eclogitic rocks have subsequently been retrogressed to amphibolite facies. Glaucophane is present in some samples. A qualitative P,T path has been constructed, which involves essentially isobaric cooling after crystallization at $P \leq 8kb$ and a temperature of about $1150^{o}C$.

The exposed basic intrusions are subordinate in extent compared to the mangeritic intrusions emplaced into the migmatitic and supracrustal gneisses at 1.8 Ga. The mangeritic rocks are coarsely crystalline rocks which are only rarely foliated. Some of the mangerites, especially the Raftsund Mangerite, show magmatic sedimentation features and well-developed cumulate textures. The Raftsund Mangerite is concentrically zoned on the basis of mafic phase assemblages. The Hopen Mangerite/Charnockite has an apatite-rich layered gabbroic margin adjacent to the basement gneisses. The mangerites are clearly intrusive into the basic intrusions on Vestvågøy, but the relationship on Flakstadøy and Langøy is uncertain, but mangeritic or charnockitic veins in the basic intrusions on these islands suggest that the mangerites are the later intrusions. The mangerites, especially Sund, Hopen, Engenykken and parts of Raftsund, contain abundant xenoliths, some of which resemble parts of the migmatitic gneisses. The xenoliths are typically well-rounded, zoned and may have reaction rims; all are more mafic than the host mangerite, suggesting that they may be restites after removal of leucocratic melts.

The mangerites show a wide range in chemical compositions (e.g. SiO_2 = 51-77%); the widest range is for the South-West Lofoten Mangerite, the smallest is for the Raftsund Mangerite. The mangerites, Hopen and Raftsund in particular, are extremely Fe-enriched. They have high K_2O concentrations and most are shoshonitic in composition. The chemistry of the mangerites, by analogy with Phanerozoic andesites, suggests that they were formed in the orogenic zone of a continental region. The evolution of the mangerite magmas has been modelled using variation diagrams to compare their compositions to those of possible end-member components contributing to the magmas, and to the composition of possible early fractionating phases. These diagrams are consistent with the formation

of the mangerites from anatexis of average 2.7 Ga migmatitic and 2.1 Ga supracrustal gneisses, and also of more acid gneisses, and from a mantle-derived basic magma. The variation diagrams are consistent with the magmas formed from these components subsequently evolving by fractional crystallization of pyroxenes, olivine, apatite, ilmenite and magnetite. These minerals are concentrated in the marginal gabbro to the Hopen intrusion.

Rb-Sr whole-rock isochron ages of 1.73 to 1.81 Ga have been obtained for the Sund-Ølkona, Hopen, Austvågøy Grey and Raftsund Mangerites. An older imprecise age was obtained for the South-West Lofoten Mangerite due to the wide scatter of data-points. Initial $^{87}Sr/^{86}Sr$ ratios range from 0.7032 to 0.7048 which implies that the mangerites were derived from either upper mantle or granulite-facies depleted lower crust.

The narrow range in Sm/Nd ratios for the mangeritic intrusions effectively rules out dating by Sm-Nd whole-rock isochron methods. Sm-Nd T_{CHUR} and T_{DM} model ages have been calculated and are shown on evolution diagrams. ε_{Nd} values at 1.8 Ga preclude an origin for the mangerites by crystallization from a magma derived solely from the mantle at 1.8 Ga. The model age, T_{DM}, of a mangerite sample can be used to estimate the proportions of different components present in that sample.

In the $^{207}Pb/^{204}Pb$ v. $^{206}Pb/^{204}Pb$ plots for the mangerites, a number of whole-rock and alkali feldspar separates plot to the left of the 1.8 Ga first-stage isochron. A single-stage evolution for these rocks until 1.8 Ga (i.e. in the mantle) is therefore totally impossible. These unradiogenic samples have been used to constrain μ, for the source of an ancient crustal component to between 7.85 and 8.1. The Pb isotope compositions have been explained using a model in which most of the Pb in the mangerites has evolved in three stages: in the mantle until 2.7 Ga,

in the migmatitic gneisses from 2.7 Ga until 1.8 Ga, and in the mangerites subsequent to 1.8 Ga. The second component in the model is Pb derived from the mantle at 1.8 Ga. The linear array of data in the $^{207}Pb/^{204}Pb$ v. $^{206}Pb/^{204}Pb$ diagram is a consequence of the different μ values for the two components in the interval 2.7 - 1.8 Ga being reflected in the magmas resulting from mixing different proportions of those components. "Plumbotectonics" confirms that the mangerites were derived from upper mantle and lower crustal sources. Apparent μ_1 values for individual samples can be used to estimate the total contribution of Pb from lower crustal sources to those samples in the same way as T_{DM} can be used to estimate the total contribution of Nd from the lower crust. Both have also been used to estimate total mass contributions from the lower crust. Mixing calculations for Nd suggest that the maximum contribution by mass from the 2.7 Ga crust to the mangerites is 60-80%. Fractional crystallization of mantle-derived magmas would progressively raise Nd concentrations in the residual magmas. Hence if mixing of crust- and mantle-derived magmas took place after such fractionation, a much smaller contribution by mass is required from the mantle-derived component. Pb too would be concentrated in residual magmas after fractional crystallization. If it were concentrated to as much as 20ppm prior to mixing, all except three samples would require a contribution by mass from the 2.7 Ga lower crust of over 50%.

Correlations of isotopic parameters derived from Rb-Sr (apparent initial $^{87}Sr/^{86}Sr$), Sm-Nd (T_{DM}), and U-Pb (apparent μ_1) systems to express the relative proportions of crust and mantle contributions to the mangerites show that the crust-mantle source model for the mangerite is fully self-consistent. The wide variations in isotope ratios in three systems and in chemistry, for the South-West Lofoten Mangerite, are all correlated. This strongly suggests that the scatter of data about the Rb-Sr isochron for this intrusion is due to variations in initial $^{87}Sr/^{86}Sr$ ratios rather

than to metamorphic disturbances. The samples with the largest mantle contribution for the South-West Lofoten Mangerite are from the side of the islands where the Moho is shallowest.

The mangerites crystallized at about 1000°C and 11kb. Retrograde reactions have subsequently taken place.

Anatectic veins and dykes of mangeritic and charnockitic composition are present in the structurally deepest parts of the Lofoten-Vesterålen province. In-situ anatexis has only been seen in one locality (Festvåg, Austvågøy) in the late Archaean migmatites, but is common in the early Proterozoic supracrustal gneisses where it is commonly associated with the emplacement of basic veins and dykes. The most convincingly in-situ anatectic melt, from Ølkona, Moskenesøy, is of intermediate composition. Its field relations, petrography and T_{DM} value suggest that this melt is derived solely from supracrustal gneisses. These gneisses are isotopically similar to mantle-derived magmas at 1.8 Ga, hence part of the "mantle-derived" component calculated for the mangerites is due to assimilation of juvenile continental crust. The estimates for total crustal contribution by mass to the mangerite intrusions are minimum values.

The anatectic melts are chemically similar to the mangeritic intrusions. High K_2O concentrations suggest derivation of the melts by reworking of continental crust.

Nd evolution diagrams are consistent with formation of the Sund-Ølkona and South-West Lofoten Mangerites by coalescing melts similar to those of mangeritic composition on Moskenesøy and Flakstadøy. The charnockite veins on Vestvågøy have chemical and isotopic similarities to the Flakstadøy Charnockite.

Despite the subordinate rôle of basic magmas in the material composition of the mangeritic intrusions, the input of heat associated with emplacement of basaltic magmas into the lower crust caused the crustal anatexis which formed the mangerite magmas. The transition from extrusive igneous activity in a subsiding basin to intrusive igneous activity and crustal thickening with reduced crustal extension is probably associated with closure of the ocean which resulted in the Sveco-Karelian orogeny. Uplift and erosion following mangerite emplacement may have resulted from this orogeny. The anatectic melts on Moskenesøy are at the deepest structural level in the province and the batholithic Raftsund Mangerite at the shallowest. Igneous processes at different levels in the crust can be studied in the Lofoten-Vesterålen Province.

Dolerite dykes intruded the Lofoten-Vesterålen province shortly after the mangerites were emplaced.

The Lødingen Granite is a very heterogeneous intrusion which is slightly, but distinctly younger, than the Raftsund Mangerite. (A 1.65 Ga Rb-Sr whole-rock isochron has been obtained for the Lødingen Granite). The granite probably formed in response to the emplacement of the mangerite which caused raising of crustal isotherms. T_{DM} model ages for the granite show that it contains a crustal component significantly older than 1.65 Ga. Sm-Nd isotopic parameters are similar to those for the mangerites .

The Leknes Group metasediments contain a mantle-derived component which was extracted from the mantle at, or later than, 1.6Ga, i.e. new crust was added to the Lofoten-Vesterålen province after the emplacement of the mangerites.

The granite pegmatites in the southern part of Moskenesøy are the youngest recognized rock-forming event in the Lofoten-Vesterålen province.

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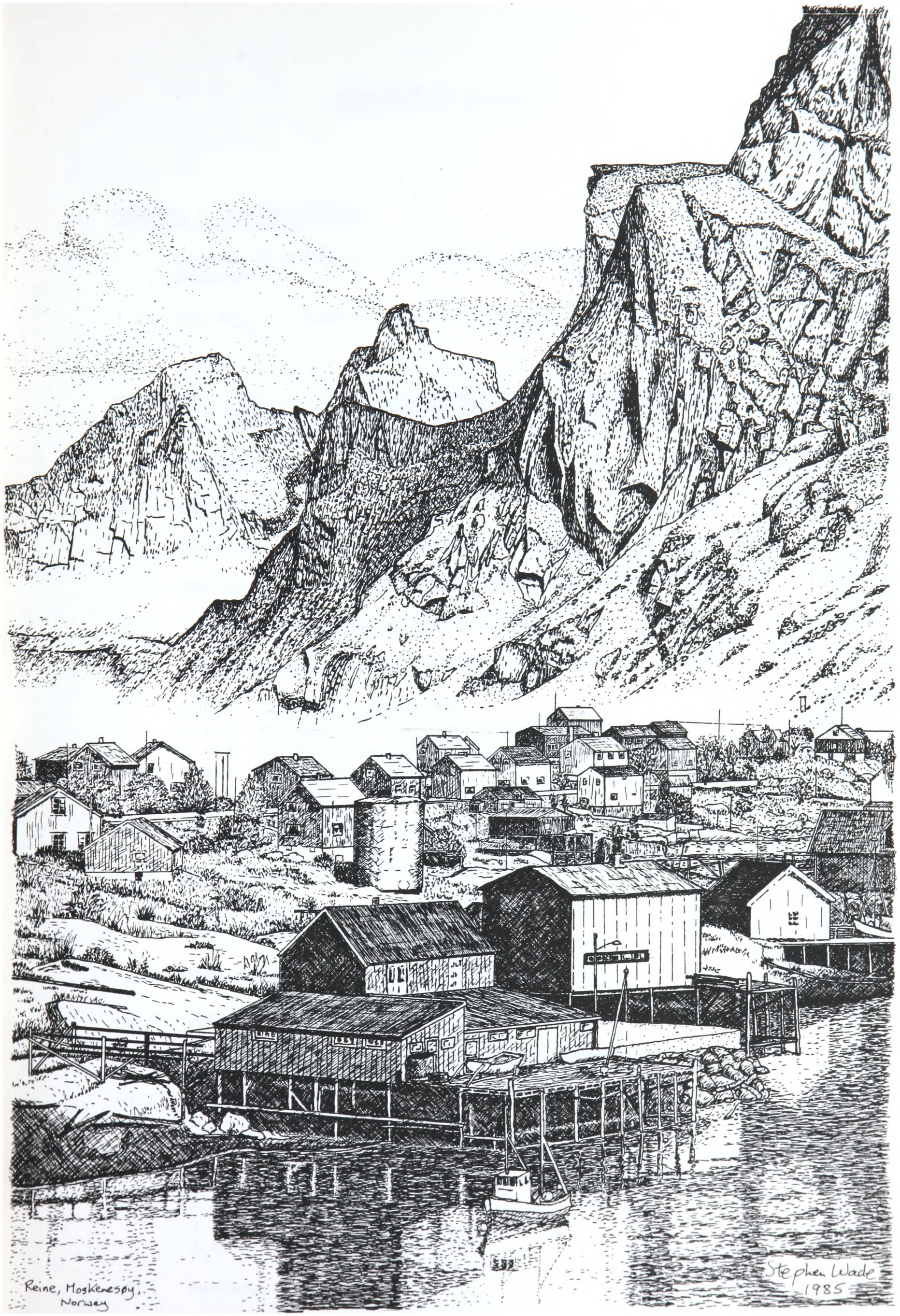
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Stephen Wade
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When we compare the present life of man with that time of which we have no knowledge, it seems to me like the swift flight of a lone sparrow through the banqueting-hall where you sit in the winter months. ... This sparrow flies swiftly in through one door of the hall, and out through another. ... Similarly, man appears on earth for a little while, but we know nothing of what went on before this life, and what follows.

THE VENERABLE BEDE 673-735

[History of the English Church and People, 2. Ch.13, transl. L. Sherley-Price]

CHAPTER 1. INTRODUCTION Pages 1-8.

CHAPTER 1. INTRODUCTION

1.1 Lofoten-Vesterålen - Its People, Lifestyle and Geographical Setting

Geographical Setting

The Lofoten-Vesterålen province is situated in northern Norway, in the counties of Nordland and Troms. The area of study is contained between 67°45'N and 69°00'N latitude, i.e. between about 100km and 250km north of the Arctic Circle, and between 12°45'E and 16°00'E latitude (Fig.1.1).

The Lofoten Islands are an archipelago extending from the Raftsund in the North-east, 120km to the Moskenstraum in the South-west. The small islands of Vaerøy and Røst which lie south of the Moskenstraum are not considered in this study. Lofoten consists of the islands of Moskenesøy (South-west end), Flakstadøy, Vestvågøy, Gimsøy and Austvågøy, and also the south-west tip of Hinnøy across the Raftsund, together with numerous smaller islands. Vesterålen consists of the islands of Hadseløy, Langøy (the third largest in Norway), Andøy,and the west of Hinnøy, Norway's largest island, plus many smaller islands. The study area also includes Lødingen, in the South-east of Hinnøy, and Hamarøy, a peninsula on the mainland, across the Vestfjord, 40km South-east of Svolvaer.

The Vestfjord separates the Lofoten Islands from the mainland. It is 200km long, and 90km wide at its mouth, but narrows towards the North-east to 15km wide off Lødingen. To the east, the Vestfjord passes into Ofotfjord which gives ships access to the rail-head town of Narvik; to the north it passes into Tjeldsundwhich separates Hinnøy from the mainland; to the south is Tysfjord.

To the north of the Lofoten-Vesterålen islands lies the Norwegian Sea. This is connected to the Vestfjord by channels between each of the

islands, which are so narrow that the Lofoten Islands, viewed from across the Vestfjord, appear as one continuous land-mass; this, combined with the precipitous mountains of the Lofoten Islands, had led to the archipelago being dubbed the "Lofoten Wall".

The landscape of Lofoten-Vesterålen was shaped during the Pleistocene glaciation. The area contains a wealth of glacial features including cirques, aretes, corries and fjords. Remnant icefields are preserved in the Trolltind Mountains (1161m) on Austvågøy and on Møysalen (1266m) in South-west Hinnøy - the highest peaks in the area. Many of the peaks on Lofoten reach over 700m and are angular; few peaks on Vesterålen exceed this height and most have a more rounded appearance. Moraines were deposited on Andøy, Langøy and North Hinnøy at the end of the last glacial epoch; these deposits are responsible for the extensive areas of marshland which characterize these areas today.

Numerous deeply-incised fjords separate the islands and almost subdivide them into smaller ones. Most fjords in Lofoten-Vesterålen trend North-South or North-east-South-west.

Climate and Seasons

The Gulf Stream leads to a surprisingly moderate climate in Lofoten-Vesterålen. The natural barrier of the "Lofoten Wall" protects the Vestfjord side of the islands from the northerly winds and often produces a marked contrast in weather across the islands. The warmest months in Lofoten-Vesterålen are July and August with average daily temperatures of 12-14°C; during the winter months the average temperature is -1 to -2°C. The range of precipitation in the area is 714 to 1523mm (28-60 ins) (Helland, 1897, p.37). In the north of the study area, in Vesterålen, the Midnight Sun is visible from approximately May 20 until July 24; in

contrast, the sun is never above the horizon between the end of November and mid-January. The Northern Lights or aurora borealis are visible at this time. More tourists see the Midnight Sun than see the Northern Lights.

Population

The islands were inhabited at least as far back as the Stone Age, when the islanders lived by fishing and hunting and then migrated seasonally. The regional parliament was held in Kabelvåg, Austvågøy in Viking times and into the early Middle Ages. The population increased dramatically in the latter half of the 19th Century as the fish trade with the south of Norway grew in importance and people migrated north.

The present population is: Lofoten 27,000 (Svolvaer 4100), Vesterålen 35,000 (Andenes 4200; Sortland 4000; Stokmarknes 3000; Melbu 2000). Most of the settlement of the Lofoten Islands has taken place on a coastal strip of flat land between the mountains and the sea, three-quarters of the population living on the Vestfjord side of the islands.

Communications

The roads on and approaching the islands are rapidly being improved; thirty years ago the sparcity of roads on the islands meant that travel by car was not feasible. Today the E6, Rome to Kirkenes highway is sealed as far north as Narvik (and much further generally) and only one ferry crossing remains. Access may be had to the Lofoten Islands via the Skutvik-Svolvaer ferry and to Vesterålen via either the Bognes-Lødingen ferry or via Riksvei 19 which passes through most of the islands, from Narvik to Å in the South of Moskenesøy. Ten years ago there were seven ferries between Narvik and Å; now there are only two. As well as tunnelling, bridge-building and surface improvements on Rv19, other

major road-construction schemes in the area in the past decade have included: a bridge to the fishing village of Henningsvaer; a road and tunnel through the mountains of Central Langøy; the Svolvaer by-pass; a road on the eastern side of Raftsund; and a road to Lonkan, Western Hinnøy to Rv19. All of the road improvements, plus extensions to airfields, have exposed fresh sections which are of great benefit to those who have to carry out extensive sampling for geochemical studies in such a mountainous and moss-covered terrain.

Despite the recent improvements in roads, Lofoten-Vesterålen is still dependent on sea-borne transport. As well as regular ferry services on the major routes, less frequent boat services operate to serve the communities of the smaller islands. The coastal steamers supply North Norway from Bergen and Trondheim; they call daily at Stamsund, Svolvaer, Stokmarknes, Sortland, Risøyhamn and Harstad. Other freight is shipped to the islands from the railheads of Bodø and Narvik. The line to Bodø goes south to Oslo; the Narvik line goes east to Stockholm - its principal purpose is the transport of ore from the Swedish iron mines at Kiruna and Gällivare.

The islands are served by airfields at Leknes, Svolvaer, Skagen on Langøy, Andenes on Andøy, and Evenes between Harstad and Narvik.

Trade and Industry

Between January and April of each year, cod migrate from the Barents Sea to the shallow coastal banks near the Lofoten Islands to spawn. The Lofoten Fisheries have been important since the time of the Sagas and are still the centre of the local economy. The yield per year has varied between 19.5 and 146 million Kg, each cod weighing, on average, 3.5kg. The fish processing industry (including cod roe

'caviar') is also important. In recent years fish breeding has become prominent.

Farming, principally animal farming, is a valuable industry across the whole area. Other important commercial activities include service industries, mercantile shipping, construction and tourism.

Tourism

The Pleistocene glaciation has produced some spectacular scenery in the Lofoten-Vesterålen province, which is the principal reason for the every-growing number of tourists in the area during the summer months. The razor-sharp peaks rising to 100m from deep, blue fjords produce some interesting challenges for photographers and rock-climbers alike. The Midnight Sun and the Lofoten fisheries also attract tourists.

The highlights of the tourists' itinerary include: Å, at the end of the road on Moskenesøy, sporting the world's shortest place-name; Reine and Hamnøy on Moskenesøy, for the classic combination of mountains, fjords and fishing villages; Raftsund, Trollfjord and the Trolltind Mountains, and also Reka on Langøy for stunning views and challenging climbs.

1.2 The Present Study

The Lofoten-Vesterålen province of north Norway is a high-grade metamorphic terrain consisting of two generations of basement gneiss and a series of intrusions, emplaced in the mid-Proterozoic. These intrusions are of intermediate and basic compositions; the intrusions with intermediate compositions are more areally extensive than the basic rocks at the present level of erosion, and occupy about half of the area of the Lofoten-Vesterålen terrain. These intermediate rocks are

mangerites and charnockites; they are rocks of monzonitic and granitic composition (Streckeisen 1976) respectively, which have crystallized at granulite-facies and as such contain modal ortho-pyroxene. Malm and Ormaasen (1978) sub-divided the mangeritic rocks thus:

Basic mangerite	:	<56 weight per cent SiO_2
Mangerite	:	56-62 weight per cent SiO_2
Acidic mangerite	:	62-67 weight per cent SiO_2
Charnockite	:	>67 weight per cent SiO_2

The basic rocks are anorthosites, norites, troctolites and gabbros. Griffin et al. (1978), in their major review paper on research in the Lofoten-Vesterålen province, deduced that an origin for the mangerites by extensive anatexis of the observed basement rocks was excluded by the low initial $^{87}Sr/^{86}Sr$ [or $(^{87}Sr/^{86}Sr)_o$] ratios of the mangerite. Subsequently it has been recognised that depleted, lower crustal granulites can have low Rb/Sr ratios, identical to those of the upper mantle, and hence rocks derived by extensive anatexis of ancient granulite-facies gneisses would have low $(^{87}Sr/^{86}Sr)_o$, indistinguishable from rocks derived from the upper mantle (Spooner and Fairbairn, 1970; Taylor et al., 1984).

Griffin et al. (1978) also presented Pb-isotope data from the largest mangerite pluton in Lofoten-Vesterålen (the Raftsund Mangerite) which showed that a two-stage model was not feasible for this intrusion and that at least some of its lead was probably derived from ancient lower crustal rocks, depleted in U.

The re-evaluation of the significance of low $(^{87}Sr/^{86}Sr)_o$ values and the evidence from the limited amount of Pb-isotope data necessitated

further isotopic studies to elucidate the nature of the origin of the mangeritic rocks of Lofoten-Vesterålen.

Unlike Sr isotopes, both Pb and Nd isotopes can distinguish conclusively between a lower-crustal and an upper-mantle origin for igneous rocks (Taylor et al., 1984; Moorbath and Taylor, 1985). Most of the work in this thesis consists of the use of these isotopic systems to answer the question: do the mangeritic rocks of Lofoten-Vesterålen represent an addition of crustal material from the mantle or re-working of ancient continental crust? This is part of the general question of crustal addition versus crustal re-working in the Proterozoic.

The basic rocks are associated both spatially and temporally with the mangeritic rocks. An understanding of the nature of this relationship is essential to fully understand the origin of the mangerites. Are the basic and mangeritic rocks comagmatic? What are their contact relationships and their relative ages? Are the areas occupied at surface by the two magma types proportional to their total volumes? If the mangerites are produced by extensive anatexis of the lower crust, could the basic rocks have provided the thermal input to cause the melting?

After the emplacement of the basic and mangeritic rocks, a heterogeneous granite was emplaced in the east of the area around Lødingen. At the start of this work, the age of this granite was poorly constrained and the high $(^{87}Sr/^{86}Sr)_o$ led Griffin et al. (1978) to propose a crustal anatectic origin. This study aimed to produce a more reliable age and to re-examine the relationship of the granite to the earlier mangerites.

Chapter 1: a general background to the area; an introduction to the present study.

Chapter 2: an introduction to the geology and structure of Lofoten-Vesterålen as it was understood before this study, with most information taken from the synthesis and summary by Griffin et al. (1978).

Chapter 3: presents field observations, petrography, major element geochemistry and Nd and Sr isotope data for the late Archaean migmatites.

Chapter 4: presents data for the early Proterozoic supracrustals.

Chapter 5: contains field observations, petrography, major element geochemistry and Nd, Pb and Sr isotope data for the mid Proterozoic basic intrusions.

Chapter 6: presents data for the mangeritic rocks and develops models to explain their origin.

Chapter 7: presents key field observations demonstrating crustal anatexis as the mode of origin of stringers, dykes and sheets, together with associated petrography, major element geochemistry and Nd, Pb and Sr isotope data.

Chapter 8: presents field observations, petrography, major element geochemistry, Nd, Pb and Sr isotope data for lithologies younger than the mangerites.

Chapter 9: conclusions on the origin of the mangeritic rocks and other crust-forming events in Lofoten-Vesterålen.

CHAPTER 2. THE GEOLOGY, GEOCHRONOLOGY AND STRUCTURE OF THE LOFOTEN-VESTERÅLEN PROVINCE

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CHAPTER 2. THE GEOLOGY, GEOCHRONOLOGY AND STRUCTURE OF THE LOFOTEN-VESTERÅLEN PROVINCE

2.1 Introduction

The gneiss complex of Lofoten-Vesterålen has been divided into two principle units (Griffin et al., 1978): a late Archaean migmatite complex, apparently derived by anatexis of a supracrustal sequence, and an overlying series of early Proterozoic supracrustal rocks. The gneiss complex also includes the Gullesfjord granitic gneiss on Hinnøy, which intruded the migmatites 2600-2700 Ma ago (Griffin et al., 1978). The gneiss complex was regionally metamorphosed to granulite facies over most of the area (Heier, 1960; Glassley, 1975) at about 1830 Ma ago (Griffin et al., 1978).

During and shortly after the peak of the granulite-facies metamorphism, the gneisses were intruded by a series of plutons, which now occupy about half the area of Lofoten-Vesterålen. Most of the intrusions crystallized with granulite-facies mineral assemblages, and consist of charnockites, mangerites, gabbros, anorthosites and ultramafics. Shortly after emplacement, these intrusions were cut by a dolerite dyke swarm (Misra and Griffin, 1972). This was followed by the emplacement of the Lødingen Granite on Hinnøy (Griffin et al., 1978; Andresen and Tull, 1983).

The Leknes Group of pelitic and semipelitic metasediments and amphibolites occurs on Vestvågøy (Tull, 1977) and Vaerøy, south of Moskenesøy (Griffin and Taylor, 1978). They were metamorphosed to amphibolite facies at about 1150 Ma when there was a widespread retrogression of the older granulite-facies rocks (Griffin et al., 1978).

The Caledonian orogeny had little effect on Lofoten-Vesterålen, being

limited to minor thrust faulting, partial resetting of K-Ar systems, and the intrusion of some large granitic pegmatites on Moskenesøy. Fig. 2.1 is a simplified geological map of Lofoten-Vesterålen after Griffin et al. (1978).

2.2 Geophysics

Important information on the structure of Lofoten-Vesterålen has been obtained from geophysical studies of the area.

On the Bouguer anomaly map of northern Norway (Norges Geografiske Oppmåling, 1960), the Lofoten-Vesterålen area is characterized by an elongate positive gravity anomaly, with a peak of 138 mgal at the southern end, centred on Hamnøy, Moskenesøy (Fig. 2.2). This anomaly, and another to the west, are part of a nearly continuous belt of gravity highs extending to the south-west along the continental margin to the Precambrian Lewisian Complex of north-west Scotland (Talwani and Eldholm, 1972).

The rocks at the surface of Lofoten-Vesterålen are exceptionally dense (Svela, 1971): the granulite-facies migmatites average 2.86 $g.cm^{-3}$ and the early Proterozoic supracrustal rocks average 2.70 $g.cm^{-3}$. The fact that the mangerites tend to occur in the early Proterozoic supracrustal rocks or near the migmatite/supracrustal contact may be explained by the density contrast between the magma and the country rocks. The mangerite magmas would have densities of 2.60 - 2.70 $g.cm^{-3}$ (Griffin et al., 1978) and hence would have considerable buoyancy while in the granulite-facies migmatites, but much less when they had risen into the overlying supracrustal rocks. Despite the high density of these rocks, an upwarping of the Moho, along an axis running to the west of the island chain from Hamnøy to Stø, on the northern tip of Langøy, is required to explain the observed gravity (Kjenes, 1970; Svela, 1971). In addition,

it is necessary to postulate a plug-like extension of mantle material, or a large basic body in the lower crust, to explain the anomaly peak at Hamnøy.

A reversed seismic refraction profile from Hamnøy to Stø (Fig. 2.2) confirmed the presence of an upwarp of the Moho and yielded evidence of a shallow (11km) Conrad discontinuity (Sellevoll and Thanvarachorn, 1977). There is also a discontinuity at about 20km depth, at which Vp increases from 6.6 to 7.1.

The gravity and seismic data imply that the depth to Moho increases from only about 25km between Stø and Hamnøy along the western side of the islands, to 35-37km on the eastern side of Hinnøy (Kjenes, 1970; Svela, 1971; Sellevoll and Thanvarachorn, 1977).

These studies show that:

(i) Lofoten-Vesterålen is underlain by rocks that are more dense than typical continental crust;

(ii) the Conrad discontinuity is near the surface on Moskenesøy, in the south-west of the region; and

(iii) the crust in the west of the region is unusually thin for an Archaean terrain.

A deep section through the earth's crust is exposed at the surface in the Lofoten-Vesterålen province, with the deepest part exposed along the west coast of Langøy and Moskenesøy.

2.3 The Late Archaean Migmatites, c. 2700Ma

The migmatite complex crops out over large areas of Moskenesøy, Langøy and Hinnøy, generally at the deepest crustal level exposed in Lofoten-Vesterålen. The migmatites on Langøy and Hinnøy are heterogeneous on outcrop scale and the more mobile parts clearly enclose and intrude the less mobile parts although Griffin et al. (1978) state that it is difficult to establish a clear distinction between palaeosome and neosome. Griffin et al. (1978) infer from field relations, that these migmatites formed when a heterogeneous series of rocks, layered on a scale of up to metres, was heated to temperatures at which some of the more leucocratic parts melted. This anatexis was accompanied by folding and boudinage but did not involve a significant juvenile igneous component.

Green and Jorde (1971) reported that the migmatites crop out over most of Moskenesøy. Most of this group consists of intermingled monzonitic and subordinate dioritic gneiss. The monzonitic gneiss is characterized by large (2-3cm) porphyroblasts of mesoperthite giving the rock an augen gneiss texture. The migmatites in the northern part of the island have been extensively retrograded from granulite facies.

On south-west Hinnøy, the migmatites enclose a large area of mafic Archaean supracrustal gneisses. The contact
can seldom be located to ± 100m (Griffin et al., 1978).
Other outcrops believed to be part of the Archaean migmatite complex include the leucocratic gneiss mapped by Romey (1971) on north-west Flakstadøy and the migmatites of Laukvik, northern Austvågøy (Griffin, 1971) and at Festvåg and Kabelvåg, south-west Austvågøy (Tveten, 1978).

An orthopyroxene isograd runs across eastern Langøy and south-west

Hinnøy, sub-parallel to two isograds defined by mineral parageneses in carbonate rocks (Glassley, 1975). The migmatites are in granulite facies to the west of the orthopyroxene isograd. There are no structural differences between the migmatites on either side of the isograd suggesting that the granulite-facies assemblages were developed in the rocks after they had been migmatised. The metamorphic reactions will be considered later.

Chemical Composition

Chemical analyses from the migmatite complex have been published in Heier and Thoresen (1971) and Griffin et al. (1978). The data from Griffin et al. for migmatites from Moskenesøy, Langøy and Hinnøy is reproduced here as Table 2.1.

Heier and Thoresen (1971) concluded that the migmatites are, on average, of intermediate composition and that there is no major variation in the chemistry of the migmatites across the orthopyroxene isograd. Griffin et al. (1978) stated that the analyses presented in their work (Table 2.1 in this work) probably gave a fair picture of the range of compositions of the migmatites on each island, but that the average values were more likely to be inaccurate due to sampling bias. Nevertheless, marked differences in average SiO_2 content can be seen for the migmatites of the three islands considered. The migmatites at Vikan in West Langøy are more silicic than the others and those from Moskenesøy are more basic. The Hinnøy migmatites are intermediate in composition with high Na+K and are very similar to the average of all the migmatites analysed. The more silicic composition of the Vikan migmatites is quite similar to that of other Archaean migmatites, but the Lofoten-Vesterålen migmatite complex as a whole is less silicic and more alkali-rich than other gneiss terrains of similar age (Griffin et al., 1978).

Griffin et al. (1978) and Krogh (1977) infer that the original Archaean supracrustal complex consisted largely of volcanic and volcaniclastic rocks prior to the anatexis that formed the migmatite complex. The average composition is similar to that of Cenozoic K-rich andesites of the Chilean Andes but with a higher $Na_2O + K_2O$ and lower CaO.

Isotopic Data

A Pb/Pb whole-rock isochron age of 3410 ± 70Ma was published by Taylor (1975) for the granulite-facies migmatitic gneisses at Vikan on west Langøy. This date was inferred to be the age of the migmatite-forming event, as U-Pb whole-rock systems are unlikely to remain closed during anatexis. Jacobsen and Wasserburg (1978) published T_{CHUR} Nd model ages for three Vikan samples, the oldest of which gave T_{CHUR} = 2640 ± 30Ma, and interpreted the Vikan gneisses to be part of the world-wide crustal-accretion differentiation superevent (Moorbath and Taylor, 1981) at about 2600Ma. Griffin et al. (1978) reinterpreted the Pb-Pb data due to some anomalous features of the isotope systematics. The high μ_1 ($^{238}U/^{204}Pb$) value of 8.9 implied for the source region of the Vikan gneisses using a two-stage model, and the absence of any particularly non-radiogenic Pb isotope compositions, are unusual when compared to Pb-Pb data from other Archaean terrains. Determinations of the U and Pb concentrations in the Vikan gneisses show that the present range of Pb-isotope compositions cannot have resulted from a single-stage evolution of Pb-isotope compositions from a uniform initial Pb-isotope composition over any reasonable period of time.

Griffin et al. (1978) and Moorbath and Taylor (1981) have shown that the alignment of Pb-Pb data reported by Taylor (1975) could have resulted from addition of the protoliths of the Vikan gneisses from mantle to crust at ∿ 2700Ma with a range of U/Pb ratios mostly character-

istic of upper continental crust, followed by severe U-depletion during granulite-facies metamorphism at ∿ 1800Ma. This interpretation accounts for the low, fairly uniform present-day $^{238}U/^{204}Pb$ ratios of the Vikan gneisses. Hence the alignment of data is now interpreted as a 'transposed palaeo-isochron' (Moorbath and Taylor, 1981).

A Pb/Pb isochron age of 2685 ± 65Ma (Table 2.3, after Griffin et al., 1978) for quartzo-feldspathic migmatites from four localities on NW Hinnøy, east of the area affected by the 1800Ma granulite-facies metamorphism, also provided clues to the currently accepted model of three-stage Pb-isotope evolution for the Vikan gneisses which were affected by the 1800Ma granulite-facies metamorphism. The protoliths of the NW Hinnøy amphibolite-facies gneisses were derived from a mantle source with a μ_1 value of ∿7.8, which is similar to μ values obtained from most quartzo-feldspathic gneisses of the cratons around the North Atlantic (Moorbath and Taylor, 1981).

Jacobsen and Wasserburg (1978) analysed two NW Hinnøy amphibolite-facies gneiss samples for Nd; both had T_{CHUR} = 2670 ± 30Ma, within error of the T_{CHUR} model ages of three of the four samples analysed from the granulite-facies migmatites of Langøy and hence concluded that Sm and Nd were not fractionated as a result of the 1800Ma granulite-facies metamorphism.

The Rb-Sr whole rock isochrons for the Vikan and NW Hinnøy gneisses give ages younger than the Pb/Pb ages: 2245 ± 150Ma for the Vikan gneiss with an initial $^{87}Sr/^{86}Sr$ ratio $[(^{87}Sr/^{86}Sr)_o]$ of 0.7126 ± 0.0011 and 2360 ± 185Ma, $(^{87}Sr/^{86}Sr)_o$ = 0.7080 ± 0.0029 for the NW Hinnøy gneisses (Table 2.3). The discrepancy has been explained by Griffin et al. (1978) as due to partial redistribution of Rb and Sr in the 1800Ma metamorphism.

The Gullesfjord granitic gneiss is a unit of granodioritic composition within the NW Hinnøy amphibolite-facies gneisses. Griffin et al. (1978) report a Rb-Sr whole-rock isochron age of 2715 ± 120Ma with $(^{87}Sr/^{86}Sr)_o$ = 0.6952 ± 0.0056. If it is accepted that $(^{87}Sr/^{86}Sr)_o$ should be > 0.700 then the true age of the Gullesfjord granitic gneiss is ∿2590Ma.

Jacobsen and Wasserburg (1978) reported a T_{CHUR} Nd model age of 2040 ± 40Ma for a monzonitic granulite from Reine, Moskenesøy and concluded that the protoliths of the Moskenesøy migmatites are distinctly younger than, and not related to, the protoliths of the migmatites of Langøy and Hinnøy.

Summary

The geological and geochronological data suggest that the Langøy and Hinnøy migmatites represent products of a single crust-forming event; the protoliths of the migmatites formed at ∿2700Ma. The single Nd model age from Moskenesøy suggests that at least some of the Moskenesøy migmatites are younger than those in Vesterålen. The Rb-Sr and U-Pb, but not the Sm-Nd isotopic systems of the migmatites have been disturbed by the 1800Ma granulite-facies metamorphism, which caused depletion of U and Rb.

2.4 The Early Proterozoic Supracrustal Rocks, c.2100-1800Ma

These rocks outcrop on all of the major islands in the study area and make up most of the gneiss complex on Flakstadøy, Vestvågøy and Austvågøy. The supracrustals (or paragneisses) are more fine-grained, more finely-foliated and more felsic than the rocks of the Archaean migmatites. Their fine-scale layering and thin aphitic stringers have led the supracrustals to be referred to as the Veined and Layered Gneisses on Moskenesøy and Flakstadøy (Green and Jorde, 1971, Romey, 1971). Unlike the chaotically heterogeneous late Archaean migmatites, the early Proterozoic supracrustals do not appear to have undergone extensive anatexis.

Most of the supracrustals are fine-grained quartz, two-feldspar gneisses, but quartz-rich plagioclase gneisses are dominant on Austvågøy and more basic rocks occur on Hinnøy. Rocks of indisputably metasedimentary origin occur throughout Lofoten-Vesterålen in the early Proterozoic supracrustal sequence. Marbles, graphite schist, calc-silicates, and iron formations are restricted to the early Proterozoic gneisses. Cross-bedding, quartzite conglomerates and possible pyroclastic bedding have been recognized locally and jasper is preserved in the Vestpolltind iron-formation on Hinnøy (Krogh, 1977). This is despite the isoclinal folding of the gneisses which has resulted in the original depositional layering becoming discontinuous. On Moskenesøy and Langøy, the early Proterozoic supracrustals occur as large scale isoclinal infolds in the migmatite complex. Green and Jorde (1971) described the contacts as sharp and concordant on Moskenesøy, but their nature is ambiguous.

Chemical Composition

Chemical analyses from Griffin et al. (1978) for the early Proterozoic supracrustal gneisses are reproduced here in Table 2.2.

The average composition of the supracrustal rocks is intermediate (excepting the limestones, iron formations and graphite schists) and the differences from the migmatites are not statistically significant. Griffin et al. (1978) stated that the average composition is approximated by the andesites of the Chilean Andes, but that it is more alkaline. Excluding the basic compositions (SiO_2 < 54%)), the remainder of the supracrustals have a similar bulk composition to Chilean rhyodacites.

Krogh (1975, 1977) worked on the banded iron formations in Lofoten-Vesterålen and concluded that they represent Mn-rich oxide facies and Mn-poor carbonate, oxide and silicate facies, metamorphosed in granulite facies. There are similarities to the lower grade leptite iron ores of central Sweden, which are also Svecofennian(1600-1800Ma) in age, and which are considered to be related to keratophyres, tuffs and tuffites (Magnusson, 1970). Krogh (1977) inferred from the high heat flow in Lofoten-Vesterålen at this time, that the supracrustal sequence formed in a back-arc basin, associated with volcanism above a subduction zone.

Isotopic Data

Griffin et al. (1978) reported a Rb-Sr whole-rock isochron age of 1830 ± 35 Ma with $(^{87}Sr/^{86}Sr)_o$ of 0.7037 ± 0.0005 (Table 2.3) from six localities in the supracrustal gneisses from Austvågøy and Langøy and interpreted it as referring to the time of granulite-facies metamorphism of these rocks. The low $(^{87}Sr/^{86}Sr)_o$ and present Rb and Sr content of these rocks suggest that their protoliths could not have been deposited much more than 250Ma before metamorphism. The high K/Rb ratios (Table 2.2, after Griffin et al., 1978) indicate that Rb was depleted during the granulite-facies metamorphism, hence the pre-metamorphic period was probably shorter than 250Ma. Griffin et al. (1978) suggested that the regional homogeneity of $(^{87}Sr/^{86}Sr)_o$ in these supracrustals is

also consistent with a volcanic origin.

Summary

The early Proterozoic supracrustal rocks are largely of volcanic origin and are compositionally similar to Cenozoic K-rich continental orogenic suites. The iron formations are probably related to this volcanism; the limestones and graphite- and sulphide-bearing schists may have formed in a shallow marine environment associated with the early stages of the volcanic activity.

2.5 Regional Metamorphism, c.1830-1700Ma

Most of the Lofoten-Vesterålen province was metamorphosed to granulite facies at 1830Ma. In the east of the area, in parts of Langøy and Hinnøy the metamorphic peak produced amphibolite-facies assemblages. The two areas are separated by an isograd which indicates the first appearance of orthopyroxene on the up-grade, granulite-facies side. The isograd has been located petrographically on Langøy by Heier (1960) and its extension onto SW Hinnøy has been determined by Glassley (1975). In many areas, the appearance of orthopyroxene is accompanied by a change in colour in the feldspars from pink and grey to dark-brown and green in the granulite facies; this change is clearly visible in the field. The abundance of hornblende and biotite decreases westward from the orthopyroxene isograd on southern Langøy. As has been mentioned earlier, severe Rb- and U-depletion accompanied the 1830Ma granulite-facies metamorphism.

Glassley (1975) mapped two isograds in the carbonate rocks of the region (Fig. 2.3, after Griffin et al., 1978) just west (up-grade) of the orthopyroxene isograd. The lower isograd corresponds to the prograde reaction Hornblende → Diopside + Spinel (1) and the higher isograd to the

prograde reaction Diopside + Dolomite → Forsterite + Calcite (2). Glassley (1975) examined ten carbonate bodies, on Gimsøy, Austvågøy and Langøy, which occur as pods and lenses with dimensions of a few tens of metres; 80 to 90 per cent of the rock volume is carbonate material while discrete grains of silicates and spinel occurring in knots or bands, or disseminated throughout the body, account for the remainder. The isograds are defined on the basis of mineral assemblages in the cores of the zoned carbonate bodies, and on the sequence of mineral assemblages from core to rim. Reaction (2), which takes place at a higher T and lower CO_2/H_2O ratio in the fluid phase, has occurred in the outer parts of bodies which are downgrade from the first isograd. Hence the present positions of the carbonate isograds have been influenced by kinetic factors, especially the rate at which water has diffused from the country rocks into the cores of the pods, as well as by temperature.

Griffin et al. (1978) reported that experimental data at P_{CO_2} = 3 and 5kb suggest that the temperature of the higher grade reaction, Diopside + Dolomite → Forsterite + Calcite is greater than 750°C at 10kb. This temperature must be regarded as a minimum temperature for both the amphibolite and granulite facies, since this reaction has gone to the right wherever the CO_2/H_2O ratio reached suitable values.

CO_2 is the main component of fluid inclusions in quartz grains of the gneisses on either side of the orthopyroxene isograd. There is no evidence that this isograd is marked by a sudden increase in the CO_2/H_2O ratio of the metamorphic fluid phase. High density (up to 1.19 $g.cm^{-3}$) CO_2 inclusions are common throughout the terrain; exploded fluid inclusions, indicating formation at very high pressures, are also common (Olsen, 1978).

Table 2.4 (after Griffin et al., 1978) attempts to constrain the P,T conditions that existed during and after the 1830Ma metamorphism. These attempts have met a number of problems. Although high-grade assemblages have survived (cpx-opx, opx-gt), exchange reactions have clearly continued during post-metamorphic cooling, so that the use of published geothermometers often does not yield realistic metamorphic temperatures. A single sample analysed using different geothermometers often yields markedly different temperatures, implying either a lack of equilibrium, equilibration of different mineral pairs at different temperatures during cooling, or errors in the geothermometers (Krogh, 1977).

Relict cordierite-bearing assemblages on Langøy and Hinnøy show that the early stages of the 1830Ma metamorphism reached high temperatures at low pressures (750-850°C, 3-4kb; Olsen, 1978). However, data from several workers using a variety of methods (Table 2.4) suggests that the maximum metamorphic conditions over most of the granulite facies area reached 850-950°C, 9-12kb. These conditions are consistent with the presence of sillimanite on Langøy (Heier, 1960), corundum + quartz at Vestpolltind (Krogh, 1977) and medium pressure granulite facies assemblages (cpx + opx + plag) (Fig. 2.4, after Griffin et al., 1978).

The observed formation of kyanite after sillimanite (Krogh, 1977) and the development of high-pressure granulite-facies assemblages in reaction coronas (Griffin and Heier, 1969) would be expected during cooling to about 770°C at 9-10kb, and are consistent with the pressures and temperatures calculated using other mineral assemblages (Table 2.4). These values indicate that the post-metamorphic uplift was slower (>6mm. 100 yr.$^{-1}$) than the burial that followed the initial low-pressure heating (>12mm. 100 yr.$^{-1}$) (Fig. 2.4).

'Crystallization temperatures' estimated for some of the mangerite intrusions are only 100-150°C greater than the peak metamorphic temperatures (Table 2.4, Fig. 2.4). The regional temperature was probably raised and the thermal peak associated with the metamorphism prolonged, by the heat associated with the massive late- to post-tectonic intrusion of magmas; hence the P,T conditions produced granulite-facies conditions over most of Lofoten-Vesterålen until at least 1700Ma ago.

2.6 Intrusive Activity, c.1800-1700Ma

2.6.1 The Basic Intrusions

The Proterozoic basic intrusions occupy a much smaller area of Lofoten-Vesterålen at the present level of erosion than the mangerite series with which they are temporally and spatially associated. Probably the earliest event in the 1800-1700Ma series of intrusions was the emplacement of a number of small gabbroic bodies which occur throughout Lofoten-Vesterålen. Three islands, Moskenesøy, Flakstadøy and Langøy have areally extensive complexes of basic igneous rocks.

The Moskenesøy Anorthosite (Green and Jorde, 1971)

South of Å on Moskenesøy, a dome-like body of anorthosite is exposed in the core of a broad E-W anticline in the early Proterozoic supracrustal gneisses. Locally the structure of the gneisses is disrupted at the contact with the anorthosite and the two lithologies are interleaved. Gneisses near the contact, and included within the anorthosite, are more mafic than elsewhere, and rare earth element data suggest that a low melting fraction may have been removed from these rocks. Pegmatitic feldspar-rich pods are developed in the gneisses immediately surrounding the anorthosite and might represent the product of the suggested partial melting. The evidence for partial melting of the gneisses suggests that

the anorthosite was considerably hotter than the gneisses when it was emplaced.

The Moskenesøy Anorthosite is purple-grey and very coarse-grained, (1-10cm) and consists of plagioclase (An_{35}-An_{50}) some of which is anti-perthitic. Subordinate phases include orthopyroxene, clinopyroxene and magnetite. The orthopyroxene is often overgrown by clinopyroxene. Retrogression coronas of amphibole around pyroxenes, and garnet-biotite around magnetite occur.

Green and Jorde (1971) suggested two possible mechanisms for producing the Moskenesøy Anorthosite. Firstly, by crystal accumulation from a gabbroic magma. Secondly, the anorthosite may represent the crystalline residuum after extraction of a partial melt from the supra-crustal gneisses. Evidence against the first suggestion includes the absence both of a gabbroic margin and of cumulus textures in the Moskenesøy Anorthosite. The high Bouguer anomaly under Moskenesøy might indicate the presence at the base of the crust of a large ultramafic residuum complementary to the anorthosite, but the anomaly could also be interpreted as having been produced by a shallow Moho-discontinuity (Sellevoll, 1967). If the anorthosite represents a residuum after partial melt extraction, then the melt fraction would seem to be almost completely lost; only a few monzonitic dykes, especially well exposed at Å in the south of Moskenesøy, could possibly represent such a melt. Nonetheless, Green and Jorde (1971) favoured the latter view, stressing the more mafic character, low Rb and high K/Rb ratios of the Moskenesøy gneisses compared to those in other parts of Lofoten-Vesterålen and interpreting this evidence as suggesting that a major partial melting event may have taken place.

The Flakstadøy Basic Complex (Romey, 1971)

A layered complex of basic rocks occurs as a belt across the south-eastern part of Flakstadøy. The complex is about 40km^2 in area, dips east and overlies a body of mangerite to the west. Xenoliths of the early Proterozoic supracrustal gneisses are found within the complex, hence the basic complex is intrusive into the gneisses.

Romey (1971) sub-divided the complex into three zones:

Zone C: upper zone of coarse-grained anorthositic norite, noritic anorthosite and anorthosite.

Zone B: layered sequence of iron-ore, norite, troctolite and anorthosite.

Zone A: lower zone of gabbro and norite.

Plagioclase shows a wide range in composition in each of the zones but averages An_{54} in the whole complex. Hypersthene, olivine and iron-ore are the main mafic phases, but olivine is only stable in Zone B.

Zone A has a weak compositional layering and igneous lamination defined by preferred orientation of (010) faces of plagioclase crystals. The occurrence of trough-banding and igneous lamination in Zone B is strong evidence that convection currents operated during the crystallization of the magma. No layering has been observed in Zone C, in which the rocks become more anorthositic from base to top.

Romey (1971) interpreted the Flakstadøy Basic Complex as the product of differentiation of a gabbroic magma by processes of crystal settling and convection-flotation. Hence the lowest zone, Zone A, consists of cumulus crystals of mafic phases and plagioclase crystals too large to be

supported by the convection currents. Zone C formed simultaneously, developing downwards from the roof. Smaller plagioclase crystals were carried upwards, in upwelling convection currents, to the roof where they were trapped and where they were enlarged by adcumulus growth. Small volumes of trapped liquid crystallized to produce the small proportion of mafic minerals present in Zone C. Zone B, in which olivine is stable, is of cumulus origin with well-developed rhythmic layering. The presence of dykes from Zone B intruding Zone C indicates that Zone B solidified later than the overlying Zone C.

The anorthosites on Moskenesøy and Langøy may have a similar origin, by flotation from a differentiating gabbroic magma, but this cannot be proved in the absence of the associated layered norites, troctolites and gabbros.

The Eidsfjord Anorthosite, Langøy (Heier, 1960)

This massive, grey, coarsely crystalline anorthosite crops out on the south side of Eidsfjord in southern Langøy. At the east end of its outcrop, the Eidsfjord Anorthosite has been thrust over the red augen gneisses, described by Heier (1960); which are interpreted as retrograded mangeritic rocks (Griffin et al., 1978), and the anorthosite rests on a 40m thick mylonite zone. No other contacts between the anorthosite and gneisses and a mangerite are observed, hence the relationship of the anorthosite to its surrounding rocks is unclear.

There are some similarities in the petrography of the Eidsfjord and Moskenesøy Anorthosites. Plagioclase composition in the Eidsfjord intrusion is in the range An_{45} to An_{55}. Minor orthopyroxene is overgrown by clinopyroxene and the latter also occurs as primary crystals. Locally the anorthosite is more gabbroic in composition; occasional bands

and lenses of pyroxenite are also present.

The Eidet-Hovden Gabbro, Langøy(Heier, 1960)

This gabbroic intrusion occupies most of the peninsula north and west of Eidet in the north-west of Langøy and an area of about 25 km^2. The contact between the gabbro and the early Proterozoic supracrustal gneisses is nowhere exposed, but the two lithologies are found within a few metres of each other and the gneisses show no contact metamorphism or deformation associated with the emplacement of the gabbro. Generally the banding in the gneisses is sub-parallel to the margin of the intrusion, but locally it is truncated by it. The gabbro is highly altered and deformed at its margin with the gneisses but is fresh and undeformed away from the margin. Excellent layering is developed in the centre of the intrusion. Ultrabasic lenses and bands occur within the gabbro, some of which grade into magnetite-ilmenite ore deposits which are elongate parallel to the igneous layering of the gabbro. These observations suggest that the Eidet-Hovden Gabbro may have formed by crystal settling from a gabbroic magma, and hence it may be comparable to the Flakstadøy Basic Complex.

As well as the contact with the supracrustal gneisses across the isthmus near Eidet, the gabbro also has contacts with two mangerites, Hovden and Engenykken,which are discussed later. Heier (1960) observed veins and dykes of monzonitic composition cutting the gabbro, which suggests that the mangerites post-date the gabbro. However, the gabbro is finer-grained near the contact with the Hovden Mangerite, possibly representing a chill zone, which would imply that the monzonitic veins and dykes are rheomorphic back-veins formed by contact melting.

Fine-to-medium-grained dykes of varying composition have been

recognised in the Eidet-Hovden Gabbro, the strike of which is generally parallel to the gneiss-gabbro contact and also to the banding in both the gneiss and the gabbro. The relationship between these dykes and the mangerites are unclear.

The conformity of the banding in the gneisses and the gabbro and the contact between the two lithologies suggests that the Eiden-Horden Gabbro may have originated as a slightly discordant sheet, perhaps similar to the Flakstadøy Basic Complex. The fine- to medium-grained dykes may have been intruded along the banding in the gabbro.

The Bø Gabbro, Langøy (Heier, 1960)

The Bø Gabbro occupies an area of about 8 km^2 in SW Langøy. Its major primary phases are plagioclase, clinopyroxene and biotite. At its contact the gabbro is sheared, foliated and retrograded, but in its central parts the gabbro has fresh mineral assemblages with an ophitic texture.

A zone of retrogression extends north-north-east from the gabbro in the adjacent gneisses and also locally affects the Bø Mangerite. Heier (1960) attributed this to the emplacement of the gabbro but the extension of the zone of retrogression for several kilometres away from the gabbro suggests that it is more likely to be associated with one of the later regional retrogressive events.

The relationship between the gabbro and the mangerite is not determinable due to both lithologies being strongly sheared in the contact zone. Similar strong shearing has been noted around the margins of the Eidet-Hovden Gabbro, the Eidsfjord Anorthosite and the Moskenesøy Anorthosite.

Granodioritic and red granitic pegmatites intrude the Bø Gabbro; the latter are probably related to the young red granites on Langøy, described by Heier (1960) as produced by anatexis of the gneisses and as late syn-kinematic.

Heier thought that the Bø Gabbro consists of two lenticular bodies dipping steeply to the west.

Other Basic Intrusions

Numerous small basic and ultramafic intrusions occur in Lofoten-Vesterålen in addition to the intrusions described in detail above.

Tull (1977) reported the presence of many small gabbroic bodies on Vestvågøy. Most are only a few hunderd square metres in area but a rather homogeneous gabbro near Borge occupies 14 km^2; this gabbro is intrusive into the early Proterozoic supracrustal gneisses but its relationship to the mangerite body around it is unknown. However, several of the other basic intrusions show clear evidence that they pre-date the emplacement of the mangerite. Small basic bodies in the Ramberg-Vestresand area (Tull, 1977, Plate 1) ranging from gabbroic diorite to norite, have been brecciated by the emplacement of mangerite (Tull, 1977, Fig. 5). Two small gabbroic bodies, west of Haugen in NE Vestvågøy, are also intruded by mangerite. Tull also reports the occurrence of small dioritic bodies near Smedvik, Smorten, Moland and north-west of Leknes. The development of granoblastic clinopyroxene around primary poikilitic clinopyroxene in some of these basic intrusions is regarded as a product of granulite-facies metamorphism and hence these intrusions are regarded as pre-metamorphic. A similar texture was seen in the Djupfjord Gabbro on Mosekensøy by Green and Jorde (1971).

The basic rocks of Vestvågøy have been affected by retrogression and deformation and are now mainly amphibolite. The Fuglehuk intrusion in North Moskenesøy and parts of the gabbro at Laukvik in North Austvågøy have also been retrogressed to amphibolites.

Partial retrogression of the mineral assemblages of the gabbros has commonly produced aggregates of amphiboles and plagioclase; examples include the Ølkona Spotted Gabbro and Ertneset gabbro-ultramafic intrusions on Moskenesøy (Green and Jorde, 1971) and parts of the Laukvik Gabbro (Griffin, NGU Report 1971). The Laukvik Gabbro pre-dates the Aust vågøy Grey Mangerite which contains blocks of the gabbro.

Green and Jorde (1971) noted several ultramafic intrusions with gabbroic margins which occur in the gneisses close to the Moskenesøy Anorthosite. These intrusions are cut by monzonitic veins and dykes which Green and Jorde suggested may have resulted from partial melting of the gneisses due to the emplacement of gabbroic magma, and hence they may be regarded as rheomorphic back-veins.

2.6.2 The Mangerite-Charnockite Series

The Lofoten-Vesterålen mangerites are rocks of monzonitic composition which have crystallized in granulite-facies conditions and hence have high-grade metamorphic mineral assemblages. Primary mesoperthite (an intergrowth of a K-rich and a Na-rich feldspar where the phases are present in equal amounts) and two pyroxenes are essential constituents. Plagioclase is usually present and either olivine or quartz may occur, as well as hornblende and biotite. Ilmenite and magnetite are the usual ore minerals. Zircon, apatite and pyrite are common accessory minerals.

The range in silica content for mangerites is 52-66%. Within the mangerite intrusions some lithologies occur which are more acidic than this;

such rocks are called charnockites. They have a granitic composition but with a granulite-facies mineral assemblage, i.e. essential quartz, feldspar and orthopyroxene.

The mangerites of Lofoten-Vesterålen are generally coarse-grained and dark-coloured, usually dark brown or green, when in the granulite facies. When retrograded the mangerites are a light colour, usually grey or pink.

The mangeritic intrusions occupy about 2100 km^2 (Malm and Ormaasen, 1978) or half of the Lofoten-Vesterålen terrain.

The Sund-Ølkona Mangerite (Green and Jorde, 1971; Romey, 1971; Ormaasen, 1976)

This is the westernmost mangeritic body in Lofoten-Vesterålen. On the eastern coast of Moskenesøy, about 4km North-north-east of Hamnøy, and at a locality marked as Ølkona by Green and Jorde (1971) is an outcrop of mangerite occupying an area of about 1km^2. This is probably an extension of the mangerite that crops out around Sund in South-west Flakstadøy and which extends over an area of about 4km^2. The Ølkona Mangerite on Moskenesøy is dark brown and coarse-grained and it is bounded by strongly sheared late Archaean monzonitic and dioritic gneisses (Green and Jorde, 1971). Romey (1971) reported that foliation and diffuse layering is more common in the Sund Mangerite than in the larger mangeritic body further east on Flakstadøy, especially near their contacts with the veined and layered (early Proterozoic supracrustal) gneisses. The Sund-Ølkona Mangerite is similar to the rocks of the Hopen Massif (Ormaasen, 1976), including the presence of charnockite in the Sund outcrop.

The Flakstadøy Charnockite (Romey, 1971; Ormasen, 1976)

The Flakstadøy Charnockite is restricted to a 2 km^2 outcrop west of Bø, in the central part of the island. It is surrounded by early Proterozoic supracrustal gneisses and is a coarse-grained quartz-rich rock with megacrysts of mesoperthite up to 10mm across (Ormaasen,1976). Ormaasen saw a resemblance between the Flakstadøy Charnockite and charnockitic rocks at Engenykken and Bø in West Langøy, but also noted that primary biotite and secondary garnet are abundant in the Flakstadøy body but not at Engenykken and that the outcrop in the Bø Mangerite has abundant primary hornblende while it is only present as an accessory mineral in the Flakstadøy Charnockite.

The South-West Lofoten Mangerite (Romey, 1971; Ormaasen,1976; Tull, 1977)

The Flakstadøy Mangerite is areally the dominant mangeritic rock on Flakstadøy. It lies immediately to the west of the Flakstadøy Basic Complex and underlies it. It crops out in a band of maximum width of 3km and extends the length of the island (about 18km) in a north-north-east orientation. The Flakstadøy Mangerite, together with all of the mangeritic rocks on Vestvågøy and Gimsøy, comprises the South-west Lofoten Mangerite. The Vestvågøy Mangerite crops out over about two-thirds of the island's area and is divided into two groups: the brown mangerite and the spotted mangerite. The main outcrop of the spotted mangerite is in an east-west band from north of Stamsund to Leknes. The South-west Lofoten Mangerites are characteristically dark-brown, coarse-grained and contain mesoperthite megacrysts. They are very similar from island to island (Romey, 1971). The spotted mangerite (or mangerite orthogneiss, Tull, 1977) is distinguished from the brown mangerite by the degree of hydrous retrograde metamorphism. All gradations exist between these metamorphically altered rocks and the massive brown mangerite. The spotted mangerites contain equilibrium mineral assemb-

lages characteristic of upper amphibolite facies. These orthogneisses can only be distinguished from the supracrustal paragneisses by their homogeneity and by their gradation into massive igneous rocks. Their mineralogy is the same consisting of quartz, graphic microcline, calcic oligoclase, biotite, hornblende, epidote, sphene, magnetite and garnet (Tull, 1977). The main effect of the retrogression from granulite- to amphibolite-facies mineral assemblages is the replacement of pyroxenes by aggregates of biotite, amphibole and almandine garnet. This, together with the change in colour of feldspars from dark to light, produces the characteristic spotted appearance of the retrograded mangeries. Mesoperthite recrystallizes to grid microcline and plagioclase and titaniferous magnetite exsolves some of its titanium to form sphene rims.

The mangerites on Vestvågøy intrude the early Proterozoic supracrustal gneisses and the small basic intrusions (Tull, 1977). Romey (1971) was unsure of the relationship between the Flakstadøy Basic Complex and the underlying Flakstadøy Mangerite, especially around West Nesland where the contact is gradational, however he concluded that the basic complex was probably intruded into the mangerite. The gradational contact and fine-scale interfingering between the veined and layered (early Proterozoic supracrustal) gneisses and the Flakstadøy Mangerite led Romey (1971) to suggest that the mangerite formed by partial melting of the gneisses, possibly related to the emplacement of the basic complex.

The Hopen Mangerite and Charnockite (Ormaasen, 1976, 1977)

The Hopen Massif occupies the south-western corner of Austvågøy, an area of about 15 km^2. Some of the most extreme topography (942m peak about 1000m from the coast) in Lofoten-Vesterålen, restricts field-work in the area. Ormaasen (1976, 1977) did not observe obvious intrusive

relations between the mangeritic rocks and the surrounding gneisses and hence the border between these units is somewhat arbitrarily drawn, indeed Ormaasen's border is about 1000m south of Tveten's (1976) on the western coastline. Ormaasen (1976, 1977) reported magmatic layering east of Festvåg in the south of the Hopen Massif.

The Hopen mangeritic rocks are usually massive, unfoliated and medium to coarse-grained with mesoperthite crystals up to 10mm across. Retrograde hydrous metamorphism increases to the south and south-west, but does not severely affect the mineral assemblages.

Ormaasen (1976, 1977) divided the Hopen Massif on the basis of modal silica and of feldspar and orthopyroxene textures. The northern third of the massif is charnockitic with modal quartz up to 60%, but typically 15-25%. To the south, the mangeritic rock has less than 10% modal quartz. Ormaasen (1976, 1977) found that the dominant primary mafic minerals in the Hopen Massif are clinopyroxene and an opaque mineral and that the total content of mafic minerals rarely exceeds 10%. Ormaasen (1976, 1977) determined the temperature of crystallization to be 1000 ± 25°C at a pressure of 12 ± 1kb i.e. a depth of about 35km.

Ormaasen's (1976, 1977) observation of secondary orthoferrosilite as a reaction product between fayalitic olivine and quartz in the more acidic parts of the Hopen intrusion, supported Griffin and Heier's (1969, 1973) deduction of an isobaric cooling inferred from garnet coronas between plagioclase and olivine (in anorthosites) and between plagioclase and pyroxenes (in mangerites).

The Austvågøy Grey Mangerite (Ormaasen, 1976)

The island of Austvågøy is almost bisected by Austnesfjord and Slöverfjord. The grey mangerite crops out over the central parts of the western half of the island, occupying an area in excess of 200 km^2. It is intrusive into the early Proterozoic supracrustal gneisses and the Laukvik Gabbro and is itself cut by the Raftsund Mangerite. The outcrop of the Austvågøy Grey Mangerite is cut by a major fault, the Olderfjord Fault. The contact between mangerite and gneiss is off-set by about 4km in the west and by about 600m in the east. Ormaasen has suggested that the mangerite to the north of the fault represents a higher level in the intrusion than that to the south, on the basis of higher U and Th contents in the north and a more acid composition.

The grey mangerite has undergone extensive hydrous retrogression, especially north of the Olderfjord Fault where the deformation is also more intense with the mangerite tightly infolded with the gneisses. The contacts between the mangerite and the gneisses are obscured by the deformation and recrystallisation, as is the primary magmatic texture.

The Raftsund Mangerite (Griffin et al., 1974)

The Raftsund Mangerite is the largest intrusive body in Lofoten-Vesterålen. It underlies the eastern half of Austvågøy and the south-western part of Hinnøy, as well as many of the islands between Austvågøy and Hinnøy, including Store Molla and Lille Molla. Similar rocks in the eastern part of Hadseløy (Devaraju and Heier, 1974) and across the Vestfjord on Hamarøy (Malm, 1976) are taken to be parts of the same intrusion (Griffin et al., 1974). Hence the Raftsund Mangerite covers an area of about 2000km^2.

The Raftsund Mangerite is clearly intrusive into the surrounding

early Proterozoic supracrustal gneisses and also cuts the Austvågøy Grey Mangerite. The Hamarøy Quartz-Rich Mangerite (Malm, 1976) and the Lødingen Granite intrude the Raftsund Mangerite.

Magmatic sedimentary structures are exceptionally well-developed and can be observed in many localities in the Raftsund Mangerite (see Figs. 3-5, Griffin et al., 1974). Three types of layering have been described by Griffin et al. (1974):

1. Centimetre-scale layering: 1-2 cm thick mafic bands alternating with 10-20 cm of normal mangerite;
2. Coarse-irregular layering: a planar structure is defined by layers and pods of mafic minerals up to 20cm thick; many of these layers are planoconvex in section and resemble shallow trough fillings;
3. Fine-grained layering: seen in four localities along strike from one another; a fine-grained band 0.5 - 2 m thick is similar in mineralogy to the normal mangerite but has a grain size of 0.5 - 1.0mm compared to 1-5mm for the normal mangerite; the layer has a sharp mafic base, grading upwards to a more felsic top; the zone may consist of several such rhythmic sequences.

Griffin et al. (1974) have divided the Raftsund Mangerite into three zones on the basis of olivine and pyroxenemineralogy:

1. Clinopyroxene-orthopyroxene (inverted pigeonite) zone;
2. Clinopyroxene-orthopyroxene-olivine zone;
3. Clinopyroxene-olivine zone.

The Zone 1 primary mafic mineral assemblage is found largely at the margins of the intrusion and Zone 3 in the core.

No roof to the intrusion has been observed, but Zone 1 rocks have

been recovered from the 1000m peaks west of the Raftsund, so perhaps the roof lay not far above these peaks. The contacts generally dip inwards; the dips of the layering are shallow towards the centre of the intrusion. Griffin et al. (1974) proposed that the mangerite may be an extensive sheet, possibly 3km thick. Conditions of crystallization have been estimated by Krogh (1977): temperature of crystallization of 1050°C and a pressure of 11 ± 1kb implying a depth in excess of 30km.

The Eidsfjord Mangerite (Heier, 1960; Ormaasen, 1976)

The Eidsfjord Mangerite crops-out on the southern side of Eidsfjord in southern Langøy. The red augen gneisses (Heier, 1960) to the east of the Eidsfjord Mangerite may be retrograded equivalents, possibly contiguous with similar outcrops on Hadseløy (Devaraju and Heier, 1974). No field evidence has been observed concerning the relationship of the Eidsfjord Mangerite to the Eidsfjord Anorthosite which crops-out to the west of the mangerite. Griffin (NGU Report 1971) has reported the occurrence of rare inclusions of the country rock gneisses in the mangerite, generally as dark, streaked-out schlieren.

The fresh mangerite is coarse-grained with large megacrysts of mesoperthite. Ormaasen (1976) compared the Eidsfjord Mangerite to that at Bø in western Langøy: the textures are similar, but the mineralogy is slightly different as neither biotite nor hornblende are primary minerals in the Eidsfjord Mangerite and its megacrysts are mesoperthite compared to true perthite in the Bø Mangerite. The retrograded mangerites (Heier's Red Augen Gneisses) have amphibolite-facies mineral assemblages and augen of K feldspar are red in colour.

The Bø Mangerite (Heier, 1960; Ormaasen, 1976)

The Bø Mangerite occurs in the south of western Langøy. Heier (1960) mapped it as a single body extending from Straumsjøen in the north, to Bø in the south. Ormaasen (1976) considers that its outcrop had been exaggerated and that two geographically separate units are present: the Søberg Mangerite between Søberg, Straumsjøen and Hole, and to the south-east, the Steine Mangerite between Gimstad and Bø. The Søberg Mangerite crops-out about 2km south of the Vikan Gneiss (Taylor, 1975; Griffin et al., 1978) outcrop and is emplaced into similar late Archaean gneisses. The Steine Mangerite is in contact with the Bø Gabbro, late Archaean gneisses and a young red granite (Heier, 1960). Ormaasen (1976) reported that the contacts between the mangerites and the gneisses are gradational and that inclusions similar to the basic parts of the gneisses are common, especially in the Søberg Mangerite.

The Bø Mangerite is massive or weakly foliated; it has megacrysts of alkali feldspar perthiteup to 40mm across. Ormaasen (1976) has reported the presence of magmatic layering at a locality in the Søberg Mangerite. He also recognised mineralogical differences between the Søberg and Steine bodies: the former has a greater antiperthite to perthite ratio and also a higher content of mafic minerals.

The Engenykken Mangerite (Heier, 1960; Ormaasen, 1976)

The $1km^2$ outcrop is almost completely surrounded by sea. It crops-out on a peninsula and an island on the western coast of the larger peninsula in north-west Langøy which is underlain by the Eidet-Horden Gabbro. The border between the Engenykken Mangerite and the Eidet-Horden Gabbro corresponds to the narrow isthmus that links the small peninsula to the rest of north-west Langøy. The border is strongly sheared, subvertical and strikes north-south. The gabbro adjacent to

the contact is fine-grained, as it is next to the Hovden Mangerite to the north. The grain size of the Engenykken Mangerite increases, and the content of ferric minerals decreases, away from the contact with the gabbro. The mangerite adjacent to the sheared contact, has a foliation parallel to the contact.

The typical mangerite in this intrusion is homogeneous and weakly foliated with mesoperthite megacrysts up to 15mm across and is similar to the Steine Mangerite in hand specimen. At the south-western end of the peninsula the mangerite is a quartz-rich variety with an irregularly variable grain size. The content of mafic minerals is very low but the content of basic inclusions ($< 0.5m^2$), similar to those in the Søberg Mangerite, is much greater than in the typical rock further east (Ormaasen, 1976).

The Hovden Mangerite (Heier, 1960; Ormaasen, 1976)

The Hovden Mangerite is surrounded by the Eidet-Hovden Gabbro on all sides except to the north where it disappears into the sea. Heier (1960) observed that the gabbro became finer towards the contact with the mangerite. Ormaasen (1976) confirmed this for the southern part of the mangerite, but found that this was not as pronounced to the north. The banding of the gabbro strikes parallel to the contact with the northern part of the Hovden Mangerite and dips steeply beneath it. The veins and dykes of monzonitic composition observed in the gabbro by Heier (1960) imply that the mangerite post-dates the gabbro, but the possible chill zone in the gabbro would imply that the gabbro is the later intrusion and that the veins of monzonitic composition were formed by contact melting.

Ormaasen (1976) observed no inclusions in the Hovden Mangerite. He divided the intrusions into two parts; that in the south has a

greater content of mafic minerals and its contact with the gabbro follows the contour lines and thus it may be a flat-lying sheet above the gabbro; the mangerite to the north sharply cross-cuts the contours and may have the form of a vertical dyke.

Chemical Composition

Table 2.5 shows the whole-rock major-element chemistry of the Lofoten-Vesterålen mangerites after Ormaasen (1976). High alkalis and a high Fe/Mg ratio characterize the mangerites. Griffin et al. (1974) noted that the overall composition for the Raftsund Mangerite is that of an intermediate rock, but with lower Al_2O_3 and CaO and higher $Na_2O + K_2O$ than a typical alkaline monozonite or granodiorite. They stressed that the chemical analysis of a cumulate rock represents the cumulus phases, plus trapped intercumulus material and may be only distantly related to the composition of the magma from which the rock formed.

The mangeritic rocks of Lofoten-Vesterålen can be divided into two groups with different Fe/Mg ratios (Fig. 2.5, after Griffin et al., 1978; Malm and Ormaasen, 1978). The group with lower Fe/Mg ratios may be older than those with higher Fe/Mg ratios based on field relations. The mangerites of Moskenesøy, Flakstadøy, Vestvågøy and Langøy and the Austvågøy Grey Mangerite belong to the low-Fe/Mg series, while the Hopen mangeritic rocks and the Raftsund Mangerite (including its extensions onto Hamarøy and Hadseløy) belong to the high-Fe/Mg series (Malm and Ormaasen, 1976). The two series show otherwise similar chemistry and are alkalic-calcic in nature (Fig. 2.6; after Griffin et al., 1978).

Green et al. (1972) analysed anorthosites, mangerites and granulites from Lofoten-Vesterålen and published rare-earth element (REE) distributions and K/Rb ratios. They concluded that the anorthosites and mangerites are

not comagmatic, which agreed with Romey's (1971) conclusions reached from field relations between the two rock-types on Flakstadøy. The argument against a comagmatic origin produced by Green et al. (1972) is based largely on the positive Eu anomalies in both the anorthosites and mangerites. Early fractionation of plagioclase from a magma to form the anorthosites would have depleted the residual magma in Eu: if the mangerites were formed by crystallization of this residual magma, they would be expected to have negative Eu anomalies.

Malm and Ormaasen (1978) investigated the REE distributions of acid mangerites and charnockites and found either no Eu anomalies or negative anomalies, and a general REE enrichment relative to the mangerites. (All the mangeritic rocks analysed by Green et al. are mangerites (s.s.) i.e. 56-62 weight % SiO_2). The positive Eu anomalies reported for the mangerites by Green et al. suggest that accumulation of feldspar may have played an important part in the crystallization of these rocks as well as for the anorthosites. If the mangerites are at least partially cumulate in origin, then their parental magma(s) may have had a less marked positive Eu anomaly than the mangerites of Green et al. They could even have had negative Eu anomalies complementary to the positive Eu anomalies of the anorthosites. This argument of Malm and Ormaasen's weakens Green et al.'s case against comagmatic origins. The former authors also emphasised that if the area of outcrop of anorthosites versus mangerites reflects the relative volumes of these lithologies, then it is clear that the accumulation of small volumes of anorthosite would have had only a minor effect on the residual magma. Thus, Malm and Ormaasen (1978) proposed that the anorthositic and mangeritic intrusions were derived from a common precursor, possibly a gabbroic magma at the base of the crust.

Green et al. (1972) also analysed gneiss interleaved with the

Moskenesøy Anorthosite and found it to be depleted in light REE relative to gneisses away from the anorthosite. They attributed this to melting of the gneiss in the contact zone at the time of emplacement of the anorthosite. Late-stage monzonitic veins and dykes on Moskenesøy with a high K/Rb ratio and a linear fractionated REE pattern may represent a part of this melt.

Rocks similar to the mangerites of Lofoten-Vesterålen are common as post-tectonic intrusions in the North Atlantic craton (Bridgwater and Windley, 1973); they are often associated with anorthosites when they are inferred to be at deep crustal levels. The 'rapakivi granites' of Finland and Greenland have similar chemistry to the Lofoten-Vesterålen mangerites (Griffin et al., 1978).

Isotopic Data

Rb-Sr whole-rock isotopic studies of the mangerite series show that they intruded in the period 1800-1700 Ma (Table 2.6, after Griffin et al., 1978). Griffin et al. took the low $(^{87}Sr/^{86}Sr)_o$ ratios (0.7030 to 0.7052) as evidence that the mangerites were derived from upper mantle magma sources, not by extensive anatexis of the basement rocks, but they attributed the variation in the initial ratios to involvement of crustal material in the origin of some of the mangerites.

Pb isotopic compositions in the Raftsund Mangerite also suggest some involvement of lower crustal material in the genesis of these magmas (Griffin et al., 1978). The present day Pb-isotopic compositions of most of the Raftsund Mangerite samples are less radiogenic than any plausible upper mantle Pb-isotopic compositions at 1700Ma (age determined by Rb-Sr method), hence a simple two-stage model for the isotopic evolution of Raftsund Mangerite Pb is ruled out and at least some of the Pb was probably

derived from a U-depleted lower crustal source (Taylor,1974).

2.6.3 The Young Red Granites, Langøy (Heier, 1960)

A number of unfoliated, coarse-grained, red granites occur in the granulite-facies terrain of Langøy, ranging in size from the Torset Granite (about 20km^2) to small patches of only a few square metres. In places the red granites have been seen to incorporate gneiss xenoliths, but they are generally homogeneous.

There are no contact metamorphic effects associated with the emplacement of the red granites. Heier (1960) attributed this to either the granites pre-dating regional metamorphism or the granites intruding the gneisses while the latter were still hot. The lack of foliation, and the presence of biotite rather than orthopyroxene as the usual mafic mineral in the granites, suggests that the latter interpetation should be preferred. Heier suggested that the granitic melts formed only slightly below the present surface by partial melting of gneisses during the high-grade metamorphism and that they intruded to their present level late in the tectonothermal event: for example, a red granite cross-cuts the Steine Mangerite in western Langøy.

A Rb-Sr whole rock errorchron age of 1705 ± 80 Ma and an $(^{87}Sr/^{86}Sr)_o$ of 0.7057 ± 0.0036 has been obtained for the Torset Granite in northern Langøy (Brueckner, 1971). This date is best interpreted as the age of intrusion.

2.6.4 Dolerite Dykes (Misra and Giffin, 1972)

Two dolerite dyke swarms intrude the gneisses and mangerites on Austvågøy and south-western Hinnøy. These dolerites are low in SiO_2 and high in TiO_2 and alkalis (Misra and Griffin, 1972). They are similar to many basalts from continental margins and tensional zones; equivalent dolerites have been found in many localities along the Norwegian coast in gneisses of 1600-1800 Ma age.

The first dolerite dyke swarm strikes NE-SW and cuts both the gneisses and the Raftsund Mangerite. The dykes are vertical, 1-3m wide, and are extensively back-veined where they intrude the mangerites suggesting that the mangerites were still hot when the dolerites were emplaced. The dykes of the second set are more numerous than the first; they are vertical, strike E-W, and are up to 10m wide. They have chilled margins, but have also melted the mangerites and gneisses and hence are back-veined on a limited scale.

Corona textures, formed during cooling, suggest that the dolerite dykes were emplaced at intermediate crustal levels (Misra and Griffin, 1972).

2.6.5 Summary

A series of intrusions of a wide range of compositions were emplaced into the gneisses of Lofoten-Vesterålen at 1800-1700Ma. Rocks of mangeritic composition are dominant at the present level of erosion. The formation of the mangerites apparently involved a limited amount of lower crustal material. The genetic relationships between the mangeritic magmas and the parent magmas of the gabbroic-anorthositic rocks are uncertain. The young red granites of Langøy appear to post-date the mangerites. Two swarms of dolerite dykes intruded the gneisses and mangerites of Austvågøy and south-western Hinnøy while the country rocks were still very hot.

2.7 The Lødingen Granite c. 1600 Ma

The Lødingen Granite crops-out in southern Hinnøy over a large area, east of the Raftsund Mangerite. It is heterogeneous, even on outcrop scale, with lenses and bands of mafic inclusions (basic dykes?), but it isdominantly granitic in composition. Its heterogeneity, the presence of a marked foliation and the common occurrence of K-feldspar porphyroblasts distinguish the Lødingen Granite from the young red granites of Langøy. The foliation may have been developed in the 1100 Ma metamorphism described below.

The Lødingen Granite intrudes the Raftsund Mangerite near Offersøy, on the Vestfjord coastline; further north the two units are separated by amphibolite-facies gneisses which may have been derived by retrograde metamorphism of the Raftsund Mangerite (Griffin et al., 1978).

Some large, relict gabbro bodies occur within the Lødingen Granite and are intruded by it. They have strongly foliated and amphibolitized borders.

Isotopic Data

A five-point Rb-Sr whole-rock errorchron age of 1380 ± 80 Ma and an $(^{87}Sr/^{86}Sr)_o$ of 0.7118 ± 0.0015 is given for the Lødingen Granite at Nes in Taylor (1974) and Griffin et al. (1978). They suggested that the high $(^{87}Sr/^{86}Sr)_o$ might reflect an origin for the granite by derivation from either the Raftsund Mangerite or the early Proterozoic supracrustal rocks.

Andresen and Tull (1983) published an 11-point Rb-Sr whole-rock isochron age of 1644 ± 36 Ma (MSWD = 3.4) and an $(^{87}Sr/^{86}Sr)_o$ of 0.70780 ± 0.00074 for the Lødingen Granite from samples collected along the road on the western side of the Tjeldsund. They also attribute the

high $(^{87}Sr/^{86}Sr)_o$ to derivation of the granite from the mangerites and gneisses which crop-out to the west.

2.8 Late Proterozoic Events c.1200-1000 Ma

2.8.1 The Leknes Supracrustal Group

The youngest supracrustal rocks in Lofoten-Vesterålen are a thin group of metasediments which include quartzite, meta-pelites, amphibolite and marble, all metamorphosed in lower amphibolite facies. The pelites have a paragenesis of staurolite-kyanite-garnet-biotite-muscovite-plagioclase-quartz.

These rocks have been called the Leknes Group by Tull (1978) after their main outcrop area of about $20km^2$ around Leknes, in south-west Vestvågøy. Metasediments on Vaerøy, to the south-west of Moskenesøy, have been correlated with the Leknes Group, by Griffin and Taylor (1978).

A poorly-defined Rb-Sr whole-rock date of 1140 ± 135 Ma with $(^{87}Sr/^{86}Sr)_o = 0.7052 \pm 0.0021$ has been obtained for the rocks near Leknes. The scatter of data points may reflect variation in $(^{87}Sr/^{86}Sr)_o$ among the samples as may be expected for a suite of metasediments (Griffin et al. 1978). On Vaerøy, mica schists and meta-arkoses from a single outcrop define two parallel arrays (slope c.950 Ma) with different $(^{87}Sr/^{86}Sr)_o$ (Griffin and Taylor, 1978). These dates are interpreted as the time during low-grade metamorphism at which large scale movement of intergranular fluids ceased.

Tull (1978) reported that the Leknes Group on Vestvågøy was emplaced tectonically against the south-west Lofoten Mangerite. This thrusting preceded the earliest recognizable folding and metamorphism of the schists.

After emplacement, the schists on Vestvågøy were folded together with the mangerites in tight meso- and macroscopic similar folds. This folding has also been recognised in the northern parts of Flakstadøy and Austvågøy. The isoclinal folding is confined to a belt 60-70 km long, up to 20 km wide, and sharply bounded at least on its southern side. This southern boundary to the zone of deformation and retrogression is substantially defined by the Olderfjord fault line.

The trace of the Olderfjord Fault is a prominent feature on the geological map of Lofoten-Vesterålen (Fig. 2.1). It can be mapped across the width of the western half of Austvågøy, striking ENE-WSW. Vertical movements have produced an offset of the contact between the Austvågøy Grey Mangerite and the supracrustal gneisses. To the north of the fault the mangerite has been retrograded to amphibolite facies, whereas to the south it has largely escaped the effects of this metamorphism and retains granulite-facies mineral assemblages. The fault line does not cut the Raftsund Mangerite to the east but it can be extended to the west along the southern coast of Vestvågøy, separating the three southern lobes of Stamsund, Mortsund and Ballstad from the rest of the island. These lobes have the best-preserved granulite-facies mineral assemblages on the island. The trace of the fault has not been recognised on Flakstadøy and Moskenesøy.

The main period of movement on the Olderfjord Fault is believed to have been during the late Proterozoic metamorphic and deformational event.

2.8.2 Metamorphism and Deformation

Synchronous with the folding, the Leknes Group was metamorphosed to lower amphibolite facies and both the granulite-facies supracrustal rocks and parts of the mangerites on Vesvtågøy were retrograded to amphibolite-facies mineral assemblages. The retrograde changes include, in general order of appearance: (1) breakdown of mesoperthite to plagioclase and microcline with a change in colour from brown or green to pink or grey; (2) oxidation of primary biotite to produce secondary garnet; (3) oxidation of magnetite to haematite; (4) hydration of primary pryoxenes to produce biotite and amphibole. Tull (1978) demonstrated the association of the retrogression of the basement, the prograde metamorphism of the Leknes Group, and the isoclinal folding event. Olsen (1978) estimated T = 550-600°C and P = 5-6kb for the retrograde metamorphism on eastern Langøy.

These retrograde effects are widespread in Lofoten-Vesterålen, even where the isoclinal folding event cannot be recognised (Griffin and Heier, 1969). A low-Fe/Mg mangerite has been altered to red 'augen gneiss' on south-east Langøy and Hadseløy and the Raftsund Mangerite to a biotite-garnet orthogneiss on Hadseløy, and around Lonkan on Hinnøy (Heier, 1960; Devaraju and Heier, 1975). Mineral separates from the early Proterozoic sequence give ages of 900-1200 Ma (Table 2.2, after Griffin et al., 1978) which are interpreted as reflecting the retrograde metamorphism.

The retrogression of the granulite-facies rocks of Lofoten-Vesterålen must have involved the introduction of large quantities of water. Locally some of this may have been derived by the metamorphism of Leknes Group sediments (Tull, 1978), but the source of most of the water is not known.

2.9 The Caledonian Orogeny c.400Ma

The northern Norwegian belt of Caledonian deformation lies just east of the Lofoten-Vesterålen area. Studies of other 'basement gneiss' areas further south along the coast have shown that these basement gneisses were extensively involved in the Caledonian deformation (Wilson and Nicholson, 1973; Nicholson and Rutland, 1969). Lofoten-Vesterålen is in the middle of the orogenic belt in pre-drift reconstructions of the Caledonides. Caledonian structures in north Norway face east (Gee, 1975) while those in eastern Greenland face west (Henriksen and Higgins, 1978). Nonetheless, this area shows evidence of only limited Caledonian effects. Mineral ages show a range of ages (Table 2.2) (Broch, 1964; Heier and Compston, 1969; Tull, 1978) indicating that isotopic resetting was incomplete. Griffin et al. (1978) concluded that no extensive metamorphic recrystallization occurred in Lofoten-Vesterålen during the Caledonian orogeny. Fluid inclusion studies show that temperature and pressure were low (T = ca. 300°C, P = 1-3kb) at that time (Olsen, 1978).

A group of granitic pegmatites occur in the south of Moskenesøy (Green and Jorde, 1971). They are up to 3m wide and may be followed along strike for over a kilometre. Griffin et al. (1978) quote unpublished K-Ar muscovite and biotite ages of McIntyre (1973) which show that these pegmatite dykes are Caledonian. No other intrusive rocks associated with the Caledonian orogeny have been recognized in Lofoten-Vesterålen.

In the eastern part of the province, the late Archaean Gullesfjord granitic gneiss has been thrust eastward over gneisses correlated with the Lødingen Granite. This thrust may be related to those on Langøy (Heier, 1960), Vaerøy (Griffin and Taylor, 1978) and SW Hinnøy. The Gullesfjord thrust plane is isoclinally folded by the folds identified as

Caledonian F_1 near Harstad (Gustavson, 1966). The foliation associated with this Caledonian folding penetrates the Gullesfjord granitic gneiss, but dies out to the west. The Caledonian deformation and metamorphism have not affected the majority of the Lofoten-Vesterålen province.

The Lofoten-Vesterålen basement terrain is unconformably overlain by a transgressive metasedimentary sequence on eastern Hinnøy which was described and named the Storvann Group by Bartley (1981), who proposed that it represents the westward extension of the foreland sequence developed on the Baltic craton prior to the Caledonian orogeny. Griffin et al. (1978) also concluded that Lofoten-Vesterålen was part of the European plate prior to the Caledonian orogeny. This conclusion was based on reconnaissance studies in the Hamnøy-Tysfjord region (Foslie, 1941; Carr, unpubl. data; Griffin, unpubl. data) which indicates that the Tysfjord Granite, which consists of retrograded mangerites, is continuous with basement gneisses along the Swedish border to the east. Griffin and Taylor (1978) favoured a model in which the Norwegian nappes came over Lofoten, rather than one in which they are rooted between Lofoten and the Rombak window to the east, on the basis of eastward thrusting observed on Vaerøy (SW of Moskenesøy). Tull (1977), however, favours a model in which Lofoten has been moved at least 100km to the east, relative to the Baltic Shield. Hodges et al. (1982) postulated a similar model in which subduction of the leading edge of the Baltic craton beneath Greenland was limited by plate buoyancy. Further convergence would produce imbricate thrusts in the leading edge of the Baltic craton; these are correlated with low angle westerly-dipping, reverse faults on eastern Hinnøy and west of the Rombak window. Hodges et al. (1982) propose that the Lofoten-Vesterålen terrain and windows of Precambrian basement along the Norway-Sweden border are the consequence of westerly-dipping thrusts which propagated at the climax

of the Caledonian orogeny. Bartley (1982) proposed that the involvement of the uppermost parts of the Lofoten-Vesterålen terrain was catalyzed by the release of volatiles by dehydration reactions in the adjacent cover, whereas the absence of Caledonian effects in the deeper sections of Lofoten-Vesterålen may be a consequence of the lack of such fluids in dry granulite-facies rocks (Griffin et al., 1978; Bartley, 1982).

2.10 Summary

The geological history of Lofoten-Vesterålen is summarized in Table 2.7, reproduced from Griffin et al. (1978). The main periods of new crust generation in Lofoten-Vesterålen were first at about 2700 Ma and then over a protracted period from about 2100 Ma to 1700 Ma. Major metamorphic events occurred at 2700-2600 Ma, c. 1830 Ma and 1100-900 Ma (Griffin et al., 1978). Griffin et al. (1978) reported that the first two metamorphic events involved only local anatexis and that large-scale crustal melting was limited to the production of granitic magma which formed the Lødingen Granite, dated at c.1640 Ma by Andresen and Tull (1983). The area has experienced repeated burial and uplift of about 30km amplitude.

The oldest rocks in Lofoten-Vesterålen are the migmatite gneisses of Langøy and Hinnøy, the protoliths of which were extracted from the mantle at c.2700 Ma; the protoliths of the Moskenesøy migmatites were distinctly younger (Jacobsen and Wasserburg, 1978). These protoliths have calc-alkaline chemistry and are dominantly intermediate volcanics. By analogy with modern day volcanics of similar composition, these rocks may have formed at a destructive plate margin (Griffin et al., 1978).

The second crust-forming episode began with the eruption of alkali-

rich, calc-alkaline volcanic rocks from about 2100 Ma. These rocks are dominantly dacitic to rhyolitic in composition but also include significant amounts of basic material. Today, such volcanism is most common in areas where oceanic crust is being, or has been, subducted beneath thick continental crust. This suggests that the earlier migmatites may have formed a thick crust in the early Proterozoic (Griffin et al., 1978).

The lower part of the volcanic and sedimentary pile (preserved today at surface) was metamorphosed at pressures corresponding to a depth of 30-40km. Allowing for the present depth to Moho in Lofoten-Vesterålen the continental crust may have been at least 60-75km thick during and shortly after the 1830Ma metamorphic event. The implied rapid build-up of crustal thickness >40km may be achieved in two ways:

(i) by rapid accumulation of volcanic products (Andean-type);

(ii) by collision of continental plates (Himalayan-type).

About 3km of lava and ash have been accumulated in the Andes in less than 12Ma (Pichler and Zeil, 1971). The deposition of the supracrustals occurred over 100-150Ma and a continuation of Andean-type volcanism over such a period could result in burial of the lower parts of the pile to suitable depths even allowing for erosion and compaction. Himalayan-type burial would produce a cool geothermal gradient, with high P/T ratios and blueschist to high pressure granulite-facies metamorphism (Bird et al., 1975). The relatively high geothermal gradient in Lofoten-Vesterålen suggests that the Andean analogy might be the more applicable for this area (Griffin et al., 1978). Both models involve subduction and suggest the operation of plate tectonic processes in the early Proterozoic.

Krogh (1975, 1977) suggested that the volcanic area may have been over-ridden by a colliding continental plate.

The production of a thick metavolcanic and metasedimentary sequence in Lofoten-Vesterålen was followed by the production of alkaline intermediate magmas and the intrusion of an alkali-basalt dyke swarm. Such rock associations are found in more recent continental rift zones (Griffin et al., 1978) which may have been formed by upwelling of mantle diapirs, causing crustal thinning, and by intrusion of large volumes of basic magma into the lower crust (Fairhead, 1976; Ramberg, 1976). This would imply that within the 100Ma after the c.1830 Ma metamorphism, there has been a rapid transition from a compressional to a tensional regime. The need for a rapid change could be avoided by the use of the model of Watson (1976) in which Proterozoic mobile belts may have developed under conditions of horizontal tension associated with the break-up of cratons, but Griffin et al. (1978) point out that the need to postulate a continental rift volcanism and/or crustal foundering that could bury the basement cover contact to depths of over 30km is a major drawback to the application of this model to Lofoten-Vesterålen.

The Leknes Group metasediments crop-out on Vestvågøy and Vaerøy (Tull, 1978; Griffin and Taylor, 1978). Tull (1978) described this unit as tectonically emplaced against the basement, before it was metamorphosed, which implies that the basement was near the surface 1200-1100 Ma ago. The subsequent metamorphism of the Leknes Group involved burial to c.15-20km depth.

The repeated, relatively rapid periods of burial and subsequent erosion in Lofoten-Vesterålen is in marked contrast with the tectonic evolution of the Lewisian terrain (Dickinson and Watson, 1976), although between Lofoten-Vesterålen and the Lewisian, Talwani and Eldholm (1972) have traced a nearly continuous basement high in geophysical anomalies which they interpret as Precambrian rocks.

Petrological and geochemical evidence suggests that plate tectonic processes operated in Lofoten-Vesterålen in the period 2100-1700 Ma and possibly as far back as 2700 Ma.

The Scandinavian and East Greenland Caledonides were formed by the early Palaeozoic closure of the Iapetus Ocean, which had separated the European and North American plates (Dewey, 1969; Harland and Gayer, 1971; Gee, 1975). The collision between the two plates resulted in predominantly eastward nappe transport in Scandinavia (Gee, 1975) and predominantly westward nappe transport in eastern Greenland (Henriksen and Higgins 1978). The Lofoten-Vesterålen terrain lay within this collision zone but has been relatively unaffected. Lofoten wasprobably part of the leading edge of the Baltic Shield (Griffin et al., 1978; Bartley, 1981) and may have been thrust at least 100km to the east, relative to the Baltic Shield during the Caledonian orogeny (Tull, 1977; Hodges et al., 1982). The fact that Lofoten-Vesterålen consisted of massive, dry, granulite-facies rocks may have enabled the area to escape the large-scale effects of the Caledonian orogeny.

CHAPTER 3. THE LATE ARCHAEAN MIGMATITES

c.2700Ma

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CHAPTER 3. THE LATE ARCHAEAN MIGMATITES c. 2700 Ma

3.1 Introduction

The formation of the protoliths of the late Archaean migmatite complex is the oldest crust-forming event in Lofoten-Vesterålen.

The main outcrops of the migmatite complex (on Moskenesøy, Langøy and north-west Hinnøy) correspond to the deepest crustal levels currently exposed in Lofoten-Vesterålen and so it is reasonable to assume that the migmatites underlie much of the terrain to the south and east. Griffin et al. (1978) observed that the mangeritic rocks tend to occur at or near the migmatite/supracrustal contact. Whatever the origin of the mangerites, the migmatites have played an important role: if they are mantle-derived, then any crustal-contamination would originate primarily from the migmatites and if they are derived by crustal anatexis, the migmatites would be the main source involved.

3.2 Fieldwork and Petrography

The migmatite complex crops out over large areas of Moskenesøy, Langøy and Hinnøy, and over smaller areas of Flakstadøy and Austvågøy.

The migmatites have been examined in this study at four localities:

(i) between Ølkona and Moskenes on the east coast of Moskenesøy;

(ii) at Festvåg on the south-western tip of Austvågøy;

(iii) near Sandset in central Langøy;

(iv) near Sigerfjord in northwest Hinnøy.

3.2.1 Moskenesøy

The granitic pegmatite that was mined in the Moskenes Pegmatite Quarry was intruded into the monzonitic and dioritic gneiss sequence of Green and Jorde (1971) (Fig. 3.1). The pegmatite has been mined out but the quarry walls make possible the collection of fresh samples of the gneisses. In the field these rocks cannot be distinguished from the mangerites of Lofoten-Vesterålen. Their most striking feature is the presence of blocky mesoperthite crystals up to 2cm across. WL237 is a sample of this gneiss. Stringer mesoperthite and biotite are the dominant minerals, the former showing undulose extinction and the latter kinking of the cleavage traces and recrystallization, indicating that the rock has been deformed. Subordinate clinopyroxene and opaque phases are often overgrown by quartz and garnet. Minor orthopyroxene may be rimmed by biotite, hornblende and garnet. Minor plagioclase (An_{35}) is also present. The petrography of this rock is very similar to that of the mangerites.

The gneisses adjacent to the Ølkona spotted gabbro and the Ølkona Mangerite belong to the monzonitic and dioritic gneiss sequence (Green and Jorde, 1971). The gneisses around the Ølkona Mangerite are strongly sheared. Ultramylonite zones are developed up to 10cm wide; these may continue into the mangerite. At the southern contact of the Ølkona Mangerite, the mangerite and the gneisses interfinger over at least 50m.

3.2.2 Festvåg, Austvågøy

The small area of gneisses exposed to the south of the Hopen Mangerite, around Festvåg and on the islands of Henningsvaer is of uncertain age but they are shown as belonging to the late Archaean migmatite sequence on the NGU 1:250,000 Svolvaer sheet (Tveten, 1978).

For this work, the migmatites have been studied on the headland from which the bridge to Henningsvaer begins. The border with the Hopen Mangerite is at most a few hundred metres to the north. An isochron suite consisting of samples WL549 to WL562 was collected from the new road-cuts at the foot of the bridge (see Fig. 3.2, Plates 3.1 and 3.2). These gneisses are predominantly leucocratic, with thin amphibolite stringers, typically 1-2cm thick, but continuous over many metres. Locally the leucocratic gneisses have recrystallized into coarse-grained granular pods around these mafic stringers. Mafic material also occurs as irregular pods. The foliation of the gneisses is continuous through both the felsic and mafic components indicating that they were associated before the foliation-forming event. WL549 to WL553 are from the leucocratic gneisses, in which mafic phases total less than 2%. In all cases perthite and quartz are the dominant phases, with variable amounts of plagioclase. Biotite and an opaque phase are the main mafic minerals present. Hornblende may occur in minor amounts. Myrmekite is commonly developed. Accessory apatite is present in all samples, zircon in some. WL555 to WL560 are samples from the amphibolite components of the migmatite sequence. Mafic phases total between 30% and 60% of these rocks. Green hornblende is the most abundant mafic phase in all samples except WL559 where it is present in minor amounts and biotite, two pyroxenes and an opaque phase are all more abundant. Plagioclase (An_{40}) is the felsic phase in all these samples. Biotite, an opaque, and an orthopyroxene are usually subordinate phases. Apatite may be present as an accessory phase. WL557 has garnet coronas developed around opaques. WL556 contains a 1cm wide pyroxenite vein consisting of ortho- and clinopyroxene and perthite with minor biotite, opaque and plagioclase. WL554 is a coarser-grained (ca. 5mm) sample associated with an amphibolite stringer and possibly formed by recrystallization of the leucocratic gneisses. It only has quartz present in myrmekites.

Its dominant phases are perthite, orthopyroxene, biotite and plagioclase and as such is petrographically indistinguishable from a mangerite. WL561 and WL562 are gneisses of intermediate composition; the former also has mangeritic affinities.

On the eastern side of the headland is a migmatite of rounded leucocratic blocks up to 40cm across with more melanocratic rims surrounded by a leucocratic neosome (Plate 3.3). This neosome may represent an in situ melt extracted from the palaeosome leaving a depleted, restite rim. Alternatively this could represent one igneous body intruding another shortly after the first was emplaced, causing melting of the more mobile phases from the palaeosome in contact with the newly-emplaced magma. Plate 3.1, taken about 300m west of the locality described above, provides evidence in favour of the latter hypothesis. Here, rounded palaeosome blocks are up to 2m across. A strong foliation wraps around these blocks and includes elongate rafts of gneiss parallel to the margins of the large blocks. The volume of neosome is very small compared to that of the palaeosome. It may be that the old gneiss unit was broken up at the time of emplacement of the Hopen Mangerite and that the neosomes are veins from the mangerite. This, however, would imply a gap of up to a billion years between formation of the gneisses and the injection of a neosome. Remelting of the old gneiss blocks might still be feasible after such a time interval if the neosome was injected during a granulite-facies event.

Further north on the eastern side of the headland, towards the contact with the Hopen Mangerite, elongate amphibolite pods (ancient dolerite dykes) are broken up by veins from the mangerite. The Hopen Mangerite has a layered border series against the migmatites with thick layers of gabbroic composition dipping steeply to the north.

Magmatic sedimentation features are well-developed (see Chapter 5).

3.2.3 Sandset, Langøy

An isochron suite of six samples, WL723 to WL728 was collected from the new road-cut on the south side of Skjelford, 2km south-east of Sandset in central Langøy (Fig. 3.3). These gneisses show marked compositional banding and are indisputably paragneisses. WL723-WL725 were taken from a single block and are quite distinct petrographically, despite being samples of layers only a few centimetres from each other. In each of these three samples quartz, alkali feldspar and plagioclase are the major phases but the grain size differs between the samples. WL725 is the only sample to show a foliation (defined by biotite); WL724 is the coarsest of the three samples, shows extensive recrystallization of quartz and feldspars at grain boundaries, has extensive development of myrmekite and its main mafic phase is an opaque which occurs in the interstices of the quartz and feldspars. Orthopyroxene is present in all three samples but in widely different amounts.

WL726-WL728 were also taken from a single block, about 5m from the block just described and they too show considerable compositional variation. WL726 is equigranular and is not foliated. Mafic phases (brown amphibole, orthopyroxene, biotite, clinopyroxene and opaque) comprise about 65% of the sample. Plagioclase is the felsic phase. WL727 is more coarse-grained (up to 5mm) with some recrystallization at grain boundaries and a weak foliation. Perthites and quartz are present as well as plagioclase; mafic phases comprise 20% of the sample. WL728 has pronounced cm-scale compositional banding; the mafic-rich bands are very similar to WL727, while the mafic-poor bands consist of quartz grains up to 10mm across with subordinate alkali feldspar and plagioclase and accessory opaques and apatite.

3.2.4 Sigerfjord, NW Hinnøy

Two samples, WL661 and WL662, were collected from the amphibolite-facies gneisses at Ridderset on Sigerfjord in north-west Hinnøy (Fig. 3.3). The gneisses are strongly foliated and are deformed by numerous kink-bands. Quartz pods are common. WL661 is very fine-grained (less than 1mm) with biotite defining the foliation. Quartz, biotite (with radiation haloes around zircon grains), plagioclase and opaques are the main phases. Accessory muscovite and apatite are also present. WL662 has marked compositional banding unlike WL661; this is defined by a 5mm wide band of quartz with a selvage up to 4mm wide of which green biotite constitutes about 80%. The rock between these quartz-biotite segregations is similar to WL661 except that myrmekite and microcline is present and muscovite and opaques absent.

3.3 Geochemistry

Chemical analyses for samples from the migmatite complex on Moskenesøy, Langøy and Hinnøy published in Griffin et al. (1978) are reproduced here as Table 2.1. Seven samples have been analysed for major element contents in this study: one from the Moskenes Pegmatite Quarry and two each from Festvåg, Sandset and Ridderset (Table 3.1).

The sample from Moskenesøy, WL237, is very similar to the mean composition for 22 samples from that island shown in Table 2.1. This sample could not be distinguished from the mangerites in the field, nor on the basis of petrography. Green and Jorde (1971) noted differences between the chemistry of the monzonitic gneiss and that of the intrusive mangerites: the gneiss contains higher Ca, Mg, Fe and lower Si and K. These differences are also true of WL237.

The two samples analysed from the Festvåg migmatite sequence of south-

west Austvågoy, WL549 and WL560, were chosen for their contrasting petrography; this is reflected in their chemical compositions. They have granitic and gabbroic compositions respectively.

The migmatites near Sandset in central Langøy have differences in grain size, intensity of foliation and segregation into mafic and felsic layers, but do not have the major division into two groups of widely different bulk composition that is apparent for the migmatites at Festvåg. This is reflected in the bulk composition of the two samples, WL725 and WL728, analysed from central Langøy. Although the former has only a weak foliation and the latter has centimetre-scale gneissose banding, the chemistry of the two samples is very similar. Their compositions are also very similar to the average shown in Table 2.1 for 56 migmatite samples from Langøy, i.e. granodioritic.

The two samples analysed from the northwest Hinnøy amphibolite-facies gneisses at Ridderset, WL661 and WL662, are more acid than the average of 40 samples from Hinnøy shown in Table 2.1. SiO_2 content is 66.8% and 69.5% for WL661 and WL662 respectively compared to the average value quoted of 61.7%. Correspondingly most of the other oxides are slightly lower in the two samples analysed in this study than for the average Hinnøy migmatite.

3.4 Isotopic Results

3.4.1 Nd Isotopes

Of the seven migmatite samples for which the chemical compositions have been determined, five have also been analysed using the Sm-Nd isotopic method (Table 3.2).

The two samples from near Sandset give T_{DM} model ages (using De Paolo's

(1981) depleted mantle model, see Appendix 3) within error of 2.7 Ga (Fig. 3.4) confirming that they are part of the late Archaean migmatite complex and not of the early Proterozoic supracrustal sequence which crops-out on the northern shore of Skjelfjord. The two samples from the Festvåg migmatites yield widely different T_{DM} model ages (Fig. 3.5). The granitic sample, WL549, gives a T_{DM} of 3.1 Ga, whereas the gabbroic sample, WL560, has a T_{DM} of 1.9 Ga. The sample from Moskenesøy, WL237, has T_{DM} = 2.2 Ga (Fig. 3.6). All five rocks are light rare earth element (LREE) enriched, including the gabbroic sample from the Festvåg migmatites, WL560. The LREE-enriched nature of these rocks and their Sm-Nd ratios means that they define reliable Nd model ages. Jacobsen and Wasserburg (1978) also analysed a migmatite sample from Reine on Moskenesøy and found that it gave a significantly younger model age (T_{CHUR} = 2.0 Ga) than the migmatites from Langøy and Hinnøy (T_{CHUR} = 2.6 - 2.7 Ga). They concluded that the protoliths of the Moskenesøy migmatites are distinctly younger and not related to the Langøy and Hinnøy migmatites. The present study supports this conclusion.

The young model age (T_{DM} = 1.9 Ga) of the gabbroic sample from the Festvåg migmatites suggests an association of the basic components of the migmatite with the intrusive activity at 1.8 - 1.7 Ga (especially with the adjacent Hopen Massif which has a gabbroic marginal series). However, none of the mangeritic or basic rocks analysed for this study (Chapters 5 and 6) has given a depleted mantle model age as young as 1.9 Ga. The sample could conceivably be associated with the Proterozoic dolerite dyke swarms which post-date mangerite emplacement, but no Nd studies have been conducted on these rocks.

3.4.2 Sr Isotopes

Whole-rock Sr isotope analyses have been obtained from six Sandset migmatite samples (WL723 to WL728) (Table 3.3) and from thirteen Festvåg migmatite samples (WL549 to WL562, except WL557) (Table 3.4). These data do not define isochrons (Figs. 3.7 and 3.8).

T_{DM} model ages show that the protoliths of the Sandset migmatites were formed 2.7 Ga ago (Fig. 3.4). A 2.7 Ga reference isochron with $(^{87}Sr/^{86}Sr)_o$ of 0.7013 is shown in Fig. 3.7 in relation to the Rb-Sr whole-rock isotopic data for the six Sandset migmatite samples. All of the Sandset data-points lie above and to the left of the 2.7Ga isochron. This distribution could be produced by either Rb-depletion or addition of Sr at some time in the past. Within the context of the known geological history of Lofoten-Vesterålen (Table 2.1), the most plausible explanation is Rb-depletion during the granulite-facies metamorphism around 1.8 Ga ago. Assuming $^{87}Sr/^{86}Sr$ = 0.7013 (see Table 3.3) at 2.7 Ga and with measured present-day $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ values, the $^{87}Rb/^{86}Sr$ ratio between 2.7 Ga and 1.8 Ga can be calculated, as can a "depletion factor" reflecting the change in $^{87}Rb/^{86}Sr$ effected by the metamorphism (Table 3.5). The range for this value $([^{87}Rb/^{86}Sr]_{1.8-0}/[^{87}Rb/^{86}Sr]_{2.7-1.8})$ is 0.18 to 0.72. The Sr evolution diagram illustrating this model is shown in Fig. 3.9.

Fig. 3.8 shows the Rb-Sr whole-rock isotope data from the Festvåg migmatites plotted with a 2.7 Ga reference isochron for $(^{87}Sr/^{86}Sr)_o$ = 0.7013 (as for the Sandset data points in Fig. 3.7). Using an identical model to that used for the Sandset migmatites most of the samples appear to have suffered Rb-depletion at 1.8 Ga. Those samples which show an apparent enrichment in radiogenic Sr (WL556-WL560) are of gabbroic composition (except WL549), and the Nd model age for one of the samples (WL560) has shown that the mafic components of the migmatites may belong to the 1.8 Ga intrusive

event. (Note how these samples, WL555-WL560, compare to a 1.8 Ga reference isochron with $(^{87}Sr/^{86}Sr)_o$ = 0.7020 in Fig. 3.8). The range of values for the "depletion factor" for the acid and intermediate samples, WL550-WL554 and WL561 & WL562, is 0.33 to 0.90 and for WL549 it is 1.60. The Sr evolution diagram illustrating this model is shown in Fig. 3.10.

3.5 Conclusions

1. It may be impossible to distinguish some migmatites from the mangerites (notably on Moskenesøy), both in the field and petrographically, but they may be resolved by differences in chemical composition e.g. migmatites on Moskenesøy have higher Ca, Mg, Fe and lower Si and K contents than the mangerites.

2. Locally, at Festvåg, the migmatites have been heated to produce melts of intermediate composition, which may have mangeritic affinities.

3. The protoliths of the Langøy migmatites formed ca. 2.7 Ga ago; this is in agreement with Jacobsen and Wasserburg (1978), who report similar protolith formation ages for Langøy and Hinnøy migmatites in both amphibolite and granulite facies.

4. The protoliths of the Festvåg migmatites were probably formed at the same time as those of Langøy and Hinnøy, but the Festvåg migmatites now contain a younger basic component which may be associated with the emplacement of dolerite dyke swarms after the main phase of mangerite intrusion.

5. The migmatites on Moskenesøy are distinctly younger than, and not related to, those of Langøy and Hinnøy, again in agreement with the conclusion of Jacobsen and Wasserburg (1978).

6. Rb-Sr isotope systems have been disturbed in the migmatites at Sandset, Langøy and at Festvåg, Austvågøy. The present distribution of data-points is probably best explained by Rb-depletion during 1.8 Ga granulite-facies

CHAPTER 4. THE EARLY PROTEROZOIC SUPRACRUSTAL ROCKS

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CHAPTER 4. THE EARLY PROTEROZOIC SUPRACRUSTAL ROCKS 2100 - 1800 Ma

4.1 Introduction

In the Lofoten-Vesterålen province, during the early Proterozoic, a major pile of supracrustal rocks was deposited on a basement consisting of the migmatitic gneisses. Both the migmatites and the supracrustal rocks at the present level of erosion were subjected to metamorphism in granulite facies at ca. 1800 Ma. Together these two groups of gneisses constitute the country rock into which the basic and mangeritic intrusions were emplaced.

Griffin et al. (1978) postulate a density control on the depth of emplacement of the mangerite intrusions, which are often located at the interface between the denser migmatitic gneisses and the lighter supracrustal gneisses. Thus the supracrustal gneisses probably represent the highest level in the crust which could be considered as a possible source of material components for the mangerite magmas - possibly contributed by processes of partial melting or assimilation-contamination. In order to review the origins of the mangerites, then, it is necessary first to note the characteristics of possible sources of material contributions to their parent magmas, including the supracrustal gneisses considered in this chapter.

4.2 Fieldwork and Petrography

The supracrustal rocks make up most of the gneiss complex on Flakstadøy, Vestvågøy and Austvågøy and occur as isoclinal infolds in the migmatite complex on Moskenesøy and Langøy (Griffin et al., 1978).

The supracrustals have been examined in this study at the following localities (samples which have been used for geochemical and isotopic studies are described in this section):

(i) at Ølkona, and 1km to the south-west, on Moskenesøy;

(ii) near Hamnøy Bridge, Moskenesøy;

(iii) between Å and Sørvågen, Moskenesøy;

(iv) near Sund, Bø and Nesland, Flakstadøy;

(v) near Vik, west Vestvågøy;

(vi) near Ramberg and Smorten, North Vestvågøy;

(vii) in the south of Austvågøy;

(viii) near Erikstad, South Hinnøy.

Most work has been done on Moskenesøy because there the lowest part of the supracrustal sequence is exposed.

4.2.1 Ølkona, Moskenesøy

The geological map of Moskenesøy produced by Green and Jorde (1971) shows the rocks between the Ølkona Mangerite and the Ølkona Gabbro as belonging to the monzonitic-dioritic gneisses (i.e. the late Archaean migmatites), but in at least two localities, rocks belonging to the veined and layered gneisses (i.e. early Proterozoic supracrustals) are found. One is on the easternmost tip of the Ølkona headland, the other is on the headland 1km to the south-west (Fig. 4.1).

The gneisses of the Ølkona headland have a banding which strikes 012° with a dip of 75° to the west. Within the gneisses a 4m wide marble layer clearly demonstrates the supracrustal origins of the rocks at this locality. The marble layer continues along stike for 100m, before running on out to sea at both ends. The marble consists of calcite crystals (and in some samples dolomite) up to 10mm across, in a granular texture and occupying about 80% of the volume of the rock. The other phases, silicates and spinel may be disseminated through the rock or in bands a few centimetres wide. See Table 1 for a list of

parageneses of the marbles and their significance. Plate 4.1 shows the margin of the marble containing phlogopite-rich bands. Diopside and forsterite are other important silicate phases in the marble. In some samples, most of the forsterite has been serpentinized and large areas of the calcite recrystallized. Diopside has been seen overgrowing forsterite. Chondrodite and apatite are accessory phases in the marble. Another feature of the marble is that whereas bands of silicates at the margin may be continuous over a number of metres, in the centre silicate-rich bands have been broken up and are randomly orientated in relation to the margins of the marble (Plate 4.2). This is probably due to the competency contrast between silicates and carbonates. Some layers of very impure carbonate are more competent than the more pure carbonate within which they were deposited and hence are present as blocks up to 25 x 12cm within the centre of the marble band; these blocks have subsequently developed reaction rims. In thin section, many of the minerals in the marble show evidence of extreme deformation e.g. calcite recrystallization, kinking in phlogopite, and silicate layers at the margin of the marble are often boudinaged (Plates 4.1 and 4.3).

Glassley (1975) described carbonate bodies from the supracrustal gneisses on Austvågøy and Langøy. The marble at Ølkona is very similar to those described by Glassley (1975), but those which he describes are pods with dimensions of a few tens of metres, whereas this body is continuous for at least 100m and is of constant width. One of the similarities is the presence of a clinopyroxenite mantle to the carbonate bodies, which Glassley describes as being massive, dark green zones which vary in thickness from 50 to 200cm. Plate 4.1 shows a 25cm wide monomineralic clinopyroxene band at the margin of the Ølkona marble. This band consists of equigranular diopside grains, 5mm across, with recrystallization along grain boundaries. Other layers adjacent to the marble include

spinel clinopyroxenite and garnet clinopyroxenite. Further from the marble the mantles become amphibole and phlogopite rich, as was also observed by Glassley (1975). This grades into quartzo-feldspathic gneisses. The total width of the clinopyroxenite mantle at Ølkona is 4m.

The gneisses are intensely deformed and broken up in places 5-10m west of the marble layer (Plate 4.4); at these localities it appears that some melt development took place, represented today by leucocratic material between disrupted gneiss blocks (see Chapter 7).

A vein cuts quartzo-feldspathic gneiss and the clinopyroxenite mantle to the marble layer, but it cannot be traced into the marble layer itself. A sample of this vein (WL244) in the quartzo-feldspathic gneiss is shown in Plates 4.5 and 4.6. The vein is 14mm wide and contains rounded gneiss fragments of similar composition to the host rock, up to 11mm across. Smaller fragments are angular. Opposite sides of the vein match suggesting dilatational opening. The vein is probably a gas-release structure in which the larger fragments have been rounded by abrasion. It is probably related to de-carbonation reactions such as the formation of diopside and chondrodite after calcite and forsterite recognized in WL240 from the Ølkona Marble (Table 4.1, Reaction 4.2).

Early Proterozoic supracrustal rocks also crop out on a headland 1km to the south-west of the Ølkona Marble Locality; this headland is about 200m north of the northern contact of the Ølkona spotted gabbro (Green and Jorde, 1971). Clinopyroxenite layers, similar to those adjacent to the marble band 1km to the north-east, are present (Plate 4.7). These layers, which are dominated by green diopside, but which may also contain significant amounts of spinel and biotite, have been

broken into sub-rounded blocks, typically 20-50cm across. Calcite may be present in fractures, but there is no carbonate layer present at this locality. the quartzo-feldspathic gneisses are intensely deformed here and have a similar appearance to those in Plate 4.4.

Plate 4.7 shows a white-weathering rock of gabbroic-noritic composition, between the clinopyroxenite layers and veining into them. Small pods of leucocratic melt surrounded by mafic selvages occur within the deformed quartzo-feldspathic gneisses. The pale orange weathering rock in Plate 4.7 consists of up to 80% mesoperthite with minor ortho- and clinopyroxene and opaques, with accessory apatite and zircons and is similar to the leucocratic melts seen adjacent to the Ølkona marble.

4.2.2 Hammøy Bridge, Moskenesøy

The northern, Hammøy, end of the Reine-Hammøy bridge rises from a peninsula with Hammøy Harbour to the west and the Vestfjord to the east. The early Proterozoic supracrustals on the eastern side of the peninsula have been examined in detail (Fig. 4.1).

Green and Jorde (1971) described the gneisses in this area and south of Sørvågen as, "frequently containing monzonitic layers". Plate 4.8 shows coarse stringers in well-foliated gneiss; this gneiss is dominantly perthite, with minor ortho- and clinopyroxene, opaques and biotite, and calcite associated with the breakdown of clinopyroxene. Another gneiss sample, WL255, has more abundant monzonite layers and on a finer scale. The gneiss consists of layers of plagioclase, ortho- and clinopyroxene, opaques and quartz with grain size less than 1mm, comprising 60-70% of the rock; the remainder consists of alkali feldspar, plagioclase, quartz, opaques and orthopyroxene, with a grain size of 2mm.

Material representing melts is present as veins and dykelets from millimetre to metre scale. Such veins appear to be intimately associated with gabbroic material (Plates 4.9 and 4.10). Plate 4.9 shows an anastomosing network of gabbroic veins in the supracrustal gneisses. In places pinkish-grey patches of melt are seen around the mafic veins. Plate 4.10 shows thin mafic veins which are broken up by the development of a leucocratic melt. The melts at Hamnøy Bridge will be described and discussed in detail in Chapter 7.

4.2.3 Å-Sørvågen, Moskenesøy

The early Proterozoic supracrustal gneisses along the southern end of the Lofoten Highway between Å and Sørvågen (Fig. 4.1) consist of massive granulites consisting mainly of perthites with minor ortho- and clinopyroxene, and opaques, with accessory apatite, quartz and zircon, and of layered gneisses with coarse-grained leucocratic layers and more fine-grained, melanocratic layers. The layered gneiss at Tind consists of plagioclase in granular texture with ortho- and clinopyroxene, all these minerals being coloured pinkish-brown by inclusions, biotite which defines the gneissosity, opaques, and accessory garnet, hornblende and apatite. The garnet occurs as rims to the opaques. Layers parallel to the gneissic fabric are reddish-brown and consist of plagioclase and pyroxenes up to 4mm across. Pyroxenite lenses up to 1cm across are associated with the gneiss where it has the reddish-brown, coarse-grained layers.

All lithologies on Moskenesøy are affected by ultramylonites. The supracrustal gneisses at Ølkona and Hamnøy include ultramylonite zones up to 10cm wide. A thin-section of a supracrustal gneiss sample from Sørvågen includes a 2cm wide ultramylonite zone in which all minerals are crushed beyond identification, except for trails of isotropic garnets.

At the margins of the ultramylonite zone relict pyroxenes remain as augen.

The supracrustal gneisses around Å are intruded by reddish-brown sheets up to 2m wide. They are best seen immediately north of Å Camp-site. These sheets were recognized by Green and Jorde (1971), who called them the "monzonitic dykes and veins". They are discussed further in Chapter 7.

4.2.4 Sund, Bø and Nesland, Flakstadøy

Romey (1971) marked the southern contact between the Sund Mangerite and the layered gneiss (early Proterozoic supracrustal rocks) as gradational On the eastern side of the isthmus protecting Sund Harbour (Fig. 4.2), the rock has a granular texture in places and good gneissose banding in others. This banding strikes 006° and dips 45° to the west towards the Sund Mangerite. The layered gneiss at the northern contact of the mangerite is quartzo-feldspathic with its foliation defined by thin biotite-rich layers which strike 030° with a near-vertical dip.

The Flakstadøy Charnockite intrudes early Proterozoic supracrustal rocks, west of Bø, in central Flakstadøy (Fig. 4.2). On the eastern side of the charnockite intrusion the gneiss strikes 005° and dips 80° to the east; it is light grey and largely consists of feldspar and biotite-rich mafic layers. Coarse-grained 20cm wide veins cross-cut the gneiss, xenoliths of which are present in the veins. The veins strike 055°, and consist of feldspars, and dark green clinopyroxenes up to 5cm.

On the eastern side of Skjelfjord, the supracrustal gneisses are well exposed on the foreshore and the contact with the Flakstadøy

Mangerite is only about 100m inland (Fig. 4.2). Here the rocks are quite basic; mafic phases (orthopyroxene and opaques with minor clinopyroxene and biotite) make up about 25% of the volume of the rock; the remainder is plagioclase with small amounts of alkali feldspar. Foliation is defined by variations in grain size (up to 2.5mm) and by orthopyroxene-rich layers. The gneisses here, about 3km north-west of Nesland are cut by two generations of coarse, granular, feldspar-rich sheets, the earlier of which is very similar to those at Å, Moskenesøy. These will be described further, and discussed in Chapter 7.

4.2.5 Vik, West Vestvågøy

An isochron suite of nine samples, WL427 to WL435, was collected from the new road-cut at Vik, on the road to Utakleiv, and about 5km north of Leknes (Fig. 4.2). Tull (1977) mapped the rocks here as "granitic paragneiss" (!). The gneisses are quartzo-feldspathic. They are compositionally banded, and have a strong foliation defined by biotite (Plate 4.11). Rare amphibolite layers may represent pelitic sediments or volcanics. All of the samples, except WL435, were collected from a 10m section of road-cut, and mafic phases comprise at most 10% of the rock volume, except for WL435. The latter sample was collected from 200m north of the others, but in the same road-cut, and has about 15% mafics. A 3cm wide felsic band, rich in allanite and epidote cuts the gneiss at the WL435 site. Quartz, microcline and plagioclase are the main phases in all the samples. Grain size rarely exceeds 2mm. Biotite laths (which are green in some samples) define the foliation, sometimes with opaques. All samples have epidote crystals, smaller than 0.5mm, with yellow-brown cores of allanite. WL434 contains a 30 x 10mm black area with a surrounding halo which is an aggregate of allanite. Zircon and sphene are common phases in some samples (notably WL429 and WL430) and may be up to 2mm long. Accessory phases include apatite, muscovite,

rutile and spinel.

4.2.6 Ramberg and Smorten, North Vestvågøy

The quarry at Ramberg in north Vestvågøy exposes a section through the paragneisses which shows strong compositional banding. The gneiss banding strikes 130° and dips 35° to the north. The rocks in the quarry are largely amphibolites and quartzo-feldspathic gneisses. Quartzites and biotite schists also occur. A sample with finely divided quartzo-feldspathic and amphibolite layers, WL449, consists of microcline-quartz layers with grain size less than 1mm and hornblende-biotite-rich layers, in which the minerals are elongate parallel to the compositional banding, and up to 3mm in length. Minor sphene, epidote and plagioclase and accessory opaques, apatite, zircon and myrmekite are also present. A quartz-micorcline gneiss sample contains abundant calcite-filled fractures; other phases (opaques, biotite and epidote) comprise less than 1% of the rock.

A 200-500m wide band of augen gneiss, mapped by Tull (1977) as cropping out across the width of Vestvågøy, is exposed 200m south of Ramberg Quarry. Alkali feldspar augen up to 4cm across occur (Plate 4.12). The augen have perthitic texture and contain inclusions of all the minerals in the gneiss. Hornblende and biotite-rich bands define the foliation and wrap around the augen.

Many old mine shafts are present on the north coast of Vestvågøy around Smorten. The early Proterozoic supracrustal rocks were worked for iron in these mines. Plate 4.13 shows a spectacular banded iron formation. Magnetite, pyrite and bornite also occur at Smorten.

4.2.7 Southern Austvågøy

The early Proterozoic supracrustal gneisses have been investigated at a number of localities in the southern part of the western half of Austvågøy (Fig. 4.3), especially near the contacts with mangerite intrusions.

It is not possible to draw a simple contact between the Hopen mangerite intrusions and the supracrustal gneisses on the west coast, south of Rørvik. The NGU 1:250,000 Svolvaer sheet (Tveten 1978) shows the contact 1km further north than Ormaasen (1976). The gneiss along the main Å to Svolvaer road at either end of the Rørvik tunnel shows compositional banding with quartzo-feldspathic and biotite-hornblende-rich layers. In places where the content of mafic minerals is very low, no foliation is apparent. The rock in the disputed contact zone is medium- to fine-grained with a weak foliation and is intruded by a dolerite dyke swarm, parallel to the foliation. Samples from this area have charnockitic texture and mineralogy, thus lending support to Tveten's (1978) view on the position of the contact.

The contact between the Austvågøy Grey Mangerite and the supracrustal gneisses is also more complex than shown on Fig. 4.3, which is modified from Tveten (1978). Small outcrops of mangerite occur in Kabelvåg and on the road to Svolvaer. The gneiss to the north-west of Kabelvåg is quartzo-feldspathic with thin bands of biotite and hornblende and it is pink in colour. The gneiss exposed in new road-cuts west of Svolvaer is also pink. It is fine-grained, contains few mafics and has a poorly-developed foliation. Coarse bands have amphiboles up to 3cm long. Mangerite occurs in the same road-cuts.

The supracrustal rocks which crop out east of Svolvaer around the airport at Helle are pink quartzo-feldspathic gneisses with micaceous

layers. WL537 is a sample from a road-cut 6km north of Helle on the west side of Austnesfjord. It has quartzo-feldspathic and biotite-rich layers (Plate 4.14). Pyrite is visible in hand-specimen. Plagioclase and microcline are both present, the former in greater abundance and commonly saussuritized in the cores. This suggests that the plagioclase is zoned, with an albite-rich rim since sodic plagioclase does not undergo saussuritization (Deer et al., 1966). Euhedral epidote is abundant in the biotite-rich layers, is up to 1.5mm across and shows twinning and zoning. Euhedral sphene is also quite common. Accessory phases present are muscovite, opaques, apatite and calcite.

4.2.8 Erikstad, South Hinnøy

Within the Lødingen Granite near Erikstad on South Hinnøy is a very large raft of early Proterozoic supracrustal gneiss. In a road-cut it consists of pyroxenite which is weathering yellow and green due to the breakdown of the highly abundant sulphides, which are mostly pyrrhotite, although pyrite and bornite are also present (Plate 4.15). The pyroxenite is fine-grained (typically less than 1mm) and in places contains no opaque phases, but irregular bands consist of up to 70% opaques. In the hill-side, about 100m along strike from the sulphide-bearing pyroxenite, is a banded ironstone (Plate 4.16). Bands of hedenbergite and grunerite, the latter with a darker green amphibole in its core, are overprinted by opaques which may occur in aggregates up to 5mm long. This ironstone is likely to have been a haematitic chert, possibly formed in deep water.

4.3 Geochemistry

Chemical analyses for samples from the supracrustal gneisses on Moskenesøy, Austvågøy and Langøy published in Griffin et al. (1978) are reproduced here as Table 2.2. Three samples have been analysed for

major element contents in this study: one from Flakstadøy north-west of Nesland and two from Vik, west Vestvågøy (Table 4.2).

The sample from Flakstadøy, WL339, has a chemical composition very similar to the mean composition for 13 samples from the adjacent island, Moskenesøy, shown in Table 2.2. This sample is basic in composition and is similar to the composition of WL237, the Moskenesøy migmatite sample, shown in Table 3.1. Tables 2.1 and 2.2 show that the gneisses of both units are more basic on Moskenesøy than on the islands to the east.

The two samples analysed for major elements from the Vik gneiss isochron suite, (WL428 and WL432), have compositions within error of each other (Table 4.2). They have granitic compositions (SiO_2 = 75%) which are much more acid than the mean values shown in Table 2.2 and at the extreme acid limit of the range for 81 supracrustal gneisses shown in Table 2.2.

Griffin et al. (1978) inferred a volcanogenic origin for the bulk of the early Proterozoic supracrustal rocks from their geochemistry. From the $Na_2O + K_2O$ v. SiO_2 diagram of Cox et al. (1979), WL339 is at the basaltic end of the trachyandesite field; the Vik gneisses, WL428 and WL432, have the composition of rhyolites.

4.4 Isotopic Results

4.4.1 Nd Isotopes

The three supracrustal gneiss samples for which major element data are shown in Table 4.2 have also been analysed using the Sm-Nd isotopic method (Table 4.3).

The samples from the Vik gneisses, Vestvågøy, (WL428 and WL432), give T_{DM} model ages of 2.1Ga (Fig. 4.5) which is the age given for the start of the deposition of the supracrustal gneisses in Griffin et al. (1978), hence they could have been derived from the mantle and deposited at this time. These samples are extremely LREE-enriched i.e. they have very low Sm/Nd ratios, which is to be expected from their very acid composition.

The sample from Nesland, Flakstadøy (WL339) has T_{DM} = 2.3Ga (Fig. 4.6). This model age suggests that a component of the Nd may have been derived from an older crustal source, i.e. by contamination from the late Archaean migmatites at depth.

4.4.2 Sr Isotopes

Whole-rock Sr isotope analyses have been obtained from the Nesland gneiss sample, WL339, and from the nine Vik gneiss samples, WL427 to WL435 (Table 4.4).

WL339 has Rb and Sr concentrations of 4 and 992 ppm respectively and a $^{87}Sr/^{86}Sr$ ratio of 0.70311 ± 5. The only rocks in the Lofoten-Vesterålen province with lower Rb/Sr ratios belong to the Flakstadøy Basic Complex (Chapter 5).

The Vik gneiss data do not define an isochron (Fig. 4.7). Nd model ages show that the Vik gneisses could have been derived from the mantle at 2.1Ga, probably with an initial $^{87}Sr/^{86}Sr$ ratio [or $(^{87}Sr/^{86}Sr)_o$] of ca. 0.7020 (see Table 4.4). A 2.1Ga reference isochron with $(^{87}Sr/^{86}Sr)_o$ of 0.7020 is shown in Fig. 4.7 in relation to the Rb-Sr whole-rock isotopic data for the nine Vik gneiss samples. All of the Vik data-points lie below and to the right of the 2.1Ga isochron. This distribution could

have been produced by either Rb-enrichment or loss of (radiogenic) Sr at some time in the past. Within the context of the known geological history of Lofoten-Vesterålen (Table 2.1), the most plausible explanation is Rb-enrichment during the retrograde metamorphism around 1100 Ma ago. During this metamorphic event the Leknes Group sediments were metamorphosed to lower amphibolite facies and some of the granulite-facies rocks were retrograded to amphibolite-facies assemblages. At deeper crustal levels, high grade metamorphism may have caused contemporary Rb-depletion. Fluids rich in Rb could have risen to crustal levels at which K-rich phases were present and stable. Rb then could have been incorporated into the biotite of the Vik gneisses.

Assuming $^{87}Sr/^{86}Sr = 0.7020$ at 2.1 Ga, and with measured present-day $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ values, the $^{87}Rb/^{86}Sr$ ratio between 2.1Ga and 1.1Ga can be calculated, as can an "enrichment factor" reflecting the change in $^{87}Rb/^{86}Sr$ effected by the metamorphism (Table 4.3). The range for this value ($[^{87}Rb/^{86}Sr]_{1.1-0}/[^{87}Rb/^{86}Sr]_{1.8-1.1}$) is 1.15 to 1.93. The Sr evolution diagram illustrating this model is shown in Fig. 4.8. This "enrichment factor" is analogous to the "depletion factor" used in the models for the late Archaean migmatites in Chapter 3.

4.5 Discussion and Synthesis

The paragenesis of the Ølkona Marble constrains its maximum temperature of metamorphism to at least 600°C (Table 4.1, Fig. 4.4). In Griffin et al. (1978) it was suggested that the reaction Diopside + Dolomite → Forsterite + Calcite took place at temperatures in excess of 750°C at 10kb (see Table 2.4, after Griffin et al. (1978) for estimates of pressure in the Lofoten-Vesterålen province during the 1.8Ga metamorphism). In WL223 this reaction has gone to completion and hence 750°C is a minimum temperature for the granulite-facies metamorphism at

Ølkona, Moskenesøy. In the marbles in the east of the region, where the metamorphic peak was in the amphibolite facies, the reaction has also gone to the right and so 750°C is a minimum temperature of metamorphism in the amphibolite-facies, structurally higher parts of the province. The maximum temperature attained in the structurally deeper parts of the area (Moskenesøy and Langøy) are considered to have been considerably higher than this. Krogh (1975) obtained a temperature of metamorphism of 950°C for Langøy and Vestvågøy from magnetite and ilmenite. Metamorphic temperatures and pressures would have been even greater deeper than the present level of exposure. At 10kb the granite liquidus with 1% of water is at 1100°C (Patchett, 1980) so granitic melts would form in these gneisses if the temperature were raised by about 150°C. This could be achieved at considerably increased depths due to the geothermal gradient or by emplacement of basalt magmas (T = 1200°C, Patchett (1980)). The possibility that the dykes and veins of monzonitic composition present in the supracrustal gneisses, especially on Moskenesøy, are partial melts from the gneisses is discussed further in Chapter 7. The observed more basic character of the migmatitic and supracrustal gneisses on Moskenesøy and Flakstadøy could be due to the extraction of an acid partial melt.

The presence of marble bands within the early Proterozoic gneisses shows conclusively that they are supracrustal in origin. The marbles were probably deposited in shallow lagoons. The banded ironstones were probably formed by metamorphism of haematitic cherts deposited in deeper water. The presence of lagoonal or marine metasediments in a thick pile of volcanic rocks suggests the basin must have been subsiding at a similar rate to the accumulation of lava and ash. An accumulation of about 40km in less than 300 Ma is required to account for the high pressures of metamorphism at the bottom of the supracrustal pile. The high geothermal gradient in

Lofoten-Vesterålen (peak metamorphic conditions of about 800°C and 10kb imply 25-30°C. km.$^{-1}$) suggests that burial by rapid accumulation of volcanic products is appropriate for the Lofoten-Vesterålen province (rather than by continental collision). Krogh (1977) inferred from the high heat-flow, that the early Proterozoic supracrustals were deposited in a back-arc basin. This is a suitable environment for the formation of limestone and chert. Alternatively the supracrustals could have been deposited in an Andean environment on a thin (25km ?) continental margin with shallow seas.

CHAPTER 5. THE MID-PROTEROZOIC BASIC INTRUSIONS

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CHAPTER 5. THE MID-PROTEROZOIC BASIC INTRUSIONS

5.1 Introduction

The major basic intrusions of the Lofoten-Vesterålen province occur on Moskenesøy (c. 8km^2), Flakstadøy (c.40km^2) and Langøy (c. 10, 25 and 8km^2 for the intrusions at Eidsfjord, Hovden and Bø, respectively) (i.e. at the deepest crustal levels of the area). They consist of gabbros and anorthosites, with norites and troctolites present, in addition, on Flakstadøy. Smaller gabbroic bodies (typically <1km^2) are present on Vestvågøy and Austvågøy.

The basic intrusions are associated, both temporally and spatially, with the intrusions of the mangerite series which are really more extensive at the present level of erosion. However, geophysical evidence suggests that large volumes of basic rock may be present at depth in Lofoten-Vesterålen. The origins of these two intrusive series are likely to be closely related. It is a major aim of this work to elucidate the nature of this relationship:

(i) The mangerites and basic rocks may be comagmatic.

(ii) If the mangerites were derived by crustal anatexis, then the basic rocks may have provided the necessary thermal input.

5.2 Fieldwork and Petrography

Previous research on the basic intrusions of the Lofoten-Vesterålen province has been reviewed in Chapter 2. This section describes the samples that have been used in the geochemical and isotopic studies, and the localities from which they have been collected. New field evidence pertaining to the relative ages of the basic and mangeritic intrusions is presented. New petrographic evidence bearing on the pressure and temperature of metamorphism of the basic rocks is presented, and the first report

of probable eclogite-facies metamorphic phases in Lofoten-Vesterålen is made.

5.2.1 Moskenesøy

The Ølkona spotted gabbro (Green and Jorde, 1971) (see Fig. 3.1) has pronounced spotted appearance on weathered surfaces. The spots are up to 10cm across and in thin-section are seen to be due to aggregates of hornblende, individual grains up to 1cm across; intergrown with orthopyroxene and plagioclase. The plagioclase (An_{50}) is dusty due to the presence of minute inclusions and may be recrystallized along grain boundaries. Minor clinopyroxene, and accessory opaques, biotite, apatite and garnet coronas are also present. Griffin and Heier (1969) described a number of garnet-forming reactions in Lofoten-Vesterålen granulites all of which are permissible with falling T at constant P and which are therefore considered to be retrograde reactions.

Ultramylonite zones, c. 0.1mm wide, are common in WL224. The anorthosite in the south of Moskenesøy is not accessible on foot due to the topography of the island.

5.2.2 Flakstadøy

A layered complex of basic rocks occurs in the south-eastern part of Flakstadøy, dips to the east and overlies a body of mangerite to the west. Romey (1971) identified xenoliths of the early Proterozoic supracrustal gneisses in the basic complex and hence showed that the latter intrudes the former. Romey (1971) sub-divided the basic complex into three zones (Fig. 5.1).

The basic complex has been examined and sampled in a number of areas for this study (Fig. 5.1):

(i) between Nesland and the contact with the mangerite, 2km to the west;

(ii) at Nusfjord, and in the road-cuts and on the hills to the north and north-east;

(iii) in Mørkdalen, 1km west of Nusfjord;

(iv) at Napp, north-east Flakstadøy.

Samples were collected for isotopic and geochemical work from all these areas, but the road-cuts around Nusfjord provided the freshest samples.

(i) Nesland

The best layering in the Flakstadøy Basic Complex is seen in the troctolite zone (Romey, 1971) east of Nesland, at the southern end of the central lobe of the island (Fig. 5.1). Plate 5.1 shows compositional and grain-size layering. Peridotite layers up to 10cm thick grade upwards into plagioclase-rich layers, often with an accompanying reduction in grain size. In places, the plagioclase crystals are aligned parallel to the peridotite layers, but elsewhere, as on the left of Plate 5.1, the texture is subophitic. The strike of the peridotite layers is 036°, dipping 52° east. Arcuate peridotite layers resemble channel fillings in sedimentary sequences; their orientation and the graded-upwards layers suggest that the Flakstadøy Basic Complex is the right way up. Pods of porphyritic anorthosite occur within the troctolite, individual plagioclase crystals reaching 10x5 cm, within a subophitic groundmass with plagioclase crystals 1-3cm long. Gabbroic and noritic pods, with grain size in excess of 10cm, are also present.

On the foreshore, 500m south-west of Nesland, in the norite zone (Romey, 1971) very large-scale ophitic textures are developed (Plate 5.2) with plagioclase laths up to 30x5cm. Crescumulate textures have developed

in the gabbronorite: long, thin plagioclase laths, typically 10x0.5cm, grow into residual basaltic magma from layers of cumulus plagioclase. WL344 and WL345 are samples of ca. 10cm wide veins which intrude the gabbronorite; they are both charnockites, but the former is more quartz-rich, consisting of granular perthites and quartz, up to 10mm across with a single altered clinopyroxene crystal. Fine-scale perthites comprise about 80% of WL345. They are coloured pink by minute inclusions and as such are similar to the perthites of WL230 and WL232, monzonitic dykes in supracrustal gneisses at Å, Moskenesøy. Pale-green clinopyroxenes occurring in WL345 have garnet coronas as do opaques which completely enclose one clinopyroxene; this shows that clinopyroxene crystallized prior to opaques in the rocks and the subsequently falling T caused limited retrogression. Quartz occurs in the interstices of the perthite. Accessory orthopyroxene, apatite, zircon and rutile are also present. The gabbro-norite is also cut by dolerite dykes. Two joints intersect at an angle of 50°; a 30cm wide dolerite dyke following one of these joints turns to follow the other joint where they intersect, implying that the planes of weakness in the basic rock pre-date the emplacement of the dolerite.

WL311 is a sample from the norite zone west of Nesland and about 500m east of the contact with the Flakstadøy Mangerite. It was collected in an old road-cut in which most of the rock is very deeply weathered. Plagioclase, An_{48}, up to 25mm long, comprises about 85% of WL311. Olivine, up to 4mm across and altering to opaques, iddingsite and spinel, is overgrown by first ortho- and then clinopyroxene. Biotite overgrows opaques and green spinel. Zircon is present as an accessory phase.

The contact between the basic complex and the mangerite on the road west of Nesland cannot be located to better than 100m. Both lithologies

have reduced grain size near the contact. WL310 is a norite, about 200m from the mangerite, which has a grain size typically less than 2mm. WL346 was collected at the westernmost exposure of the basic complex on the Nesland road. Plagioclase, up to 3mm long, makes up about 70% of the rock. Orthopyroxene is the most abundant mafic mineral with less abundant clinopyroxene, opaques and biotite. Minor antiperthite is also present suggesting this may be a hybrid rock.

(ii) Nusfjord

The new road-cuts in and to the north of Nusfjord, provide the freshest samples of the Flakstadøy Basic Complex. They enable sampling of the anorthosite (uppermost) and troctolite zones of Romey (1971). A 10m wide dolerite dyke striking 015° and dipping 55°W cuts the basic rocks of the anorthosite zone at the southern end of Nusfjord. It has plagioclase laths, 3mm long, enclosed by clinopyroxene in a subophitic texture. Olivine, orthopyroxene, opaques and biotite are also present. Plate 5.3 shows that the rock here is, in fact, a troctolite, and that coronas are developed between olivine and plagioclase. In thin-section, a sample of this rock, WL314, shows that the corona is as described for the anorthosites of Flakstadøy by Griffin and Heier (1969) and as shown in Plates 5.5 and 5.6. Griffin and Heier (1969) described these coronas as consisting of an inner zone of fibrous orthopyroxene adjacent to the olivine, and an outer zone with spinel inclusions, oriented perpendicular to the corona boundary; garnet often appears as an isotropic brown dust at the contact between the inner and outer zones; this brown dust recrystallizes and grows into the clinopyroxene-spinel zone as large grains. Where the olivine is overgrown by pyroxenes in a magmatic reaction rim, no coronas form. Griffin and Heier (1969) used the experimental data of Green and Ringwood (1967) to infer that assuming $T \approx 850^{\circ}C$ during corona formation, the appearance of garnet indicates pressure of

about 9kb. The corona forming reaction was inferred to have taken place with decreasing T at pressures less than 9kb and to have been terminated by falling temperatures.

WL324 and WL325 are samples taken from the troctolite zone, 700m and 300m west of Nusfjord respectively, on the track to Morkdalen. In both samples plagioclase laths, up to 12mm long, comprise about 80% of the rock and olivine is the main mafic phase. Except where the olivine is mantled by pyroxene, double coronas, as described above, are developed. The plagioclases (and the orthopyroxenes of WL325) contain minute, oriented opaque needles. WL324 contains quite abundant reddish-brown biotite and opaques (with green spinel) up to 2mm across. WL325 contains accessory biotite, opaques and rutile. A 30cm wide sheared vein of charnockitic composition, WL321, strikes 084° through the troctolite at the site of WL327, 700m west of Nusfjord.

Plate 5.4 shows a polished slab of WL326, collected 1km north of Nusfjord in the troctolite zone. Plagioclase laths are up to 10cm long. The polished slab has advantages over the thin-section for such a coarse-grained rock. The thin-section of WL326 shows euhedral plagioclase with interstitial orthopyroxenes (plus minor clinopyroxene and opaques, the latter with biotite-garnet coronas). The polished slab in Plate 5.4 shows two olivines about 3cm across mantled by orthopyroxene magmatic reaction rims. It also shows that the plagioclases contain fractures in the same orientation throughout the rock unrelated to cleavage orientation. This feature indicates that the rock was subjected to stress after consolidation of the magma.

The coronas shown in Plates 5.5 and 5.6 are from WL330, a troctolite sample from 2km north of Nusfjord at the eastern end of Storvatn, the

lake in central Flakstadøy. This sample also includes coronas between biotite and plagioclase which appear to be clinopyroxene-spinel symplectites, like the outer zone of the olivine-plagioclase coronas. Dusty garnet is developed between the corona and the biotite.

Romey's (1971) anorthosite zone forms the rounded hills to the east of Nusfjord and Storvatn. There is 100% exposure on the tops of these hills but the rocks are deeply weathered. WL329 is a sample from these hills, about 800m east of WL330. Plagioclase, An_{60}, up to 15mm long, occupies about 95% of the slide and contains minute opaque needles. Strongly pleochroic orthopyroxene, opaques with associated green spinel, and accessory biotite comprise the rest of WL329.

(iii) Mørkdalen

Mørkdalen is a natural amphitheatre in the centre of Flakstadøy, 1km west of Nusfjord. Streams drain from a ring of mountains over 700m high, via three small lakes and into the sea at Nusfjord. Romey (1971) marks the contact between the Flakstadøy Mangerite and Basic Complex as exact around the western shore of the westernmost of these three lakes; his map shows an area of gabbro adjacent to the mangerite. WL324 (troctolite) and WL321 (sheared charnockite) were collected from the eastern end of the westernmost lake. Along the southern side of this lake, towards Mørkdalen and the Flakstadøy Mangerite, charnockite veins are common, striking east-west, usually only a few centimetres wide, but some may be 30cm wide. Areas of very coarse gabbros and norites are seen to the south of this lake, often showing crescumulate textures. Pyroxenite layers pass into a perpendicular feldspar rock with intercumulus mafic minerals, and then into a finer-grained gabbro, before an abrupt change to pyroxenite. Each cycle is about 80cm thick and includes feldspars up to 15cm long. At the south-west corner of the lake the gabbros and norites

are cut by shear zones (Plate 5.7) which appear to cause a change in mineralogy (Plate 5.8). Two shear zones are present, each strikes 126°, the eastern shear dips 54°E and the western shear dips 77°E. Only 2m from the western shear zone, the gabbro contains 10cm plagioclase laths: adjacent to the shear plagioclase seldom exceeds 1cm in length. Nevertheless, the transition from unfoliated to foliated rock is abrupt (Plate 5.7).

The undeformed basic rock to the south-west of the westernmost lake consists of plagioclase (An_{45}) comprising about 70% of the rock, with varying proportions of ortho- and clinopyroxene, minor biotite and opaques, with rare hornblende or garnet coronas (WL232, WL357). WL353-WL356 and WL357-WL359 represent traverses into the eastern and western shear zones respectively. WL359 is only 30cm from the undeformed gabbronorite sample, WL357. Changes in mineralogy include saussuritization of plagioclase and alteration of pyroxenes to hornblende such that albite- and hornblende-rich domains form, with small grain-size, which may be streaked out to produce a chaotically-foliated rock (Plate 5.8). Subhedral to euhedral garnets up to 2mm across form. Zoisite and glaucophane are developed in accessory amounts. Opaques are altered to rutile. In places within the shear, chaotic foliation is superceded by very strong, parallel shear fabrics, with all minerals reduced to less than 0.1mm, except for garnets, up to 3mm across, which act as augen in the hornblende-biotite-rich layers (WL356, WL360). Opaque and rutile layers may be up to 15 x 0.2mm. WL323 (a polished slab of which is shown as Plate 5.8) contains garnets which show multi-stage growth (Plates 5.9 and 5.10). Plate 5.9 shows a single garnet with a pink, inclusion-free core, a dusty inner rim and a colourless, inclusion-free outer rim. Plate 5.10 shows more typical euhedral garnets with spongy cores and idiomorphic rims. Droop (1983) described similar garnets

from the Tauern Window in the Eastern Alps. He described how such garnets grew as rings around omphacite in quartz eclogites during amphibolitization, eventually replacing the omphacite producing euhedral garnets with spongy, quartz-filled cores and idiomorphic rims and which are in equilibrium with albite and hornblende. The dominant phases of WL323 are albite, garnet, diopside and hornblende, with minor tremolite, opaques and zoisite, plus accessory biotite and rutile. Each shear zone is only a few metres wide and the samples with the highest grade assemblages, WL361, were collected only 100cm from undeformed gabbro. WL361 contains pink to orange garnets, up to 7mm across which with fine-grained clinopyroxenes and hornblende comprise over 95% of the rock. The clinopyroxene up to 1mm across, is pale-green and may be omphacitic. The remainder is largely rutile, occurring as stringers up to 15 x 1mm, altered from opaques. Accessory amounts of albite, opaques and glaucophane are present, the latter within spongy garnets. The presence of glaucophane, large amounts of rutile, spongy garnets, and probably omphacite implies that this rock and other Mørkdalen samples such as WL323, WL354, WL355, WL356 and WL359, were eclogites which have subsequently been retrograded to amphibolite facies mineral assemblages. This is the first report of eclogitic rocks in the Lofoten-Vesterålen province. 10m to the west of the western shear zone is sheared mangerite, but the contact could not be accurately located.

(iv) Napp

The northernmost exposure of the Flakstadøy Basic Complex is on the east coast of the island, north of Napp. The rock is a fine-grained (about 0.5mm grain size) gabbro with plagioclase, clinopyroxene and minor antiperthite, opaques and orthopyroxene. Garnet rims occur between opaques and plagioclase, and clinopyroxene and plagioclase (WL317). In the small quarries, north of Napp, the gabbro contains

abundant gneiss xenoliths from the early Proterozoic supracrustals, and is cross-cut by mangerite/charnockite veins up to 10cm across.

To the south of the Napp gabbro, on the south side of the valley running south-west from Napp is a porphyritic norite described by Romey (1971). Extensions to the Napp ferry terminal provide access to fresh samples of the norite. Plagioclase phenocrysts up to 15 x 5 cm are present. A sample of this rock, WL363, has an ophitic groundmass consisting of plagioclase (some with exsolved antiperthitic potassium feldspar), orthopyroxene, opaques and minor biotite, with grain size of about 0.5mm. Carlsbad and albite twinning occurs in both groundmass and phenocryst plagioclase. Within 10m of this sample of porphyritic norite is a coarse-grained (up to 7mm) gabbro with opaques comprising about 10% of the rock. Opaque-orthopyroxene and biotite-orthopyroxene symplectites and, in a single case, olivine with a double corona against plagioclase, are also present.

5.2.3 Vestvågøy

The basic intrusives of Vestvågøy occur as small bodies in several locations mostly in the northern third of the island. Tull (1977) produced evidence that they pre-date the mangerites.

The basic intrusives of Vestvågøy have been examined and sampled at the following localities (Fig. 5.2):

(i) near Borge, north Vestvågøy;

(ii) between Vik and Våje, west Vestvågøy

(iii) west of Ramberg, north Vestvågøy

(i) Borge

The Borge gabbro is the largest basic intrusive body on Vesvtågøy, with an area of about 14km^2. It is clearly intrusive into the supracrustal gneisses but its relationship with the mangerite is unknown (Tull, 1977). The gabbro is quarried for roadstone which provides constantly renewed exposures. In the quarry the gabbro is cut by a metre-wide mangerite dyke. Both the gabbro and the mangerite are cut by two charnockitic sills, each 30cm thick. These will be described further and their significance discussed in Chapter 7.

A sample of the Borge gabbro, WL465, has been collected from the roadstone quarry. Dusty plagioclase (An_{42}), up to 3mm across, makes up 50% of the rock. Minor olivine is overgrown by much more abundant clinopyroxene; both contain opaques, as needles along cleavage planes in the clinopyroxene and irregularly oriented in olivine. Biotite overgrows clinopyroxene and fine-grained garnet rims are present between mafic phases and plagioclase, especially in double coronas around olivine. There is rare alteration of clinopyroxene to hornblende in WL465, and accessory calcite.

1km to the north-west of the roadstone quarry, within 100m of the contact between the Borge gabbro and the supracrustal gneisses, the gabbro has been altered to garnet amphibolite. It contains a 15 x 10cm xenolith of the supracrustal gneisses and is intruded by a mangerite/charnockite vein. The garnet amphibolite is not foliated and the gneiss xenolith is undeformed. Garnets are up to 3mm across and have dusty cores and inclusion-free rims. A fine-grained intergrowth of albite and hornblende (although the latter may be up to 3mm across) makes up the bulk of the rock. Biotite-rich patches are common, calcite-rich patches less so. Accessory phases include zoisite, opaques and rutile.

(ii) Vik-Våje

Tull's (1977) geological map of Vestvågøy marks a basic intrusive body, a few hundred square metres in area, 1km east of Våje, west Vestvågøy. To the north, towards Vik, Tull marks the rock as 'granitic paragneiss', i.e. early Proterozoic supracrustal gneiss. The hill to the north of Tull's Våje gabbro was re-examined for this study. At least 1km north of this gabbro, basic rocks are still exposed; these may continue for a short distance to the north and east, but 1km to the north of the northernmost observed gabbroic outcrop, is the site of collection for the Vik gneiss isochron suite of samples, WL427-WL435. Most of the gabbro has been altered to garnet amphibolite.

At the northernmost basic outcrop seen, 1km south of the Vik gneiss site and at the foot of the hills about 200m east of the road, unaltered basic rocks are present in close proximity to sheared garnet amphibolites. WL437 and WL438 were collected within 30cm of each other. The former is a subophitic gabbronorite with plagioclase laths up to 5cm, which exhibit both Carlsbad and albite twinning. Clinopyroxene is less abundant than orthopyroxene and is overgrown by it. The latter is up to 10mm long, strongly pleochroic, contains opaque needles along cleavages and has hornblende rims. Garnet rims form between biotite and plagioclase. WL438 is very similar to the garnet amphibolite in the Borge gabbro, WL462. Relict ortho- and clinopyroxene are rimmed by biotite and amphibole and an outer rim of fine-grained garnet. Pseudomorphs after pyroxenes consist of hornblende, albite, biotite and opaques, and are rimmed by pink, euhedral garnets,up to 1mm across. Accessory phases include glaucophane, apatite, rutile and white mica associated with saussuritisation of plagioclase. A few metres north of the site of WL437 and WL438, the transition from gabbro to garnet amphibolite is associated with shearing, as at Mørkdalen, Flakstadøy. The transition from undeformed gabbro,

through sheared gabbro with garnet development, to garnet pyroxenite/amphibolite can be observed over a few centimetres. The shear foliation is oriented 045° dipping 54°S. The width of the garnet amphibolite cannot be determined here, but it is at least 2m. WL476 is a garnet amphibolite with a strong shear fabric, defined by biotite plates, zoisite needle and by lenses of very fine-grained diopside. Albite and hornblende are the major phases, the latter often occurring in aggregates with quartz. Accessory rutile is also present. The textures and mineralogies of this rock and WL438 are very similar to garnet amphibolites from Mørkdalen, Flakstadøy, which were inferred to be amphibolitized eclogites.

Garnet amphibolite crops out on this hill up to the 120m contour (and may continue above this height to the north-east) from the shear zone from which WL476 was collected, to the south and east to the gabbro marked on the coast road by Tull (1977). WL477 was collected about 200m south of WL476. WL477 is not foliated but 10m away the garnet amphibolite has a very strong foliation striking 032°, dipping 56°E. WL477 contains garnet up to 3mm across, occurring in aggregates up to 10mm across. Some garnets have spongy cores. Pale green hornblende comprises most of the rock; as well as garnet it also contains minor albite, quartz and biotite, and accessory opaques, rutile and sphene. Marked changes in grain size can be observed across the hill. WL478 is a sample of the coarsest garnet amphibolite seen. It is from a loose block. Garnets are up to 5mm across and have many inclusions, some garnets contain rutile, others plagioclase. After garnet and hornblende, rutile is the next most abundant phase as most of the opaque phases have altered to rutile. The textures within these rocks suggest that they were formerly eclogites whose mineral assemblages have subsequently been retrograded to amphibolite facies. The western contact of the basic intrusive body marked by Tull (1977) with the mangerite shows their relatives ages, and that

two phases of mangeritic activity have occurred. A breccia of metre-scale angular blocks of unaltered gabbro in mangerite is exposed in a road-cut. The breccia is cut by a 6cm wide mangerite/charnockite vein.

The road-cut 1km east of Våje provides the best opportunity to examine the mineralogical and textural variations in the Vik-Våje basic rock. WL440 is very similar to the undeformed basic rock collected from 1km to the north-west, WL437, except that it has more clino- than orthopyroxene and hence is a true gabbro. This gabbro to garnet amphibolite/eclogite transition can be seen to occur over a few centimetres. The relation between foliation and the transition is not, in this case, simple. At the west end of the road-cut, vertical, 10cm wide alternating bands of gabbro and garnet amphibolite occur. A foliation is developed in both lithologies, but it intersects the transition boundaries at 30° on the rock surface. Plate 5.11 figures a sample, WL442, which shows this transition. The gabbro at the extreme left of the plate (WL442D), i.e. about 5cm from the transition, is similar to samples WL437 and WL440. Plagioclase is up to 3mm across and is full of very fine, dusty inclusions. Clinopyroxene is more abundant than orthopyroxene and has minute inclusions which colour it brown, and also opaque needles along cleavage traces. The orthopyroxene is strongly pleochroic and free of inclusions. Both ortho- and clinopyroxene partially altered to hornblende and biotite. Garnet rims are common around biotite. A section through the gabbro up to the transition (WL442C) shows plagioclase full of minute, dusty inclusions, especially at the contact with mafics. Pyroxenes are extensively broken down to hornblende and biotite with fine-grained garnet rims. Opaque needles are more common in clinopyroxene. At the garnet amphibolite end of the thin-section, most pyroxenes are pseudomorphed by aggregates of hornblende, biotite and opaques, rimmed by euhedral garnets up to 1mm across. Plagioclase is altered to fine-grained albite

in which original grain boundaries are not recognisable at this end of the thin-section. In WL442B, in the garnet amphibolite area, most pyroxenes are completely pseudomorphed. White mica is present associated with the saussuritisation of plagioclase. Rutile is present in the thin-sections from the gabbroic part of WL442, but only as a few grains. In WL442B and WL442A, from the garnet amphibolite part, most of the opaques have altered to rutile. WL442A from the core of the garnet amphibolite part of WL442 is similar to WL438, the garnet amphibolite from 1km to the north-west towards Vik. Rare relict pale green clino-pyroxene remains. Euhedral garnets, many with spongy cores, are up to 2mm across. WL443 is another garnet amphiboite from within 2m of WL442. It contains no relict pyroxenes. Zoisite needles and white mica are present in albite-rich areas. Most opaques are altered to rutile. These garnet amphibolites have textures and mineral assemblages that suggest they are amphibolitized eclogites. A 20cm wide lens of pyroxenite with a granitic core occurs within the garnet amphiboilite. WL446 was collected within 20cm of WL443 and this rock is 95% pale-green clinopyroxene in a granular texture; the remainder is olivine altered to iddingsite, opaques and very rare hornblende. The core to the pyroxene, WL447, consists of fine-grained quartz and plagioclase with abundant biotite and garnet. There is a garnet-hornblende marginal zone between the pyroxenite and the granitic band. The alteration of opaque phases to rutile is a common feature in the garnet amphibolites in the Vik-Våje area as it is in Mørkdalen, Flakstadøy. Rutile is best developed in WL448, from the west end of the road-cut east of Våje. Individual rutile grains are up to 1mm across and occur in aggregates up to 5mm long. Hornblende can be seen to be replacing pale-green clinopyroxene grains volume by volume. Aggregates of acicular actinolite and plagioclase are also present. The gabbroic and garnet amphibolite parts of this road-cut are both cross-cut by charnockitic pegmatites which contain small, relict xenoliths

of garnet amphibolite. A sample of one of these pegmatites, WL441, contains over 95% alkali feldspar, up to 20mm across, with quartz occurring along feldspar grain boundaries and in small shear zones with orthopyroxene and garnet. The deformation in the pegmatite veins is brittle, whereas it is ductile in the garnet amphibolites.

(ii) West of Ramberg

Tull's (1977) geological map of Vesvtågøy marks two small gabbroic bodies, west of Ramberg, on and near the north-west coast of the island. Fig. 5 of Tull (1977) shows an igneous breccia illustrating the intrusive contact between the mangerite and one of these gabbroic bodies. From this present study, this gabbro contains area of garnet amphibolite, and shows extensive development of ductile shear fabrics: these shear zones are vertical and strike 040°.

WL455 is an altered gabbronorite. Relict ortho- and clinopyroxenes up to 3mm across remain. Orthopyroxene-opaque symplectites are also quite common, but most pyroxenes have been altered to hornblende and biotite, both of which may be up to 5mm across. Garnets, many of which are euhedral, are up to 0.5mm across and are present as incomplete rims to hornblende. Plagioclase is sausseritized to a fine-grained mass of albite and white mica. Apatite is present as an accessory phase. WL456 contains no pyroxenes. All have been altered to hornblende up to 10mm across. A late vein of plagioclase and acicular actinolite (?) cuts the rock. WL474 is foliated; the foliation is defined by alignment of biotite, zoisite and hornblende. Garnet rims are complete around many hornblende pseudomorphs after pryoxene. Glaucophane and rutile are present in this rock and also in WL458, which has no foliation and contains relict pyroxenes. WL459 is a charnockitic pegmatite vein which intrudes the gabbro. It contains quartz and perthitic feldspar grains

up to 10mm across, but they may also be less than 1mm in size, especially along the boundaries of the larger grains. A single clinopyroxene grain is partially altered to hornblende and is rimmed by garnet, as is the only opaque grain present. A single biotite grain, 3mm long, and a number of small zircons are the other phases present.

Plate 5.12 shows a polished slab of WL460. This sample is a quartzo-feldspathic rock with dark green foliae and augen of garnet and dark green minerals. In thin-section the augen are seen to consist largely of pale green clinopyroxene, up to 4mm across, with orthopyroxene exsolution lamellae and rutile inclusions. Orthopyroxene grains are less common. Pyroxenes are partially altered to hornblende and quartz. Plagioclase has been sausseritized and areas of white mica are quite common. Garnets are up to 1.5mm across but crystal faces are rarely well-developed. Biotite, opaques and sphene are also present. In contrast to the garnet amphibolite augen, the quartzo-feldspathic bulk of the rock has a well-developed foliation, defined by biotite plates, less than 1mm long. Most of the quartz and alkali feldspar have grain sizes of less than 1mm, but some quartz, alkali feldspar grains (up to 2mm across) and aggregates of biotite are preserved as augen. Single pyroxene grains altering to hornblende and quartz, up to 5mm across also occur as augen with biotite plates wrapping around them. Accessory sphene and rutile are present in both the quartzo-feldspathic and garnet amphibolite parts of the sample. In WL361 pyroxene augen in quartz and alkali feldspar are present; in this sample the foliation is defined by biotite and muscovite, and also by stringers (up to 20 x 1mm) of quartz grains, up to 2mm across. The presence of glaucophane and abundant rutile in some of the garnet amphibolites implies that they probably formerly had eclogite-facies mineral assemblages.

5.2.4 Langøy

Heier (1960) mapped three large basic intrusions on Langøy. Samples have been collected from each of these intrusions for this study in the following localities (Fig. 5.3):

(i) on the Eidet-Hovden peninsula, north-west Langoy;

(ii) near Bø, south-west Langøy;

(iii) near Grønning, Eidsfjord, south-east Langøy.

(i) Eidet-Hovden

The Eidet-Hovden gabbro occupies the whole of the peninsula north of Eidet in north-west Langøy, except for the Hovden and Engenykken mangerites. It is banded overparts of its outcrop, but this is only visible on a weathered surface. Plate 5.13 shows plagioclase-rich and pyroxene-rich layers, typically 30cm wide, near the quarry used for producing rock for the new breakwater for Hovden habour, at the extreme north of the peninsula. The banding here, and in the quarry, strikes between 120° and 150°, and dips between 42 and 72°S. The plagioclase-rich bands weather white, whereas the pyroxene-rich bands (which also contain abundant olivine) weather dark blue. The banding is cut by two reverse faults with 0.65m throw (one is seen in Plate 5.13). The quarry at Hovden harbour is in Heier's (1960) fine-grained gabbro zone and the Hovden mangerite crops out immediately to the west. In the west wall of the quarry, the gabbro is cut by a quartzo-feldspathic vein, 20cm wide, which strikes 050° and dips 70°N. It has a coarse, granophyric texture visible in hand specimen. Quartz and alkali feldspar crystallized together after green clinopyroxene and may be up to 30mm long. Opaques, sphene and rutile are accessory phases in this rock (WL720).

A good exposure of banded gabbro is present 4km SSW of Hovden on a hillock by the road, a few hundred metres west of the contact with the

Hovden mangerite. The layering strikes 095° and dips 65°N; the layering is continuous along strike, but perpendicular to strike, large areas of gabbro are clearly not compositionally layered. Along the road, to the north, the grain size of the gabbro is reduced closer to the mangerite. Near the outcrop showing banded and homogeneous gabbro, a sample, WL718, was collected from a large loose block. It is a very coarse olivine gabbro. Clinopyroxene, with abundant opaque inclusions, is up to 50mm across, and in thin-section is seen to include olivine and rutile grains up to 2mm across and minor orthopyroxene. Plagioclase (An_{65}) is interstitial to the euhedral clinopyroxenes. A few hundred metres to the south is the small mangeritic pegmatite dyke observed by Heier (1960) and located in Ormaasen (1976). In fact, a number of mangeritic/charnockitic pegmatites occur in the area. They are oblique to the banding of the gabbro. A 4cm wide pegmatite can be traced for 3.5m along a strike of 130°, dipping 32°N. 50m to the north of this pegmatite, a pair of similar dykes 3 and 10cm wide and separated by 60cm, strike 105° and dip 40°S. The 10cm wide dyke was sampled (WL717) and shows that the well-developed foliation (parallel to dyke margins) of these rocks are defined by segregations of quartz and alkali feldspar. Quartz grains up to 2mm across occur as narrow bands or lenses with fine-grained feldspar between. Other phases - clinopyroxene, opaques, orthopyroxene, epidote, sphene and rutile - total less than 5% of the rock.

Heier (1960) stated that banding in the gabbro was best developed in the central part of the gabbro between two lakes east of Nykvåg (illustrated in Heier's Fig. 19). The bands in the gabbro at this locality are only 1-10cm thick (typically 30cm thick at Hovden) and the pyroxenite bands are deeply weathered to a rust-red colour. The banding is vertical and strikes 160°.

WL708 and WL709 are from the western flank of Selvågtind, in the middle of the peninsula, near the contact with the Hovden mangerite. WL708 is a troctolite with olivine grains up to 3mm across overgrown by orthopyroxene. Symplectites of orthopyroxene and opaques which are associated with biotite and green spinel are common. Clinopyroxene is rare, has opaque needles along cleavage traces and is overgrown by orthopyroxene. Reddish-brown biotite occurs in aggregates up to 5mm across. Plagioclase, An_{55}, up to 5mm long, comprises about 50% of the rock. WL709, from about 200m north of WL708, is a peridotite from an east-west oriented pod in the basic rock. Rounded olivines, about 2mm across, altering to serpentine and opaques, make up about 80% of the rock. Minor clinopyroxene is also present. Plagioclase is interstitial. Minor amounts of green spinel and a brown amphibole, probably pargasite also occur.

A road-cut at Sandvik, 1km south-east of the Engenykken mangerite, shows similar layering to that seen in the quarry at Hovden harbour. The layers are between 1 and 30cm thick and strike 140°, dipping 75°E. A thin-section of this rock shows compositional banding. Rounded olivines, about 2mm across, altering to serpentine and opaques are the main phase in the more mafic layers. They are often enclosed by a yellow-brown amphibole that is altered to brown amphibole along cracks radiating from the olivines and caused by their alteration. These cracks also fracture the interstitial plagioclase. Biotite, orthopyroxene and spinel are the other phases present. The more felsic layers are made up chiefly of plagioclase with subordinate olivine.

(ii) Bø

The Bø gabbro occurs on the coast of south-west Langøy around the town of Bø. Heier's (1960) map shows the gabbro as extending for 4km along the coast. WL711 was collected from 1km west of the western contact with the Steine mangerite, but the road-cut from which it was taken contains only very dark, homogeneous gabbro. It has a grain size of 0.5 - 1mm. Plagioclase, some of which is antiperthitic comprises about 60% of the rock. Pale green clinopyroxene is more abundant than strongly pleochroic orthopyroxene and biotite. Opaques and apatite are accessory phases.

(iii) Eidsfjord

The Eidsfjord anorthosite crops out on the south side of the Eidsfjord, which almost makes south-east Langøy into a separate island. The anorthosite has been thrust over retrograded mangeritic rocks to the south-east.

The anorthosite is very coarse in places; near Grønning, Carlsbad twinned plagioclases, up to 20 x 4cm, are present. Albite twinning can be seen in outcrop. Finer-grained pods occur; these are richer in pyroxenes and sulphides. A sample (WL736) was collected from an orthopyroxenite pod 2km east of Grønning. The orthopyroxene is strongly pink-green pleochroic and is up to 20mm long. It is broken by kink bands into pieces 0.2 - 4mm wide. Clinopyroxene occurs in exsolution lamellae and is also recrystallized along the kink fractures. Orthopyroxene is 95% of the rock. Opaques are the next most abundant phase.

5.3 Geochemistry

Eight samples from the Flakstadøy Basic Complex, and seven from other basic intrusions from Vestvågøy and Langøy have been analysed for major element contents in this study (Tables 5.1 and 5.2).

The important points about the major element compositions of the Flakstadøy Basic Complex (Table 5.1) are:

(i) SiO_2 = 45-54%;

(ii) K_2O = 0.5-1.1%;

(iii) $Na_2O/K_2O > 2.5$;

(iv) Fe_2O_3/MgO = 1.0-3.2;

(v) CaO = 6-11%.

The sample from the Napp gabbro, WL317, is markedly more basic than the others: low SiO_2, Al_2O_3, high Fe_2O_3, TiO_2, P_2O_5. The calculated CIPW norms are in good agreement with the modal analyses; e.g. 90% normative plagioclase, minor orthopyroxene, but no olivine or clinopyroxene for the anorthosite sample, WL329; abundant clinopyroxene and opaques, and 6.6% normative orthoclase, reflected in the mode of the Napp gabbro, WL317; the presence of ortho- and clinopyroxene, apatite and orthoclase, and the absence of olivine in the sample near the contact with the mangerite at Nesland, WL346.

Variation diagrams for the eight Flakstadøy Basic Complex samples (Figs. 5.4 - 5.7) show that three samples consistently plot away from the other five. On a K_2O v. SiO_2 plot (Fig. 5.4) the Napp gabbro sample, WL317, has low SiO_2 and high K_2O; the Mørkdalen 'eclogite', WL323, has lower SiO_2 than the other samples; and WL346, from adjacent to the mangerite at Nesland, has high K_2O. The mean values for all Archaean and Proterozoic gneisses (Tables 2.1 and 2.2, after Griffin et al., 1978) and for the Flak-

stadøy Mangerite (Table 2.5, after Ormaasen, 1976) are also plotted. WL346 plots away from the main groups of points towards these possible contaminants. On a CaO v. MgO plot (Fig. 5.5), WL346 plots towards the mean values for the Flakstadøy Mangerite and for the Archaean gneisses. WL317 and WL323 have higher MgO than the group of five basic rocks, WL311, WL314, WL325, WL326 and WL329. These five samples have a wide spread parallel to the MgO axis reflecting an increase in the proportion of mafic phases present. P_2O_5 v. TiO_2 (Fig. 5.6) shows that the five typical Flakstadøy Basic Complex samples are very low in incompatible elements. The Mørkdalen 'eclogite', WL323, has high TiO_2, but low P_2O_5, presumably due to originally being rich in ilmenite which has subsequently changed to rutile. WL347 plots near to the mean value for the Flakstadøy Mangerite but has higher TiO_2. Romey (1971) marks a number of magnetite-ilmenite pods near Nesland and this could explain the high TiO_2. The Napp gabbro is rich in disseminated ore which would explain the very high TiO_2 content (2.4%) of WL317. TiO_2 v. CaO (Fig. 5.7) again shows that WL317, WL323 and WL346 have high TiO_2. The low CaO of WL346 can be modelled as due to contamination from either the Flakstadøy Mangerite or the late Archaean gneisses. The proximity of this sample to the Flakstadøy Mangerite and the presence of antiperthites suggests that incorporation of mangerites is the more likely explanation. The Napp gabbro has been seen to contain abundant gneiss xenoliths and to be cross-cut by mangerite/charnockite veins and the geochemistry of WL317 is best explained as a high-Ti gabbro contaminated by Proterozoic gneisses.

To obtain an average set of values for the major elements of the Flakstadøy Basic Complex the analyses for WL311, WL314, WL325, WL326 and WL329 have been used. These have a wide geographic range and variation in primary mafic phases. WL317 and WL346 are regarded as atypical due to their contamination, as is WL323 due to its deformation and change in mineralogy.

The average thus obtained is:

Oxide	Per Cent Abundance
SiO_2	51.5
Al_2O_3	24.3
TiO_2	0.2
Fe_2O_3	4.9
MgO	3.9
CaO	9.7
Na_2O	4.2
K_2O	0.6
MnO	0.1
P_2O_5	0.0
LOI	0.5

These values are used in Chapter 6 in modelling the geochemistry of the mangerites.

The important points from the major element compositions of the other basic intrusions from Vestvågøy and Langøy (Table 5.2) are:

(i) SiO_2 = 43-52%;

(ii) K_2O = 0.3-1.2%;

(iii) $Na_2O/K_2O > 2.5$;

(iv) Fe_2O_3/MgO = 0.5-2.3;

(v) CaO = 6-17%

(vi) the gabbro-eclogite transition is isochemical.

The gabbro-eclogite transition is clearly seen in the road-cut at Våje, west Vestvågøy. Plate 5.11 is a photograph of a sample, WL442, which shows this transition. The central 6cm was removed and the remaining

'gabbroic' and 'eclogitic' parts crushed to form WL442'G' and WL442'E' respectively. These two samples have no significant differences in chemistry (Table 5.2), showing that the transition is not controlled by chemistry. WL716, the troctolite sample from Sandvik, is the most basic sample analysed from the Lofoten-Vesterålen province; this suggests that the sample may be of a mafic-rich layer. There is good agreement between the calculated CIPW norms and the model analyses, e.g. the abundant olivine, presence of ortho- but not clinopyroxene and the abundant volatiles expressed as calcite in the norms and as brown amphibole in WL716; the very abundant clinopyroxene of WL718 and the presence of apatite in the Bø gabbro sample, WL711.

The chemistry of the samples from these smaller gabbroic bodies is more similar to the 'contaminated' and 'deformed' samples from Flakstadøy than to the samples regarded as typical. This would suggest that the smaller gabbroic bodies have assimilated larger amounts of mangeritic material than the larger Flakstadøy Basic Complex, due to their greater surface area to volume ratio.

The presence of mangeritic and charnockitic veins in the gabbros, implies that the mangerites post-date the basic rocks. Nonetheless, evidence presented in Chapters 6 and 7 suggests that the mangerites were emplaced very shortly after the gabbros and that the mangerites have actually formed by country-rock melting caused by emplacement of the basic intrusions. At the margins of the basic magmas, mixing with the monzonite melts could have taken place.

5.4 Isotopic Results

5.4.1 Sr Isotopes

Twelve samples from the Flakstadøy Basic Complex have been analysed for Sr isotope compositions. The results are tabulated in Fig. 5.8. The range in Rb content is 2 to 12ppm and in Sr content is 521 to 1234 ppm. The Rb and Sr values produce a range of $^{87}Rb/^{86}Sr$ ratio of 0.003 - 0.0518. The $^{87}Sr/^{86}Sr$ ratios range from 0.70405 - 0.70549. These data yield an isochron of 1764 ± 155 Ma (2σ), MSWD = 0.89, and an initial $^{87}Sr/^{86}Sr$ ratio of 0.70406 ± 0.00005.

The Rb concentration determinations for the Flakstadøy Basic Complex are at the detection limit for the XRF, hence it has not been possible to obtain a well-defined age from this intrusion. A better constrained age might be produced using isotope dilution methods or the Sm-Nd system.

There are no significant differences between the Rb and Sr concentrations and $^{87}Sr/^{86}Sr$ ratios for the gabbroic and eclogite parts of WL442.

Data from the other gabbros of the Lofoten-Vesterålen province are presented in Table 5.3. The range in Rb and Sr concentrations is 2 to 49ppm, and 158 to 880ppm, respectively. Rb and Sr concentrations were determined on nine samples and $^{87}Sr/^{86}Sr$ on three samples.

5.4.2 Nd Isotopes

Seven samples from the Flakstadøy Basic Complex have been analysed for Nd isotopes in this study (Table 5.4; Fig. 5.9). Sm and Nd concentrations are in the range 0.2 to 7.6ppm and 1.9 to 37ppm, respectively. $^{147}Sm/^{144}Nd$ ranges between 0.086 and 0.157, and the range for $^{143}Nd/^{144}Nd$ ratios is 0.51106 to 0.51190. These data yield an isochron age of 1803 ± 112 Ma (2σ), (MSWD = 0.60) and an initial $^{143}Nd/^{144}Nd$ ratioof 0.51106 ± 0.00009.

The Sm and Nd concentrations for the five samples analysed from the other Lofoten-Vesterålen gabbros are 0.6 to 7.0ppm and 3 to 36ppm, respectively (Table 5.4). The differences in the concentrations and isotope ratios for the gabbroic and eclogitic parts of WL442 are insignificant.

The wide range in Sm/Nd ratios for the Flakstadøy Basic Complex enables its age to be fairly well determined by a whole-rock isochron method. This range of values arises because this intrusion consists of cumulates. The troctolite samples have the lowest REE concentrations as olivine contains very little REE, and they have the lowest Sm/Nd ratios (with the anorthosites) since plagioclase tends to be LREE-enriched.

The narrow range of individual ε_{Nd} values for the Flakstadøy Basic Complex at 1.8 Ga (-4.3 to -5.1) (Fig. 5.10) is a consequence of the homogeneity of $^{143}Nd/^{144}Nd$ through the intrusion at that time, further evidenced by the good quality of fit of the isochron.

At 1.8Ga, ε_{Nd} in the depleted mantle was +3.9 (DePaolo, 1981), hence the Flakstadøy Basic Complex cannot have formed by simple extraction of magma from depleted (or undepleted) mantle at 1.8Ga. ε_{Nd} at 1.8Ga for the late Archaean gneisses of Lofoten-Vesterålen is typically -12 (Jacobsen and Wasserburg, 1978; and the present study), so assimilation of late Archaean crust, during the ascent and differentiation of magmas derived from a depleted mangle source, could account for the measured $\varepsilon_{Nd}(1.8)$ of the Flakstadøy Basic Complex.

ε_{Nd} for the other Lofoten-Vesterålen gabbros at 1.8Ga is in the range -2.2 to -3.2 (Fig. 5.11). As for the Flakstadøy Basic Complex, this may be indicative of assimilation of late Archaean crust by depleted

mantle-derived gabbroic magmas. ε_{Nd} (1.8) for the Flakstadøy Basic Complex is, on average, two ε-units lower than the other gabbros, from Vestvågøy and Langøy, perhaps because the much larger, differentiated magma body has assimilated more ancient crustal material. Other possible explanations are that the amounts of crustal material assimilated by the magmas were the same but that either:

(i) the crust assimilated by the Flakstadøy Basic Complex had a greater concentration of Nd

or (ii) the depleted mantle-derived Flakstadøy Basic Complex magma had a lower Nd concentration than the other gabbroic magmas.

Alternatively the two magma suites may have been contaminated by crustal material of different ages; the Flakstadøy Basic Complex assimilating late Archaean crust, whereas contamination with mid-Proterozoic supracrustal material may have been more extensive in the other gabbroic magmas.

5.4.3 Pb Isotopes

The full range of Pb isotope ratios for the nine samples analysed from the Flakstadøy Basic Complex presented in Table 5.5 is: 14.794 to 15.600 for $^{206}Pb/^{204}Pb$; 14.958 to 15.139 for $^{207}Pb/^{204}Pb$; and 34.678 - 35.772 for $^{208}Pb/^{204}Pb$. The nine data-points are aligned with a slope corresponding to an apparent age of about 2.6 Ga. (Fig. 5.12) This clearly has no direct age significance since both Nd and Sr isochron diagrams (Figs. 5.8 and 5.9) imply an age of emplacement of 1.8 Ga. Also the gabbro at Napp contains xenoliths of the supracrustal gneisses which have a maximum age of 2.1Ga. A model has been constructed to account for the Pb isotopic characteristics of the Flakstadøy Basic Complex in terms of contamination of the basic magmas by older continental crust (for similar models see

Moorbath and Welke (1969), Dickin (1981) and Taylor et al. (1980, 1984).

For the model, a mantle μ_1 value ($^{238}U/^{204}Pb$) of 7.9 has been chosen. This is a typical value for the North Atlantic craton (Moorbath and Taylor, 1981) and is in the range (7.7 to 8.2) predicted for the apparent μ_1 of the mantle source at 2.7Ga by the Plumbotectonics model of Zartman and Doe (1981). The Pb/Pb isochron for the North-West Hinnøy gneisses and the transposed palaeo-isochron for the Vikan gneisses both give $\mu_1 = 7.85$. A slight increase of mantle μ_1 with time is predicted by the models of Oversby (1974) and Zartman and Doe (1981), supporting the use of $\mu_1 = 7.9$ for the mantle source at 1.8Ga.

The Nd isochron in Fig 5.9 shows that the age of the intrusion of the Flakstadøy Basic Complex is 1.8Ga, but the Nd evolution diagram (Fig. 5.10) shows that these basic rocks have been contaminated by considerably older continental rocks. The $^{207}Pb/^{204}Pb$ v. $^{206}Pb/^{204}Pb$ diagram (Fig. 5.12) shows that the samples with the least radiogenic Pb plot very close to the 1.8Ga first-stage isochron. Any finite μ value for these rocks at present day would imply that at 1.8Ga these data-points plotted to the left of this isochron and hence a single-stage evolution for these rocks up to 1.8Ga (i.e. in the mantle) is not possible. Geochemistry has shown that some of the basic rocks on Flakstadøy (including WL317 from the Napp gabbro) have been contaminated more than others. In the field, xenoliths of early Proterozoic supracrustal rocks have been observed in the Napp gabbro. Griffin et al. (1978) observed that the mangerites tend to occur in the early Proterozoic gneisses or at their interface with the late Archaean migmatites. This is also true of the basic intrusions now at surface in Lofoton-Vesterålen, but geophysical evidence suggests that the region may be underlain by large volumes of basic rock (see 2.2). This implies that the rocks that the mantle-derived basic magmas have

passed through are predominantly late Archaean migmatites, 2.7 Ga old (see section 2.3), and hence contamination of these magmas with ancient crustal material would have taken place by assimilation of these migmatites. In the model in Fig. 5.12 basic magmas are derived from a mantle with μ = 7.9 at 1.8Ga and are contaminated with different amounts of 2.7Ga crustal material. Two parallel 1.8Ga isochrons are shown in Fig. 5.12, the upper isochron is for uncontaminated rocks derived from mantle with μ_1= 7.9 at 1.8Ga, the lower isochron would be produced by melting of the mean Pb isotope composition for the 2.7Ga north-west Hinnøy amphibolite-facies gneisses at 1.8Ga (Griffin et al., 1978). Still lower 1.8Ga isochrons would be produced by melting north-west Hinnøy amphibolite-facies gneisses with particularly low μ_2 values. All nine of the Pb data-points for the Flakstadøy Basic Complex plot between these two theoretical isochrons. This distance from each isochron is an indication of relative crust-mantle contributions, e.g. the Napp gabbro sample, WL317, has the largest crustal component. This is in accord with field observations and geochemical data. The linear array of Pb data-points can be explained using this model. The two end members in the model are mantle with μ = 7.9 and 2.7Ga migmatites with low μ_2 (unlike the gneisses on north-west Hinnøy, the 2.7Ga gneisses beneath Flakstadøy are in granulite facies and hence are probably severely depleted in U, having even lower μ_2 values than the 2.7Ga gneisses used in the model). Those samples plotting closest to the 1.8Gaisochron for mantle-derived rocks have a greater proportion of the relatively high μ mantle component, and less of the low μ crustal component, and hence they have high μ values and evolve to Pb isotope compositions far removed to the right of the 2.7Ga palaeo-isochron at 1.8Ga. Those samples plotting nearer to the 1.8Ga "mean crustal isochron" have a greater proportion of the low μ crustal component and hence evolve to less radiogenic compositions, little removed from the 2.7Ga palaeo-isochron at 1.8Ga.

The sources of the Pb present in a rock today can be modelled using the Plumbotectonics model of Zartman and Doe (1981). Apparent μ_1 values for the Flakstadøy Basic Complex (Table 5.5) are plotted on a μ_1 v. T diagram showing the evolution of apparent μ_1 with time in "orogene", upper crust, lower crust and upper mantle (Fig. 5.13). This model confirms that the Flakstadøy basic rocks were derived from an upper mantle source and contaminated with different amounts of material from a lower crustal source. The Napp gabbro sample, WL317, has the greatest lower crustal component in this model. WL326, from the troctolite zone north of Nusfjord (Plate 5.4) has the greatest upper mantle component - in fact, according to this model it crystallized from pristine mantle-derived magma, without any crustal contamination.

WL708, a troctolite from Hovden, Langøy, has the most radiogenic Pb composition analysed from Lofoten-Vesterålen (Table 5.5). It, too, can be modelled as a mantle-derived magma contaminated by ancient crustal rocks.

5.5 Discussion and Synthesis

5.5.1 Isotope Geochemistry

The Nd evolution diagrams for the basic rocks (Figs. 5.10 and 5.11) show that they have all been contaminated by an ancient lower continental crustal component. An estimate of the amount of crustal material assimilated by the mantle-derived magmas can be made in the same way as Patchett and Bridgwater (1984) estimated the percentage of an Archaean crustal component in rocks from the Ketilidian terrain of South Greenland. The two-component mixing equation of Pushkar et al. (1972) describing the addition of x parts of crustal material to one part of mantle-derived component reduces to:

$$x = \frac{C_{Nd}^{MD} \ (\varepsilon_{Nd}^{NC} - \varepsilon_{Nd}^{MD})}{C_{Nd}^{cc} \ (\varepsilon_{Nd}^{cc} - \varepsilon_{Nd}^{NC})} \qquad (5.1)$$

The mantle-derived material is denoted MD, cc refers to the pre-existing Archaean crust, NC to the new, 1.8Ga crust and C is the concentration. ε_{Nd}^{MD} = +4 (the ε_{Nd} of the depleted mantle at 1.8Ga in DePaolo's (1981) model). The ε_{Nd} values at 1.8Ga for the late Archaean gneisses of Lofoten-Vesterålen (Jacobsen and Wasserburg, 1978; the present study) are about -12. A larger body of data exists for similar late Archaean gneisses from the Lewisian of North-West Scotland (Hamilton et al., 1979) which suggests that a value for ε_{Na}^{cc} of -10 is appropriate. The average ε_{Nd} at 1.8Ga for the five samples used to produce the average major-element composition for the Flakstadøy Basic Complex (WL311, WL314, WL325, WL326 and WL329) is -4.6. Hence equation 5.1 reduces to:

$$x = \frac{C_{Nd}^{MD}}{C_{Nd}^{cc}} \qquad (1.6)$$

so the proportion of Archaean crust is largely dependent on the Nd concentrations in the mantle and crustal components. Table 5.6 shows the proportion of late Archaean crustal component implied for different Nd concentrations in the mantle- and crust-derived components. The average Nd concentration for the five basic samples listed above is 3.6ppm, but this is for cumulus phases which have very low concentrations of rare earth elements. The Nd concentration in the magma before fractionation would have been higher than this. Cohen et al. (1980) analysed eleven mid-ocean ridge basalt glasses, the average Nd concentration of which is 11.5ppm; the lowest value is 6ppm. The mean Nd concentration of late Archaean gneisses from the Lofoten-Vesterålen province (Green et al., 1972; Jacobsen and Wasserburg, 1978; the present study) and from the Lewisian province of North-West Scotland (Hamilton et al., 1979) is 32ppm. Contamination of the magmas

may have taken place by assimilation of mangerite/charnockite melts derived from the gneisses rather than by direct assimilation of the gneisses themselves. The Nd concentration in the melts ranges between 9 and 204 ppm (Chapter 7). Using these ranges of Nd concentrations the crustal contribution to the basic rocks may be anywhere between 5 and 67%. The most plausible scenario is for contamination of mean mid-ocean ridge basalt (11.5ppm Nd) with crustal material somewhat enriched in Nd (c. 50ppm) suggesting an average crustal contribution to the basic magma of about 20%. If the crustal contamination took place when the basic magma had fractionated, causing increase of the Nd concentration in the residual magma then the crustal contribution required to account for the Nd isotope data would be much larger.

The average ε_{Nd} at 1.8Ga for the other basic rocks analysed (WL442'G', WL442'E', WL465, WL711, WL716) is -2.9 and their average Nd content is 16.6ppm. This implies a smaller proportion of crustal contaminant than for the Flakstadøy Basic Complex e.g. for 11.5 ppm Nd in the mantle-derived component and 32ppm in the crust-derived component, 26% crustal material is required for the basic rocks of Vestvågøy and Langøy, compared to 37% for the Flakstadøy Basic Complex.

The major element compositions for the small gabbros from Vestvågøy and Langøy are similar to the compositions of the samples collected from Napp (WL317) and Nesland (WL346), and from the Flakstadøy Basic Complex adjacent to the mangerite. All these have markedly different compositions from samples collected away from the margins of the complex. Variation diagrams have shown that this could be due to incorporation of mangeritic material, restricted to the margin of the large intrusion on Flakstadøy, but pervasive in the smaller intrusions due to their larger surface area to volume ratio. Modelling from Nd evolution diagrams suggests that the

smaller gabbroic bodies contain a slightly lower proportion of ancient crustal contaminant than the Flakstadøy Basic Complex. Contamination with ancient crustal material produces a lower ε_{Nd} at 1.8Ga (since $\varepsilon_{Nd}^{cc} = -12$) whereas contamination with contemporaneous mangeritic material has little effect on the Nd isotopic composition (since ε_{Nd} at 1.8Ga for mangerites averages -5 (Chapter 6), similar to that of the basic rocks). Contamination with mangeritic material could, however, be reflected in the whole rock chemistry e.g. mangerites have much higher Fe_2O_3/MgO ratios and lower Al_2O_3 contents than the basic rocks. As predicted, the small gabbros, and WL317 and WL346 have higher Fe_2O_3/MgO ratios and lower Al_2O_3 contents than the other basic samples from Flakstadøy.

Crustal involvement in the genesis of the Flakstadøy Basic Complex can also be constrained using Pb isotopes. The $^{207}Pb/^{204}Pb$ v. $^{206}Pb/^{204}Pb$ diagram for the Flakstadøy Basic Complex (Fig. 5.12) shows that all of the Pb data lie between the 1.8Ga model isochrons produced by derivation from mantle with $\mu_1 = 7.9$ and by melting of mean North-West Hinnøy amphibolite-facies gneisses. Those data which plot closer to the 1.8Ga isochron for mantle-derived rocks have a smaller fraction of crustal Pb than those that plot close to the "mean crustal isochron". The percentage of crustal Pb in each sample can be calculated using the Lever Rule; this value is presented for nine samples in Table 5.7.

Assuming that the Flakstadøy Basic Complex is derived entirely from the mantle and from melting of 2.7Ga gneisses and provided that suitable assumptions about Pb concentrations can be made, then a similar model to that used for the Qôrqut Granite of Southern West Greenland (Moorbath et al., 1981) can be used for the Flakstadøy Basic Complex. The fraction, f, of the basic complex contributed by the late Archaean migmatitic gneisses, can be calculated:

$$f = \frac{C_M (a_M - a_B)}{C_c (a_B - a_C) + C_M (a_M - a_B)}$$

where a refers to $^{206}Pb/^{204}Pb$ ratios at 1.8Ga, and C refers to ^{204}Pb concentrations. The subscripts M,C,B, refer to mantle-derived, crust-derived and basic magma respectively. The percentage crustal contribution for each of the nine samples is tabulated in Table 5.7, using the mean concentration (total Pb = 20.2 ppm, ^{204}Pb = 0.28 ppm) from the North-West Hinnøy amphibolite-facies gneisses as the concentration for the crustal component. The crustal contribution is calculated for three Pb concentrations in the mantle-derived magma. 0.6ppm is the mean value for pristine mid-ocean ridge basalt (Cohen et al., 1980); 3.2ppm is the total Pb concentration for the Preshal Mhor basalt from the Isle of Skye with the highest calculated mantle-derived Pb component (93%) (Dickin, 1981); and 7.0ppm is from a basalt from the Southern Zone of the Andes (McHugh, 1980). Table 5.7 shows that a large proportion of the Pb in the Flakstadøy Basic Complex is derived from 2.7Ga lower crust (average 63%). Because the Pb concentration in the lower crust is much higher than that in the basaltic magma, the crustal contribution by mass is less than the fraction of the Pb derived from the crust. The greater the difference in concentrations between the basaltic magma (low Pb) and the crustal contaminant (high Pb), the smaller the proportion of crustal material required to account for the level of crustal Pb in the Flakstadøy Basic Complex. Taking 3.2ppm Pb as a suitable Pb concentration (from the sub-continental mantle-derived Preshal Mhor basalt) for a basaltic magma (and excluding WL317, which on geochemical grounds has been contaminated by mangeritic material post-emplacement) the mean crustal contribution is 22%. This is similar to the fraction of crustal material in the basic complex calculated from Nd isotopes. The very high (99%) fraction of

crustal Pb in the Napp gabbro sample, WL317, which on evidence from field observations and geochemistry has undergone late contamination by a mangerite magma (itself derived from 2.7Ga lower crust, Chapter 6), suggests that the mangerite magmas may have higher Pb concentrations than the 2.7Ga gneisses themselves. Hence, if crustal contamination occurred via mangerite melts then the material crustal contribution would be less than 22%.

5.5.2 Evolution of P-T Conditions

The recognition in the basic rocks of primary magmatic mineral assemblages, corona development, the attainment of eclogite-facies assemblages in shear zones, and subsequent retrogression to amphibolite-facies, enables a qualitative P,T,t path to be constructed (Fig. 5.14).

The Flakstadøy Basic Complex is a coarsely-crystalline, cumulate body which has crystallized from a basaltic magma. The coexistence of olivine and plagioclase limits the pressure of crystallization to < 8kb (Austrheim and Griffin, 1982) at c.1150°C (Green, 1969). The development of orthopyroxene-clinopyroxene-garnet coronas between olivine and plagioclase (Griffin and Heier, 1969) (Plates 5.5 and 5.6) occurs during either isothermal burial or isobaric cooling or some intermediate condition involving cooling during further burial; the limited extent of the reaction suggests termination by falling temperatures, so isobaric cooling is favoured. Griffin et al. (1978) state that these coronas would be expected to form during cooling to c.770°C at 9-10kb. Austrheim and Griffin (1982) infer cooling with slight pressure increase for the development of similar coronas in granulite-facies anorthosites of the Bergen Arcs of Western Norway. Continued cooling after the termination of corona formation is necessary to attain eclogite-facies conditions. Table 2.4, after Griffin et al. (1978) shows P-T estimates of 625-675°C, 8.5 - 11kb for South Hinnøy and

Langøy (Krogh, 1977; Olsen, 1978). Fig. 2.3, after Griffin et al. (1978) shows that these conditions are on the border between granulite and eclogite facies; they are also on the limit of the glaucophane stability field. Such conditions must have been attained by the Flakstadøy Basic Complex to account for the presence of glaucophane and relict eclogite-facies mineral assemblages and textures. The observed deformation must have started in the P-T stability field of the eclogite-facies paragenesis, but in the absence of the deformation, the granulite-facies paragenesis persists metastably. Austrheim and Griffin (1982) recognised a similar relationship between deformation and the development of eclogite-facies mineral assemblages in the anorthosites of the Bergen Arcs. They attributed the triggering of eclogite-forming reactions in the shear zones to:

(i) grain size reduction increasing the surface area over which reaction can occur, and to the mixing of the minerals present on a finer-scale than in the coarse-grained granulites;

(ii) surface area increase facilitates grain-boundary diffusion, which is much faster than volume diffusion;

(iii) an increase in pathways along which a fluid phase can move, promoting reaction and diffusion;

(iv) an increase in the density of crystal lattice defects, which promotes volume diffusion.

Austrheim and Griffin (1982) did not regard "tectonic overpressure" in these narrow shear zones as an important factor in the formation of eclogites.

Reactions from Austrheim and Griffin (1982) (5.2 - 5.4) and from the present study(5.5), for the operation of which there is good evidence in

the Lofoten-Vesterålen eclogites include:

albite → jadeite (in cpx) + quartz (5.2)

albite + H_2O → paragonite (in phengite) (5.3)

Fe,Ti-oxides + plagioclase + eustatite (from cpx)
→ garnet + omphacite + rutile (5.4)

$$15\ CaMgSi_2O_6 + 40NaAlSi_3O_8 + 9CaMg_5Al_4Si_6O_{24} + 26H_2O \longrightarrow 20Na_2Mg_3Al_2Si_8O_{22}(OH)_2 + 12Ca_3Al_3Si_3O_{22}(OH) + 8SiO_2$$

$$\text{diopside} + \text{albite} + \text{garnet} \xrightarrow{(H_2O)} \text{glaucophane} + \text{epidote} + \text{quartz} \qquad (5.5)$$

Reaction 5.2 is inferred as having operated from the present occurrence of albite and diopside aggregates which suggests that Reaction 5.2 was reversed causing the breakdown of the jadeite component of omphacite during retrogression. Reaction 5.3 would explain the very fine-grained areas of white mica in altered plagioclases. Reaction 5.4 would explain the progressive increase of rutile (which has the lowest molar volume of the Ti minerals) at the expense of opaque phases towards the core of the shear zones. Reaction 5.5 would explain the presence of glaucophane, associated with quartz, within spongy garnets.

An interesting feature of the eclogites of Lofoten-Vesterålen is that although their formation has clearly been triggered by shearing, those which have developed full eclogite parageneses do not have a deformational fabric preserved. This is very clear in Mørkdalen, Flakstadøy where the sheared gabbros (Plate 5.7) retain a strong foliation whereas the eclogitic rocks from the core of the shear zone (e.g. WL361) are not foliated.

WL442 (Plate 5.11) from Våje, Vestvågøy shows gabbro-eclogite transition (which is isochemical) apparently not associated with shearing. A possible explanation for this could be that as eclogite-forming reactions take place in response to deformation, the development of high-pressure phases may inhibit further shearing in that zone and the shear may propagate elsewhere. This could explain why there are two, sub-parallel shear zones in Mørkdalen, and alternating c.10cm wide gabbroic and eclogitic bands at Våje.

The garnet amphibolite augen present within sheared quartzo-feldspathic rock in the small gabbro, west of Ramberg, Vestvågøy, (Plate 5.12) is interpreted as an igneous breccia of gabbro in charnockite (see Fig. 5, Tull, 1977), in which deformation has caused the development of eclogitic assemblages (subsequently amphibolitized in the gabbroic blocks).

The field relationships, mineral assemblages, and suggested explanations for the development of eclogites in sheared basic rocks, are similar for those in Lofoten-Vesterålen and those in the Bergen Arcs described by Austrheim and Griffin (1982). The shearing in the Bergen Arcs has been dated as Caledonian from Nd analyses of mineral separates from the eclogites. The shearing is unlikely to be of a similar age in Lofoten-Vesterålen. If it were Caledonian, the shearing might be expected to be related to the low-angle, westerly-dipping, reverse fault placed between the Rombak Window and the Tysfjord-Lofoten-Vesterålen region in the model proposed by Hodges et al. (1982) (see Chapter 2.8) but the shear zones at Mørkdalen, Vik-Våje and Ramberg dip steeply to the east or are vertical. Also similar shears have not been seen in younger units in Lofoten-Vesterålen. It is more likely that the shearing is related to the Sveco-Karelian orogeny of northern Sweden dated at 1.9 - 1.75Ga (Witschard, 1984), but without isotopic dating of the eclogites, a Caledonian origin cannot be completely

ruled out.

Subsequent to the formation of the eclogites, the mineral assemblages have been retrograded to amphibolite facies. This would have involved isothermal uplift from c. 10kb. The foliation at Våje, at 30° to the gabbro and eclogite bands, must be associated with this uplift.

5.6 Conclusions

1. The Flakstadøy Basic Complex is the largest basic igneous body exposed in Lofoten-Vesterålen. It has cumulate textures and magmatic sedimentation structures. It is the only basic intrusion in Lofoten-Vesterålen to show strong compositional variations on a large scale (the Hovden gabbro has alternate mafic-rich, mafic-poor layers).

2. The basic intrusions are seen to cross-cut the basement gneisses and to contain gneiss xenoliths at Napp, Flakstadøy.

3. The basic intrusions pre-date the mangerites. This relationship can be seen in the gabbro west of Ramberg (Tull, 1977; the present study), in the Borge gabbro (the present study) and in the Hovden gabbro (Heier, 1960; the present study). No conclusive field evidence for the relative age of the Flakstadøy Basic Complex and the South-west Lofoten Mangerite has been produced.

4. Geochemistry has shown that the more marginal samples from the Flakstadøy Basic Complex, adjacent to the Flakstadøy Mangerite (part of the South-west Lofoten Mangerite) are more contaminated than samples from high levels. Field relations and variation diagrams suggest that the most likely source of this 'excess' contamination is the adjacent mangerite. The smaller gabbros, from Vestvågøy and Langøy, have similar major element compositions to the more contaminated, marginal Flakstadøy Basic Complex samples. This suggests that they

are more contaminated with mangeritic material due to their greater surface area to volume ratio.

5. It is not possible to produce a good whole-rock Rb-Sr isochron for the Flakstadøy Basic Complex due to its low Rb concentrations. The wide range in Sm/Nd ratios (which arises because the rocks are cumulates) enables the production of a Sm-Nd isochron with a slope corresponding to an age of emplacement of 1.8Ga. This range in Sm/Nd is not present in the smaller, non-cumulate gabbros.

6. The Nd evolution diagrams for the Flakstadøy Basic Complex and for the basic rocks from Vestvågøy and Langøy show that they must all have assimilated large amounts of ancient continental crust. They also show that $^{143}Nd/^{144}Nd$ was homogeneous at 1.8Ga for the Flakstadøy Basic Complex (this follows from the production of an isochron of that age) and hence the model ages have no geological age significance.

7. Modelling $^{207}Pb/^{204}Pb$ v. $^{206}Pb/^{204}Pb$ compositions for the Flakstadøy Basic Complex suggests that it formed by extraction of a basaltic magma from typical upper mantle at 1.8Ga and was contaminated by varying amounts of ancient crustal material from the 2.7Ga migmatites. The "Plumbotectonics" model supports this conclusion.

8. Mixing calculations suggest that the small gabbros on Vestvågøy and Langøy have probably been contaminated with less material derived from the Late Archaean migmatites than has the Flakstadøy Basic Complex, but the small gabbros and the margins of the Flakstadøy Basic Complex have been contaminated by mangeritic material.

9. The inferred, P,T-path for the basic rocks involves crystallization at $P < 8kb$ at c.1150°C; cooling with slight pressure increase, due

to increased overburden, to the boundary between the granulite and eclogite P,T stability fields; deformation causes parts of rocks to develop eclogite-facies mineral assemblages; isothermal uplift causes retrogression to amphibolite-facies parageneses.

CHAPTER 6. THE MID-PROTEROZOIC MANGERITIC INTRUSIONS

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CHAPTER 6. THE MID-PROTEROZOIC MANGERITIC INTRUSIONS 1800-1700Ma

6.1 Introduction

In the Lofoten-Vesterålen province, during the mid-Proterozoic a major suite of mangeritic intrusions was emplaced under granulite-facies conditions, into a basement consisting of late Archaean migmatitic gneisses and early Proterozoic supracrustal gneisses. Heier (1960) and Tull (1977) found evidence which suggested that the mangerites post-date the mid-Proterozoic basic intrusions, although Romey (1971) concluded that the Flakstadøy Basic Complex was intruded into the mangerite on that island.

Most of the mangeritic intrusions are composed of proper mangerites or their retrograded equivalents, but some of the intrusions of this suite include the more acid rock-type, charnockite. (Mangerites are monzonitic in composition, charnockites granitic:both include essential orthopyroxene - they are deep crustal intrusive rocks. Mangerites take their name from Manger ,in the Bergen Arcs of Western Norway, where the rock-type was first defined by Kolderup (1903).). At the present level of erosion, these mangeritic intrusions occupy almost half of the Lofoten-Vesterålen province, and hence their emplacement is the most important rock-forming event recognized in the region. Identification of their mode of origin is, therefore, essential for an understanding of the evolution of the crust in the region.

On the basis of low initial $^{87}Sr/^{86}Sr$ ratios Griffin et al. (1978) concluded that the mangeritic rocks could not have been derived by crustal anatexis. However, on the basis of Pb isotopes and variations in initial $^{87}Sr/^{86}Sr$ ratios for different intrusions, they also concluded that some component of crustal material may have been involved in the petrogenesis

of certain of the mangerites.

The rôle of the contemporaneous basic intrusions is also important. Green et al. (1972) ruled out a comagmatic origin on the basis of rare-earth element distributions. If crustal material was involved in the genesis of the mangeritic rocks, the high intrusion temperature of the basic plutons may have caused the required anatexis.

6.2 Fieldwork and Petrography

Previous research on the mangeritic intrusions of the Lofoten-Vesterålen province has been reviewed in Chapter 2. This section describes the samples that have been used in the geochemical and isotopic studies, and the localities from which they have been collected.

All of the major mangeritic intrusions in the Lofoten-Vesterålen province have been examined in this study:

(i) The Sund-Ølkona Mangerite, on Moskenesøy and Flakstadøy;

(ii) The Flakstadøy Charnockite;

(iii) The South-West Lofoten Mangerite, on Flakstadøy, Vestvågøy and Gimsøy;

(iv) The Hopen Mangerite/Charnockite, south-west Austvågøy;

(v) The Austvågøy Grey Mangerite;

(vi) The Raftsund Mangerite, on Austvågøy, Hadseløy, Hamarøy and Hinnøy;

(vii) The mangerites at Hovden, Engenykken, Steine, Søberg and Eidsfjord on Langøy.

6.2.1 The Sund-Ølkona Mangerite

This is the westernmost mangeritic intrusion in the Lofoten-Vesterålen province. It occupies an area of about $1km^2$ at Ølkona, the easternmost point of Moskenesøy (Fig. 3.1) and about $4km^2$ around Sund in South-west Flakstadøy (Fig. 5.2). The mangerite at Ølkona is emplaced into late Archaean migmatites, which are strongly sheared. Ultramylonite zones are present in the migmatites and the Ølkona mangerite (Plate 6.1). At the southern contact between the mangerite and the migmatites, the two lithologies interfinger over at least 50m. The Ølkona Mangerite is dark brown or green, coarse-grained (typically 5mm) and has a granular texture. Mesoperthite is the dominant mineral and can constitute over 90% of the rock. Other important phases include clinopyroxene, orthopyroxene, opaques and antiperthitic plagioclase. Apatite is present in accessory amounts. Garnet occurs in coronas to opaques and clinopyroxene, sometimes with an inner rim of quartz around clinopyroxene. Hydrous alteration of pyroxenes has produced hornblende rims. No primary quartz was observed in any mangerite from Ølkona. Pods of coarse-grained pegmatitic mangerite occur within the Ølkona mangerite; they contain blocky mesoperthite grains up to 2cm across and elongate green clinopyroxenes up to 6cm long. Xenoliths of more mafic material, possibly from the dioritic parts of the late Archaean migmatites, are quite common in the mangerite. The xenoliths are well-rounded and often have diffuse margins; the largest xenolith seen is 80 x 35cm (Plate 6.2). A number of dolerite dykes trending 074°, up to 2m wide, cut the mangerite. One of these dykes also cuts an ultramylonite zone in the mangerite. These dykes have an east-west trend, as does the second dolerite dyke-swarm on Austvågøy and South-west Hinnøy which Misra and Griffin (1972) suggested intruded the mangerites shortly after emplacement, hence the above relationship constrains the ultramylonites to a similar age.

The Sund Mangerite is emplaced into early Proterozoic supracrustal gneisses at the southern end of the western lobe of Flakstadøy. Its outcrop is elongate NNE-SSW. It is not possible to locate the western margin of the mangerite, north-west of Sund. The mangerite has an outcrop appearance different to that at Sund. It is lighter, finer-grained and has a foliation in places. In thin-section, this mangerite has abundant quartz, up to 20 modal per cent in WL302. Grain size is typically less than 0.5mm. Mesoperthite is the dominant mineral, but in WL302 it has recrystallized to plagioclase and orthoclase grains. Reactions between mafic phases and feldspars are common. Reactions in WL303 include:

orthopyroxene + plagioclase ⟶ garnet + quartz (Plate 6.3)

opaques + plagioclase ⟶ garnet + quartz

In WL302, pyroxenes have been replaced by psuedomorphs of hornblende, quartz and biotite, as recognised by Griffin and Heier (1969) (Fig. 14). Accessory zircon and rutile are also present.

Further east, away from the marginof the mangerite, more coarse-grained mangeritic rock crops out, similar to that at Ølkona, except in that it is quartz-rich. Mesoperthite, up to 1.5mm across, and quartz may make up over 95% of the rock. The remainder consists of opaques and pyroxenes , which are partially or completely pseudomorphed. Calcite is present in pseudomorphs in WL333. Garnet rims to mafic phases are common. Accessory phasesinclude apatite, epidote and rutile. Xenoliths are present throughout the Sund Mangerite: they are well-rounded, have diffuse margins, and rarely exceed 20cm across. They are all more mafic than the host mangerite. They may represent restites, after the extraction of leucocratic melts, or they may have survived as xenoliths much better than leucocratic rock-types immersed in the magma: the latter type might

have become fully assimilated. Alternatively, it may be that xenoliths of more monzonitic composition are much more difficult to detect in the field, due to lack of visual contrast with the host intrusive rock. The xenoliths are fine-grained (less than 1mm); they lack any perthite, and have 50% mafic minerals. WL332 and WL336 are from centimetre-scale xenoliths. They do not contain primary quartz, and they have retrograde reactions:

$$\text{clinopyroxene} + \text{plagioclase} \ (+H_2O) \rightarrow \text{hornblende} + \text{garnet} + \text{quartz} \quad \text{(WL332)}$$

$$\text{opaques} + \text{mesoperthite} \rightarrow \text{biotite} + \text{garnet} \quad \text{(WL336)}$$

WL366, from a metre-scale xenolith has abundant primary quartz in granular texture with plagioclase and orthoclase. Pale green clinopyroxene, biotite, which defines a foliation in the rock, and opaques are the mafic phases present. Rare garnet coronas occur around opaques; no major retrogression of the rock has occurred. This xenolith, large enough that the core is unaffected by interaction with the mangeritic magma, closely resembles the dioritic gneiss from Moskenesøy described by Green and Jorde (1971).

6.2.2 The Flakstadøy Charnockite

The Flakstadøy Charnockite has a $2km^2$ outcrop, west of Bø, in the central lobe of Flakstadøy (Fig. 5.2 and 6.1). It is emplaced into strongly foliated early Proterozoic supracrustal gneiss. The charnockite is more fine-grained than the typical Sund-Ølkona sample and is clearly quartz-rich. 20mm mesoperthite phenocrysts were observed in the field, and up to 4mm in WL306 in thin-section, but most crystals are less than 1mm. Quartz constitutes about 30% of this sample, and with mesoperthite makes up about 98%. Minor opaques, overgrown by biotite and garnet, orthopyroxene, and apatite are the other phases present. The sample is

cut by shear zones. Charnockitic veins up to 20cm wide cut the gneisses around the main charnockite body; these veins include blocks of the adjacent gneiss and are pegmatitic, with green clinopyroxenes up to 5cm long.

6.2.3 The South-West Lofoten Mangerite

All of the mangeritic rocks of Vestvågøy and Gimsøy, with the largest mangeritic body on Flakstadøy, are known collectively as the South-west Lofoten Mangerite (Fig. 6.1). Its total outcrop area is about 500km^2. It is intruded into early Proterozoic supracrustal gneisses. It also intrudes the small gabbroic bodies on Vestvågøy (Tull, 1977), but its relation to the Flakstadøy Basic Complex is uncertain, although Romey (1971) suggested that the mangerite was probably intruded by the basic complex. The mangerite on Vestvågøy has undergone hydrous retrograde metamorphism in places, notably in an east-west band from north of Stamsund to Leknes.

The largest mangeritic intrusion on Flakstadøy, the Flakstadøy Mangerite, occupies an area of about 50km^2, elongate NNE-SSW across the centre of the island. It is bounded to the west by supracrustal gneisses and to the east by the Flakstadøy Basic Complex beneath which the mangerite dips to the east. Near the contact between the basic and mangeritic rocks, rocks of mixed origin occur. WL346, from west of Nesland, is a basic rock with some mangeritic affinities and is described in Chapter 5. No sharp contacts between the main outcrops of basic and mangeritic rocks were observed on Flakstadøy during this study, but charnockitic veins, up to 30cm wide, were observed intruding the basic rocks between Nusfjord and Mørkdalen (Chapter 5). Mangeritic and charnockite veins intruding the supracrustal gneisses, north-west of Nesland, indicate that the mangerite post-dates these rocks too (Chapter 4). The mangerite typically

weathers dark brown or green but in places, especially south-east of Flakstad, it weathers pink, indicating hydrous retrogression. Plagioclase and mesoperthite up to 2cm across are seen in hand specimen. In thin-section these are clearly the dominant minerals. Plagioclase is much more abundant than mesoperthite in the samples from Myrland, on the northern tip of the island (WL315 and WL316), whereas primary plagioclase is absent from WL319 from Kilan, in the centre of the outcrop of the intrusion. Minute opaque needles are abundant in the feldspars of WL331 from within 100m of the basic complex, north of Nusfjord. This sample contains no primary quartz; all the other samples contain accessory amounts, and WL319 contains abundant myrmekite at mesoperthite grain boundaries. Maximum grain size varies between 4 and 10mm. The abundance of mafic phases ranges from less than 5% (WL319) to 20-25% for the Myrland samples and for WL309, from west of Nesland, less than 100m from basic rocks. The dominant mafic phases are orthopyroxene, clinopyroxene, opaques and biotite. Orthopyroxene is always the most abundant except for WL319, in which hydrous retrogression to biotite has occurred. This reaction:

$$\text{orthopyroxene} + \text{orthoclase} \ (+H_2O) \rightarrow \text{biotite}$$

is also observed in WL307, from north-west of Nesland. Biotite occurs as a primary mineral in three of the six rocks examined: WL315 and WL316 from Myrland and WL307, i.e. those samples in which mafics are most abundant. Hornblende and garnet are common secondary minerals; the following reactions were observed in WL315:

$$\text{opaques} + \text{mesoperthite} \ (+H_2O) \rightarrow \text{hornblende} \ (+ \text{biotite} + \text{garnet})$$

$$\text{biotite} + \text{plagioclase} \rightarrow \text{hornblende} + \text{orthoclase}$$

$$\text{orthopyroxene} + \text{mesoperthite} \ (+H_2O) \rightarrow \text{hornblende}$$

$$\text{opaques} + \text{plagioclase} \rightarrow \text{garnet} + \text{quartz}$$

$$\text{biotite} + \text{plagioclase} \rightarrow \text{garnet} + \text{orthoclase}$$

Apatite is present as an accessory phase in all samples, epidote in some.

The South-West Lofoten Mangerite occupies about two-thirds of the area of Vestvågøy, on which island it is subdivided into the Vestvågøy Brown Mangerite and the Vestvågøy Spotted Mangerite (the latter is the mangerite which has suffered extensive hydrous retrograde metamorphism. The boundaries drawn between the two mangerites on Vestvågøy (Tull, 1977) are simplified: Fresh, brown mangerite can be collected in spotted mangerite zones and vice versa. WL402 is from the retrogressed band west of Leknes and shows minor psuedomorph development after pyroxene:

$$\text{orthopyroxene} + \text{mesopherthite } (+H_2O) \rightarrow \text{biotite} + \text{quartz} + \text{hornblende}$$

which is almost complete in WL404 and WL416 from brown mangerites area at Mortsund and Haugen, at the south-west and north-east ends of the islands respectively. Plagioclase is much more abundant than mesoperthite in WL416 whereas mesoperthite is the dominant mineral in the other rocks examined. The mesoperthite has exsolved/unmixed into plagioclase and orthoclase domains during recrystallization to amphibolite-facies assemblages (Plate 6.4) as described by Griffin and Heier (1969). Plate 6.5 shows a typical hornblende-quartz-biotite pseudomorph after pyroxene (WL404). In other samples the retrogressive reactions of pyroxenes were incomplete when terminated by falling temperatures. Plate 6.6 shows a narrow hornblende and quartz corona between orthopyroxene and perthites. Plate 6.7 shows biotite plates, up to 0.1mm long growing perpendicular to the margin of an opaque crystal. Quartz and biotite are present as primary phases in some of the Vestvågøy mangerites. Biotite is the main primary mafic phase in WL416, which has similar petrography to WL315 and WL316 from Myrland, Flakstadøy. Apatite is a ubiquitous accessory; sphene, zircon and rutile also occur.

The only xenoliths seen in the South-West Lofoten Mangerite are on the road west from Stamsund. They are well-rounded, fine-grained and darker than the mangerite host rock. The largest xenolith, 50x25cm (Plate 6.8) was sampled (WL471). An equigranular (c. 0.5mm) texture of plagioclase, orthoclase and pyroxenes contains aggregates of pyroxenes, biotite and hornblende up to 2mm across, which define a weak foliation. Opaques are present as accessory minerals. Mafic phases comprise c. 40% of the xenolith, which probably belongs to the early Proterozoic supra-crustal gneisses.

6.2.4 The Hopen Mangerite/Charnockite

The Hopen Massif, consisting of mangerites and charnockites, occupies an area of about 15km^2 on the south-western corner of Austvågøy (Fig. 6.2). The extent of the mangeritic rocks is limited to the north by the early Proterozoic supracrustal rocks and to the south by the Festvåg migmatites at least part of which is late Archaean in age. The migmatites are described in Chapter 2. Near the contact with the Hopen Mangerite, the migmatites are broken up by veins from the mangerite. The Hopen Mangerite has a layered border series against the migmatites, which consists of gabbroic rock, with pyroxenite and magnetite layers up to 40cm thick (Plate 6.9) which dip steeply (c. 70°) to the north. Magmatic sediment-ation features, such as planar (Plate 6.10) and trough cross-bedding are common. Shear zones have rotated the pyroxenite layers parallel to the shearing implying that the deformation occurred while the rock was partially crystalline. Samples of the gabbroic rock show that grain size rarely exceeds 2mm, but mafic minerals occur in aggregates producing the spotted appearance of the rock. Opaques and apatites crystallized first (with accessory olivine in WL547) and are overgrown by clinopyroxene and minor orthopyroxene. Biotite is an important primary phase, hornblende less so; hornblende overgrows biotite in places. Plagioclase has undergone slight

alteration to calcite and epidote. The accumulation of these cumulus minerals (clinopyroxene, orthopyroxene, olivine) supports the idea that the gabbroic rocks are part of an early-formed marginal series of the Hopen intrusion. WL548 contains about 10 modal per cent apatite, which may be up to 15 x 3mm. Apatite is evidently a cumulus phase in the Hopen intrusion. WL544 consists of alternating clinopyroxenite and hornblendite layers with variable, but minor, amounts of biotite, opaques, plagioclase and apatite. No similar gabbroic rocks have been observed on the northern margin of the Hopen intrusion, but a very large fallen block on the road 1km south-west of Rørvik consists of a breccia of gabbro (blocks up to 1m across) in mangerite. This has presumably fallen from the mountains to the south-east where a marginal gabbroic series may crop out. The position of the border between the Hopen intrusion and the supracrustal gneisses on the west coast, differs by 1km on the maps of Ormaasen (1976) and Tveten (1978). As discussed in Chapter 4, fieldwork for this study supports the more northerly position for the contact, as favoured by Tveten (1978).

The rocks of the Hopen Massif are typically granular, with green feldspars and blue-grey quartz up to 10mm, with minor opaques and pyroxenes which have undergone partial hydrous retrogression (Plate 6.12). Pyrite has been identified in the field. On the foreshore between Rørvik and Djupfjord, the mangeritic rocks are often foliated, which may lead to their confusion with the supracrustal gneisses. The foliation is defined by alignment of mafic minerals in bands, a few millimetres wide. The foliation strikes east-west, parallel to the doleritic dykes, described by Misra and Griffin (1972), which range from 0.1 - 2.5m wide. These dykes contain slivers of mangeritic rock and have anastomosing margins with the foliated mangerite/charnockite. The foliated mangerite with numerous east-west dolerite dykes is restricted to the foreshore between

the northern limits of the mangerite as proposed by Ormaasen (1976) and Tveten (1978) - i.e. in supracrustal gneisses according to Ormaasen, and in mangeritic rocks according to Tveten. A number of samples have been collected from this foliated rock including WL508, WL510 and WL575. Quartz (up to 50% of the rock, in WL575), mesoperthite and minor plagioclase, up to 4mm across comprise 80-90% of these rocks. Alignment of the mafic phases into bands defines the foliation in these rocks. The primary mafic phases are orthopyroxene, clinopyroxene and opaques (and minor hornblende in WL508). Hydrous retrogression of pyroxenes, producing coronas, is common:

$$\text{clinopyroxene} + \text{plagioclase } (+H_2O) \longrightarrow \text{hornblende } (+ \text{ quartz})$$

$$\text{orthopyroxene} + \text{plagioclase } (+H_2O) \longrightarrow \text{hornblende } (+ \text{ quartz})$$

Opaques are rimmed by biotite or garnet. Accessory phases include apatite, zircon, sphene and rutile.

The area of foliated mangeritic rock and abundant dolerite dykes is bounded to the south by a dolerite body about 100m wide which is back-veined by mangeritic rock. To the south of this dolerite, towards Djupfjord, no dolerite dykes are seen, the mangerite/charnockite is coarser and is rarely foliated, e.g. WL565, Plate 6.12. WL565 is atypical of the Hopen mangeritic rocks in that hornblende-quartz-biotite pseudomorphs are more common than relict pyroxenes. The granular mangeritic rocks are similar petrographically to those with a foliation. Quartz is up to 30% of the rock (WL514) and mafics are 5-10%. The mangerite/charnockite near to the Festvåg Migmatites on the west coast contains abundant xenoliths and is foliated. Plates 6.13 and 6.14 show a large block of mafic rock being broken up by mangeritic rock. Plate 6.14 shows how slivers of the xenolith are being incorporated by the mangerite which is foliated parallel to the margin of the xenolith. WL567 (Plate 6.15) is

a sample of the mangerite-xenolith margin. The xenolith (WL568) has a grain size of less than 0.1mm and consists of plagioclase, hornblende, biotite and accessory apatite and has a strong foliation defined by the hornblende and biotite. It is similar to the amphiboites in the migmatites at Festvåg especially WL557. Along the contact between the xenolith and the host mangerite is a green pyroxene-rich band. The foliation in the host mangerite is defined by concentrations of pyroxenes into bands. This is also true for WL517 and WL518, other foliated mangeritic rocks from road-cuts just north of Festvåg. Plate 6.16 shows a rounded xenolith, 25 x 20cm, from 1km north of Festvåg in mangeritic rock which has no foliation.

6.2.5 The Austvågøy Grey Mangerite

The Austvågøy Grey Mangerite occupies the centre of the western half of Austvågøy. Its total area is about 200km^2 and it is bisected into a north and a south part by the Olderfjord Fault. North of the fault the mangerite has undergone extensive hydration metamorphism and is tightly infolded with supracrustal gneisses. To the south retrogression and deformation is less severe. The Austvågøy Grey Mangerite south of the fault has been examined in this study around Svolvaer and Kabelvåg (Fig. 6.2). All of the samples described and analysed fall outside the grey mangerite as shown by Ormaasen (1976), but inside the grey mangerite as shown by Tveten (1978). The supracrustal gneisses and the mangerite are both exposed in the new road-cuts in west Svolvaer. Pink and grey quartzo-feldspathic gneisses are present, some of which have coarse, granular layers within the foliation or cross-cutting it. These may be anatectic veins or veins from the mangerite. Some of the mangerite is brown and equigranular, and resembles the Vestvågøy Brown Mangerite; other parts have pale grey feldspars and dark aggregates of mafic phases, similar to the Vestvågøy Spotted Mangerite. Mesoperthite is the dominant

mineral in the Austvågøy Grey Mangerite. Plagioclase is less abundant. Primary quartz up to 7mm is present in all of the rocks, and represents 30% of the mode of WL526, from a road-cut between Kabelvåg and Svolvaer. Mafic phases total from 5% (WL526) to 20% (WL529 and WL530, from the new road-cuts in west Svolvaer). Primary magmatic hornblende is the main mafic mineral in WL525 from 1km north-east of Kabelvåg) and WL529; it sometimes has a thin biotite reaction rim and overgrows orthopyroxene and clinopyroxene. Clinopyroxene and opaques are the next most abundant mafic phases, the former with hornblende and quartz coronas and the latter rimmed by biotite, hornblende or garnet. Orthopyroxene is a minor phase except in WL526 and WL533 (from 1km east of Svolvaer), where it is as abundant as clinopyroxene; in WL526 it contains exsolution lamellae of clinopyroxene. As well as from the development of coronas, retrogression is also indicated by the recrystallization of mesoperthite into plagioclase and orthoclase especially in WL522, from Kabelvåg, and in WL529. Apatite is an accessory in all of the Austvågøy Grey Mangerite samples. Sphene and zircon also occur, as do white micas associated with the breakdown of plagioclase, and myrmekite.

6.2.6 The Raftsund Mangerite

The Raftsund Mangerite is the largest intrusion in the Lofoten-Vesterålen province, underlying an area of about 1000km^2. Its main outcrop is on the eastern part of Austvågøy and the south-western part of Hinnøy. Some of the mangeritic rocks of Hadelsøy and Hamarøy are believed to represent extensions of the Raftsund Mangerite (Fig. 6.3). Griffin et al. (1974) recognised a zonation of the intrusion on the basis of primary mafic phases: clinopyroxene - orthopyroxene, clinopyroxene - orthopyroxene - olivine and clinopyroxene - olivine. In general the first is found in the outer parts of the intrusion and the last in the core. Griffin et al. (1974) estimated that the intrusion crystallized at 900° ±

100°C at 8 ± 1kb (20-25km), and suggested that it is c.3km thick.

The Raftsund Mangerite is typically dark brown or green, granular, and composedchiefly of blocky mesoperthite crystals up to 10mm (Plate 6.17). Plagioclase is common, up to 20mm long (WL602, WL604, WL605). Pyroxenes are up to 10mm. In a few places the mangerite has undergone extensive hydration retrogression. There, the feldspars are pale grey or white, and the pyroxenes are altered to hornblende and biotite which occur in aggregates with opaques and garnets, giving the rock a spotted appearance (Plate 6.18) similar to the retrograded mangerites on Vestvågøy and Austvågøy. Magmatic-sedimentary features are well developed in the central parts of the intrusion. Plate 6.19 shows a graded channel which scours into normal homogeneous mangerite; this is from the north side of Tenfjord on the eastern (Hinnøy) side of Raftsund. A similar graded mangerite, 4km to the south-west near Digermulen was collected as WL618 (Plate 6.20, the top is to the right). The lower 7cm is rich in mafic phases, shows grading and has a sharp bottom. The felsic top to this sample is 23cm thick in outcrop, where normal, coarser mangerite is in sharp contact with it. 10cm above this sharp top is a 40 x 30cm block of fine-grained rock. These features of this structure are similar to those described for "fine-grained layering" in Griffin et al. (1974). "Centimeter-scale layering" of Griffin et al. (1974) is commonly observed along the Raftsund. Mafic bands about 2cm thick alternating with 20-30cm thick bands of normal mangerite are common at Tengelfjord, at the north end of the Raftsund, and similar cm-scale layers are seen to be cross-bedded at Alteret, north of Digermulen at the south end of Raftsund. At Kongsmarken at the north end of Raftsund, mafic segregations define pods and lenses within normal mangerite. Pyroxenes are aligned parallel to the long axes of the pods suggesting that the pods define primary igneous lamination. Coarse pegmatites intrude the normal mangerite. At Finkirken a quartz-rich

pegmatite contains 2cm allanite crystals with associated radiating cracks, resulting probably from volume expansion of the allanite caused by cumulative lattice damage resulting from Th and U radioactive decay rendering the crystal structure metamict. At Storneset a pegmatite layer rich in blue quartz occurs with pyroxenes up to 2cm and elongate amphiboles up to 8cm long.

The main primary minerals of the Raftsund Mangerite are mesoperthite and plagioclase. The former may make up 90% of the rock (e.g. WL540), but the two feldspars are more commonly present in roughly equal amounts (e.g. WL606). Recrystallization results in exsolution of mesoperthite to plagioclase and orthoclase (Plate 6.1) (Griffin and Heier, 1969). Griffin et al. (1974) defined the zonation of the mangerite on the basis of the mafic silicate assemblages. They found that clinopyroxene and orthopyroxene occur as the primary mafic silicate phases in the margins of the intrusion; clinopyroxene and olivine in the core of the intrusion; and clinopyroxene, orthopyroxene and olivine occur between the other zones. The distribution of these assemblages in the rocks examined in the present study is shown in Fig. 6.3. This agrees broadly with Griffin et al. (1974), but rocks belonging to the zones other than clinopyroxene - olivine have been recognized in the centre of the intrusion. All five samples from Hamarøy belong to the "marginal" clinopyroxene - orthopyroxene zone. In this zone the pyroxenes often contain exsolution lamellae; Plate 6.22 shows broad exsolution lamellae of clinopyroxene in orthopyroxene in WL605 from Raften. Plate 6.23 shows orthopyroxene as a magmatic overgrowth to pale green clinopyroxene, which contains earlier-formed opaques (WL538; south of Fiskebol, Austvågøy). Magmatic hornblende often overgrows earlier-formed pyroxenes (e.g. WL617). Olivine is commonly altered to magnetite and a brown anisotropic phase but the pseudomorphs can be recognized by their surrounding orthorpyoxene-garnet

coronas (e.g. WL617) (Griffin and Heier, 1973). Olivine has broken down to opaques and granular orthopyroxene in some samples (e.g. WL609, WL625), as described by Griffin and Heier (1969). In those samples in which olivine grains are present, they have high (third order) birefringence, which suggests that they are fayalitic in some samples (e.g. WL609, WL625). The clinopyroxenes in the two pyroxene-olivine and clinopyroxene-olivine zones are typically rather dark green, which suggests that they are Fe-rich; they also contain abundant oriented opaque needles (e.g. WL625, WL626). In the two-pyroxene-olivine zone, the orthopyroxene commonly occurs as overgrowths to clinopyroxene, as in the two-pyroxene zone, but it also occurs between clinopyroxene and olivine, suggesting that it is a magmatic reaction product (e.g. WL605 from Tengelfjord). Olivine crystallized prior to the pyroxenes; it is overgrown by clinopyroxene in WL625 and WL626.

Accessory phases in the Raftsund Mangerite include apatite, zircon, sphene, rutile, quartz, epidote, calcite and muscovite. In the typical, unretrograded mangerites, mafic phases are 5-20% of the rock and grain size is 0.5-8mm. In the retrograded mangerites, mafics may constitute up to 25% of the rock (e.g. WL101, WL610). All five mangerite samples from Hamarøy have retrograded assemblages. Plate 6.18 shows WL610, from Tennstrand, with the typical appearance of retrograded mangerite. During retrogression, mesoperthite has recrystallized to exsolve plagioclase and orthoclase. Plagioclase is partially saussuritised producing epidote, calcite and white mica. Hornblende-quartz-biotite (and sometimes calcite) pseudomorphs after pyroxenes occur in aggregates with pink garnets, apatite, and sphene which overgrows opaques, (radially in WL610) to produce the characteristic spotted appearance. WL105, from Hamarøy, shows a pale-green clinopyroxene, 1.5mm across, with an inner corona of hornblende and quartz, and a complete outer corona of hornblende. Within 5mm of this is a complete hornblende-quartz pseudomorph after pyroxene with a

surrounding ring of subhedral sieve-like garnets, indicating that retrograde reactions advance to different extents within the same thin-section. Most of the unretrograded mangerites contain hornblende coronas to pyroxenes, exsolution of mesoperthite, and/or garnet rims to opaques and pyroxenes, all of which are reactions permissible with falling temperature at constant pressure - retrograde reactions (Griffin and Heier, 1969).

WL607, from Brubrekken, is from a 5 x 2m fine-grained area in normal mangerite. The feldspars-plagioclase, orthoclase and minor perthites - are usually less than 1mm across. Mafic phases are c.35% of the rock, which is chiefly orthopyroxene and primary biotite (up to 4mm across), with minor opaques and clinopyroxene. This is either a xenolith or a fine-grained biotite-rich mangerite pod.

WL650, from Tenfjord, is from a mafic-rich layer within an otherwise homogeneous section of mangerite. It consists of rounded olivine and clinopyroxene grains, typically 3mm across, which together total c.90% of the rock with an interstitial opaque phase, and minor apatite with accessory orthopyroxene, hornblende and biotite. This is a cumulus layer of the primary mafic phases from the Raftsund Mangerite.

The only mangeritic intrusions with more abundant xenoliths than Raftsund, are the Sund Mangerite, the Hopen Mangerite/Charnockite and the Engenykken Mangerite on Langøy. Xenoliths are particularly easy to see in the Raftsund Mangerite around Kongsmarken, at the northern extremity of the road along Raftsund, where they stand out against the white feldspars of the retrograded mangerite. Most are well-rounded, some are zoned, some have reaction rims and most are less than 10cm across (Plate 6.24); all these facts suggest that the xenoliths have undergone partial assimilation

by the mangerite.

Two xenoliths have been petrographically studied. WL614A is a centimetre-scale xenolith from Finkirken. It is fine-grained (0.2-1mm) with plagioclase and orthoclase in a granular texture. Mafic phases are c.35% of the xenolith; the dominant mafic phase is primary hornblende, up to 2mm long, which overgrows minor orthopyroxene. Biotite is also common; zircon is present in accessory amounts. WL649 is from a 100x 60cm xenolith from Tenfjord. Plagioclase, less than 1mm across constitutes 50% of the rock, mafic phases the other 50%. The mafic phases are hornblende, biotite, clinopyroxene and orthopyroxene, up to 1.5mm across, with accessory opaques and zircons. It is cut by a charnockitic vein with grains up to 5mm across. This xenolith resembles WL568 from the Hopen intrusion which in turn was compared to WL557, an amphibolite from the Festvåg migmatites.

The eastern contact of the Raftsund Mangerite is with the Lødingen Granite (Figs. 6.3 and 7.1). At Offersøy road-junction, the contact is exposed in a road-cut. A 6m wide block of retrograded mangerite is intruded by 2-3cm wide granitic veins, one of which separates a 40x10cm block of mangerite from the main body. The granite-mangerite contact is sharp, indicating that the Lødingen Granite intruded the Raftsund Mangerite after the latter had solidified.

6.2.7 The Eidsfjord Mangerite

The Eidsfjord Mangerite crops-out on the southern side of Eidsfjord in southern Langøy (Fig. 6.4) and underlies an area of about 12km^2. A sample of the mangerite, WL732, was collected south of Grønning. Mesoperthite megacrysts, up to 3mm across, occur in a matrix of mesoperthite and plagioclase of grain size less than 1mm. Mafic phases constitute

c. 20% of the rock and are dominantly orthopyroxene, clinopyroxene and opaques. Minor primary biotite is present, contrary to the observation of Ormaasen (1976). Accessory apatite is present.

WL733 is a sample of a fine-grained (less than 1mm) mafic-rich (c. 35% of the rock) xenolith from the same road-cut as WL732. The dominant mafic phases are orthopyroxene, clinopyroxene and opaques with minor biotite. Plagioclase is much more abundant than orthoclase. Apatite is present in accessory amounts.

6.2.8 The Bø Mangerite

The Bø Mangerite consists of two geographically separate units in the south of Western Langøy (Fig. 6.4). The Steine Mangerite lies to the south-east of the Søberg Mangerite and occupies about 40km^2; the Søberg Mangerite underlies an area of about 10km^2. In samples from around Steine (WL713, WL714) the fine-scale perthites and plagioclase, with grain size up to 5mm across and myrmekites at grain boundaries, constitute about 95% of the rock. Biotite and opaques are the main mafic phases. Pyroxenes are pseudomorphed by biotite, quartz and opaques in WL714. Primary hornblende is present in WL713, but not WL714. Apatite and zircon occur as accessories, the latter in biotite and surrounded by radiation haloes.

A sample from the Søberg Mangerite, WL715, contains perthite megacrysts up to 15mm across, in a matrix of plagioclase and perthite grains less than 2mm across. Mafic phases total about 15% of the rock (c. 5% in the Steine Mangerite) and consist of orthopyroxene, clinorpyoxene, opaques and biotite. Apatite and quartz occur as accessory phases.

6.2.9 The Engenykken Mangerite

Two mangerites crop-out on the Eidet-Hovden peninsula in the north of Western Langøy. The smaller of the two, underlying an area of about $1km^2$ on a peninsula, is the Engenykken Mangerite (Fig. 6.4). The contact between the mangerite and the Eidet-Hovden Gabbro corresponds to the narrow isthmus linking the peninsula to the mainland. WL737 is from the south-western end of the peninsula where the mangerite has an irregularly variable grain size and a high content of basic inclusions, as reported by Ormaasen (1976). This sample shows coarser and finer banding, with the former having grain size up to 7mm and about 15% mafic phases and the latter having grain size less than 1.5mm and mafic phases about 25% of the rock. In both the coarser and finer bands orthopyroxene and clinopyroxene are the dominant mafic phases with less abundant opaques, secondary biotite and apatite. The fine-grained areas are very similar to the xenolith from the Eidsfjord Mangerite, WL733, except that it has minor primary biotite. The irregularly variable grain size in this part of the Engenykken mangerite is probably due to a high content of partially assimilated mafic xenoliths.

6.2.10 The Hovden Mangerite

The larger of the two mangerites which crop-out on the Eidet-Hovden peninsula, is the Hovden Mangerite, which underlies an area of about $5km^2$, around Hovden on the northern tip of the peninsula and to the south (Fig. 6.4). The reduction in grain-size of the surrounding (except to the north) Eidet-Hovden Gabbro adjacent to the mangerite, which was noted by Heier (1960), has been observed to the west of the quarry at Hovden Habour. Plate 6.25 shows a coarsely granophyric sample from a 20cm wide vein of charnockitic composition (WL720) (see Chapter 5) which intrudes the gabbro in the western face of the Hovden Harbour quarry, only a few tens of metres from the Hovden Mangerite. Other veins of mangeritic composition have been

observed in the gabbro about 4km WSW of Hovden. These veins suggest that the mangerite post-dates the gabbro, as is also the case on Vestvågøy, but if the fine-grained gabbroic margin was a chill zone then the gabbro post-dates the mangerite and the veins were formed by contact melting.

Two samples from the Hovden Mangerite, WL703 and WL707, 2km and 1.5km WSW of Hovden respectively, contain megacrysts of ribbon mesoperthites, up to 10mm across, similar to the Eidsfjord and Søberg Mangerites. The megacrysts contain biotite, opaques and apatite, and constitute about 50% of the rock. The matrix is mesoperthite, plagioclase and orthoclase up to 0.5mm across. Mafic phases make up about 5% of the rock. Secondary hornblende and biotite occur as rims to pyroxenes and opaques.

6.3 Geochemistry

Thirty-eight samples from the mangeritic intrusions of the Lofoten-Vesterålen province have been analysed for major element contents in this study (Table 6.1 - 6.7). Analyses for three xenoliths in mangerites are presented in Table 6.8.

The important points to make regarding the major element compositions of the mangerites as a whole, and compared to the major element compositions of the basic intrusions (Table 5.1 and 5.2) are:

(i) SiO_2 = 51-77% (for basics, SiO_2 = 43-54%);

(ii) K_2O = 1.6-9.0% (for basics, K_2O = 0.3 - 1.2%);

(iii) $Na_2O/K_2O < 2$, except for WL202, WL315, WL410 (for basics, $Na_2O/K_2O > 2.5$);

(iv) Fe_2O_3/MgO = 1.5-63 (for basics, Fe_2O_3/MgO = 0.5-3.2);

(v) CaO = 0.8-9% (for basics, CaO = 6-17%).

On the basis of the Malm and Ormaasen (1978) sub-division of mangeritic rocks (Chapter 1.2), charnockites (SiO_2 > 67%) are present in the Sund Mangerite, the Flakstadøy Charnockite, the Hopen Mangerite/Charnockite and the Austvågøy Grey Mangerite. Basic mangerites (SiO_2 < 56%) only occur in the South-West Lofoten Mangerite, the Eidsfjord Mangerite, and the fine-grained sample, possibly a xenolith, in the Raftsund Mangerite, WL607. Excluding WL607, the Raftsund Mangerite shows the least chemical variation of any of the mangerites, despite its size and its layering. The Hopen intrusion shows the most extreme chemical evolution, having both the most acid (WL575), and the highest Fe_2O_3/MgO ratio (WL570) samples in all the Lofoten mangeritic suite. WL570 is from the locality at Hopen at which Ormaasen (1977) reported the occurrence of orthoferrosilite, and in the norm (Table 6.4) this sample shows extreme Fe-enrichment in the mafic phases and almost pure albitic plagioclase.

On a plot of K_2O v. SiO_2 (Peccerillo and Taylor, 1976; Ewart, 1982) (Fig. 6.6) most of the mangerites are classified as having the compositions of shoshonites and banakites i.e. basaltic andesites (SiO_2 = 52-56%) and andesites (SiO_2 = 56-63%) with high K_2O (> c.3%). In basaltic andesites of the shoshonite series common phenocryst phases include plagioclase, augite, Fe-Ti oxide, biotite, hornblende, ± olivine, ± orthopyroxene, and less commonly, sanidine (Ewart, 1982). These are all primary phases in the Lofoten-Vesterålen mangerites (orthoclase rather than sanidine, and abundant orthopyroxene due to crystallization in granulite-facies conditions). Ewart (1982) recognised a contrast in the K-content of andesitic rocks of the orogenic zones of continental regions and of island arcs. A much greater proportion of high K or shoshonitic magmas occurs in continental orogenic regions, especially in the western USA and Mediterranean region. The proportion of K-enriched magmas is greater in the dacitic and rhyolitic compositional groups of the continental

regions; the difference in proportions of low K to high K magmas with basaltic composition, between the continental and island arc environment, is minimal. Ewart (1982) concluded:

> "The interpretation of this writer is that the increase in K (and related elements) must be sought in processes of reworking of continental crust by repeated intrusion, assimilation, and anatexis. This applies especially to intermediate and silicic magmas of the continental and Andean orogenic regions. The intrusions providing the necessary heat for these processes are assumed to rise from the region of the Benioff zone and the overlying peridotite wedge into a crust that may itself have been produced in a geologically short time span by an earlier episode of orogenic magmatism."

A number of variation diagrams are presented for the thirty-eight chemical analyses of Lofoten-Vesterålen mangerites (Figs. 6.5 - 6.14). Each of the diagrams includes points for the average late Archaean migmatite and average early Proterozoic supracrustal gneiss from Griffin et al. (1978) (Table 2.1 and 2.2 in this work). They also include the average Flakstadøy Basic Complex composition (derived in Chapter 5.3) and the average composition of the two rhyolitic supracrustal gneisses from Vik, Vestvågøy (Table 4.1). These represent some of the possible source materials for the production of the mangeritic magmas. The composition of the Flakstadøy Basic Complex is taken to approximate the composition of a possible mantle-derived component, and the gneisses represent possible crustal sources. The composition of the average pyroxenite from Cox et al. (1979) is used to represent cumulus mafic phases which may have crystallized early from the mangeritic magma.

Na_2O v. SiO_2 (Fig. 6.5) shows all thirty-eight mangerite samples. The four possible source compositions considered have Na_2O of about 4% for variable SiO_2 content. All of the mangerite samples have Na_2O greater than or equal to 4%. This trend above the "end-member mixing-line" plots

on a vector away from the average pyroxenite composition implying that higher Na_2O contents in mangerites may result from crystal fractionation of mafic phases. Mixing involving the Vik rhyolitic supracrustal gneiss end-member could explain the presence of the high-SiO_2 mangeritic rocks. WL202 from Ølkona and WL517 from Hopen have the highest Na_2O content and in that respect are the most fractionated mangerites.

The corresponding plot of K_2O v. SiO_2 (Fig. 6.6) is similar to Na_2O v. SiO_2 except that the "end-member mixing-line" has a steep positive slope. Data-points plot all along this line between the Vik and basic rock end-members but most plot on a vector diametrically away from the average pyroxenite composition through the compositions of mean Archaean and mean Proterozoic gneisses. WL204 and WL210 from Ølkona have the highest K_2O content and hence using this parameter are the most evolved of the mangerite samples. Note the high K_2O content of the mangerites as a whole; they could thus be classified as shoshonitic and banakitic in composition (but not in texture + mode of occurrence).

The mangerites show good negative correlation between CaO and SiO_2 (Fig. 6.7). They plot on a vector diametrically away from the average pyroxenite composition implying that the mangerite magmas evolved by early crystallization of pyroxenite phases. Those mangerites (WL315, WL316 and WL410 from the South-West Lofoten Mangerite) which are less evolved (higher CaO, lower SiO_2) than the mean Proterozoic gneiss composition must have a large component of Ca derived from rocks similar to the Flakstadøy Basic Complex i.e. mantle-derived. The rocks on this plot which have evolved furthest, to low CaO-high SiO_2 are the charnockites from Sund, Hopen and the Austvågøy Grey Mangerite.

The mangerites also have a negative correlation between MgO and SiO_2

(Fig. 6.8), but the trend is convex towards the origin implying that MgO-rich phases are fractionated early from the magma. MgO is less than 1% for all mangeritic rocks with SiO_2 > 63%. This curvature is probably due to the early crystallization of olivine with subsequent crystallization of clinopyroxene which less markedly reduced MgO in the residual magma. WL315, from the South-West Lofoten Mangerite, is the least evolved mangerite sample on this variation diagram.

The variation diagram Fe_2O_3 v. SiO_2 (Fig. 6.9) shows a negative correlation for the mangerites. Those mangerites (WL309 and WL315 from the South-West Lofoten Mangerite, WL607 from Raftsund and WL732 from Eidsfjord) which are less evolved (higher Fe_2O_3, lower SiO_2) than the mean Proterozoic gneiss composition must have a large component of high Fe_2O_3 - low SiO_2 material. The Flakstadøy Basic Complex has low Fe_2O_3 (4.9%), but prior to fractionation the basic rocks may have had higher Fe_2O_3: the average gabbro composition of Cox et al. (1979) contains 10.6% total Fe oxides. The small basic bodies on Vestvågøy and Langøy have an average Fe_2O_3 content of 10.7%.

The variation diagram Fe_2O_3 v. TiO_2 (Fig. 6.10) shows a good positive correlation through the origin. Since Fe is a compatible element (shown by decrease in Fe_2O_3 with evolution of the magmas, Fig. 6.9) Ti is also a compatible element: more evolved magmas having lower TiO_2 contents. This is compatible with the observation that opaques (probably ilmenite and magnetite) occur with cumulus pyroxenes and olivine.

The variation diagram for P_2O_5 v. TiO_2 (Fig. 6.11) also has a good positive correlation (the slight displacement from the origin may be due to calibration difficulties with the XRF) which indicates that P is also a compatible element, i.e. P-bearing phases are fractionated from the magma.

This is in accord with the occurrence of early-crystallized (? cumulus) apatite within pyroxenes and olivine, and with the occurrence of the very high apatite content in the marginal gabbroic series of the Hopen intrusion (apatite up to 15mm long in WL548). WL309 (South-West Lofoten Mangerite), WL607 (Raftsund) and WL732 (Eidsfjord) have P_2O_5 > 1%, TiO_2 > 1.9% which are much higher than the P_2O_5 and TiO_2 contents of the average pyroxenite. This is because the phases that have fractionated from the mangerite magmas are those of typical pyroxenite plus apatite and ilmenite.

Figs. 6.12 and 6.13 are both CaO v. MgO variation diagrams: Fig. 6.12 is for all thirty-eight mangeritic analyses; Fig. 6.13 is for six analyses from the Raftsund Mangerite. Both diagrams show that CaO and MgO are positively correlated with the trend convex towards the CaO axis. The average pyroxenite composition has CaO and MgO both greater than 10%, so progressive removal of pyroxenite phases from the mangerite magmas would produce an evolution trend with positive slope towards the origin from the original magma composition. The highest CaO and MgO contents for the mangerites (except for the very primitive South-West Lofoten Mangerite sample, WL315) are very close to the mean composition of early Proterozoic gneisses. The curvature of the trend can be explained by the early crystallization of olivine and plagioclase which causes a marked decrease in MgO in the residual liquid, but less markedly so in CaO. Subsequently, as clinopyroxene crystallizes, the CaO content of the magmas decreases markedly. This curvature of the evolution trend is shown very clearly for the mangeritic intrusion with the best-developed cumulate features, the Raftsund Mangerite (Fig. 6.13), within which olivine is overgrown by clinopyroxene (Section 6.2.6), supporting the sequence of crystallization offered in explanation of the curvature above.

These variation diagrams suggest that the mangerite magmas were

derived primarily by anatexis of typical late Archaean and early Proterozoic gneisses, with some samples containing significant components derived ultimately from the mantle, or from rhyolitic gneisses similar to those at Vik, Vestvågøy. These magmas evolved by crystallization of apatite, ilmenite and pyroxenite phases to higher Si, K, Na, lower Ca, Fe, Mg, Ti, P magmas. The most evolved mangeritic samples are the charnockites from Sund, Ølkona and the Austvågøy Grey Mangerite, and the most primitive are WL309, WL315, WL316 and WL410 from the South-West Lofoten Mangerite, WL607 from the Raftsund Mangerite, and WL732 from the Eidsfjord Mangerite.

6.4 Isotopic Results

6.4.1 Sr Isotopes

Sr isotope data for the mangeritic intrusions of the Lofoten Vesterålen province are tabulated and plotted in Figs. 6.14 to 6.18.

Five samples from the Sund-Ølkona Mangerite have been analysed (Fig. 6.14). They have a range in Rb and Sr concentrations of 9 to 124ppm and 57 to 590 ppm, respectively. The range in $^{87}Rb/^{86}Sr$ ratios is 0.0324-4.425 and for $^{87}Sr/^{86}Sr$ ratios is 0.70511-0.81974. These five points define an isochron of 1774 ± 42Ma (2σ), (MSWD = 4.11) with an initial $^{87}Sr/^{86}Sr$ [or $(^{87}Sr/^{86}Sr)_o$] ratio of 0.70422 ± 27. This compares to an age of 1800 ± 50Ma (2σ) and $(^{87}Sr/^{86}Sr)_o$ of 0.70333 ± 58 from a seven-point isochron published in Griffin et al. (1978).

The single sample from the Flakstadøy Charnockite has a higher Rb concentration than all of the Sund-Ølkona samples: 163ppm. The Sr concentration is 235ppm; $^{87}Rb/^{86}Sr$ is 2.093; and $^{87}Sr/^{86}Sr$ is 0.75722.

Fourteen samples from the South-West Lofoten Mangerite (six from Flakstadøy, eight from Vestvågøy) have been analysed (Fig. 6.15). Their ranges of Rb and Sr concentrations are 20 to 155ppm and 263 to 973ppm, respectively. $^{87}Rb/^{86}Sr$ ranges between 0.0521 and 1.778; $^{87}Sr/^{86}Sr$ between 0.70447 and 0.74887. These data yield a whole-rock isochron age of 1893 ± 128Ma (2σ), (MSWD = 24.3) and an initial $^{87}Sr/^{86}Sr$ ratio of 0.70336 ± 43. This compares with an age of 1735 ± 25Ma (2σ) and $(^{87}Sr/^{86}Sr)_o$ of 0.70411 ± 16 from a 14-point isochron published in Griffin et al. (1978).

The nine samples analysed from the Hopen Mangerite/Charnockite (Fig. 6.16) have Rb and Sr concentrations in the range 38 to 93ppm and 39 to 326ppm respectively. The range in $^{87}Rb/^{86}Sr$ ratios is 0.4155 to 11.493 and for $^{87}Sr/^{86}Sr$ is 0.71581 to 0.99700. These data define an isochron with an age of 1811 ± 33Ma (2σ), (MSWD = 3.37) and an initial $^{87}Sr/^{86}Sr$ ratio of 0.70480 ± 48. The age published in Griffin et al. (1978) for this intrusion is 1790 ± 50Ma (2σ) with $(^{87}Sr/^{86}Sr)_o$ of 0.70456 ± 72 (9 points).

Six Sr isotope analyses are presented for the Austvågøy Grey Mangerite in Fig. 6.17. The range in Rb and Sr concentrations is 30 to 149ppm and 106 to 454ppm respectively. $^{87}Rb/^{86}Sr$ has a range from 0.2199 to 3.437; $^{87}Sr/^{86}Sr$ has a range of 0.71005 to 0.79414. These data yield an age of 1808 ± 70Ma (2σ), (MSWD = 10.90) and an initial $^{87}Sr/^{86}Sr$ ratio of 0.70411 ± 67. The corresponding age in Griffin et al. (1978) is 1805 ± 20Ma (2σ), with $(^{87}Sr/^{86}Sr)_o$ of 0.70318 ± 32 (9 points).

Twenty-one samples from the Raftsund Mangerite have been analysed (Fig. 6.18). Thirteen of these samples are from Hinnøy, three from Austvågøy, and five from the mangerite on Hamarøy which is believed to be

a part of the Raftsund Mangerite. The range in Rb and Sr concentrations for these twenty-one samples is 27 to 75ppm and 55 to 424ppm, respectively. The range in $^{87}Rb/^{86}Sr$ ratios is 0.2043 to 5.095 and for $^{87}Sr/^{86}Sr$ is 0.71007 to 0.83162. These data yield an age of 1731 ± 18Ma (2σ), (MSWD = 2.24) and an initial $^{87}Sr/^{86}Sr$ ratio of 0.70475 ± 12. Griffin et al. (1978) published separate isochron data for the main body and the Hamarøy part of the Raftsund Mangerite. For the main body, the age is 1705 ± 30Ma (2σ) and $(^{87}Sr/^{86}Sr)_0$ is 0.70514 ± 34 (12 points), and for the Raftsund Mangerite on Hamarøy the age is 1710 ± 60Ma (2σ) and $(^{87}Sr/^{86}Sr)_0$ is 0.70462 ± 48 (8 points).

Rb-Sr data for the Langøy Mangerites are presented in Table 6.9.

Rb-Sr data for three xenoliths in mangerites are presented in Table 6.10 The ages for the mangerites produced in this study are not significantly different from those presented in Griffin et al. (1978) except for the South-West Lofoten Mangerite. There is considerable scatter about the isochron in Fig. 6.15, giving rise to the poor precision of the measured age. This scatter could be caused by either subsequent disturbance of the Sr isotopes by metamorphism [as suggested for the migmatites at Fesvtåg and Sandset (Chapter 3) and for the Vik gneisses (Chapter 4)] or by variations in initial $^{87}Sr/^{86}Sr$ ratios. The wide variations in petrography (Chapter 6.2.3) and chemistry (Tables 6.2 and 6.3) for this mangerite suggest the latter explanation may be valid.

The Raftsund Mangerite is confirmed as the youngest mangeritic intrusion in the Lofoten-Vesterålen province. It shows little variation in Rb/Sr ratios, just as it also showed little chemical variation.

The greatest variation in Rb/Sr ratios is found in those intrusions which contain charnockites i.e. Sund-Ølkona, Hopen and the Austvågøy

Grey Mangerite, since the charnockites tend to have high Rb/Sr ratios.

6.4.2 Nd Isotopes

Thirty-five samples from the mangerite intrusions of the Lofoten-Vesterålen province have been analysed for Nd isotopes in this study (Table 6.11). Excluding the anomalous Flakstadøy Charnockite sample, WL307, the full range of values is as follows: Sm and Nd concentrations are in the ranges 2.6 to 19.2ppm and 13.7 to 111.3ppm, respectively; $^{147}Sm/^{144}Nd$ ranges little, between 0.095 and 0.119, and $^{143}Nd/^{144}Nd$ varies from 0.51109 to 0.51172.

The narrow ranges of $^{147}Sm/^{144}Nd$ values (compared to that of 0.086 - 0.157 for the Flakstadøy Basic Complex, Table 5.4) effectively rules out Sm-Nd whole-rock isochron dating for the mangerite intrusions. Furthermore, variations of the initial $^{143}Nd/^{144}Nd$ ratios (calculated at about 1.8 Ga) in some of the intrusions, particularly the South-West Lofoten Mangerite, are sufficiently large to further preclude whole-rock isochron dating by the Sm-Nd method for the mangerites. It is also worthy of note that the South-West Lofoten Mangerite yielded a poorly fitted Rb-Sr whole-rock isochron. The combined evidence of large variations in major element chemistry of this intrusion, together with large apparent variations in initial Sr and Nd isotopic compositions, suggest that the possibility of real heterogeneities in the Sr and Nd isotopic compositions of parent magmas is an issue worthy of serious attention.

The charnockites from the Sund, Hopen, and especially the Flakstadøy Charnockite intrusion have lower $^{147}Sm/^{144}Nd$ ratios than the mangerites themselves.

Nd evolution diagrams for the mangeritic intrusions (Figs. 6.19 -

6.25) show the small range of $^{147}Sm/^{144}Nd$ values by the sub-parallelism of the evolution lines. Only the Hopen and Raftsund intrusions show convergence of evolution lines at 1.8 - 1.7Ga, and hence most individual intrusions show a range of initial ε_{Nd} values (ages for initial Nd isotopic composition calculations have been taken from Rb-Sr geochronological data). The Nd isotopes were not homogeneous at the time of intrusion, but variations in the model ages of T_{CHUR} and T_{DM} for the individual samples have an important significance for the interpretation of petrogenesis.

At 1.8Ga, the ε_{Nd} in the depleted mantle was +3.9 (DePaolo, 1981). Hence the mangerites (ε_{Nd} at 1.8Ga = +0.5 to -7.5) cannot have formed by simple extraction of magma from depleted (or, in most cases even undepleted) mantle at 1.8Ga. ε_{Nd} at 1.8Ga for the late Archaean gneisses of the Lofoten-Vesterålen province is typically -12 (Jacobsen and Wasserburg, 1978; the present study), so assimilation of late Archaean crust, during the ascent and differentiation of magmas derived from a depleted mantle source, could account for the measured ε_{Nd} (1.8) values such as those of the mangerites. Since the ε_{Nd} (1.8) values of mangerites are all greater than ε_{Nd} (1.8) for the late Archaean gneisses (and some are greater than -1.5 which is the ε_{Nd} (1.8) value for the early Proterozoic supracrustal gneisses at Vik) the mangerites cannot have been derived by crustal anatexis without any component from the depleted mantle.

The range in T_{DM} for the mangerites is 2.06 - 2.64Ga. (This parameter, T_{DM}, is a model age indicating the time of extraction of a rock derived from the model depleted mantle of DePaolo (1981).). In the case of the mangerites of the Lofoten-Vesterålen province, however, the interpretation of a T_{DM} model age as approximating rock-formation age is manifestly inappropriate. Rather, in cases where magmas have been derived from two or more sources, e.g. mantle and lower crust, T_{DM} is

similar in significance to T_{CR}, which is used for sediment samples to determine the mean crustal residence time of the rocks in the drainage basin of a large river system (Miller and O'Nions, 1984; Goldstein et al., 1984). (T_{CR} does not refer to a specific geological event, but rather is used to average the total crustal residence age of all of the rocks sampled by the sedimentary processes acting in the river system). Similarly, in this case, T_{DM} does not denote a specific geological event e.g. magma extraction from the mantle, but rather it is a composite age, averaging the crustal residence age of all the components sampled in magmatic processes. Thus T_{DM} can be used to estimate the proportion of material derived from different sources of known age and chemical character which contribute to the magma. For the mangerites of the Lofoten-Vesterålen province, possible source materials include 2.7Ga migmatitic gneisses, and 1.8Ga depleted mantle. The contribution from the c.2.1Ga supracrustal gneisses is likely to be small as the mangerites tend to occur at the migmatite/supracrustal contact or only slightly above it (Griffin et al., 1978). Qualitatively, those mangerites with T_{DM} close to 2.7Ga, i.e. Hopen, Austvågøy Grey and Eidsfjord, have the greatest ancient crustal contributions, and those with T_{DM} close to 1.8Ga, i.e. South-West Lofoten, Steine and Hovden, have the greatest mantle-derived contributions. No T_{DM} values fall outside of the range 1.8-2.7Ga; if any did, the simple model for the derivation of the mangerites from these two principal sources would be invalidated.

The Raftsund Mangerite shows a very narrow range in T_{DM} values, (Table 6.11, Fig.6.24) from 2.42 to 2.54 Ga, which, using this model, would indicate that it is a very homogeneous body. In contrast, the South-West Lofoten Mangerite shows a wide range in T_{DM} values (2.06-2.54 Ga) (Table 6.11, Fig. 6.21) and is thus the most heterogeneous of the mangerites. This is in agreement with the major element data (Section 6.3). Sm-Nd

isotopic data will be discussed further, after the presentation of Pb isotopic data.

6.4.3 Pb Isotopes

The full range of Pb isotope ratios for the forty-seven whole-rock analyses presented in Table 6.12 is: 14.383 to 17.620 for $^{206}Pb/^{204}Pb$; 14.864 to 15.307 for $^{207}Pb/^{204}Pb$; and 34.269 to 38.044 for $^{208}Pb/^{204}Pb$.

The full range of Pb isotope ratios for the eleven alkali feldspar mineral separates presented in Table 6.12 is: 14.280 to 16.031 for $^{206}Pb/^{204}Pb$; 14.867 to 15.165 for $^{207}Pb/^{204}Pb$; and 34.279 to 36.931 for $^{208}Pb/^{204}Pb$.

The whole-rock data-points are aligned with a slope corresponding to an apparent age of about 2.6Ga, as for the Flakstadøy Basic Complex. This clearly has no direct age significance since Rb-Sr whole-rock isochron diagrams (Figs. 6.14 to 6.18) imply an age of emplacement of 1.8Ga; Nd evolution diagrams for the Hopen (Fig. 6.22) and Raftsund intrusions (Fig. 6.24) show a convergence of evolution lines at about 1.8Ga, and the mangerites intrude the c. 2.1Ga supracrustal gneisses and at least some of the 1.8Ga basic intrusions. A model has been constructed to account for the Pb isotopic characteristics of the mangerites.

The model, as with the model for the Flakstadøy Basic Complex, has a typical mantle μ_1 value, for the North Atlantic craton, of 7.9 (Moorbath and Taylor, 1981). According to the Plumbotectonics model of Zartman and Doe (1981) the mantle source at 2.7Ga has $\mu_1 = 7.7 - 8.2$. μ_1 is in any case restricted to a narrow range of possible values by the Pb isotope compositions of the alkali feldspar separates from the mangerite intrusions. Because alkali feldspars contain only insignificant amounts of uranium,

the Pb isotope ratios do not evolve very much from the time of closure of the alkali feldspar, hence the Pb isotope ratios of alkali feldspars should generally approximate to the initial Pb isotopic composition at, in this case, 1.8Ga. In this model, most of the Pb has evolved in three stages: in the mantle until 2.7Ga, in the migmatitic gneisses from 2.7Ga until 1.8Ga, and in the mangerites subsequent to 1.8Ga. Three-stage Pb isotopic evolution equations are as follows:

$$^{206}Pb/^{204}Pb = a_o + \mu_1 (e^{\lambda_1 T} - e^{\lambda_1 t_1}) + \mu_2 (e^{\lambda_1 t_1} - e^{\lambda_1 t_2}) + \mu_3 (e^{\lambda_1 t_2} - e^{\lambda_1 t_3})$$

where $a_o = 9.307$, $\lambda_1 = 0.155125 \times 10^{-9}.a^{-1}$. $T = 4.57 \times 10^9$a., and for the model outlined here, $t_1 = 2.7 \times 10^9$a., $t_2 = 1.8 \times 10^9$a., $t_3 = 0$a. Similar equations exist for $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$. μ_1, μ_2 and μ_3 are all unknowns but by assuming values for μ_1, μ_2 and μ_3 can be calculated. If either μ_2 or μ_3 values are computed as negative then manifestly the μ_1 value chosen is inappropriate. Hence the least radiogenic Pb isotopic analyses, i.e. especially the alkali feldspar separates, can be used to limit the possible values of μ_1. The lower limit for μ_1 is defined by the whole-rock $^{206}Pb/^{204}Pb$ ratio for WL204: μ_3 is negative when μ_1 is 7.8, but positive when $\mu_1 = 7.85$. The upper limit is defined by the alkali feldspar $^{206}Pb/^{204}Pb$ ratio for WL516: μ_2 is negative when μ_1 is 8.2, but positive when μ_1 is 8.1. Therefore μ_1 for the three-stage model is constrained to be between 7.85 and 8.1.

Fig. 6.26 shows the interpretational model for the mangerites in the $^{207}Pb/^{204}Pb$ v. $^{206}Pb/^{204}Pb$ diagram. Using this model, four whole-rock and four alkali feldspar analyses plot to the left of the 1.8Ga first-stage isochron (see Figs. 6.27, 6.29 and 6.30). A single-stage evolution for these rocks until 1.8Ga (i.e. in the mantle) is therefore totally impossible. The model in Fig. 6.26 shows a mantle evolution curve with $\mu_1 = 7.9$, between 4.57Ga and 0Ga. A 2.7Ga palaeo-isochron at

1.8Ga represents the position of the isochron for 2.7Ga gneisses e.g. Vikan and North-West Hinnøy at 1.8Ga, i.e. at the time of emplacement of the mangerites. The Vikan and North-West Hinnøy gneiss Pb/Pb palaeo-isochron centroids are shown, and the upper and lower limits for the spread of the North-West Hinnøy data. The 1.8Ga first-stage isochron and the Geochron are plotted from the Bulk-earth or primordial Pb isotopic composition at 4.57Ga (given by Canyon Diablo troilite Pb) to 1.8Ga and 0Ga respectively. Parallel 1.8Ga isochrons are shown for rock suites hypothetically derived from single-stage mantle with $\mu_1 = 7.9$ at 1.8Ga, and from melting of Vikan, and North-West Hinnøy gneisses of mean Pb isotope composition at 1.8Ga. The 1.8Ga isochron for rocks hypothetically produced by melting Vikan gneisses has Pb too radiogenic to provide the Pb isotope compositions of the mangerites (see Figs. 6.27 to 6.32). Most of the mangerite Pb data lies between the isochrons which would have been produced by melting mean North-West Hinnøy gneisses, and mantle with $\mu_1 = 7.9$, at 1.8Ga. All of the mangerite Pb data lie between isochrons produced by melting of the least radiogenic North-West Hinnøy gneisses, and mantle with $\mu_1 = 7.9$ at 1.8Ga. The same construction lines are shown on each of the diagrams illustrating the data for the individual mangerites (Figs. 6.27-6.32) and for the Flakstadøy Basic Complex (Fig. 5.12).

The linear array of data points can be explained using a model similar to that used to interpret the Flakstadøy Basic Complex Pb isotopic data in Section 5.4.3. The two end-members in this model have very different $^{238}U/^{204}Pb$ ratios. The mantle has $\mu_1 = 7.9$ whereas the 2.7Ga granulite-facies migmatitic gneisses have very low μ values (mean $\mu_2 = 4.0$, range 1.1 to 7.5). Those samples plotting closest to the 1.8Ga isochron for derivation from the model single-stage mantle have a large proportion of high mantle components and less of the low μ crustal component. Hence they themselves have relatively high μ values for their final stage of Pb

isotopic evolution (1.8Ga to present) and thus they evolve to fairly radiogenic Pb isotope compositions far from the 2.7Ga palaeo-isochron at 1.8Ga. Those samples plotting near to, or below the 1.8Ga isochron for derivation from average 2.7Ga crust have a greater proportion of the low μ crustal component and hence evolve to less radiogenic compositions, close to the 2.7Ga palaeoisochron.

The $^{207}Pb/^{204}Pb$ v. $^{206}Pb/^{204}Pb$ diagram for the Sund-Ølkona Mangerite (Fig. 6.27) shows two whole-rock and one alkali feldspar mineral separate data-point less radiogenic than the 1.8Ga first-stage isochron, demonstrating that these rocks must contain a component derived from U-depleted crust older than 1.8Ga. The Sund-Ølkona Mangerite also includes the most radiogenic mangeritic rock from the Lofoten-Vesterålen province: WL302, a charnockite from Sund.

Fig. 6.28 shows the Pb data for the South-West Lofoten Mangerite and also the single data-point for the Flakstadøy Charnockite. The South-West Lofoten Mangerite, on geochemical grounds the most varied of the mangerites, shows the widest range of Pb isotope compositions for any mangerite (excluding WL302 from Sund-Ølkona). The most basic samples (WL315, WL316, WL410 and WL416) have the most radiogenic Pb isotope compositions for the mangerite and plot close to the 1.8Ga isochron for rocks derived from the mantle, implying that they have a smaller crustal component than the other samples in this mangerite body. The Flakstadøy Charnockite samples, WL306, plots below the 1.8Ga "mean crustal isochron" but has evolved to a Pb isotope composition far from the 2.7Ga palaeo-isochron. This suggests that it has been derived from a low μ lower crustal source but has been enriched in U during anatexis to produce a higher μ_3.

The Hopen Mangerite/Charnockite has the least radiogenic average Pb

isotope composition of the mangerites (Fig. 6.29). One whole-rock and three alkali feldspar mineral separate data-points are less radiogenic than the 1.8Ga first-stage isochron, demanding an ancient crustal component for these rocks. The Hopen data-points plot on or below the 1.8Ga "mean crustal isochron" suggesting that the lower crustal component for these rocks was extremely depleted in uranium.

One whole-rock data-point for the Austvågøy Grey Mangerite is less radiogenic than the 1.8Ga first-stage isochron (Fig. 6.30). The data for the Austvågøy Grey Mangerite form an array parallel to the data for the Hopen intrusion but plotting higher on the diagram; this suggests that the crustal component for the Austvågøy Grey Mangerite was less strongly depleted in uranium than the crustal source of the Hopen intrusion, and consequently it may have been derived from a higher crustal level.

The Raftsund Mangerite is the largest mangeritic intrusion in the Lofoten-Vesterålen province,but it also has the smallest chemical variation. Fig. 6.31 shows that, except for the sample from Hamarøy,it shows the smallest variation in Pb isotope compositions of any mangerite. The sample from the mangerite on Hamarøy, WL101, is much more radiogenic than any of the other fourteen samples from the main body of the Raftsund Mangerite. This casts some doubt on the theory that the Hamarøy Mangerite is a part of the Raftsund Mangerite.

The two samples from the Hovden Mangerite have Pb isotopic compositions within analytical error of each other, and more radiogenic than the majority of the mangerites (Fig. 6.32).

The sources of the Pb present in a rock today can be modelled using "Plumbotectonics" of Zartman and Doe (1981). The full range of apparent

μ_1 values for the mangerites (7.54 to 7.86) are plotted on a model μ_1v.T diagram, showing the evolution of model μ_1 with time in "orogene", upper crust, lower crust and upper mantle (Fig. 6.33). The model μ_1 values cover a slightly wider range than those for the Flakstadøy Basic Complex (7.62 to 7.80; Fig. 5.13), but the diagram, nonetheless, confirms that the mangerites were derived in part from an upper mantle source and contain different proportions of material derived from a lower crustal source. The Hopen Mangerite/Charnockite has the largest proportion of lower crustal component, and the more basic samples from the South-West Lofoten Mangerite (WL315, WL316, WL410 and WL416) have the smallest lower crustal component: indeed the model necessitates a small component from the "orogene" or "Plumbotectonics" upper crust source for these samples.

6.5 Discussion and Synthesis

6.5.1 Isotope Correlation Diagrams

The Nd and Pb isotope characteristics of the mangerites have been explained using a model involving their derivation from a mixture of 2.7Ga depleted lower crustal gneisses and 1.8Ga mantle-derived magmas. Griffin et al. (1978), while excluding an anatectic origin for the mangerites on the basis of their low initial $^{87}Sr/^{86}Sr$ ratios, stated that they showed enough variation to suggest the involvement of crustal material in the genesis of some of the mangerites. There are parameters in each of the three isotope systems used (Rb-Sr, Sm-Nd, U-Th-Pb) which are sensitive indicators of the relative proportions of lower crust and upper mantle involved in magma genesis. For the Rb-Sr system, the relevant parameter is the initial $^{87}Sr/^{86}Sr$ ratio, since the Rb/Sr ratio of upper mantle is very low compared to that for the crust as a whole. However, the lower crust also has very low Rb/Sr ratios, but the presence of significant initial $^{87}Sr/^{86}Sr$ ratio variation between the mangerites suggests that this parameter may, nonetheless, be sensitive to varying material contributions

from the upper mantle and the lower crust. Rather than simply comparing the $(^{87}Sr/^{86}Sr)_o$ ratios of isochrons for the different mangerites, the apparent $(^{87}Sr/^{86}Sr)_o$ ratios for individual samples can be calculated if it can be assumed that there has been no post-intrusive disturbance to Rb-Sr systems. Such an approach may be used to resolve whether the scatter on the isochron diagram for the South-West Lofoten Mangerite (Fig. 6.15) is due to metamorphic disturbance or initial $^{87}Sr/^{86}Sr$ ratio variations, since if the latter were true a correlation might be expected with Nd and Pb parameters sensitive to crust-mantle contributions, whereas, although a correlation between Sr and Pb parameters might be expected following metamorphic disturbance since Rb and U might behave similarly, correlations with Nd parameters would be much less likely to result from metamorphic processes. Apparent initial $^{87}Sr/^{86}Sr$ ratios may be calculated for each sample by assuming their age (1.8Ga), fitting a line of slope corresponding to that age, and finding its intercept on the $^{87}Sr/^{86}Sr$ axis, i.e.:

$$(^{87}Sr/^{86}Sr)_{1.8} = (^{87}Sr/^{86}Sr)_{MEAS} - (e^{\lambda t_{1.8}} - 1)\ (^{87}Rb/^{86}Sr)_{MEAS}$$

In the Nd model (Section 6.4.2) it was postulated that when magmas had been derived from the two or more sources, the proportion of whose contributions vary, the values of the model age T_{DM} is a reflection of the proportions in which the end-members have contributed to the magma.

In the Pb model (Section 6.4.3) it was postulated that those samples which plot closer to the 1.8Ga isochron for mantle-derived rocks have a higher proportion of mantle-derived Pb than those which plot closer to the 1.8Ga isochron produced for hypothetical melting of average North-West Hinnøy amphibolite-facies gneisses. The parameter describing the relationship of the Pb compositions of the individual samples to the

two model isochrons is the apparent μ_1. This is obtained by calculating the intercept between the Geochron and a line of slope corresponding to the Sr isochron ages for the mangerite intrusion (see footnote to Table 6.12) fitted through each data point. The apparent μ_1 values calculated in this way are tabulated for the mangerites (Table 6.12) and for the basic rocks (Table 5.5).

A plot of apparent $(^{87}Sr/^{86}Sr)_{1.8}$ v. T_{DM} (Fig. 6.34) for mangerites and basic rocks shows a positive correlation as predicted by the model (the higher the T_{DM} model age, the larger the proportion of material from 2.7Ga crust; and the higher the apparent $(^{87}Sr/^{86}Sr)_{1.8}$ the greater the proportion of the magma that was derived from a crustal source).

A plot of apparent μ_1 v. T_{DM} (Fig. 6.35) for mangerites and basic rocks shows a negative correlation as predicted by the model (the higher the apparent μ_1, the greater the proportion of the high-μ component, i.e. mantle-derived magma).

Only WL514, a Hopen charnockite sample, deviates markedly from the broad positive correlation trend for apparent initial $^{87}Sr/^{86}Sr$ v. T_{DM} (Fig. 6.34), which is slightly convex towards the T_{DM} axis. This curvature implies that addition of crust to the mantle-derived end-member (low T_{DM}, low $(^{87}Sr/^{86}Sr)_{1.8}$) has a large effect on T_{DM} but little effect on $(^{87}Sr/^{86}Sr)_{1.8}$ and hence the Sr/Nd ratio for the mantle-derived end-member is greater than that for the crust-derived end-member. This is as expected since the lower crust is light rare earth element enriched relative to the upper mantle and hence has higher Nd concentrations. The Flakstadøy Basic Complex has a very narrow range of apparent $(^{87}Sr/^{86}Sr)_{1.8}$ and a wide range in T_{DM} values; both of these facts are due to the homogeneity of the basic magma at its time of crystallization.

The negative correlation between apparent μ_1 and T_{DM} (Fig. 6.35) is slightly convex away from the origin. This implies that addition of mantle-derived material to a magma of substantially lower crustal origin has a less marked effect on the Nd parameter than on the Pb parameter, suggesting that the mantle-derived component may have a rather high Pb concentration. This observation might imply that mixing of crustal melts with mantle-derived liquids may only occur after extensive earlier fractionation of the mantle-derived magma, concentrating up the less compatible element, Pb, more effectively, perhaps, than the Nd. The Flakstadøy Basic Complex does not show correlation between apparent μ_1 and T_{DM}, since its Nd isotopes were homogenized at 1.8Ga. Hence <u>its</u> T_{DM} values do <u>not</u> indicate relative crust and mantle contributions to the magma. WL306, the Flakstadøy Charnockite sample, has a very low apparent μ_1 of 7.56. (μ_1 is regarded as anomalous rather than T_{DM} since the latter correlates to $(^{87}Sr/^{86}Sr)_{1.8}$). The low apparent μ_1 for WL306 is clearly seen in the $^{207}Pb/^{204}Pb$ v. $^{206}Pb/^{204}Pb$ diagram (Fig. 6.28); this was explained in Section 6.4.3 as due to its derivation from a low-μ lower crustal source, during the melting of which the resulting magma became enriched in uranium, giving rise to the present observed radiogenic $^{206}Pb/^{204}Pb$ ratio. Alternatively the contamination could have taken place in a similar way to that of the Nûk magmas of West Greenland (Taylor et al., 1980), for which a trace-element enriched fluid phase released from dehydrating older continental crust was proposed. Mixture of mantle-derived magmas with a fluid phase enriched in uranium and unradiogenic Pb might explain the character of the Pb isotopes for the Flakstadøy Charnockite. The extraction of such fluids may have been at least partially responsible for the 1.8Ga depletion in uranium in the 2.7Ga gneisses.

WL514, a charnockite from the Hopen intrusion has anomalously high

$(^{87}Sr/^{86}Sr)_{1.8}$ and a somewhat low apparent μ_1 for its T_{DM} value. Its mantle-derived component may have been contaminated by a fluid phase extracted from a gneiss with radiogenic Sr and unradiogenic Pb.

The samples with the lowest T_{DM} values are from the South-West Lofoten Mangerite. Could the wide range in chemical composition, T_{DM} values and apparent μ_1 be correlated with the wide range in apparent $(^{87}Sr/^{86}Sr)_{1.8}$ ratios? If they are, this would show that the scatter about the Sr isochron for this mangerite (Fig. 6.15) is due to original variation in $(^{87}Sr/^{86}Sr)_{1.8}$ ratio and not to metamorphic disturbance. Fig. 6.36 shows apparent $(^{87}Sr/^{86}Sr)_{1.8}$ v. T_{DM} for the South-West Lofoten Mangerite. Ten points are plotted, five from Flakstadøy and five from Vestvågøy. The same ten samples are plotted on apparent μ_1 v. T_{DM} (Fig. 6.37) and on CaO% v. T_{DM} (Fig. 6.38) diagrams. Each diagram shows a correlation between T_{DM} and the other parameter plotted. In each diagram WL315, WL316 and WL410 plot as a separate group (squares); these are the most basic samples from the South-West Lofoten Mangerite (Tables 6.2 and 6.3) and are from the north-west coasts of the islands (Fig. 6.1); they have low T_{DM} and apparent $(^{87}Sr/^{86}Sr)_{1.8}$ and high apparent μ_1 and CaO content. The samples from the south of the islands (WL307, WL319, WL331, WL401, WL404 and WL405) plot as a group (spots) with high T_{DM} and apparent $(^{87}Sr/^{86}Sr)_{1.8}$ and low apparent μ_1 and CaO content (except for WL319 which has low $(^{87}Sr/^{86}Sr)_{1.8}$). WL416, from the north-east of Vestvågøy, near Haugen is intermediate between these two groups (triangle). All these parameters correlate in the way predicted by the model. Those samples which have a greater crustal contribution have older T_{DM} values, higher apparent $(^{87}Sr/^{86}Sr)_{1.8}$ ratios, lower apparent μ_1 values and lower CaO contents (and higher SiO_2 contents). There is even a geographical relationship to the relative crust and mantle contributions to the magma suggesting that the South-West Lofoten

Mangerite could be mapped using the above parameters. The upwarping of the Moho along an axis running west of the island chain, proposed by Griffin et al. (1978) to explain the observed gravity (Kjenes, 1970; Svela, 1971) could explain this geographical relationship to the relative contributions of crust and mantle to the magma, since this would imply that the crust is thinner beneath WL315, WL316 and WL410, and hence the mantle source is closer, than beneath the samples to the south.

These correlations between parameters from the Sr, Nd and Pb isotope systems, major element chemistry and perhaps geographical distribution, demonstrate that the scatter about the Sr isochron is not random but, rather, is due to real initial $^{87}Sr/^{86}Sr$ ratio variations, and further, that the South-West Lofoten Mangerite did not crystallize from a single, homogeneous magma.

6.5.2 Isotope Geochemistry

An estimate of the relative proportions of crust and mantle derived material in the mangerites can be made using Nd isotopes in the same way as was done for the Flakstadøy Basic Complex in Section 5.5.1 and Fig. 5.6. The equation (Puskhar et al., 1972) describing the addition of x parts of crustal material to one part of mantle-derived component reduces to:

$$x = \frac{C_{Nd}^{MD} (\varepsilon_{Nd}^{NC} - \varepsilon_{Nd}^{MD})}{C_{Nd}^{CC} (\varepsilon_{Nd}^{CC} - \varepsilon_{Nd}^{NC})} \qquad (6.1)$$

The mantle-derived material is denoted MD, CC refers to the pre-existing Archaean crust, NC to the new 1.8Ga (mangerite) crust and C is the concentration. $\varepsilon_{Nd}^{MD} = +4$ (the ε_{Nd} of the depleted mantle at 1.8Ga in DePaolo's (1981) model). $\varepsilon_{Nd}^{CC} = -10$ (from the Lewisian province of North-

West Scotland (Hamilton et al., 1979) see Chapter 5.5.1). The average ε_{Nd} at 1.8Ga for the thirty-five mangerite samples analysed for Nd isotopes is -4.8. This is very similar to the value for the Flakstadøy Basic Complex (-4.6) but this value ranges between +0.5 and -7.5 for the mangerites. Equation 6.1 reduces to:

$$\text{For } \varepsilon_{Nd}^{NC} = +0.5 \qquad x = \frac{C_{Nd}^{MD}}{C_{Nd}^{CC}} . \quad (0.3)$$

$$\text{For } \varepsilon_{Nd}^{NC} = -4.8 \qquad x = \frac{C_{Nd}^{MD}}{C_{Nd}^{CC}} . \quad (1.7)$$

$$\text{For } \varepsilon_{Nd}^{NC} = -7.5 \qquad x = \frac{C_{Nd}^{MD}}{C_{Nd}^{CC}} . \quad (4.6)$$

Table 6.13 shows the percentage of late Archaean crustal component implied for different Nd concentrations in the crustal derived component assuming a Nd concentration of 11.5ppm in the mantle-derived component (mean Nd concentration in MORB glasses, Cohen et al., 1980). According to Nd isotopes, the mangerites with the more negative ε_{Nd} values at 1.8Ga i.e. Hopen, Austvågøy Grey, Eidsfjord and Sund-Ølkona, may have a contribution by mass, from the 2.7Ga gneisses, of about 60-80%. The samples with the highest ε_{Nd} values at 1.8Ga, i.e. WL315, WL316 and WL410 from the South-West Lofoten Mangerite, have a lower crustal contribution of as little as 10%.

Fractional crystallization of the original mantle-derived magmas prior to mixing with crustal derived melts would tend progressively to raise the concentration of Nd in the residual liquid (and the concentration of Pb). Thus if mixing of crust- and mantle-derived magmas were to take

place only after prolonged crystal fractionation of the parent mantle-derived liquid, a much smaller mass contribution of mantle derivation would be required to impart the mantle Nd (and Pb) isotopic characteristics to mangerite melts. Thus the estimates of mass contribution from upper mantle sources to the mangerites-based on a Nd concentration of ca. 11.5 ppm - are likely to be a maximum limit.

Crustal involvement in the genesis of the mangerites can also be constrained using Pb isotopes in a similar way to that used in Section 5.5.1 for the Flakstadøy Basic Complex. The percentage of crustal Pb in each sample can be calculated using the Lever Rule between the 1.8Ga isochron for mantle-derived rocks and the 1.8Ga isochron produced by anatexis of mean 2.7Ga lower crust; this value is presented for forty-seven mangerite samples in Table 6.14. Those points which plot below the 1.8Ga mean crustal isochron (all Hopen samples, two Austvågøy Grey Mangerite samples, one from Sund-Ølkona and the Flakstadøy Charnockite sample) give crustal Pb values in excess of 100%. This suggests that these samples were derived from crust with extremely low μ_2 values. No sample plots below the 1.8Ga model isochron produced by anatexis of the least radiogenic North-West Hinnøy amphibolite-facies gneiss sample.

Assuming that the mangerites are derived entirely from the mantle and from melting of 2.7Ga gneisses and provided that suitable assumptions about Pb concentrations can be made, then a similar model to that used for the Qôrqut Granite of southern West Greenland (Moorbath et al., 1981) can be used for the mangerites (as it was for the Flakstadøy Basic Complex in Chapter 5.5.1). The fraction, f, of the mangerites contributed by the late Archaean migmatitic gneisses, can be calculated:

$$f = \frac{C_M(a_M - a_R)}{C_C\,(a_R - a_C) + C_M\,(a_M - a_R)}$$

where a refers to $^{206}Pb/^{204}Pb$ ratios at 1.8Ga, and C refers to ^{204}Pb concentrations. The subscripts M,C,R refer to mantle-derived, crust-derived and the resulting mangeritic magmas respectively. The percentage crustal contributions by mass to each of the forty-seven mangerite samples is tabulated in Table 6.14 using the mean concentration (total Pb = 20.2ppm, ^{204}Pb = 0.28ppm) from the North-West Hinnøy amphibolite-facies gneisses as the concentration for the crustal component. The crustal contribution by mass is calculated for four Pb concentrations in the mantle-derived magma. As well as the three values used in Chapter 5.5.1 for the Flakstadøy Basic Complex, a total Pb concentration of 20ppm is used. This latter figure is a hypothetical Pb concentration for an extremely Pb-enriched magma, which has evolved from an original basaltic parent liquid by extensive crystal fractionation.

The samples which plot below the 1.8Ga isochron for derivation from average 2.7Ga crust have crustal Pb percentages in excess of 100 and thus do not produce meaningful results in the calculations but clearly for a given crustal μ_2, these samples (all of Hopen, WL202 from Ølkona, WL306 from the Flakstadøy Charnockite and WL522 and WL525 from the Austvågøy Grey Mangerite) have the greatest crustal contribution by mass; also they are derived from crust with a very low μ_2 value. For any given Pb concentration in the mantle component, the greatest variation in crustal contribution by mass is found in the South-West Lofoten Mangerite. Relatively little variation is found in the Raftsund Mangerite and the Austvågøy Grey Mangerite. As was argued for the Nd isotopes, fractional crystallization of the original mantle-derived magmas prior to mixing with crustal melts would progressively concentrate Pb in the residual liquid. The more fractionated the basaltic magma at the time of mixing, the smaller the mantle-derived mass contribution required to impart the mantle Pb isotope characteristics to the mangerite melts. The column in Table 6.14 for a Pb concentration of 20ppm in the mantle-

derived component (i.e. equal to the Pb concentration in the crust-derived component) shows that if mixing took place with a magma extremely enriched in Pb after extensive crystal fractionation, then all mangerite samples (except WL315, WL316, WL410 and WL416) would need a contribution of less than 50% by mass from the mantle source.

6.5.3 Evolution of P-T Conditions

The pressure and temperature of crystallization of the Hopen and Raftsund intrusions have been determined using magnetite and ilmenite (for temperature), and olivine and orthopyroxene (for pressure). The Hopen Mangerite/Charnockite crystallized at 1000 ± 25°C and 12 ± 1kb (Ormaasen, 1976) and the Raftsund Mangerite crystallized at 1050°C and 11 ± 1kb (Krogh, 1977).

The anhydrous reactions in which thin coronas of garnet, usually with quartz, form around pyroxenes and opaques adjacent to feldspars have been interpreted by Griffin and Heier (1969) as retrograde reactions occurring with falling T at constant P. Those rocks which contain biotite-quartz-hornblende pseudomorphs after pyroxene rimmed by garnets have also suffered an addition of water.

6.6 Conclusions

1. The mangeritic intrusions occupy almost half of the Lofoten-Vesterålen province, and hence their emplacement is the most important rock-forming event recognized in the region. They consist of coarsely crystalline, massive rocks. Mesoperthite and plagioclase typically constitute over 80% of the rocks. Quartz is absent from many of the mangerites, but may total up to 30% of the mode. Clinopyroxene and orthopyroxene are the main mafic silicates; olivine is common in some parts of the Raftsund Mangerite. Hornblende and biotite are primary phases in some rocks.

Opaques are present in accessory to minor amounts in all mangerites. Accessory phases include apatite, zircon and rutile.

2. Some of the mangerites show magmatic-sedimentary features and well-developed cumulate textures. The Raftsund Mangerite can be subdivided on the basis of its primary mafic silicate assemblage. The Hopen intrusion has a layered gabbroic margin, rich in apatite, adjacent to the basement gneisses.

3. The mangerites are intrusive into the late Archaean migmatites, the early Proterozoic supracrustal gneisses and at least some of the mid-Proterozoic basic intrusions. The age relationships between the South-West Lofoten Mangerite and the Flakstadøy Basic Complex and the Hovden Mangerite and the Eidet-Hovden Gabbro are uncertain, but in each case mangeritic veins have been found intruding the basic rocks suggesting that the mangerites post-date the basic intrusions.

4. Xenoliths, some of which are very similar to the dioritic gneiss on Moskenesøy are present in all of the major mangerite bodies and abundant in the intrusions at Sund, Hopen and Engenykken. Most are well-rounded, some are zoned and some have reaction rims; all of the xenoliths are more mafic than the host mangerite, suggesting they may represent restites after the extraction of leucocratic melts.

5. The mangerite series has a wide range of SiO_2 contents (51-77%). The South-West Lofoten Mangerite shows the widest range in chemical composition within a single intrusion and the Raftsund Mangerite has the smallest range. Some of the mangerites are extremely Fe-enriched, especially Hopen and Raftsund. The mangerites have high K_2O and most are shoshonitic in composition. By comparison to Phanerozoic andesites, the chemistry of the mangerites is consistent with formation in the orogenic zone of a continental region. Ewart (1982) concluded that

concentrations in potassium, similar to that in the mangerites of the Lofoten-Vesterålen province could only be caused by reworking of continental crust by repeated intrusion, assimilation and anatexis.

6. The evolution of the mangerite magmas can be modelled using variation diagrams to compare the composition of the mangerites to the compositions of possible end-members and possible early fractionating phases. These diagrams are broadly consistent with the formation of the mangerite magmas from anatexis of average 2.7Ga and 2.1Ga gneisses, and also of more acid gneisses, and from a mantle-derived basic magma. The magmas formed from these end-members subsequently evolved by fractional crystallization of pyroxenes, olivine, apatite, ilmenite and magnetite. These minerals are found concentrated in the marginal gabbro to the Hopen intrusion.

7. Rb-Sr whole-rock isochron ages for the mangerites are in the range 1.73 to 1.81 Ga. An older imprecise age was obtained for the South-West Lofoten Mangerite due to the wide scatter of the data-points. Initial $^{87}Sr/^{86}Sr$ ratios range from 0.7032 to 0.7048 implying that the mangerites were derived from either upper mantle or granulite-facies depleted lower crust.

8. The narrow range in Sm/Nd ratios effectively rules out Sm-Nd whole-rock isochron dating for the mangerites.

9. The ε_{Nd} values at 1.8Ga, and unradiogenic Pb isotope compositions, preclude an origin for the mangerites by crystallization from a magma derived solely from the mantle at 1.8Ga.

10. Models for the Nd and Pb isotope evolution of the mangerites are consistent with their derivation from anatexis of 2.7Ga lower crust and from 1.8Ga mantle-derived magmas, mixed in varying proportions. The "Plumbotectonics" model supports this conclusion. Both T_{DM} and apparent

μ_1 can be used to estimate the proportion of material derived from mantle and crustal sources. The Hopen and Austvågøy Grey mangerites have the largest ancient crustal contribution whereas the South-West Lofoten Mangerite has the smallest ancient crustal contribution and also shows the widest range in component proportions.

11. The least radiogenic Pb isotopic compositions in the mangerite suite set important constraints on the μ_1 value of the mantle-source of the late Archaean crust which contributes the crustal-derived component of the mangerites: μ_1 values must be in the range 7.85 to 8.1.

12. Correlations of isotopic parameters derived from Rb-Sr, Sm-Nd, and U-Pb systems to express the relative proportions of crust and mantle contributions to the mangerites indicate that the combined crustal-mantle source-model for the mangerites is fully self-consistent. The trend in the μ_1 v. T_{DM} plot suggests that mixing of crustal melts with mantle-derived liquids may only occur after extensive earlier fractionation of the mantle-derived magma, causing concentration of Pb.

13. The wide variations in isotope ratios in three systems, and in chemistry, for the South-West Lofoten Mangerite, are all correlated. This strongly suggests that the scatter about the Rb-Sr isochron for this intrusion is due to initial $^{87}Sr/^{86}Sr$ ratio variation rather than to later metamorphic disturbance. Those samples from the South-West Lofoten Mangerite with the largest mantle contribution were taken from closer to the axis of the upwarp in the Moho than those with a small mantle contribution (i.e. from closer to the mantle source).

14. Mixing calculations for Nd suggest that the maximum contribution from 2.7Ga gneisses to the mangerites, is 60-80% by mass. Fractional crystallization of mantle-derived magmas would progressively raise Nd concentrations in the residual magmas. Hence mixing of crust- and mantle-

derived magmas after fractionation of the mantle-derived magma would require a much smaller contribution by mass from the mantle-derived component. Fractional crystallization would also concentrate Pb in the residual magma (isotope correlation diagrams suggest that this is, in fact, the case), and reduce the required mass contribution from the mantle-derived component. If Pb was concentrated to as much as 20ppm in the mantle-derived magma prior to mixing, all samples (except three from the South-West Lofoten Mangerite) would require a contribution by mass from a lower crustal source of over 50%.

15. The mangerites crystallized at about 1000°C and 11kb with granulite-facies mineral assemblages. Retrograde reactions have taken place; primarily in response to falling temperature (pressure remained fairly constant), and the addition of water.

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CHAPTER 7. THE MANGERITIC AND CHARNOCKITIC ANATECTIC MELTS

7.1 Introduction

In Chapter 6 it was proposed, on the basis of Sr, Nd and Pb isotopes, major element geochemistry, and the presence of abundant xenoliths, that ancient lower crust was the largest contribution by mass to mangerites. The only evidence to date of in situ anatexis of the gneisses of the Lofoten-Vesterålen province was presented by Green and Jorde (1971). They reported monzonitic veins and dykes up to 2m wide on Moskenesøy cutting the veined and layered (i.e. early Proterozoic supracrustal) gneisses especially between the anorthosite contact and Å, and only rarely intruding the monzonitic-dioritic (i.e. late Archaean migmatitic) gneisses. Green and Jorde (1971) interpreted these veins and dykes as partial melts of the gneiss. They suggested that the dioritic and quartz dioritic gneisses of Moskenesøy are of a suitable composition to have formed the Moskenesøy Anorthosite as a residuum by deep crustal anatexis, and that the monzonitic dykes and veins represent a small part of the complementary more acid melt. They stated that the bulk of this more acid magma probably intruded at higher levels.

For this study further evidence of crustalanatexis has been sought, as well as the answer to three questions:

(i) do the monzonitic veins and dykes represent a partial melt of the late Archaean and/or early Proterozoic gneisses?

(ii) if so, what caused the anatexis?

(iii) could these veins and dykes represent the deepest remnants of a large magma body which intruded at higher structural levels in the crust, equivalent to the mangerites present to the east?

7.2 Fieldwork and Petrography

This section describes the samples that have been used in the geochemical and isotopic studies of the anatectic melts, and the localities from which they have been taken. As well as those from Å, Moskenesøy first described by Green and Jorde (1971), veins and dykes of mangeritic and charnockitic composition have been described, for the first time, from:

(i) Hamnøy, Moskenesøy

(ii) Ølkona, Moskenesøy

(iii) Nesland, Flakstadøy

(iv) Våje and Borge, Vestvågøy

(v) Festvåg, Austvågøy

Some veins of mangeritic and charnockitic composition are clearly radiating from a mangerite intrusion e.g. those around the Flakstadøy Charnockite, and those in the gabbro around the Hovden Mangerite, Langøy: but at most of the localities listed above, clear field evidence shows that the veins and dykes are derived from the gneisses rather than intruded into them.

7.2.1 Å, Moskenesøy

The early Proterozoic supracrustal gneisses around Å (Fig. 4.1) are intruded by reddish-brown sheets up to 2m wide. The best examples seen in this study are immediately north of Å camp-site. Those in Plate 7.1 are discordant to the foliation of the gneisses at an angle of 25°, but they have a wide range of orientations; those at Å predominantly strike NW-SE. WL230 was collected from one of the sheets in Plate 7.1. It consists of about 80% feldspar which is perthitic on a very fine scale. Opaques, ortho- and clinopyroxene are the main mafic

phases. Orthopyroxene overgrows clinopyroxene, and opaques are rimmed by garnet. Apatite is quite abundant and is up to 2mm long. Accessory biotite, muscovite and spinel are also present. At the north end of Å, similar sheets are present. They are brick-red in colour and contain dark green clinopyroxene crystals up to 3cm long. In thin-section (WL232) the perthitic feldspars are coloured by abundant minute inclusions. Similar dykes have been observed south of Åvann, the lake to the west of Å, where there are two sets of coarse-grained dykes, one cross-cutting the other. There is also a later set of veins north of Å camp-site, up to 30cm wide, which cross-cut the reddish-brown sheets. They are very coarse-grained, with feldspars up to 5cm across, and are pale grey.

7.2.2 Hamnøy Bridge, Moskenesøy

The early Proterozoic gneisses from the peninsula at the northern end of the Reine-Hamnøy bridge are described in Section 4.2.2. This is the locality at which Green and Jorde (1971) described the gneiss as, "frequently containing monzonitic layers". WL255 is a sample of this gneiss. Green, fine-grained (less than 1mm) dioritic gneiss makes up 60-70% of the rock; the remainder consists of coarser-grained (2mm), pink monzonitic layers up to 3mm wide. Plate 4.8 shows coarsely crystalline layers of monzonitic composition up to 30mm wide, concordant with the foliation of the host dioritic gneiss. These are regarded as low melting products of anatexis which have moved no further than a few centimetres from their parent material.

The development of material representing melts appears to be intimately associated with gabbroic material (Plates 4.9 and 4.10). Is the gabbro a restite after the development of the monzonitic melts or has it been emplaced into the gneisses leading to their anatexis?

Plate 4.10 shows thin mafic veins broken up by the development of a leucocratic melt in the supracrustal gneisses. These gabbroic veins have sharp margins against the gneiss and they are irregularly oriented and hence are not related to the foliation of the gneiss. This again suggests that the gabbroic veins are a cause of the melting rather than a result of it. More substantial melt veins (e.g. Plate 7.2) have thin, discontinuous mafic selvages and are discordant to the gneiss foliation. Plates 7.3 and 7.4 are of a polished slab (WL247) through fine-grained, quartzo-feldspathic gneiss which includes mafic lenses, 1cm across, with coarse-grained pinkish-grey melts. The mafic layers consist of 60-95% clinopyroxene with plagioclase and opaques; the clinopyroxene is coloured reddish-brown by minute inclusions. The melts consist of perthites up to 12mm across and up to 95% by volume with minor ortho- and clinopyroxene, opaques and biotite, and accessory apatite and calcite.

A coarse-grained, red-brown melt, 5m wide, and concordant to the gneiss foliation for about 6m along strike, becomes discordant and widens toat least 11m. The foliation of a 30 x 10cm gneiss xenolith in the melt is parallel to that of the adjacent layered gneiss, which contains numerous monzonite layers. Plates 7.5 and 7.6 show a pinkish-grey melt, up to 2m wide which is 8m long, interfingering with the gneiss at either end. Thin slivers of gneiss are included within the melt.

7.2.3 Ølkona, Moskenesøy

The supracrustal gneisses from this locality at Ølkona, including a 4m wide marble band, have been described in Section 4.2.1. The gneisses are intensely deformed (Plate 4.3) at this locality, and some melt development has taken place, represented by leucocratic material between disrupted and rotated gneiss blocks (Plate 4.4). About 50m to the west of the northern end of the marble layer, leucocratic melts have developed along

the gneiss foliae (Plate 7.7). These melts are up to 0.3m wide while concordant with the gneiss foliation, but in Plate 7.7, they also cross-cut the gneiss and coalesce with one another. The melts weather white but are greenish-brown on a fresh surface. A sample of this melt, WL212, consists of between 80 and 90% mesoperthite, of grain size from less than 1mm to 10mm. Plagioclase (An_{40}), biotite and opaques are minor phases, the latter two phases rimmed by garnet, indicating limited retrogression has taken place. Apatite and zircon are present as accessory phases.

Another leucocratic melt, about 2m wide, cross-cuts the gneisses including the marble layer and its pyroxenite mantle (Section 4.2.1). It contains rounded blocks (but with some angular corners) of pyroxenite up to 23 x 12cm (Plate 7.8) which resemble the rocks of the mantle to the marble. WL216 is a sample of this leucocratic melt and contains 80 to 90% plagioclase and orthoclase, some of which is perthitic, up to 10mm across. Microfractures filled with a carbonate phase (from the marble) are common. Plate 7.9 shows a sample (WL217) of the leucocratic melt which contains a pyroxenite xenolith. Orthoclase is coloured pink by minute inclusions, but less strongly than the melts around Å. Green clinopyroxene, up to 10mm, is the only mafic phase in this leucocratic melt. The pyroxenite xenolith in Plate 7.9 is largely monomineralic pale green clinopyroxene, typically 3mm, but it also contains a biotite and plagioclase-bearing band and very rare calcite and opaque phases. The 15mm wide fine-grained layer of the xenolith shown in Plate 7.9 consists of very pale green clinopyroxene, plagioclase and orthoclase with minor amounts of biotite and opaques and accessory apatite, with a grain size of about 1mm.

The supracrustal gneisses on the headland 1km to the south-west

(Fig. 4.1) have a similar appearance to those described above: highly deformed and broken up by the development of leucocratic melts. The pale orange weathering rock in Plage 4.7 (WL259) consists of 80% perthites, up to 5mm. Clinopyroxene, orthopyroxene and opaques are the main mafic phases present. Apatite and zircon are accessory phases. The white weathering rock in Plate 4.7 is a norite (WL258) which disrupts the dark green pyroxenite layer. The monzonite and norite may represent a partial melt and its complementary restite; the dark green pyroxenite may be the residuum from a gabbro intrusion which caused the partial melting.

7.2.4 Nesland, Flakstadøy

The supracrustal gneisses on the eastern side of Skjelford (Fig. 4.2) are cut by two generations of coarse, granular, feldspar-rich sheets (Plate 7.10), the earlier of which is very similar to the red monzonite sheets at Å, Moskenesøy. The earlier sheets are up to 2m wide, discordant to the gneiss foliation at a small angle, and trend north-south, parallel to the gneiss-mangerite contact a few hundred metres to the east. They weather pinkish-orange but are greenish-brown on a fresh surface. A sample of one of these sheets, WL340, has a grain size typically less than 1mm but some mesoperthite and plagioclase crystals are up to 6mm. Mafic phases comprise about 25% of the rock and consist of orthopyroxene and opaques with accessory clinopyroxene. Accessory apatite occurs as grains up to 1mm across. The gneisses and the red sheets of mangeritic composition are cross-cut by a 0.2m wide dykelet of charnockitic composition (lower part of Plate 7.10). WL341 is a sample of this dykelet. It consists of mesoperthite and quartz present in equal amounts and up to 10mm across. Other phases (opaques, orthopyroxene, zircon and rutile) total about 1%.

No in situ anatexis is present at this locality and the charnockite dykelet may possibly have intruded from the Flakstadøy Mangerite, but the similarity of the earlier mangerite sheets to those around Å, Moskenesøy, and their orientation parallel to the gneiss-mangerite content suggests that they are probably not derived from the Flakstadøy Mangerite, but that rather they are melts extracted from gneisses at deeper levels by anatexis.

7.2.5. Våje and Borge, Vestvågøy

Leucocratic pegmatites are emplaced into the gabbros at Våje and Borge on Vestvågøy. They are probably derived from the surrounding South-West Lofoten Mangerite rather than being due to a discrete anatectic event.

The gabbro near Våje, in which gabbro-eclogite transitions associated with shear deformation have been recognised (described in Section 5.2.3), is intruded by leucocratic pegmatites, which are also affected by the shear deformation and hence pre-date it. A sample of this pegmatite, WL441, consists of about 95% orthoclase, some of which is perthitic, up to 20mm across. Quartz grains are up to 3mm, but most occurs in fine-grained patches along feldspar grain boundaries and in small shear zones with accessory orthopyroxene and garnet.

Two generations of leucocratic dykes intrude the gabbro in the quarry near Borge (Fig. 5.2, Plate 7.11). The earlier, 2m wide, sub-vertical dyke is cross-cut by two 0.2m wide, sub-horizontal sills which have razor-sharp contacts with the gabbro and the earlier dyke. A sample, WL467, from the lower sill in Plate 7.11 consists of about 80% orthoclase and quartz. Orthopyroxenes, up to 10mm, and rimmed by hornblende and garnet, and a single 7mm opaque grain, are the main mafic phases. A carbonate

phase is present in small shear zones. Euhedral zircons, up to 2mm, are quite abundant, as are allanites up to 3mm, which are metamict and surrounded by cracks radiating into the feldspars.

7.2.6 Festvåg, Austvågøy

The one example of possible anatexis seen in the late Archaean migmatites occurs at Festvåg in South-West Austvågøy (Fig. 3.2) (Section 3.2). There, thin amphibolite stringers, typically 1-2cm thick, cut the leucocratic migmatites. Adjacent to these the migmatites have recrystallized into coarse-grained granular pods.

7.3 Geochemistry

Chemical analyses for eight of the mangeritic and charnockitic anatectic melts described in Section 7.2 are presented in Table 7.1. Two are from Å and two from Ølkona on Moskenesøy. The two generations of melts near Nesland, Flakstadøy, have both been sampled and analysed, as have leucocratic melts which intrude the gabbros at Våje and Borge on Vestvågøy. They show a wide range in composition . The melts which cut the Vestvågøy gabbros, and the later dykelet at Nesland, are charnockites. The most basic melts are the petrographically similar red mangeritic sheets from Å (WL230, Plate 7.1) and Nesland (WL340, Plate 7.10). These samples have higher Fe, Mg, Ca, Mn and Ti and lower K than the other melts. K_2O content is greater than 6% in all melt samples except WL230 and WL340 and is nearly 13% in WL216 from Ølkona, Moskenesøy. The quote from Ewart (1982) in Section 6.3 is especially appropriate for these veins and dykes; he could only envisage such concentration of potassium by a process of reworking of continental crust by repeated intrusion, assimilation and anatexis.

How does the range in chemistry for the anatectic melts compare to

that for the mangerite intrusions?

(i) SiO_2 = 54-75% (for mangerites SiO_2 = 51-77%)

(ii) K_2O = 3.3-12.8% (for mangerites K_2O = 1.6-9.0%)

(iii) Na_2O/K_2O = 0.12-1.8 (for mangerites Na_2O/K_2O = 0.37-2, except for WL202, WL314, WL410 which have $Na_2O/K_2O > 2$)

(iv) Fe_2O_3/MgO = 1.3-18 (for mangerites Fe_2O_3/MgO = 1.5-63)

(v) CaO = 0.5-4.8% (for mangerites CaO = 0.8-7.8%)

The chemistry of the mangerites and the melts are quite similar, but the melts scatter regardless of the fractionation trends observed in variation diagrams for the mangerites (Section 6.3). This is as expected because the anatectic dykes and veins represent melts from restricted volumes of gneisses and hence will have a wide range of compositions which are not interrelated whereas, although the mangerites also show a wide range in composition, they are related by fractionation from magma composed largely of coalesced anatectic melts which partially homogenised in the magma.

7.4 Isotopic Results

7.4.1 Sr Isotopes

Sr isotope data for the mangeritic and charnockitic anatectic melts of the Lofoten-Vesterålen province are tabulated in Table 7.2. The same eight samples have been analysed for Sr isotopes as for major element compositions. They have a range in Rb and Sr concentrations of 15 to 274ppm and 48 to 802ppm respectively. The range in $^{87}Rb/^{86}Sr$ ratios is 0.0584 - 1.044 and for $^{87}Sr/^{86}Sr$ is 0.70618 - 0.73027, although the Rb and Sr concentrations for WL467 suggest that it would probably extend these ranges higher. The Rb concentration of 274ppm for WL467, the charnockite vein from the Borge gabbro is higher than for any mangerite

analysed: the Flakstadøy Charnockite sample, WL306, has the highest Rb concentration of the mangerite suite, at 163ppm. The apparent initial $^{87}Sr/^{86}Sr$ ratios at 1.8Ga will be used in Section 7.5.1 in isotope correlation diagrams.

7.4.2 Nd Isotopes

Seven samples of anatectic melts have been analysed for Nd isotopes in this study (Table 7.3). They are the same samples as analysed for Sr isotopes and major element compositions, excluding WL441, the charnockite vein from the Våje gabbro. The range in Sm and Nd concentrations is 1.0 to 26.6ppm and 8.7 to 204ppm respectively; $^{147}Sm/^{144}Nd$ ranges between 0.039 and 0.112, and $^{143}Nd/^{144}Nd$ varies from 0.51055 to 0.51136. WL341 and WL467 (charnockite veins from Nesland and Borge, respectively) have lower $^{147}Sm/^{144}Nd$ ratios than any of the mangerite intrusions except for the Flakstadøy Charnockite. WL341 has lower concentrations of Sm and Nd than any mangerite sample but quartz and feldspars constitute about 99% of the rock and they contain low concentrations of rare-earth elements. WL212, a mangerite melt from Ølkona, Moskenesøy, has a much lower $^{147}Sm/^{144}Nd$ ratio than even the Flakstadøy Charnockite. WL467, the charnockite from the Borge gabbro, has higher Sm and Nd concentrations than any sample from a mangerite intrusion. This is not surprising since the rock contains abundant allanite and zircon, both of which can contain substantial concentrations of rare earth elements.

Nd evolution diagrams for the anatectic veins and dykes (Figs. 7.1 and 7.2) show the evolution lines for the veins and dykes from Moskenesøy superimposed on the range of Nd evolution lines for the Sund-Ølkona Mangerite (from Fig. 6.19), and the evolution lines for the veins and dykes from Flakstadøy and Vestvågøy are superimposed on the range of Nd evolution lines for the South-West Lofoten Mangerite (from Fig. 6.21).

The Nd evolution line for the Flakstadøy Charnockite (from Fig. 6.20) is also shown on the latter diagram (Fig. 7.2).

Three of the four Moskenesøy anatectic melts have Nd evolution lines parallel to the Nd evolution lines for the Sund-Ølkona Mangerite (Fig. 7.1). They have a wider range in ε_{Nd} values at 0Ga and 1.8Ga, and a greater spread of T_{DM} values than the Sund-Ølkona Mangerite: this is to be expected since veins derived from localised areas of gneiss would have greater isotopic diversity than larger mangerite intrusions formed by the coalescing of melts from many anatectic veins and dykes. WL216, from Ølkona, has an older T_{DM} than WL230 and WL232 from Å, suggesting that it has a larger component of its Nd contributed by the 2.7Ga gneisses than the veins and dykes from Å. If anatexis was caused by the emplacement of basic rocks then the melts developed nearer to the basic intrusion (in this case the anorthosite south of Å) might be expected to have a larger contribution from the mantle-derived basic rocks than those developed further away.

The mangerite vein from Ølkona, WL212 (Plate 7.7), has the lowest $^{147}Sm/^{144}Nd$ ratio of any mangeritic rock or anatectic vein analysed. This requires retention of the heavy rare earth elements in minerals such as garnet, zircon or hornblende in the residuum gneiss after the extraction of the melt. This melt from Ølkona is the most convincingly in-situ anatectic melt analysed. It has a T_{DM} model age of 2.16 Ga (Table 7.3) which is very similar to the T_{DM} model ages for early Proterozoic supracrustal gneisses from Vik, Vestvågøy (2.08 and 2.11 Ga, Table 4.3). Hence Nd isotope compositions support the field observation that this anatectic melt from Ølkona, WL212, could be totally derived from the early Proterozoic supracrustal gneisses in which it occurs.

The Nd isotopic evolution line for WL340, the mangerite dyke from Nesland, is parallel to the evolution lines for the South-West Lofoten Mangerite (Fig. 7.2) but just outside the range for that intrusion. It is difficult to see how the Nd evolution line for this dyke could lie outside the full range for the South-West Lofoten Mangerite if it were an apophysis from it, whereas small melts with a wide range of Nd isotope compositions could have been combined into the South-West Lofoten Mangerite to produce the present range of Nd evolution lines. The charnockites intruded into the gneisses at Nesland (WL341) and the gabbro at Borge (WL467), have substantially more negative ε_{Nd} values than the South-West Lofoten Mangerite both at 0Ga and at 1.8Ga, but their Nd isotopic evolution lines are close and parallel to that for the Flakstadøy Charnockite. The Rb and Sr concentrations also suggested a strong similarity between the Flakstadøy Charnockite and the charnockite intruding the gabbro at Borge, Vestvågøy. Sm-Nd isotopic data will be discussed further, after the presentation of Pb isotopic data.

7.4.3 Pb Isotopes

The range of Pb isotope ratios for the five whole-rock analyses (two each from Ølkona and Nesland, one from Å) presented in Table 7.4 is: 14.825 to 16.421 for $^{206}Pb/^{204}Pb$; 15.015 to 15.295 for $^{207}Pb/^{204}Pb$; and 34.719 to 36.289 for $^{208}Pb/^{204}Pb$. These data are plotted on a $^{207}Pb/^{204}Pb$ v. $^{206}Pb/^{204}Pb$ diagram in Fig. 7.3 using the same interpretational model as used for the mangerites (Fig. 6.26). All of the points (except WL341, the charnockite vein from Nesland) plot just above the 1.8Ga model isochron for derivation from average lower crust, and just to the right of the 1.8Ga first-stage isochron. Any finite μ at the present day for the monzonite sheet at Å, WL230, would imply that at 1.8Ga this data-point plotted to the left of this isochron, and hence a single-stage Pb isotopic evolution model for this rock up to

1.8Ga (i.e. in the mantle) is not possible. WL341, the charnockite from Nesland, has a comparatively radiogenic Pb isotope composition, slightly above the 1.8Ga model isochron for rocks derived from mantle with $\mu_1 = 7.9$. Apart from this sample, the anatectic melts have Nd and Pb isotopic compositions consistent with their formation from different proportions of material derived from 2.7Ga lower crust and material derived from mantle with $\mu_1 = 7.9$ at 1.8Ga.

The relative importance of different possible sources of the Pb present in a rock today can be estimated using the "Plumbotectonics" model of Zartman and Doe (1981). The five apparent μ_1 values for the anatectic melts are plotted on the model μ_1 v. T diagram, together with the evolution curves for model μ_1 through time in "orogene", upper crust, lower crust and upper mantle (Fig. 7.4). The diagram confirms that the mangeritic anatectic melts (WL212, WL216, WL230, WL340) were derived from lower crustal and upper mantle sources. The charnockite vein from Nesland, WL341, apparently requires a component from the "orogene" or "Plumbotectonics" upper crust source added to an upper mantle component.

7.5 Discussion and Synthesis

7.5.1 Isotope Correlation Diagrams

The data for the anatectic melts are plotted on $(^{87}Sr/^{86}Sr)_{1.8}$ v. T_{DM} (Fig. 7.5) and apparent μ_1 v. T_{DM} (Fig. 7.6) diagrams, and are thus compared to the data for the mangerites, the trends for which have been described in Section 6.5.1. The melts do not show the same, clear correlations between the Rb-Sr, Sm-Nd and U-Pb isotopic systems as the mangerites. This is probably because the melts are derived from small volumes of gneisses and hence have isotopes characteristic of those gneisses. In a large magma body, which may effectively average out local variations

in isotopic compositions due to coalescing and mixing of many locally generated anatectic melts, integrated together in a single magma chamber, extreme variations in isotopic characteristics would be much reduced.

7.5.2 Isotope Geochemistry

The relative proportions of crust and mantle derived material in the melts can be estimated using Nd isotopes in the same way as was done for the Flakstadøy Basic Complex in Section 5.5.1 and Table 5.6 and for the mangeritic rocks in Section 6.5.2 and Table 6.13.

The occurrence of the anatectic melt veins and dykes in the early Proterozoic supracrustal gneisses strongly suggests that they may have made substantial contributions by mass to the anatectic melts (especially WL212, Section 7.4.2). The isotopic character of this juvenile crustal component will be similar to that for the mantle-derived component. Hence the "mantle-derived" contribution in the melts is more likely to represent the juvenile crustal contribution with or without a mantle-derived contribution.

The equation and parameters used are as detailed in Section 6.5.2 except for the ε_{Nd} of the "new crust" at 1.8Ga. The average ε_{Nd} at 1.8Ga for the seven anatectic melts analysed is -5.9. The extreme values are -3.1 (for WL232 from Å) and -8.6 (for WL467 from Borge). For these values, equation 6.1 reduces to:

$$\text{For } \varepsilon_{Nd}^{NC} = -3.1 \qquad x = \frac{C_{Nd}^{MD}}{C_{Nd}^{CC}} . \quad (1.0)$$

$$\text{For } \varepsilon_{Nd}^{NC} = -5.9 \qquad x = \frac{C_{Nd}^{MD}}{C_{Nd}^{CC}} . \quad (2.4)$$

$$\text{For } \varepsilon_{Nd}^{NC} = -8.6 \qquad x = \frac{C_{Nd}^{MD}}{C_{Nd}^{CC}} \quad (9.0)$$

Table 7.5 shows the percentage of late Archaean crustal mass contribution to the melts implied for different Nd concentrations in the crustal derived component assuming a Nd concentration of 11.5ppm in the mantle-derived (or juvenile crust-derived) component (mean Nd concentration in MORB glasses, Cohen et al., 1980). According to Nd isotopic evidence, the average anatectic melt may have a contribution, by mass from the 2.7Ga gneisses of about 40-65%; the charnockite in the Borge gabbro may have a crustal contribution of 70-90%. As was argued in Section 6.5.2, prolonged fractional crystallization of the mantle-derived magma would progressively concentrate Nd in the residual magma. Mixing of crust-derived material with a Nd-enriched residual magma would require a much smaller mass contribution of mantle derivation to impart the mantle Nd (and similarly, Pb) isotopic characteristics to the anatectic melts. Also, part of the mantle-derived character of the Nd isotope composition of the anatectic melts is very probably derived from the juvenile crust. Thus the percentage mass contributions of crust-derived material presented in Table 7.5 - based on an upper mantle source Nd concentration of 11.5ppm - are likely to be a <u>minimum</u> limit.

Pb isotopes can be used to constrain the crustal involvement in the genesis of the anatectic melts in a similar way to that used in Section 5.5.1 for the Flakstadøy Basic Complex and in Section 6.5.2 for the mangeritic rocks. The percentage of crustal Pb in each of the five anatectic veins and dykes analysed has been calculated using the lever rule between the 1.8Ga model isochrons for mantle-derived rocks and for anatexis of mean 2.7Ga lower crust. These values are presented in Table 7.6. The mangeritic veins and dykes from Å, Ølkona

and Nesland have between 50 and 90% ancient crustal Pb. The charnockite vein from Nesland plots slightly above the 1.8Ga model isochron for mantle-derived rocks and hence has a negative concentration of (lower) crustal Pb. The fraction of the anatectic melts contributed by the 2.7Ga lower crust can be calculated using the model of Moorbath et al. (1981) (see Sections 5.5.1 and 6.5.2). The crustal contribution by mass is calculated using the same four Pb concentrations in mantle-derived (or juvenile crust-derived) magma as in Section 6.5.2. The μ_1 v. T_{DM} correlation diagram for the mangerites (Section 6.5.1, Fig. 6.35) suggested that crustal melts only mixed with mantle-derived magmas after the latter had undergone extensive fractionation, causing concentration of Pb. Table 7.6 shows that with a concentration of 20ppm in the residual mantle-derived magma at the time of mixing, the mass contribution of the crustal component to the mangeritic veins and dykes would be 50-90% (compared to 30-75% if the Pb concentration in the mantle-derived magma was similar to that in Andean basalt (7ppm)).

7.5.3 The Origin of the Mangerites

Evidence from fieldwork (xenoliths; anatectic veins and dykes), from geochemistry (gneisses more basic at deeper structural levels; the importance of the 2.7Ga gneiss composition in variation diagrams for mangerites) and from isotope studies(coherent models for the Nd and Pb isotopic evolution of the mangerites, for which petrogenetically important isotopic parameters correlate with each other and with Sr isotopes; modelling of variations within an individual mangerite intrusion; mixing calculations) all suggest that a large mass contribution from the late Archaean migmatitic gneisses is present in the mangerites. A smaller contribution by mass from a mantle-derived magma and/or from juvenile supracrustal gneisses is also necessary to account for the geochemical and isotopic characteristics of the mangerites.

What is the mechanism by which large volumes of intermediate composition magmas can be derived by anatexis of the granulite-facies lower crust?

A number of models have proposed that crustal thickening by basaltic underplating or sill-injection can result in anatexis of the lower crust, producing more acid magmas. In certain tectonic environments these magmas have predominantly intermediate compositions.

McKenzie (1984) proposed a possible mechanism for epeirogenic uplift. Major ocean island volcanoes e.g. Hawaii, Iceland and the Azores are widely held to be generated by hot upwelling regions of the mantle, generated by instabilities on the lower thermal boundary layer between upper and lower mantle convecting layers. Hence the distribution and rate of magma production from these sources should be independent of the distribution of continental and oceanic lithosphere. If all the magma derived from such sources were to be erupted onto the Earth's surface, basalt should comprise about 75% of the continental crust of 30km thickness. McKenzie (1984) explained the lack of such volumes of basalt in the upper crust by proposing that picritic magmas could form sills in the lower crust or at the crust-mantle interface depending on the density of the magmas and the lower crust. Large intrusions of basic magma could provide the heat source to generate more acid magmas in the lower crust by anatexis. Uplift and erosion following the emplacement of basic magmas could explain the dominance of high-grade metamorphic rocks in the surface outcrop of Archaean terrains.

Hildreth (1981) postulated that melting temperature in the lower crust are attained almost exclusively by intrusion and crystallization of mantle-derived magma, which is principally basaltic. He produced models for magmatism under stress conditions favouring marked crustal

extension and where tectonic stress is subordinate. In each stress regime, any silicic magmas formed are derived from partial melting caused by basaltic magmas. Petrogenetic models may involve then the mixing of anatectic melt with the mantle-derived magmas to produce hybridized rocks. Evidence for the rôle of the basaltic magmas includes the mafic forerunners observed in most batholiths and volcanic provinces.

> "Protracted intrusion of the crust by such forerunners may be the key process in initiating upward mobilization of large crustal diapirs. The objection that mafic forerunners are generally quite subordinate to felsic granitoids in exposed plutonic terranes is overruled by the very nature of the model; upward penetration of basalt will be progressively retarded by the crustal melting and diapirism it promotes. The diapirs must consist not only of partially molten crustal rocks but of a network of basaltic bodies intruding and hybridizing with them. Seen in this light, the modified restite model also resolves the debate over the origin of mafic inclusions in granitoids. The inclusions are, at the same time, remnants of injected basalt (Eichelberger and Gooley, 1977) and refractory residues of the source material (White and Chappell, 1977) because the mushy diapiric source from which the grano-diorite-adamellite suite separates is itself a hybrid mixture of crustal and mantle contributions."
>
> (Hildreth, 1981)

This would explain the scarcity of basic magmas, compared to the mangeritic rocks, at the present level of erosion in the Lofoten-Vesterålen province. It would also support the interpretation of the geophysical evidence (Section 2.2) in terms of large volumes of basic magma present at depth. The extraction of partial melts from lower crustal gneisses renders the restite material more basic, e.g. Moskenesøy and Flakstadøy, and increases their density, thus impeding the upward penetration of later basaltic magmas. The mafic xenoliths in the mangerites are derived from both the lower crustal gneiss source material and from the mantle-derived basic magmas.

> "If intrusion of the crust by basaltic magma is long-lasting, extensive partial melting of crustal rocks in response to crystallization of the basalt is inevitable. The resulting local weakening of the crust tends to focus and promote further intrusion and hybridization of basalt. Feedback between dike injection and partial melting can apparently lead eventually to local ascent of increasingly silicic hybrid magmas along paths that become better established with time; ultimately, if the affected crustal domain is voluminous enough, wholesale mobilization of the focal zone can take place on a batholithic scale."
>
> (Hildreth, 1981)

The Raftsund Mangerite is the largest mangeritic intrusion in Lofoten-Vesterålen. It is also the most chemically and isotopically homogeneous and the youngest. The "affected crustal domain" in Lofoten-Vesterålen is very large and the Raftsund Mangerite represents the climatic mobilization of the focal zone of the gneisses pervaded by basaltic intrusions.

In Hildreth's (1981) first model, involving extension of the lithosphere, upward transport of both crustal melts and mantle-derived basic magmas is favoured. The lithosphere may be thinned, or not, depending on the relative importance of stretching compared to basaltic underplating.

> "In the second model a nonextending lithosphere is less accommodating, and magma transport is a fitful process of percolation slow enough that fractionation and hybridization in a plexus of lenticular conduits (cf. Shaw, 1980) lead to a predominance of intermediate magmas; the crust and lithosphere may be thickened significantly by the addition at depth of great volumes of basalt."
>
> (Hildreth, 1981)

This second model, in which tectonic extension, if any, is subordinate and shallow, could apply to the Lofoten-Vesterålen province, in which the magmas generated have intermediate compositions. The

thickened crust of Lofoten-Vesterålen at about 1.8Ga could result from a combination of a build-up of an extensive volcanic pile from 2.1Ga and from basaltic underplating.

Patchett (1980) calculated the distance over which partial melting would be expected around basalt sheets of different widths at different depths, and after different periods of time. Melts can be produced over a wide zone, but they are always partial, rather than total, melts. At 10km depth, basalt sheets can develop partial melts from zones up to 10-20% of their own thickness. For a 10m wide basalt sheet, this is achieved in less than a year, but it takes 2000 years for a 1000m wide basalt sheet. The thickness of the partial melt zone is greater at granulite-facies depths, but the uncertainty in the calculation is greater at such depths. Generation of intrusions of the size of the Raftsund Mangerite (about 1000 km^2) clearly requires intrusion of very large volumes of basaltic magma into the lower crust.

Experimental petrology has shown that syenitic magmas can be produced in thickened crust (650°C, 13kb) (Wyllie, 1984). Monzonites might be produced under similar conditions.

The anatectic veins and melts described in Section 7.2 are all located in the supracrustal gneisses except for the limited melting seen in the Festvåg migmatites on Austvågøy. The supracrustal gneisses have a maximum age of 2.1Ga, but Nd and Pb isotopes require a large component of crust older than this i.e. from the 2.7Ga migmatitic gneisses. Why is there little evidence of melting of the late Archaean gneisses? The lithological similarities between the massive, monzonitic gneisses and the mangerites might make identification of melts in the gneisses difficult, whereas the volcanic and sedimentary supracrustal gneisses

are strongly banded and have a wide range of compositions. Most of the veins and dykes that are present in the supracrustal gneisses, are not necessarily derived from those gneisses. The exceptions to this are at Ølkona and Hammøy Bridge on Moskenesøy, where good evidence for melting at the presently exposed level is seen (Section 7.2). Most of the anatexis would have taken place at crustal levels deeper than that which is currently exposed, where the density of basaltic intrusions is greatest.

The supracrustal gneisses were probably accumulated in a back-arc basin or a thin, partially submerged continental margin (Section 4.5) associated with high heat flow. In Hildreth's (1981) model for magmatism where there is marked crustal extension, as in a back-arc basin, large volumes of rhyolite are extruded along with basaltic lavas peripheral to the area into which mantle-derived basic magmas are intruded. In this situation, where extension is important, the basin would subside probably at a similar rate to that at which it was being filled. The igneous style changes from extrusive to intrusive and the crust thickens when crustal extension ceases. This could happen in a marginal sea when the ocean closes and the continent collides with the island arc. It is in this environment, the second model of Hildreth (1981), where partial melting of the lower crust by intrusion of basic magmas and hybridization between the partial melts and the basic magmas forms predominantly intermediate magmas, as was the situation in the Lofoten-Vesterålen province in the period 1.8 - 1.7Ga. The proposed back-arc basin, subduction-related magmatism and subsequent continent-continent collision must have climaxed in an orogeny. The Sveco-Karelian orogeny is dated at 1.9 - 1.75Ga in north Sweden (Witschard, 1984) and indeed the Archaean history of this area is similar to that of the Lofoten-

Vesterålen province. Perhaps the Sveco-Karelian orogeny caused not only the development of eclogites by shear deformation (Section 5.5.2) but also the uplift and erosion of Lofoten-Vesterålen and the associated retrograde metamorphism of the mangerites and the basement gneisses from granulite-facies.

The presence of anatectic melts and large basic intrusions at the deepest structural levels of the Lofoten-Vesterålen province (Moskenesøy, Flakstadøy and Langøy) and mangerite intrusions at higher levels culminating in the 1000km^2 Raftsund Mangerite at the highest structural level seen, provides a section through the crust illustrating the magmatic processes outlined above. At the deepest level exposed (South Moskenesøy) no mangeritic intrusions are present but a large anorthosite intrusion occurs, as do anatectic melts. The deepest, large mangeritic intrusion, the South-West Lofoten Mangerite is very heterogeneous and represents the coalescence of many anatectic melts and hybridization, with the mantle-derived basaltic magmas (especially samples WL315, WL316 and WL410; see Section 6.5.1). The younger, most homogeneous, and structurally shallowest mangeritic intrusion, the Raftsund Mangerite, represents the wholesale mobilization of the zone affected by extensive basaltic dyke intrusion as described in the model of Hildreth (1981). It is the most homogeneous and has the best-developed cumulates and magmatic-sedimentary features of the mangerites since it crystallized from a single batholithic magma. This would also explain the precise age for the Raftsund Mangerite compared to, for example, the South-West Lofoten Mangerite, which as well as having variations in initial isotope ratios may have crystallzied from a number of smaller magmas at different times in the period 1.8 - 1.7Ga. The more basic character of the gneisses on Moskenesøy and Flakstadøy, compared to the gneisses on the

islands at higher structural levels, suggests that substantial volumes of partial melt have been extracted from them. It is possible that the geology of these islands resembles that at depth below the Raftsund Mangerite today, and that a large intrusion, comparable to the Raftsund Mangerite, formerly existed above the presently-exposed level of Moskenesøy.

7.6 Conclusions

1. Anatectic veins and dykes of mangeritic and charnockite composition are found on the islands of the Lofoten-Vesterålen province which are at the deepest structural levels.

2. All of the examples of in-situ anatexis are found in the supracrustal gneisses, except for that in the Festvåg migmatites.

3. The most convincingly in-situ anatectic melt (WL212) from Ølkona, Moskenesøy, is of intermediate composition (61% SiO_2). Field relations, petrography and its T_{DM} value suggest that this melt is derived solely from supracrustal gneisses. This observation supports the experimental petrology of Wyllie (1984) which suggests that the first-formed magmas are syenitic at P ≈ 13kb.

4. The chemistry of the melts is similar to that of the mangerites including very high K_2O concentrations which suggests that the melts were derived by reworking of continental crust.

5. Nd isotope evolution diagrams are consistent with the formation of the Sund-Ølkona and South-West Lofoten Mangerites by coalescence of melts similar to those of mangeritic composition on Moskenesøy and Flakstadøy. The charnockite veins on Vestvågøy have similarities to the Flakstadøy Charnockite.

6. The demonstration of the importance of the early Proterozoic supra-crustal gneisses, which are isotopically similar to mantle-derived magma in the anatectic melts, implies that the percentage crustal contributions by mass for the melts, calculated for Nd and Pb isotopes, are minimum estimates. Part of the "mantle-derived" component calculated for the mangeritic intrusions in Section 6.5.2 is probably also due to the assimilation of juvenile crust. Hence, the estimates for crustal contribution by mass to the mangerite intrusions are minimum values.

7. The mangerites formed by anatexis of 2.7Ga lower crustal gneisses due to the rise in temperature associated with the emplacement of basaltic magmas. Their formation and crustal thickening due to basaltic underplating, took place in the absence of crustal extension.

8. Uplift and erosion, subsequent to emplacement of the mangerites, may be associated with the Sveco-Karelian orogeny of North Sweden.

9. Crust at different structural levels is exposed across the Lofoten-Vesterålen province. This provides a section through the development of the mangerites, from in-situ anatexis of lower crustal gneisses in the deepest parts, to an intrusion of batholithic proportions in the shallowest parts of the province.

CHAPTER 8. POST-MANGERITE ROCK-FORMING EVENTS

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CHAPTER 8. POST-MANGERITE ROCK-FORMING EVENTS

8.1 Introduction

A very small proportion of the Lofoten-Vesterålen province is underlain by rocks which post-date the emplacement of the mangerites. The four rock-forming events which come into this category are:

(i) Dolerite dykes: intruded into the mangerites while they were still hot; dolerite samples from Austvågøy were analysed by Misra and Griffin (1972).

(ii) The Lødingen Granite: on South Hinnøy, dated at 1644Ma by Andresen and Tull (1983).

(iii) The Leknes Group: underlies parts of Vestvågøy and Vaerøy (Tull, 1977; Griffin and Taylor, 1978). These metasediments are 1140Ma old.

(iv) Granite pegmatites: restricted to the southern part of Moskenesøy; they are Caledonian in age (Griffin et al., 1978).

8.2 The Dolerite Dykes

8.2.1 Fieldwork and Petrography

The dolerite dykes intrude the gneisses and mangerites on Austvågøy and South-West Hinnøy. The first dykes strike NE-SW; they are usually between 1 and 3 metres thick, and cut by later more numerous E-W dykes up to 10m across. The east-west trending dykes are abundant in the charnockite rock in the Hopen intrusion (Section 6.2.4). The foliation of the charnockite is parallel to the margins of the dolerites. The dykes there are from a few centimetres to 4m wide (Plate 8.1) and include narrow slices of the host charnockite. Melting of the charnockite has clearly taken place at the contact with the dolerite (Plate 8.2) and the charnockite back-veins the dolerite, indicating that the dykes were intruded into the charnockite when the latter was still hot. Hence the dykes post-date the

mangeritic intrusions by only a short period of time.

A small number of dolerite dykes have been recognized on Moskenesøy and Flakstadøy. A 10m dyke strikes 015° and dips 55°W in the Flakstadøy Basic Complex at Nusfjord. The contact is sharp. A sample of the dyke, WL313, consists of 3mm plagioclase laths, clinopyroxene and olivine in sub-ophitic texture. Olivine is overgrown by clinopyroxene-orthopyroxene-spinel-garnet coronas as it is in the Flakstadøy Basic Complex. This shows that the dyke experienced the same retrograde reactions as the rocks of the basic complex. Several dolerite dykes of the east-west swarm, up to 2m thick, intrude the mangerite at Ølkona, Moskenesøy (Section 6.2.1).

A number of dykes intrude the Raftsund Mangerite. A large dyke of the earlier NE-SW set strikes 015° and dips 82°W on the north side of Tenfjord. It contains blocks of pink feldspathic gneiss on which pegmatite quartz and muscovite have developed. The dyke itself is a plagioclase-phyric lamprophyre (Plate 8.3). Plagioclase phenocrysts are up to 45 x 22mm. Their cores are sericitized but not their rims (WL611-WL613). This is characteristic of zoned plagioclase, Na-rich rims being resistant to sericitization. White mica and epidote occur in the plagioclase cores. The phenocrysts comprise about 10% of the rock. The groundmass consists of biotite and plagioclase grains less than 0.5mm across. A carbonate phase occurs in the groundmass and in the cores of plagioclase. Accessory apatite is present.

8.3 The Lødingen Granite

8.3.1 Fieldwork and Petrography

The Lødingen Granite is a large, heterogeneous intrusion in the east of the Lofoten-Vesterålen province (i.e. at the shallowest structural level) on South Hinnøy (Fig. 8.1). It is intrusive into the Raftsund Mangerite. In a road-cut at the road-junction to Offersøy, the contact between the Raftsund Mangerite and the Lødingen Granite is exposed (Section 6.2.6). This granite outcrop is over 1km west of the contact on the map of Tveten (1978). The contact is sharp, unlike that between the dolerite dykes and the mangeritic rocks. This indicates that the mangerites were not so hot when the granite was emplaced as when the dolerite dykes intruded the mangerites, and hence the granite post-dates the dolerite dykes. Most of the intrusion is granitic in composition, consisting of quartz, microcline and plagioclase in varying, but broadly similar, amounts. These phases typically constitute over 95% of the rock, have a grain size of 0.5-2mm and a granitic texture. Microcline porphyroblasts are common (e.g. WL629). The main mafic phase is green biotite which occurs in aggregates, often with opaques and sphene. The biotite commonly contains pleochroic haloes due to the presence of zircon. Epidote and white mica are present in most samples, associated with the sericitization of plagioclase. Often the rims of the plagioclase crystals are not altered, indicating that they are zoned. Some of the epidote in WL634 (3km north of Rinøyvåg) has allanite cores. WL634 is typical of the Lødingen Granite: pink in colour, aggregates of mafic phases up to 5mm across, and is illustrated in Plate 8.4. Strongly pleochroic (blue-green to straw-yellow) hornblende is present in some of the more leucocratic granite samples (e.g. WL631 and WL632 from 2km south of, and at, Rinøyvåg and WL665 from north of Kanstad road junction), but absent from others (e.g. WL629 from Offersøy and WL634 from north of Rinøyvåg). In many localities

the granite is more melanocratic (e.g. WL671, Plate 8.5, from 1km south of Kanstad road-junction). K-feldspar porphyroblasts, up to 10mm across, occur in a finer-grained groundmass which may contain up to 50% mafic phases. The phases present are the same as in the leucocratic granites. Sphene is up to 1mm across, is associated with opaque phases and is the predominant phase in some mafic aggregates. Thin pegmatites, typically 10cm wide, intrude the more melanocratic granites. WL673 is a sample of such a pegmatite which intrudes the granite 1km south of Kanstad junction, illustrated in Plate 8.5. It consists of plagioclase, quartz and microcline, up to 5mm across, which together comprise greater than 99% of the rock. A carbonate phase is the next most abundant mineral. A few grains of biotite, an opaque and epidote are also present. Chlorite and apatite are also present in accessory amounts in some of the granite samples.

Large rafts of mafic material are included in the Lødingen Granite. These are usually fine-grained amphibolites (e.g. WL677 and WL678 from 2km south of Kanstad junction) consisting of green biotite, quartz, plagioclase and microcline, present in roughly equal amounts, with minor hornblende and accessory zircon. Many of the amphibolites have the form of sills or dykes and have sharp margins against the granite, suggesting that they may have been emplaced as dolerites. WL687 is a foliated biotite gabbro. Biotite plates and plagioclase laths, grain size 0.5mm, define the foliation. Clinopyroxene and subordinate orthopyroxene are present in the rock, concentrated in layers. A very large enclave, over 100m across occurs in the granite at Erikstad. It consists of banded ironstones and sulphide-rich pyroxenites (Section 4.2.7).

8.3.2 Geochemistry

Five samples from the Lødingen Granite have been analysed for their major element compositions in this study (Table 8.1). The wide range in SiO_2 contents (60-78%) and modal quartz (1-39%) reflects the heterogeneity of the intrusion. The granite samples have Fe_2O_3/MgO = 6.71 (excluding WL677, the amphibolite xenolith from the quarry 2.5km south of the Kanstad road-junction, which has Fe_2O_3/MgO = 1.4), showing extreme Fe-enrichment similar to that of the mangerites. WL666, from 3km north of Kanstad road-junction is a mafic-rich granite similar to WL671, illustrated in Plate 8.5 and has a very high Fe-content. The other three samples analysed are pink, more felsic granites.

8.3.3 Isotopic Results

8.3.3.1 Sr Isotopes

Sr isotope data for the Lødingen Granite are tabulated and plotted in Fig. 8.2. Twenty samples have been analysed. They have a range in Rb and Sr concentrations of 70 to 233 ppm and 27 to 593 ppm respectively. The range in $^{87}Rb/^{86}Sr$ ratios is 0.4012 to 18.116 and for $^{87}Sr/^{86}Sr$ ratios is 0.71376 to 1.39686. These twenty points define an isochron of 1644 ± 89Ma (2σ), (MSWD = 69.15) with an initial $^{87}Sr/^{86}Sr$ ratio of 0.70485 ± 190. This compares to an age of 1644 ± 36Ma (2σ) and initial $^{87}Sr/^{86}Sr$ ratios of 0.70780 ± 74 from a ten-point isochron published by Andresen and Tull (1983). The twenty points in the isochron produced in this study are from samples of all the lithologies in the Lødingen Granite: WL629, WL631, WL632, WL634, WL665, WL670, WL675, WL683, WL684, WL685 and WL690 are pink, felsic granites (all similar to WL634 illustrated in Plate 8.4); WL666, WL671, WL676 and WL681 are more mafic granites (WL671, Plate 8.5); WL672 and WL673 are granitic pegmatites; WL679 is a fine-grained granite; and WL677 and WL678 are amphibolite xenoliths.

Exclusion of particular samples does not alter the precision of the isochron.

8.3.3.2 Nd Isotopes

Five samples from the Lødingen Granite have been analysed for Sm-Nd isotopic composition in this study (Table 8.2). Three of the samples are pink, felsic granites (WL629, WL634 and WL690), one is a mafic-rich granite (WL666) and one is an amphibolite inclusion (WL677). The full range of values is as follows: Sm and Nd concentrations are in the ranges 6.7 to 46.0ppm and 39.0 to 306.7ppm respectively; $^{147}Sm/^{144}Nd$ ranges between 0.077 and 0.104, and $^{143}Nd/^{144}Nd$ ranges between 0.51107 and 0.51150.

Variations of the initial $^{143}Nd/^{144}Nd$ ratios (calculated at about 1.65 Ga) are sufficiently large to preclude whole-rock isochron dating by the Sm-Nd method for the Lødingen Granite. The Nd evolution diagram for the Lødingen Granite (Fig. 8.3) shows the sub-parallelism of the evolution lines and the corresponding range in T_{DM} Nd model ages (2.13 - 2.39 Ga) which imply that the Lødingen Granite contains a crustal component significantly older than 1.65 Ga. The youngest model age for the five samples is for the amphibolite inclusion, WL677. The ranges of model ages, $^{147}Sm/^{144}Nd$ and $^{143}Nd/^{144}Nd$ ratios are within those for the mangerites (Section 6.4.2).

8.4 The Leknes Group

8.4.1 Fieldwork and Petrography

The Leknes Group underlies an area of about $20km^2$ around Leknes, Vestvågøy (Tull, 1977) (Fig. 6.1). Metasediments on Vaerøy, to the south-west of Moskenesøy have been correlated with the Leknes Group by Griffin and Taylor (1978).

The Leknes Group metasediments are well-exposed in the extensions to the Leknes Airport. The main lithology is a kyanite-garnet-mica schist (WL421 (Plate 8.6) and WL422). The kyanite occurs as blue laths up to 20 x 8mm which define the schistosity. Minor pale yellow staurolite is also present. Garnet is up to 2mm across and subhedral, it is occasionally elongate parallel to the foliation. Muscovite and biotite plates are also parallel to the foliation. Quartz and plagioclase occur in layers in the schist. WL424, from the carpark of a Telecommunications company at Leknes Airport, is an amphibolite. Hornblende comprises over 50% of the mode. Subordinate phases are quartz, plagioclase, garnet and sphene.

8.4.2 Geochemistry

Two samples, WL421 and WL424, from the Leknes Group have been analysed for major elements in this study (Table 8.3). The presence of abundant kyanite in WL421 is reflected in the high Al_2O_3 content (22.1%) and the high normative corundum content (17.3%). WL424, the amphibolite sample, has a lower Al_2O_3 content than WL421 and a much higher CaO content. The Fe_2O_3/MgO ratios are 4.2 and 1.9 for WL421 and WL424 respectively.

8.4.3 Isotopic Results

8.4.3.1 Sr Isotopes

Sr isotope data for three samples from the Leknes Group are presented in Table 8.4. The range in Rb and Sr concentrations are 43 to 58ppm and 175 to 327ppm respectively.

8.4.3.2 Nd Isotopes

Nd isotope data for WL421 and WL424, from the Leknes Group are presented in Table 8.5 and plotted in Fig. 8.4. The T_{DM} Nd model ages for both of these samples are within error of 1.6Ga. These model ages

prove that a mantle-derived crustal addition to the Lofoten-Vesterålen province has been made subsequent to the emplacement of the mangerites at 1.8Ga. Nonetheless this model age is much older than the 1140 ± 135Ma Rb-Sr whole-rock isochron age for the Leknes Group (Griffin et al., 1978). This implies that the Leknes Group is, in fact, older than 1140Ma and/or that it has been derived from rocks with a previous crustal history.

8.5 The Caledonian Granite Pegmatites

8.5.1 Fieldwork and Petrography

A group of pegmatite dykes of granitic composition occurs on the southern end of Moskenesøy. K-Ar dating of muscovite and biotite from these dykes has yielded Caledonian ages (Griffin et al., 1978) (Table 2.3). Plate 8.7 shows a 5m wide granite pegmatite north of Å campsite. In thin-section this pegmatite, WL231, consists of microcline, plagioclase and quartz grains up to 12mm, with a granitic texture, and minor biotite and opaques. This dyke can be followed from the Moskenes Pegmatite Quarry to the south of Å. K-feldspar crystals up to 60 x 30cm, and plagioclase up to 30 x 20cm are present. The pegmatite contains large rafts of gneiss (up to 150 x 60cm). The foliation of the gneiss rafts is often oblique to that in the country rock gneisses. A later foliation parallel to the margin of the pegmatite often overprints the original foliation in the gneiss rafts and is well-developed in the pegmatite in places. The pegmatite strikes 062° and dips 78°W. It cuts earlier red monzonitic sheets with an offset between the two sides of up to 1m. The pegmatite typically has a very coarse, red K-feldspar-rich core 1-2m wide which has no foliation (Plate 8.7). This core contains many crystals larger than 10 x 10cm, whereas the white, plagioclase-rich margin contains no crystals larger than 3 x 3cm. The margins, which may also be 1-2m wide, are foliated parallel to the contact with the gneisses, and contain abundant elongate

slivers of gneiss. The coarser, unfoliated, inclusion-free core may have been intruded after the marginal parts. Very coarse quartz cores are also present, the largest single crystal measuring 100 x 70cm.

Three granite pegmatites intrude the late Archaean migmatitic gneisses at the north end of Djupfjord Bridge near the gabbro outcrop. They each strike about 060° and dip 70°W. The largest, easternmost pegmatite is 1.5m wide and consists of coarsely crystalline K-felspar, plagioclase and quartz. Red garnets up to 15mm across occur in the feldspars. Thick books of biotite and minor muscovite and tourmaline are also present.

8.6 Discussion and Synthesis

Rock-forming events in the Lofoten-Vesterålen province subsequent to the emplacement of the mangerites at 1.8Ga account for only a small proportion of the rocks exposed at the surface at the present. The dolerite dykes immediately post-date the emplacement of the mangerites. Misra and Griffin (1972) concluded that they are similar to many basalts from continental margins and tensional zones. The dolerite dykes may have been derived from the same magmas as the 1.8Ga basic intrusive bodies. The dykes occur at higher structural levels than the large basic intrusions. This could reflect an increasing extensional component in the area after 1.8Ga allowing basic magmas to ascend to higher structural levels.

The Rb-Sr isochron for the Lødingen Granite produced in this study yields the same age as that published by Andresen and Tull (1983), showing that the emplacement of the Lødingen Granite is slightly, but distinctly, younger than the age of emplacement of the mangerites. The age of the Lødingen Granite obtained in this study (1644 ± 89Ma) is similar to that for the Torset Granite on Langøy (1665 ± 80Ma) (Brue , 1971) (Table 2.6).

The T_{DM} Nd model ages (and the somewhat high initial $^{87}Sr/^{86}Sr$ ratio imply that the granites, like the mangerites, contain significant amounts of ancient crustal material. There are some chemical similarities between the Lødingen Granite and the adjacent Raftsund Mangerite, including high Fe_2O_3/MgO ratios, suggesting that the Lødingen Granite may represent a remobilised mangerite which ascended to a higher structural level or an anatectic magma formed due to the emplacement of the very large intrusive body, the Raftsund Mangerite.

The Leknes Group metasediments have T_{DM} Nd model ages of 1.6Ga which shows that a new addition of mantle-derived material has been made to the Lofoten-Vesterålen province subsequent to the emplacement of the mangerites. Tull (1977) suggested that the Leknes Group was tectonically emplaced onto the South-West Lofoten Mangerite.

8.7 Conclusions

1. There have been four rock-forming events in the Lofoten-Vesterålen province subsequent to the emplacement of the mangerites at 1.8Ga.

2. A suite of dolerite dykes was intruded in the region shortly after the emplacement of the mangerites, possibly in response to increased extension across the area.

3. The Lødingen Granite was the last major rock-forming event in the Lofoten-Vesterålen province. It was emplaced shortly after the mangerites, with which it shares some chemical and isotopic characteristics. A large ancient crustal component is required by the Nd isotopes. The Lødingen Granite was probably formed by crustal anatexis in response to the emplacement of the Raftsund Mangerite or by remobilization of a mangerite itself.

4. The Leknes Group metasediments have a crustal history prior to their Rb-Sr whole-rock age. Nevertheless their T_{DM} Nd model ages demonstrate that a new addition of mantle-derived material has been made to the Lofoten-Vesterålen province after the emplacement of the mangerites at 1.8Ga.

5. The granite pegmatites from the southern part of Moskenesøy represent the youngest rock-forming event in the Lofoten-Vesterålen province.

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CHAPTER 9

CONCLUSIONS

1. Previous work in the Lofoten-Vesterålen province has shown that it consists, almost exclusively, of Precambrian rocks with granulite and upper amphibolite facies mineral assemblages. Basement gneisses were formed in two events in the late Archaean and the early Proterozoic, and were intruded by basic magmas and very large mangeritic intrusions at about 1.8Ga.

2. Previous geophysical work has shown that the area is probably underlain by large volumes of basic rocks.

3. It is often not possible to distinguish monzonitic gneisses of late Archaean age from the mangerites, petrographically or in the field. Chemical compositions can be used to resolve them. The protoliths of these migmatitic gneisses formed 2.7Ga ago, but the migmatites at Festvåg contain a younger basic component and the migmatites on Moskenesøy are distinctly younger than those on Langøy, Hinnøy and Austvågøy. The Rb-Sr isotopic systems of the late Archaean migmatites were disturbed during the subsequent granulite-facies metamorphism at 1.8Ga.

4. The gneisses of early Proterozoic age are supracrustal in origin, confirmed by the presence of marble layers. Their most likely environment ofdeposition is in a back-arc basin which subsided at a similar rate to the rate of infill from the more voluminous lavas interspersed with limestones and haematite cherts. The gneisses of Moskenesøy and Flakstadøy are more basic than those from higher structural levels in the Lofoten-Vesterålen province, which could be due to the extraction of more acid melts. Small volumes of such melts are present in the early

Proterozoic supracrustal gneisses of Moskenesøy. These gneisses were extracted from a depleted mantle source at about 2.1Ga and their Rb-Sr isotope systems were disturbed during the retrograde metamorphism at about 1.1Ga.

5. The late Archaean and early Proterozoic gneisses were intruded by a series of basic intrusions during the mid-Proterozoic. Some of these intrusions are small gabbros, other are larger, have cumulate textures, magmatic sedimentation structures and contain anorthosites. These basic rocks were intruded at 1.8Ga but contain a significant ancient crustal component. The smaller gabbros have been contaminated with a smaller late Archaean migmatite-derived component than the Flakstadøy Basic Complex, but the smaller gabbros and the margins of the Flakstadøy Basic Complex contain material similar to the mangerites i.e. they have been contaminated by the first-formed anatectic melts. The basic intrusions crystallized at $P \leq 8kb$ at about $1150^{o}C$ and cooled to the granulite-eclogite boundary. Shear deformation caused parts of the basic rocks to develop eclogite-facies parageneses which were subsequently retrogressed to amphibolite facies.

6. Intrusions of mangeritic and charnockitic composition underlie about half of the Lofoten-Vesterålen province. In places they are clearly intrusive into the basic intrusions. The mangeritic rocks show magmatic sedimentation structures and well-developed cumulate textures in places. The mangerites were intruded between 1.8 and 1.7Ga ago. Formation solely from 1.8Ga mantle-derived magmas is not possible. A model has been constructed which explains the isotopic compositions of the mangeritic rocks in terms of formation by anatexis of late Archaean migmatites and mixing with 1.8Ga mantle-derived magmas and juvenile supracrustal gneisses. The late Archaean component is in excess of 50%

by mass for almost all of the mangerite samples. The presence of abundant, partially assimilated xenoliths supports the hypothesis that the mangerites are largely crustally-derived. The chemical composition of the mangerites is also suggestive of an anatectic origin and subsequent fractional crystallization by removal of pyroxenes, olivine, magnetite, ilmenite and apatite. The crust-mantle mixing model has been used to explain the wide chemical and isotopic variations in the South-West Lofoten Mangerite. The scatter of data about the Rb-Sr isochron for this intrusion is due to initial $^{87}Sr/^{86}Sr$ variations. The samples with the largest mantle-derived component are from the north-west side of the islands where the Moho is nearest the surface.

7. Anatectic veins and dykes of mangeritic and charnockitic composition are present in the structurally deepest parts of the Lofoten-Vesterålen province. In-situ anatexis is rarely seen in the late Archaean gneisses, but is common in the early Proterozoic gneisses where it is often associated with injection of basic veins and dykes. In places, melts are developed from early Proterozoic supracrustal gneisses without any material contribution from 1.8Ga mantle-derived basic rocks. This would suggest that the juvenile component in the mangerites may be very small. Nevertheless the thermal input from basaltic underplating was the cause of the anatexis of parts of the lower crust. The transition from extrusive igneous activity in a subsiding basin to intrusive igneous activity and crustal thickening with a large reduction in crustal extension is probably associated with the closure of the ocean which resulted in the Sveco-Karelian orogeny, which could also have caused the uplift and erosion, subsequent to the emplacement of the mangerites.

Crust at different structural levels is exposed across the Lofoten-Vesterålen province. Emplacement of basic magmas and in-situ anatexis

can be observed at the deeper levels, to the north and west. The South-West Lofoten Mangerite is a relatively heterogeneous mangerite, representing the coalescing of many smaller anatectic melts. The Raftsund Mangerite, at the highest structural level in the area, in the south-east, represents mobilization of a zone affected by extensive basaltic dyke injection. The first-formed magmas from crustal anatexis at P > 10kb have alkalic intermediate composition.

8. Dolerite dykes intruded the Lofoten-Vesterålen province shortly after the emplacement of the mangeritic rocks. The Lødingen Granite is a very heterogeneous intrusion which is slightly, but distinctly, younger than the Raftsund Mangerite. The granite was probably formed by crustal anatexis in response to the raising of crustal isotherms due to the emplacement of the batholithic Raftsund Mangerite. The Leknes Group metasediments contain a mantle-derived component which was extracted from the depleted mantle at, or after, 1.6Ga. The granite pegmatites in the southern part of Moskenesøy represent the youngest rock-forming event recognised in the Lofoten-Vesterålen province.

9. The major tectono-thermal event at 1.8Ga involved anatexis of the basement gneiss of the regions (Chapters 3 and 4) following the raising of isotherms due to the emplacement of basic magmas (Chapter 5). Examples of the immediate products of this anatexis are present at the deepest structural levels of the Lofoten-Vesterålen province.

10. The Lofoten-Vesterålen province is a natural "laboratory" in which "experiments" have been carried out deep in the continental crust. These natural "experiments" lend credence to theories developed from experiments conducted in the laboratory. The anatectic melt at Ølkona, Moskenesøy, which has probably been derived solely from pre-existing

continental crust, is of intermediate composition : experimental petrology (Wyllie, 1984) has suggested that the first-formed magmas at P ≈ 13kb are syenitic.

Wyllie's work (e.g. 1977, 1979, 1982, 1984) has examined the relative contributions to magma genesis from mantle, subducted oceanic crust, and deeper continental crust using experimental petrology. A detailed review of the experiments relating to the rôle of crustal anatexis in magma genesis is presented in Wyllie (1977). Fig. 9.1 (after Wyllie, 1977 and 1984) shows the approximate P-T ranges for liquids of various compositions generated from continental gneisses, with 2% H_2O in pore fluids. At high pressure, liquids with low SiO_2 are produced. In thickened crust, the near-solidus liquid approaches syenitic composition (Huang and Wyllie, 1984).

Evidence has been presented in this work (e.g. Section 7.5.3) that the igneous events in Lofoten-Vesterålen in the period 2.1 - 1.7Ga are subduction-related. Wyllie (1984) considered the different styles of magma genesis at subduction zones with different temperatures in the subducting oceanic slab and the overlying mantle wedge. For ocean-ocean convergent boundaries, Maaløe and Petersen (1981) have suggested that "super-hot" mantle can be produced by extreme induced convection cells in the mantle wedge, which would raise the temperature of the subducted oceanic crust at 100km depth to 1250°C to 1450°C. Maaløe and Petersen (1981) have suggested that similar conditions may occur beneath an ocean-continent plate boundary, because similar primary andesitic magmas have been recognised contaminated by continental crust. Fig. 9.2 (after Wyllie, 1984) shows the zones in which melting is expected to occur in at an ocean-continent plate boundary for different thermal conditions. The experimental basis for these diagrams was

reviewed by Wyllie (1979, 1982) and Wyllie and Sekine (1982). For a warm mantle (Figs. 9.2c and 9.2d), induced convection brings the region for potential melting of the mantle above the greenschist (DG) and amphibole dehydration fronts. The amount of melting of the mantle periditote caused by the metasomatic fluids could be enhanced in the region N by H_2O from dissociation of metamorphic amphibole in the mantle as it crosses the amphibolite-out boundary. A warm mantle model is most applicable to magmatic processes in the early Proterozoic at which time the concentrations of K, U and Th were much greater than their present levels.

Underplatingof magmas beneath the continent could generate additional magma from the lower crust at H and composite magmas could rise into the region B forming batholiths, perhaps with associated volcanism at the surface. Wyllie (1984) concluded that with magmas of different compositions derived from different sources at different depths, there are many opportunities for contamination, hybridization and magma mixing in multi-stage processes. The present study provides support for this conclusion and also for the conclusion that the first-formed melts deep in the crust are of intermediate composition, tending towards alkaline.

The work conducted in the Lofoten-Vesterålen province for the present study supports Hildreth's (1981) model of protracted intrusion of the lower continental crust by basic magmas causing upward mobilization of large diapirs of intermediate compositions, in regions where extension, if any, is subordinate and shallow. This causes crustal thickening. See Section 7.5.3 for a detailed comparison of Hildreth's model to the magmatic processes in the Lofoten-Vesterålen province.

Localized melting of the crust has occurred around thin mafic veins and dykes as postulated by Patchett (1980).

Magma ascent by stoping may be important in the upper crust and causes continued modification of the magma during ascent (Oxburgh and McRae, 1984); the present study has shown (from the locally abundant xenoliths in the magmas) that such a process may also be important in the lower crust. Oxburgh and McRae (1984) suggested that the stoped blocks that may be observed in intrusive igneous bodies today as assimilated xenoliths will have a sharp upper size limit, those that exceeded that size having sunk through the chamber before becoming melted. Oxburgh and McRae (1984) recognised two kinds of xenolith in the Adamello Complex: those that are clearly related to the country rock and those that are recognized as mafic patches within the igneous host. The former tend to be angular, of any size, and have sharp margins and are restricted to the marginal zone of the complex. The latter rarely exceed 30cm in diameter, are usually rounded to sub-rounded, have diffuse margins, are much more abundant that the former type, and occur anywhere in the complex. Plates 6.13 and 6.14 show xenoliths in the Hopen intrusion in the Lofoten-Vesterålen province which clearly belong to the former category: angular, sharp margins, of many sizes, at the margin of the intrusion, and clearly not removed very far from their original position in the host rock. In the Lofoten-Vesterålen province mangerite intrusions, xenoliths of the latter category are much more abundant (Figs. 6.2, 6.8, 6.15, 6.15 and 6.24). These similarities show that similar magmatic assimilation processes took place in the lower crust in the Lofoten-Vesterålen province in the Precambrian as in the Adamello Complex emplaced in the upper crust during the Tertiary.

Experimental studies of convection in crystallizing systems together with theoretical analyses suggest that convective separation of liquid from cumulus crystals is the dominant process of fractionation in magmas (Sparks et al., 1984). Crystallization along the margins of a chamber can generate highly fractionated magmas without requiring large amounts of crystallization, since the removal and concentration of chemical components affects only a small part of the magma. Sparks et al. (1984) proposed that selective removal of heavy or light components into the cumulus crystals is the basic driving force for magmatic convection. Highly fractionated magmas were generated along the margins of the Hopen intrusion by the crystallization of apatite, and pyroxenite phases. Such fractionation could have been the driving force for the convection which produced the magmatic sedimentation features, especially well-developed in the Hopen and Raftsund Mangerites.

Studies of crust-forming processes are less numerous for the Proterozoic than for Archaean. The present study has shown that new additions from the mantle to the crust have taken place in the period 2.1 to 1.8Ga. A change in the tectonic environment from extensional to one in which extension was minor or absent, caused a change in igneous styles from extrusive to intrusive, resulting in crustal thickening. Large-scale crustal reworking resulted from continued addition of mantle-derived basic magmas to the lower crust. In general, crustal reworking has increased in importance in the Proterozoic relative to the Archaean (Taylor et al., 1984). The intrusions of the Ketilidian province of South Greenland have similar ages to the mangerites of the Lofoten-Vesterålen province. Most of the Ketilidian is new crust at 1.8Ga (Patchett and Bridgwater, 1984), but crustal reworking has occurred in the Border Zone where the Ketilidian Mobile Belt overlaps the Archaean

craton (Kalsbeek and Taylor, 1985).

The present study supports the operation of plate tectonics in the Proterozoic, and relates the mid-Proterozoic events in the Lofoten-Vesterålen province to the Sveco-Karelian orogeny.

The processes recognized in the Lofoten-Vesterålen province are probably in operation today at depth in regions of thickened continental crust, such as The Andes: for deep crustal situations, then,

"The Past is the key to The Present"

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APPENDIX 1: Sample Localities

A comprehensive list of sample localities for samples collected in 1982-84 in Lofoten-Vesterålen. Map sheets and grid references refer to AMS 1:50,000 sheets.

The sample number prefix WL stands for Wade-Lofoten

Of the three digits used for each sample, the first denotes the island from which the sample was taken and the latter two digits are sequential.

First digit	Island
1	Hamarøy
2	Moskenesøy
3	Flakstadøy
4	Vestvågøy
5	Austvågøy
6	Hinnøy
7	Langøy

Sample No.	Map Sheet	Grid Ref.	Locality Description
Hamarøy			
WL101	Hamarøy	WR171510	Rocky spur, 200m easy of road to Halsen
WL102	Hamarøy	WR220541	Road-cut, 500m SE of Hamsund junction
WL103	Hamarøy	WR179492	Quarry, west of road, 500m south of Halsen junction
WL104	Hamarøy	WR138448	Foreshore, 300m NW of Skutvik ferry terminal
WL105	Hamarøy	WR140447	Quarry, 100m W of Skutvik ferry terminal
Moskenesøy			
WL201	Lofotodden	VR242415	Foreshore, 4km N of Hamnøy
WL202	Lofotodden	VR241417	Road-cut, 4km N of Hamnøy
WL203	Lofotodden	VR241417	Road-cut, 4km N of Hamnøy
WL204	Lofotodden	VR238424	Road-cut, 5km N of Hamnøy
WL205	Lofotodden	VR238425	Road-cut, 5km N of Hamnøy
WL206	Lofotodden	VR238425	Road-cut, 5km N of Hamnøy
WL207	Lofotodden	VR238426	Road-cut, 5km N of Hamnøy
WL208	Lofotodden	VR237426	Road-cut, 5km N of Hamnøy
WL209	Lofotodden	VR238424	Road-cut, 5km N of Hamnøy
WL210	Lofotodden	VR241418	Road-cut, 4km N of Hamnøy
WL211	Lofotodden	VR241419	Road-cut, 4km N of Hamnøy
WL212-223	Lofotodden	VR242415	Fore-shore, 4km N of Hamnøy
WL224	Lofotodden	VR233402	Road-cut, 2.5km N of Hamnøy
WL225	Lofotodden	VR221386	Loose block, NE end of bridge, Hamnøy
WL226	Lofotodden	VR233403	Road-cut, 2.5km N of Hamnøy
WL227-229	Lofotodden	VR196348	Hummocky ground between road and shoreline, 2km S of Moskenes
WL230	Lofotodden	VR155309	Quarry (car park), Å camp-site
WL231	Lofotodden	VR155309	Quarry (car park), Å camp-site
WL232	Lofododden	VR158315	Road-site outcrop, between Å and Tind
WL233	Lofotodden	VR158315	Road-side outcrop, between Å and Tind
WL234	Lofotodden	VR159316	Road-cut, Tind
WL235	Lofotodden	VR159316	Road-cut, Tind
WL236	Lofotodden	VR174334	Road-side outcrop, between Sørvågen and Moskenes
WL237	Lofotodden	VR184332	Quarry (disused) for Moskenes Pegmatite
WL238	Lofotodden	VR184332	Quarry (disused) for Moskenes Pegmatite

WL239-244	Lofotodden	VR242415	Fore-shore, 4km N of Hamnøy
WL245	Lofotodden	VR149311	Hill-side, 500m W of Å
WL246-255	Lofotodden	VR221386	Fore-shore, NE end of Bridge, Hamnøy
WL256-259	Lofotodden	VR237408	Fore-shore, 3km N of Hamnøy
Flakstadøy			
WL301	Moskenesøy	VR256452	Gully in isthmus east of Sund Harbour
WL302	Moskenesøy	VR248456	Road-cut, 0.8km NW of Sund Harbour
WL303	Moskenesøy	VR248457	Road-cut, 1.0km NW of Sund Harbour
WL304	Moskenesøy	VR248458	Foreshore, 1.2km NW of Sund Harbour
WL305	Flakstad	VR279537	Hil-side, 1.3km W of Bø
WL306	Flakstad	VR282538	Hill-side, 1.3km W of Bø
WL307	Flakstad	VR276472	Hill-side (steep), 100m E of Skjelford-Nesland road
WL308	Flakstad	VR274463	Road-cut, Skjelford to Nesland road
WL309	Flakstad	VR274446	Road-side outcrop, Nesland peninsula
WL310	Flakstad	VR276444	Road-cut, Nesland peninsula
WL311	Flakstad	VR281445	Road-cut, Nesland
WL312	Flakstad	VR301551	Road-cut, 1km SE of Flakstad church
WL313	Flakstad	VR311478	Road-side outcrop, Nusfjord
WL314	Flakstad	VR311477	Road-cut, Nusfjord
WL315	Flakstad	VR324623	Road-cut, 200m E of Myrland
WL316	Flakstad	VR342613	Road-side outcrop, 3km SE of Myrland
WL317	Flakstad	VR353589	Quarry, 200m NE of Napp ferry
WL318	Flakstad	VR307578	Road-cut, 1km NW of Vareid road junction
WL319	Flakstad	VR310552	Foreshore, 2km S of Varied road junction
WL320	Flakstad	VR290529	Road-cut, 1km S of Bø
WL321	Flakstad	VR303481	Rocky slop between two small lakes
WL322	Flakstad	VR299480	Hummocky ground, 100m west of small lake
WL323	Flakstad	VR299480	Hummocky ground, 100m west of small lake
WL324	Flakstad	VR303481	Rocky slope between two small lakes
WL325	Flakstad	VR308422	Track-side outcrop, west of Nusfjord
WL326	Flakstad	VR311490	Road-cut, 11m N of Nusfjord
WL327	Flakstad	VR315498	Hill-side stream (steep), E of Storvatnet
WL328	Flakstad	VR318498	Hill-top (300m), E of Storvatnet
WL329	Flakstad	VR321499	Hill-top (300m), E of Storvatnet
WL330	Flakstad	VR312501	Road-cut, E of Storvatnet, 2km N of Nusfjord

WL331	Flakstad	VR304505	Road-cut, N of Storvatnet, 3km N of Nusfjord
WL332-335	Moskenesøy	VR248457	Road-cut, 1km NW of Sund Harbour
WL336	Moskenesøy	VR252454	School playground, Sund
WL337	Moskenesøy	VR254446	Road-cut, peninsula, east of Sund Harbour
WL338	Moskenesøy	VR255451	Road-cut, peninsula, east of Sund Harbour
WL339-341	Flakstad	VR273457	Foreshore, Skjelford to Nesland road, 1.5km N of southernmost point
WL342	Flakstad	VR287448	Road-cut, 200m SW of Nesland
WL343-345	Flakstad	VR285445	Cliff-top, 500m SW of Nesland
WL346	Flakstad	VR275445	Road-side outcrop, Nesland peninsula
WL347-351	Flakstad	VR312490	Road-cut, N end of Nusfjord, 1km N of Nusfjord village
WL352	Flakstad	VR311477	Road-side outcrop, Nusfjord
WL353-361	Flakstad	VR301479	Hill, south of small lake, 1km W of Nusfjord
WL362	Flakstad	VR349575	Road-cut, 300m S of Napp, across harbour
WL363	Flakstad	VR349585	Road-cut, 300m S of Napp, across harbour
WL364	Flakstad	VR353589	Quarry, 200m NE of Napp ferry
WL365	Flakstad	VR353589	Quarry, 200m NE of Napp ferry
WL366	Moskenesøy	VR255448	Road-cut, peninsula, east of Sund Harbour

Vestvågøy

WL401	Flakstad	VR365549	Road-cut, 2km S of Sund
WL402	Flakstad	VR369576	Road-side outcrop, 1km N of Sund
WL403	Flakstad	VR437531	Cliff behind fish-racks, E end of Mortsund
WL404	Flakstad	VR433533	Road-cut, N end of Mortsund
WL405	Flakstad	VR470551	Road-cut, N end of Ure
WL406	Stamsund	VR517573	Road-cut, 1km west of Stamsund
WL407	Flakstad	VR487610	Road-cut, 100m S of Storfjord road junction
WL408	Flakstad	VR468611	Road-cut, 2km W of Storfjord road junction
WL409	Flakstad	VR468611	Road-cut, 2km W of Storfjord road junction
WL410	Flakstad	VR371668	Hill-side (steep) above road, 1.5km W of Utakleiv
WL411	Flakstad	VR390649	Road-cut, 1km SW of Vik
WL412	Stamsund	VR542645	Road-cut, east side of Raalsfjord
WL413	Stamsund	VR591648	Road-cut, 3km east of Valberg
WL414	Stamsund	VR631683	Road-cut, 1km S of road-junction to Malnes
WL415	Kvalnes	VR652726	Road-cut, between Smorten and Gimsøy Bridge
WL416	Kvalnes	VR635749	Road-cut, 500m E of Haugen
WL417	Eggum	VR430719	Road-cut, 2km SE of Unstad
WL418	Kvalnes	VR531762	Quarry, Ramberg
WL419	Kvalnes	VR530761	Road-cut, 200m south of Ramberg Quarry

WL420	Kvalnes	VR594787	Fore-shore, 4km SE of Kvalnes
WL421-425	Flakstad	VR426610	Road and car part exposures, Leknes Airport and Telecom Co.
WL426	Flakstad	VR396613	Foreshore, 2.5km NE Lilleeidet Ferry Terminal
WL427-435	Flakstad	VR390649	Road-cut, 1km SW of Vik
WL436-439	Flakstad	VR390638	Road-cut and hill-side exposure, 1km south of Vik samples
WL440-448	Flakstad	VR396633	Road-cut, 1.5km east of Våje (WL427-435)
WL449-454	Kvalnes	VR531762	Quarry, Rambert
WL455-461	Kvalnes	VR524759	Hillocks, on either side of rough road to Haukland
WL462-464	Eggum	VR478740	Road-cut, opposite outlet of Borgfjord into Norwegian Sea
WL465-469	Eggum	VR485731	Quarry, 2km NW of road junction at Bøstad
WL470	Kvalnes	VR652718	Iron mine tip, Smorten
WL471	Stamsund	VR517573	Road-cut, 1km west of Stamsund
WL472	Kvalnes	VR652718	Irone mine tip, Smorten
WL473	Kvalnes	VR652718	Iron mine tip, Smorten
WL474	Flakstad	VR635749	Road-cut, 500m E of Haugen
WL475	Flakstad	VR426610	Road-cut, Leknes Airport
WL476	Flakstad	VR391637	Hill-side, 1km south of Vik gneiss samples
WL477	Flakstad	VR391635	Hill-side, 1.2km S of Vik gneiss samples
WL478	Flakstad	VR391634	Loose block, 1.3km S of Vik gneiss samples
WL479	Flakstad	VR392634	Road-cut, 1km east of Våje
WL480	Flakstad	VR396633	Road-cut, 1.5km east of Våje

Austvågøy

WL501	Stamsund	VR680663	Road-cut, 600m W of road junction at Rørvik
WL502	Stamsund	VR680663	Road-cut, 600m W of road junction at Rørvik
WL503	Stamsund	VR745654	Road-cut, south end of Orsnesvågen
WL504	Stamsund	VR745654	Road-cut, south end of Orsnesvågen
WL505	Stamsund	VR732669	Road-cut, opposite road junction, to Hopen
WL506	Stamsund	VR708668	Tunnel entrance (east), NW of Hopen
WL507	Stamsund	VR703665	Tunnel entrace (west), NW of Hopen
WL508	Stamsund	VR678656	Road-cut, 1km SW of Rørvik
WL509	Stamsund	VR676652	Road-cut, 1.5km SW of Rørvik
WL510	Stamsund	VR676651	Road-cut, 1.6km SW of Rørvik
WL511-513	Stamsund	VR675651	Fore-shore, 100m W of road, 1.6km SW of Rørvik
WL514	Stamsund	VR678644	Road-cut, through rocky spur, N of Djupfjord
WL515	Stamsund	VR674635	Road-cut, 1.2km N of Festvåg
WL516	Stamsund	VR674635	Road-cut, 1.2km N of Festvåg

WL517	Stamsund	VR672630	Road-cut, 600m N of Festvåg
WL518	Stamsund	VR672625	Road-cut, 200m NW of Festvåg ferry terminal
WL519	Kabelvåg	VR771690	Lake-side exposure, SW of Store Kongsvatnet
WL520	Kabelgåg	VR771683	Road-side exposure, 500m S of Store Kongsvatnet
WL521	Kabelvåg	VR771671	Quarry, beside inlet 1km W of Kabelvåg
WL522	Kabelvåg	VR785669	Quarry, across road from Kabelvåg harbour
WL523-525	Kabelvåg	VR794677	Quarry/road-cut 1km NE of Kabelvåg church
WL526	Kabelvåg	VR803682	Road-cut, 2km NE of Kabelvåg church
WL527-530	Kabelvåg	CR811694	Quarry/road-cut, 500m west of Svolvaer tunnel
WL531	Kabelvåg	VR789703	hill-side, 200m from track, NE of Store Kongsvatnet
WL532	Kabelvåg	VR801692	Road-cut, 100m south of Lille Kongsvatnet
WL533	Kabelvåg	VR832702	Road-cut, 100m east of tunnel, East Svolvaer
WL534	Kabelvåg	VR852704	Quarry, east end of tunnel, Helle Airfield
WL535	Oddvaer	VR861712	Quarry/road-cut, 200m N of Husvågen Causeway
WL536	Oddvaer	VR816741	Quarry/road-cut, peninsula E of Vatterfjordpollen
WL537	Oddvaer	VR880768	Road-cut, 2km S of Sildpollen, W of Austnesfjord
WL538	Oddvaer	VR930865	Road-cut, 5km S of Fiskebøl, W of Storerfjord
WL539	Oddvaer	VR935881	Road-cut, Bretvik, 3km S of Fiskebøl
WL540	Oddvaer	VR936896	Quarry/road-cut, 1.5km S of Fiskebøl
WL541-548	Stamsund	VR682622	Foreshore section, east side of headland east of new Henningsvaer Bridge
EL549-562	Stamsund	VR679623	Road-cut, west side of headland on approach to new Henningsvaer Bridge
WL563	Stamsund	VR672629	Road-cut, 500m N of Festvåg
WL564	Stamsund	VR672629	Road-cut, 500m N of Festvåg
WL565	Stamsund	VR679643	Road-cut, through rocky spur, N of Djupfjord
WL566	Stamsund	VR679643	Road-cut, through rocky spur, N of Djupfjord
WL567	Stamsund	VR673631	Road-cut, 700m N of Festvåg
WL568	Stamsund	VR673631	Road-cut, 700m N of Festvåg
WL569	Stamsund	VR731649	Foreshore, 100m S of old Hopen
WL570	Stamsund	VR731651	Hillock, beside track NE from Hopen
WL571	Stamsund	VR731651	Hillock, beside track NE from Hopen
WL572	Stamsund	VR676647	Foreshore, 100m W of road, 800m NW of Djupfjord
WL573	Stamsund	VR676647	Foreshore, 100m W of road, 800m NW of Djupfjord
WL574	Stamsund	VR677645	Roadside outcrop, 400m NW of Djupfjord
WL575	Stamsund	VR677655	Foreshore, 50m W of road, 1.2km SW of Rørvik
WL576	Stamsund	VR677655	Foreshore, 50m W of road, 1.2km SW of Rørvik
WL577-579	Kabelvåg	VR677655	Loose block, hillside, N side of road, 2km NE of Kabelvåg church

WL580	Kabelvåg	VR811694	Quarry/road-cut, 500m W of Svolvaer tunnel
WL581	Kabelvåg	VR811694	Quarry/road-cut, 500m W of Svolvaer tunnel
WL582	Oddvaer	VR861741	Quarry/road-cut, peninsula E of Vatterfjordpollen
WL583	Stamsund	VR672629	Road-cut, 500m N of Festvåg
WL584	Oddvaer	VR931872	Road-cut, 4km S of Fiskebøl, W of Sloverfjord

Hinnøy

WL601	Våje	WR082913	Road-cut, NE of Kongsmarkvatnet, 1km SE of Kongselven
WL602	Våje	WR075915	Road-cut, 200m S of Kongselven Junction
WL603	Våje	WR067902	Road-cut, north end of Tengelfjord
WL604	Våje	WR057896	Road-cut, south end of Tengelfjord
WL605	Våje	WR047879	Quarry, on inside of sharp bend, NW of Raften
WL606	Våje	WR047869	Road-cut, 1km SW of Raften
WL607	Våje	WR048869	Road-cut, N end of Brubrekken
WL608-610	Våje	WR038859	Road-cut, 1km N of Tennstrand
WL611-613	Våje	WR031843	Roadlcut, north side of Tenfjord
WL614	Våje	WR028829	Road-cut, Finkirken
WL615	Våje	WR022815	Road-cut, 1km south of Finkirken
WL616	Våje	WR016808	Road-cut, 1.5km S of Finkirken
WL617	Våje	WR013804	Road-cut, 1.6km S of Finkirken
WL618	Oddvaer	WR012804	Road-cut, 2km NE of Digermulen Church
WL619	Oddvaer	WR006796	Road-cut, 2km NE of Digermulen Church
WL620	Oddvaer	WR006796	Road-cut, 1.8km NE of Digermulen Church
WL621	Oddvaer	WR005795	Road-cut, 1km NE of Digermulen Church
WL622	Oddvaer	WR002789	Road-cut, 1.5km SE of Digermulen Church
WL623	Våje	WR005769	Quarry, west of Pundslett, Årstein
WL624	Våje	WR021765	Road-cut, 1.5km N of Pundslett
WL625	Våje	WR026778	Road-cut, Steinsvik
WL626	Våje	WR040769	Road-cut, Økneshamn, E side of Øksfjord
WL627	Våje	WR138832	Road-cut, Fornes, E side of Øksfjord
WL627	Våje	WR174879	Quarry, 1.5km SE of Vestbygd Church
WL629	Våje	WR216806	Road-cut, 200m W of junction to Offersøy
WL630	Våje	WR253797	Road-cut, 200m W of junction to Offersøy
WL631	Lødingen	WR253797	Quarry/road-cut, 6km NE of Offersøy
WL632	Lødingen	WR292822	Quarry, Rindbø, S side of Hestfjord
WL633	Lødingen	WR303879	Road-cut, 2km N of Dalheim

WL634	Lødingen	WR303879	Road-cut, 2km N of Dalheim
WL635-638	Våje	WR082913	Road-cut, NE of Kongsmarkvatnet, 1km SE of Kongselven
WL639	Våje	WR076917	Hill-side, by road to Kongselven Jetty
WL640	Våje	WR076917	Hill-side, by road to Kongselven Jetty
WL641	Våje	WR067920	Road-cut, north end of Tengelfjord
WL642	Våje	WR061896	River, under bridge, south end of Tengelfjord
WL643	Våje	WR055873	River, unver bridge, Raften
WL644	Våje	WR038859	Road-cut, north end of Brubrekken
WL645-647	Våje	WR028829	Road-cut, north side of Tenfjord
WL648	Våje	WR034828	River, under bridge, east end of Tenfjord
WL649-653	Våje	WR034828	Road-cut, east end of Tenfjord
WL654	Våje	WR022815	Road-cut, Finkirken
WL655	Våje	WR016808	Road-cut, 1km south of Finkirken
WL656	Våje	WR016808	Road-cut, 1km south of Finkirken
WL657	Våje	WR013804	Road-cut, 1.5km south of Finkirken
WL658	Oddvaer	WR006796	Road-cut, 2km NE of Digermulen Church
WL659	Oddvaer	WR006796	Road-cut, 2km NE of Digermulen Church
WL660	Våje	WR021765	Quarry, west of Pundslett, Årstein
WL661	Sortland	WS241127	Road-cut, 4km SE of Sigerfjord Church
WL662	Sortland	WS241127	Road-cut, 4km SE of Sigerfjord Church
WL663	Gullesfjord	WS338153	Road-cut, 1km SW of Gombogen
WL664	Gullesfjord	WS320140	Road-cut, 2km N of junction, W of Gullesfjord
WL665-669	Lødingen	WR364980	Quarry, 2.8km N of Kanstad road-junction
WL670-674	Lødingen	WR359941	Quarry, 1.3km S of Kanstad road-junction
WL675-680	Lødingen	WR357931	Quarry, 2.5km S of Kanstad road-junction.
WL681	Lødingen	WR351920	Road-cut, 3.9km S of Kanstad road-junction
WL682	Lødingen	WR351920	Road-cut, 3.9km S of Kanstad road-junction
WL683	Lødingen	WR348917	Road-cut, Bugten, 4.3km S of Kanstad junction
WL684-686	Våje	WR293797	Road-cut, 200m W of junction to Offersøy
WL687	Lødingen	WR313903	Road-cut, 5km N of Dalheim, W of Høkfjord
WL688	Lødingen	WR342907	Road-cut in isolated hillock, 1.5km N of Erikstad
WL689	Lødingen	WR342908	Hill-side exposure, W of road, 1.5km N of Erikstad
WL690	Lødingen	WR361861	Quarry, Nes, 6km SW of Lødingen
WL691	Lødingen	WR342907	Road-cut in isolated hillock, 1.5km N of Erikstad

Langøy

WL701	Nykvåg	VS804306	Hillock, 1.5km NE of Nykvåg
WL702	Nykvåg	VS806309	Road-cut, 1.8km NE of Nykvåg
WL703	Nykvåg	VS812323	Road-cut, around Tussen, 2km S of Hovden
WL704	Nykvåg	VS822342	Road-cut, by jetty, Hovden Harbour
WL705	Nykvåg	VS819342	Quarry, SW corner of Hovden Harbour
WL706	Nykvåg	VS816341	Hill-side exposure, 300m SW of Hovden Harbour
WL707	Nykvåg	VS813328	Promontory, 400m from road, 1.5km S of Hovden
WL708	Nykvåg	VS818301	Hillside exposure, west flank of Selvågtind
WL709	Nykvåg	VS819303	Hillside exposure, west flank of Selvågtind
WL710	Stokmarknes	VS777214	Track-side outcrop, N of Vikanfjord
WL711	Stokmarknes	VS801107	Road-cut, 2km SW of Bø Church
WL712	Stokmarknes	VS785208	Road-cut, head of Vikanfjord, Straumsjøen
WL713	Stokmarknes	VS791111	Road-cut, inside of bend in road, SW of Steine
WL714	Stokmarknes	VS781109	Cutting for harbour buildings, SW of Steine
WL715	Stokmarknes	VS771185	Hummocky ground, 300m S of Skarvågen junction
WL716	Nykvåg	VS786269	Road-cut, Sandvik
WL717	Nykvåg	VS802303	Hillock, 1.2km NE of Nykvåg
WL718	Nykvåg	VS804306	Loose block, 1.5km NE of Nykvåg
WL719-721	Nykvåg	VS819342	Quarry, SW corner of Hovden Harbour
WL722	Stokmarknes	VS959255	River, under causeway, S end of Auenfjord
WL723-728	Nykvåg	VS990300	Road-cut, S side of Skjelfjord, 1km NW of Skjelfjordeid
WL729-731	Stokmarknes	VS978172	Road-cut, 300m SW of Grønning Church, S side of Eidsfjord
WL732	Stokmarknes	VS963157	Road-cut, Mosanden, S side of Eidsfjord
WL733	Stokmarknes	VS963157	Road-cut, Mosanden, S side of Eidsfjord
WL734	Stokmarknes	VS992177	Road-cut, 300m NE of Slåtnes, S side of Eidsfjord
WL735	Stokmarknes	VS999180	Road-cut, 1km NE of Slåtnes, S side of Eidsfjord
WL736	Stokmarknes	VS999180	Road-cut, 1km NE of Slåtnes, S side of Eidsfjord
WL737	Stokmarknes	VS776276	Peninsula, 2km south of Nykvåg

Appendix 2

Mineral assemblages in the Rocks of the Lofoten-Vesterålen Province

The minerals present in each rock are listed in approximately their order of abundance. Key to the abbreviations used:

ab	albite	ksp	K-feldspar
act	actinolite	musc	muscovite
all	allanite	myrm	myrmekitic intergrowth
ap	apatite	ol	olivine
bi	biotite	opx	orthopyroxene
cc	calcite	ore	opaques
chl	chlorite	parg	pargasite
chon	chondrodite	phl	phlogopite
cpx	clinopyroxene	plag	plagioclase
di	diopside	qz	quartz
do	dolomite	ru	rutile
ep	epidote	sp	spinel
fo	forsterite	sph	sphene
gl	glaucophane	tr	tremolite
grun	grunerite	zr	zircon
gt	garnet	zs	zoisite
hb	hornblende		
hd	hedenbergite		

Appendix 2: Petrography

Petrography of the Late Archaean Migmatites

Moskenes, Moskenesøy

WL237: Mafic phases c.25% of the rock. Stringer mesoperthites up to 5mm across. Rare plagioclase (An_{35}). Biotite is the dominant mafic; it commonly shows undulose extinction, as does mesoperthite, indicating deformation. Subordinate clinopyroxene and opaque minerals with minor orthopyroxene and accessory apatite and garnet. Opaques rimmed by biotite and/or garnet. Minute garnet and quartz rim opaques, biotite and clinopyroxene. Opx rimmed by biotite, hornblende and garnet.

plag - Ksp - bi - cpx - ore - opx - ap - gt - qz - hb

Festvåg, Austvågøy

WL549: Mafic phases total <1%: opaques, and very rare biotite, zircon and apatite. Granular texture, up to 5mm grain size. Mesoperthite and quartz c.95% of the rock. Minor plagioclase.

Ksp - plag - qz - ore - bi - ap - zr

WL550: Mafics c.2% of the rock: hornblende rimmed by biotite, opaques rimmed by biotite. Grain size up to 2mm. Less quartz than WL549. Mafics define foliation.

WL551: As WL550.

WL552: Mafics <1% of the rock. Myrmekite up to 10mm across.

Ksp - plag - qz - ore - bi ap

WL553: Mafics c.2% of the rock. 5mm wide quartz vein cuts the rock.

Ksp - plag - qz - bi - ore - ap

WL554: Mafics c.10% of the rock. Fine-scale perthites. Quartz only present in myrmekite, which is up to 5mm across. Opaques overgrown by orthopyroxene. Opaques, hornblende and opx have biotite coronas.

Ksp - plag - opx - bi - hb - qz - ap - zr

WL555: Mafics c.50% of the rock. Green hornblende with abundant opaque inclusions. Opaques overgrown by biotite and hornblende rims. Plagioclase An_{35}. Grain size c.1mm.

plag - hb - bi - ore - ap

WL556: Mafics c.50% of the rock. Green hornblende; orthopyroxene; plagioclase, An_{40}; and biotite (more abundant than in WL555). Grain size c.1mm.

plag - hb - bi - opx - ore - ap

Pyroxenite vein, up to 10mm wide, crosses slide. Pink orthopyroxene overgrown by more abundant clinopyroxene. Interstitial mesoperthite and rare plagioclase.

cpx - ksp - plag - opx - ore - bi - plag

WL557: Mafics c.60% of the rock: biotite and hornblende with subordinate orthopyroxene, overgrown by hornblende. Garnet coronas common around opaques. Accessory apatite. Biotite defines foliation. Grain size <1mm.

hb - plag - bi - ore - opx - ap - gt

WL558: As WL557, except WL558 has no garnet coronas.

hb - plag - opx - ore - bi - ap

WL559: Mafics c.30% of the rock. Grain size <0.5mm. Biotite defines foliation.

plag - opx - bi - cpx - ore - hb

WL560: As WL558. Aggregates of hornblende up to 2mm across.

hb - plag - bi - ore - opx - cpx

WL561: Mafics c.20% of the rock. Grain size c. 1mm. Biotite defines foliation.

plag - ksp - bi - ore - opx - cpx - ab - hb

Band through slide has mesoperthite up to 5mm across, cpx 2mm.

ksp - plag - cpx - ore - ap - bi - qz - zr

WL562: Mafics c.5%, occur in clots of biotite, orthopyroxene, clinopyroxene and opaques. Grain size up to 3mm.

ksp - plag - qz - opx - cpx - bi - ore - ru - sph - ap

Ridderset, North-West Hinnøy

WL661: Mafics c.25% of the rock: mainly biotite with pleochroic haloes related to zircon. Biotite defines foliation. Grain size <1mm.

qz - bi - plag - ore - musc - zr - ap

WL662: Mafics c.20% of the rock. It has compositional banding unlike WL661. 5mm wide band of quartz, grain size 0.5-4mm. 2-4mm selvage to quartz band of c.80% green biotite with quartz, plagioclase, apatite and zircon. Most of the rock is similar

to WL661 but it also contains microcline and myrmekites and has no opaques.

qz - bi - plag - ksp - zr - ap

Sandset, Central Langøy

WL723: Mafics c.15%, grain size 0.1-1mm. Opaques and orthopyroxene are the dominant mafics, with accessory biotite. No foliation.

ksp - qz - plag - ore - opx - ap - bi

WL724: Mafics c.10%, grain size up to 6mm. Extensive development of myrmekite and recrystallization of quartz and feldspars at grain boundaries. The dominant mafic mineral is an opaque in interstices of feldspars and quartz. Sphene associated with opaques. Accessory orthopyroxene. No foliation.

ksp - qz - plag - ore - opx - sph

WL725: Mafics c.10%, grain size up to 4mm. Altered orthopryoxene and opaques are the dominant mafic phases, with minor biotite. Some of the alkali feldspar is perthitic. Weak foliation.

ksp - qz - plag - opx - ore - bi - ap

WL726: Mafics c.60%, grain size 0.5-1mm. Dominant mafic phases are brown amphibole, orthopyroxene and biotite, with minor clinopyroxene and accessory opaques. No foliation.

plag - hb - opx - bi - cpx - ore - ap

WL727: Mafics. c.20%, grain size up to 5mm, some recrystallization at grain boundaries of quartz and feldspars. Pleochroic orthopyroxene. Some of the alkali feldspar is perthitic. Weak foliation.

ksp - qz - plag - opx - ore - bi - zr - ap

WL728: Pronounced compositional banding, mafic-rich and mafic-poor.

Mafic-rich: 10mm wide, mafics c.20%, very similar to WL727.

ksp - qz - plag - opx - ore - bi - ap

Mafic-poor: 20mm wide, mafics <1%, quartz up to 10mm across.

qz - ksp - plag - ore - ap

1mm wide shear zone, oblique to foliation.

Petrology of Early Proterozoic Supracrustal Rocks

Ølkona Marble Locality, Moskenesøy

WL212: 80-90% mesoperthite, from <1mm to 10mm. Occasional plagioclase (An_{40}). Biotite and opaques of ~1mm dimensions - fine-grained garnet rims. Accessory apatite and zircon.

ksp - plag - bi - ore - ap - zr - gt

MONZONITE DYKE (Green and Jorde, 1971)

WL213: Monomineralic clinopyroxenite. 5mm grains with fine-grained granular cpx recrystallized along margins and cleavage traces.

cpx PYROXENITE

WL215: 80% granular calcite 2-5mm across. Phlogopite is present in aggregates up to 10mm across and as individual grains, most are highly deformed. Diopside and chondrodite are also present.

cc - phl - di - chon

PHLOGOPITE MARBLE

WL214: Fine-grained altered rock with relict cpx up to 2mm across. Some bi and gt may be 2mm, but most are <0.5mm.

cpx - hb - bi - plag - gt - ore

GARNET AMPHIBOLITE

WL216: Plagioclase and orthoclase occasionally perthitic up to 10mm across. Also green clinopyroxene. Microfractures common, filled with calcite.

plag - ksp - cpx - cc

MONZONITE DYKE (Green and Jorde, 1971)

WL217: Ksp with very fine-scale perthitic texture, also plagioclase and green cpx, 5-10mm grain size; similar to WL216.

ksp - plag - cpx - op

ANATECTIC MANGERITE VEIN

but grades into pyroxenite layer via 15mm wide band of 1mm cpx and feldspar grains, where cpx is pale green. Mineralogy of this band:

plag - cpx - ksp - bi - ore - ap

Pyroxenite layer has 3mm pale green fpx grains, nonmineralic in parts but it also contains a biotite and plagioclase bearing band. Very rare calcite and opaques.

cpx - bi - plag - ore - cc

PYROXENITE

WL219: ~1mm grains of pale green amphibole, biotite and pleochroic opx. Bright red rutile is more common than opaque oxide phase.

hb - bi - opx - ru - ore

GRANULITE-FACIES HORNBLENDITE

WL220: Green cpx grains up to 2mm across with dusty opaque inclusions. Opaques rimmed by hornblende and fine-grained intergrowths of bi - hb - gt - qz - ksp are common.

cpx - gt - ore - hb - bi - qz - ksp

GARNET PYROXENITE

WL221: 1-2mm very pale green cpx in granular texture. Anhedral green spinel overgrows cpx. Fractures across rock infilling with calcite.

cpx - sp - cc

SPINEL PYROXENITE

WL222: Layered into carbonate and silicate bands. Silicate layer is granular hornblende up to 5mm across with cpx up to 3mm across and rare opx, biotite and garnet are common in parts.

hb - cpx - bi - gt - opx

HORNBLENDITE

Carbonate layer is largely fine-grained calcite and diopside with pods of hb - qz intergrowth from pyroxene break-down, but there are patches of granular calcite about 1mm across.

cc - cpx - hb - qz

DIOPSIDE MARBLE

WL223: Granular calcite up to 3mm across, recrystallised in parts, overgrown by rounded to sub-rounded olivines, 1-2mm across, most of which are completely altered to serpentine. Minor phlogopite

cc - fo - phl

FORSTERITE MARBLE

WL240: Similar to WL223, except the calcite is more extensively recrystallized, dolomite occurs, few olivines are serpentinized, and diopside is present. The latter overgrows olivines and some is deformed. Minor phlogopite. Chondrite up to 3mm across.

cc - fo - di - do - chon - phl

FORSTERITE DIOPSIDE MARBLE

WL241: Similar to WL240, except diopside is less common and dolomite and phlogopite more so, calcite recrystallizes around serpentinised olivines with which an opaque phase is associated. Chrondrite.

cc - fo - do - phl - di - chon - ore

FORSTERITE MARBLE

Contains a pyroxene-rich band, cpx of 1-2mm with phlogopite and fine-grained, recrystallized calcite.

cpx - phl - cc

PYROXENITE

WL244: Diopside and alkali feldspar with granular texture, calcite and zircon also present, cross-cut by 14mm wide explosion breccia. Not strictly brecciated since fragments (up to 11mm across) are rounded. Fragments have similar composition to host rock. Also some small very low relief fragments of analcite (?) 0.5mm wide margins to the brecciated part, paler than the rest of the matrix. Opposite sides match, suggesting dilatational opening. Probably a gas release structure carrying fragments for some distance, rounded by abrasion.

EXPLOSION BRECCIA

1km SW of Ølkona Marble Locality, Moskenesøy

WL256: Pyroxene-rich and plagioclase-rich bands. The former has pale green diopside up to 5mm across, the latter has plagioclase crystals of similar size and with curved albite twin lamellae. The margin between the two bands contains intensely deformed biotites with fine-grained garnet rims and rare hornblende.

cpx - plag - bi - hb - ore - gt

DEFORMED GABBRO

WL257: Biotite-rich and spinel-rich bands. The former has biotite laths up to 3mm long with opx and cpx; minor phases are ore, green spinel, apatite and plagioclase.

bi - opx - cpx - ore - ap - sp - plag

BIOTITE PYROXENITE

The spinel-rich band consists of green spinel up to 3mm across and orthopyroxene. Accessory apatite, and calcite in fractures.

opx - sp - ap - cc

SPINEL PYROXENITE

WL258: Plagioclase (An_{55}) up to 7mm across; strained extinction, curved twin lamellae. Anhedral opx, strained reddish-brown biotite. quite common zircon and apatite.

plag - opx - bi - ap - zr

NORITE

WL259: Perthites up to 5mm across are 80% of the slide. Cpx, opx and ore are the main mafic phases with accessory ap and zr.

ksp - opx - cpx - ore - ap - zr

MONZONITE (Green and Jorde, 1971)

Hamnøy Bridge, Moskenesøy

WL225: Perthites up to 5mm across are 60-70% of the slide. Reddish colour to some of cpx; calcite associated with breakdown of cpx. Also minor opx and biotite, some rimming opaques. Accessory ap and zr.

ksp - cpx - opx - ore - bi - cc - ap - zr

LAYERED MONZONITIC GNEISS (Green and Jorde, 1971)

WL246: Perthites up to 12mm. Cpx and opaques are main mafic phases with minor opx and biotite. Mafic phases 5-10%.

ksp - cpx - ore - opx - bi - ap - cc

MONZONITE LAYER (Green and Jorde, 1971)

Most of the rock is plagioclase and perthite with cpx, opx, ore, bi and ap, with layering due to layers where cpx = 60% and is reddish-brown in colour.

ksp - plag - cpx - opx - ore - bi - ap

LAYERED MONZONITIC GNEISS (Greem and Jorde, 1971)

WL248: Layers of different mineralogy:

plag - opx - ore - cpx - bi

cpx - opx - ore - ap - bi

First layer has grain size of 0.5mm, second layer is up to 4mm.

LAYERED GABBROIC GNEISS (Green and Jorde, 1971)

WL250: Granular opx and cpx up to 4mm across with minor opaques and interstitial plagioclase.

opx - cpx - ore - ap - plag

LAYERED GABBROIC GNEISS (Green and Jorde, 1971)

WL252: Granular texture of cpx and opx with calcite and quartz filling interstices. Minor phlogopite and apatite. Anhedral yellow garnet present.

cpx - opx - cc - gt - qz - ap - phl

CALCITE-BEARING PYROXENITE

WL254: Quartz grains up to 5mm across in matrix of qz - plag - cpx - opx of 0.5-2mm grain size. Both cpx and feldspar have abundant inclusions producing reddish-brow- colour. Accessory rutile.

qz - plag - cpx - opx - ore - ru

LAYERED QUARTZ DIORITIC GNEISS

(Green and Jorde, 1971)

WL255: Fine-grained matrix <1mm grain size with bands of 2mm grain size. Plag - opx - cpx - ore 1 qz with accessory bi for 60-70% of slide with coarser layers of ksp - plag - qz - ore opx.

plag - opx - ksp - ore - cpx - qz - bi - ap - cc

VEINED AND LAYERED GNEISS WITH MONZONITIC LAYERS (Green and Jorde, 1971)

Å-Tind-Sørvågen, Moskenesøy

WL230: Very fine-scale perthitic feldspar, 1-3mm across. Opaque phase with exsolved green spinel and very fine-grained garnet rims. Cpx overgrown by opx. Apatite up to 2mm long. Accessory biotite and muscovite. Mafics 15-20%.

ksp - ore - cpx - opx - ap - bi - musv - sp - gt

MONZONITE DYKE (Green and Jorde, 1971)

WL231: Contact between gneiss and pegmatite

plag - bi - ksp - hb - ore - ap

LAYERED GNEISS (Green and Jorde, 1971)

Gneiss has grain size of about 1mm, pegmatite has up to 12mm grain size

ksp - plag - qz - bi - ore - myrm

GRANITE PEGMATITE (Green and Jorde, 1971)

WL232: Very fine scale perthitic feldspar up to 12mm across, coloured pink by very fine inclusions. Cpx, ore and apatite total about 5%.

ksp - plag - cpx - ap - ore

MONZONITE DYKE (Gree- and Jorde, 1971)

WL233: Perthites 0.5-2mm across

ksp - plag - cpx - ore - opx - ap - qz - zr

LAYERED MONZONITIC GNEISS (Green and Jorde, 1971)

WL235: 1-2cm plagioclase in granular texture with opx and cpx - all have pinkish colour due to inclusions. Abundant biotite defining gneissosity. Opaques all have well-developed garnet reaction rims, some have an inner rim of biotite. Minor hornblende. Layers parallel to gneissosity have plagioclase and pyroxenes up to 4mm across, and no biotite or hornblende.

plag - cpx - bi - ore - opx - gt - hb - ap

VEINED AND LAYERED GNEISS WITH MONZONITIC LAYERS (Green and Jorde, 1971)

WL236: Intensely deformed, ksp - plag - bi - ore - cpx - opx - ap - gt gneiss. 20mm wide ultramylonite zone in which all minerals are crushed to very small grains and where garnet is abundant. In marginal zone relict pyroxenes remain in augen texture.

ULTRAMYLONITE

North-West of Nesland, Flakstadøy

WL339: Mafic phases 25%. Granular texture, grain size up to 2.5mm. Plagioclase is the dominant phase, alkali feldspar rare. Orthopyroxene and opaques are the main mafic phases. Minor cpx and biotite. Foliation defined by grain size variation and opx-rich layers.

plag - opx - ore - ksp - cpx - bi

LAYERED DIORITIC GNEISS (Romey, 1971)

WL340: Mafic phases 25%. Grain size typically <1mm, but some plagioclase and mesoperthite up to 6mm across. Stringer mesoperthite more abundant than plagioclase. Recrystallization at grain boundaries. Pleochroic orthopyroxene grains up to 3mm across, with abundant opaque inclusions. Apatite up to 1mm across. Rare clinopyroxene.

ksp - plag - opx - ore - ap - cpx

MANGERITE DYKE

WL341: Mafic phases <1%. Mesoperthite and quartz up to 10mm across. Accessory orthopyroxene, opaques, zircon and rutile.

ksp - plag - qz - ore - opx - zr - ru

CHARNOCKITE DYKE

Vik, West Vestvågøy (Quartzo-feldspathic Gneisses)

WL427: Mafic phases c.3%, mostly biotite and opaques define foliation. Accessory cpx, zircon, muscovite, apatite and epidote (with allanite cores). Main phases are quartz, microcline and plagioclase. Grain size up to 1.5mm.

qz - ksp - plag - bi - ore - musc - cpx - zr - ap - ep - all

WL428: As above.

WL429: Mafic phases 5-10%. Green biotite laths up to 5mm long with zircon and sphene inclusions, zircon up to 2mm long. Well-developed foliation.

qz - ksp - plag - bi - zr - ap - sph - ore - ep - all

WL430: As above, except much more sphene and muscovite.

qz - ksp - plag - bi - sph - zr - ap - musc - sp - ep - all

WL431: As above.

WL432: Mafic phases c.2%. Dark brown biotite up to 1.5mm long. Yellow-brown allanite cores to epidotes. Accessory rutile, sphene, muscovite- opaque and zircon.

qz - ksp - plag - bi - ep - all - musc - zr - sph - ru - ore

WL433: Mafic phases c.2%. Almost as much muscovite as biotite. Accessory zircon, sphene, rutile and opaques. Epidote commonly has darker cores of allanite (syn.:orthote).

qz - ksp - plag - bi - musc - ep - all - sph - zr - ru - ore

WL434: Mafic phases. c.5%. Epidote most abundant mafic after biotite. Single opaque grain, 2mm across. Accessory apatite, zircon, rutile and muscovite.

qz - ksp - plag - bi - ep - all - ore - ap - zr - musc - ru

30 x 10mm black area in hand-specimen with surrounding halo - concentration of allanite.

WL435: Mafic phases c.15%. As with all the Vik Gneiss samples, quartz, microcline and plagioclase are the main phases, with grain size up to 2mm. Green biotite is the dominant mafic

phase and defines the foliation. Accessories phases are sphene, zircon, apatite, hornblende, opaques, epidote and allanite.

qz - ksp - plag - bi - sph - zr - ap - hb - ore - ep - all

Ramberg Quarry, North Vestvågøy

WL449: Strong compositional banding. Quartzo-feldspathic layers may have <2% mafics, amphibolite layers >60% mafics. Microcline and quartz rarely exceed 2mm grain size, but biotite and hornblende may be 3mm long and are alligned parallel to the compositional banding. Hornblende is strongly pleochroic: straw yellow to blue-green. Sphene, epidote and plagioclase are minor phases, with accessory opaques, apatite, zircon and myrmekite.

ksp - qz - hb - bi - sph - ep - plag - ore - ap - zr - myrm

LAYERED GRANITIC GNEISS

WL450: Mafics <1%. Granular quartz-microcline with many fractures up to 0.5mm wide, infilled by calcite. Mafic phases are opaques, biotite and epidote.

qz - ksp - cc - ore - bi - ep

QUARTZ MICROCLINE GNEISS

200m South of Ramberg Quarry, North Vestvågøy

WL419: Augen of alkali feldspar (perthites) up to 35mm across. The augen contain biotite, opaques, apatite, quartz, amphiboles, pyroxenes and patches of myrmekite, but these total <5% of total volume of the augen. Layers of biotite and hornblende in the gneiss wrap around the augen. Most of the gneiss is very fine-grained feldspathic quartzite. Biotite-hornblende layers also include relict ortho- and clinopyroxene rimmed by hornblende. Opaques may be rimmed by garnet. Myrmekites are common at the margin of the augen. Order of abundance of minerals, excluding the augen:

qz - hb - bi - ksp - cpx - ore - myrm - opx - gt

AUGEN GNEISS

Austnesfjord, Austvågøy

WL537: Quartz, plagioclase, and microcline, 1-2mm across in granular texture. Mafic phases c.20%, occur in bands which define foliation. The bands consist mainly of biotite with euhedral epidote up to 1.5mm across which show twinning and zoning. Euhedral sphene is up to 1mm across. Many of the plagioclase grains are sausseritized, but have unaltered rims, implying that they are zoned, with Na-rich resistant to saussuritization. Accessory muscovite, opaques, apatite and calcite.

qz - plag - ksp - bi - ep - sph - musc - ore - ap - cc

QUARTZO-FELDSPATHIC GNEISS

Erikstad, South Hinnøy

WL688: Green clinopyroxene grains, about 1mm. Many fine-grained granular pyroxene areas. Opaques (sulphides) occur in irregular bands through pyroxenite. In places the sulphides are 70% of the rock.

cpx - ore - gt - sph - plag

SULPHIDE-BEARING PYROXENITE

WL689: Bands of grunerite and hedenbergite with darker green amphibole in cores of grunerite. Grains up to 2mm. Opaques overprint amphibole and pryoxene texture.

ore - grun - hd - hb

BANDED IRONSTONE

Petrography of Mid Proterozoic Basic Intrusions

Ølkona Spotted Gabbro, Moskenesøy

WL224: Coarse-grained (up to 10mm) hornblende grains replacing orthopyroxene and minor clinopyroxene. Plagioclase (An_{50}) between mafic aggregates. Garnet coronas. Ultramylonites.

hb - opx - plag - cpx - ore - bi - ap - gt

Flakstadøy Basic Complex

Nesland

WL310: Grain size rarely exceeds 2mm. Plagioclase (An_{52}) is 60-70% of total rock. Pleochroic orthopyroxene main mafic phase, with opaques between grains. Minor secondary biotite. Garnet reaction rims to opaques.

plag - opx - ore - bi - gt

NORITE

WL311: Plagioclase up to 2.5cm long (An_{48}). Mafics c.15%. Olivine up to 4mm long overgrown by orthopyroxene and clinopyroxene. Secondary biotite overgrows opaques and associated green spinel. Olivine altering to opaques, iddingsite and spinel.

plag - opx - cpx - ol - ore - bi - sp - zr

GABBRONORITE

WL344: Granular texture of perthites and quartz up to 10mm occurs. Single clinopyroxene with opaques and rutile with garnet coronas.

ksp - qz - ore - cpx - ru - gt

CHARNOCKITE VEIN

WL345: Fine-scale perthites up to 7mm across, coloured pink by minute inclusions as for WL232 at Å, Moskenesøy. Minor quartz in interstices. Pale green clinopyroxene up to 4mm across. One cpx completely enclosed by an opaque. Accessory orthopyroxene, apatite, zircon and rutile. Garnet coronas to cpx and opaques.

ksp - cpx - qz - ore - opx - ap - zr - gt - ru

CHARNOCKITE VEIN

WL346: Plagioclase, up to 3mm across, ca. 70% of rock. Minor anti-perthite. Abundant apatite up to 1mm long. Orthopyroxene more abundant than clinopyroxene. Opaques rimmed by garnet and rarely

biotite and garnet.

plag - opx - cpx - ore - bi - ap - ksp - gt

GABBRONORITE

Nusfjord

WL314: Plagioclase (ca.80% of the rock) and olivine up to 15mm long. Orthopyroxene-clinopyroxene-spinel-garnet coronas between olivine and plagioclase as described by Griffin and Heier (1969). Olivine and opaques are overgrown by orthopyroxene and biotite.

plag - ol - opx - ore - bi - sp - gt

TROCTOLITE

WL324: Plagioclase (ca.80% of the rock) and olivine (overgrown by orthopyroxene) are primary phases. Plagioclase is up to 12mm long; minute oriented opaque needle inclusions. Symplectite of biotite, opx and opaques. Reddish-brown secondary biotite, up to 2mm across. Double coronas to olivine as for WL314. Green spinel associated with opaques. Rare clinopyroxene.

plag - ol - ore - bi - opx - sp - cpx - gt

TROCTOLITE

WL325: Plagioclase (ca.85% of the rock) and olivine (rarely overgrown by orthopyroxene, usually with double corona at contact with plagioclase). Grain size up to 12mm. Minute, oriented, opaque needles in plagioclase and orthopyroxene.

plag - ol - opx - ore - bi - sp - ru - gt - cpx

TROCTOLITE

WL326: Plagioclase (ca.75% of the rock) and orthopyroxene (with some clinopyroxene), up to 15mm. Rare opaques have biotite-garnet rims.

plag - opx - cpx - ore - bi - gt

NORITE

WL329: Plagioclase occupies 95% of slide, grains up to 15mm, An_{60}. Strongly pleochroic pink-green orthopyroxene. Opaques with exsolved green spinel. Opaque needles in plagioclase.

plag - opx - ore - sp - bi

ANORTHOSITE

WL330: Plagioclase (ca.80% of the rock) and olivine (with orthopyroxene overgrowths and very well-developed opx - cpx - sp - gt coronas). Abundant minute opaque needles in plagioclase. Also reaction between biotite and plagioclase - giving rise, probably, to a clinopyroxene-spinel symplectite.

plag - ol - opx - ore - bi - sp - gt - cpx

TROCTOLITE

WL321: Quartzo-feldspathic rock with strong hear-foliation. Augen of perthite (up to 20mm) and orthopyroxene (up to 10mm) between crush zones consisting of quartz.

ksp - qz - opx - ore - ru

CHARNOCKITE VEIN

Mørkdalen (1km west of Nusfjord)

WL322: Plagioclase (An_{45}), approx. 70% of the mode, has albite- and Carlsbad-twinning, 20 x 8mm. Clino- and orthopyroxene up to 1mm across.

plag - cpx - opx - ore - bi

NORITIC GABBRO

WL323: Very fine-grained (except garnet up to 3mm). Strong foliation in places - similar to ultramylonites in late Archaean gneisses. Green hornblende is the major fine-grained mineral. Tremolite and zoisite are present. Garnet commonly has a spongy core with minute inclusions and a massive, idioblastic rim, suggesting two stages of growth. One garnet has a pink, inclusion - free core, a dusty inner rim, and a colourless outer rim. These garnets with spongy cores are similar to those in the eastern Alps (Droop, 1983). The spongy cores may imply former presence of omphacite rimmed by garnets. Diopside-albite patches are abundant, as described for eclogitic amphibolites by Droop (1983) - jadeite component of omphacite reacts with quartz to form albite, leaving diopside component.

ab - gt - di - hb - tr - ore - zs - bi - ru

ECLOGITIC AMPHIBOLITE

WL353: Clinopyroxene up to 2.5mm across (also minor orthopyroxene) overgrown by hornblende (with minor biotite) and by later garnet. Plagioclase altered to finegrained mass of albite with hornblende and opaque inclusions.

ab - cpx - hb - gt - ore - opx - bi - ru

AMPHIBOOLITIZED GABBRO

WL354: No original igneous pyroxenes remain. Compositional banding between albite-rich and hornblende-rich domains. Zoisite formed by sausseritization of plagioclase. Pleochroic blue amphibole present - glaucophane. Ilmenite to rutile inversion present - high pressure inversion to Ti-mineral with smallest molar volume.

ab - hb - gt - di - ore - bi - ru - gl - zs

ECLOGITIC AMPHIBOLITE

WL355: As WL354, garnets c.1mm across with pink cores and colourless rims.

WL356: Very strong shear fabric; all minerals reduced to very fine grain-size except garnets, up to 3mm across, which act as augen in the hornblende-biotite-rich layers. Continuous layers of opaques and rutile up to 15mm long and c.0.2mm thick.

hb - ab - gt - ore - bi - ru - gl

ECLOGITIC AMPHIBOLITE

WL357: Relatively unaltered norite. Plagioclase up to 7mm across c.70% of rock. Quartz rods in clinopyroxene. Pink-green pleochroic orthopyroxene. Pyroxenes rimmed by hornblende. Opaques rimmed by biotite. Garnet rims quite rare - developed on biotite and less commonly on hornblende.

plag - opx - cpx - ore - bi - hb - gt - az

GABBRONORITE

WL358: Plagioclase saussuritized but individual grains still recognizable. Relict orthopyroxenes up to 3mm across. Garnets up to 2mm across.

ab - hb - opx - bi - ore - gt - cpx - zs - gl - ru

AMPHIBOLITIZED GABBRONORITE

WL359: Outline of plagioclase still visible. More rutile and less relict pyroxene than WL358.

ab - hb - gt - bi - ore - cpx - opx - zs - ru

ECLOGITIC AMPHIBOLITE

WL360: Very similar to WL356 with strong shear fabric, except WL360 has higher proportion of felsic minerals.

WL361: Very little albite in this rock (in fact a series of five slides collected over a few m^2). Fine-grained hornblende, pyroxenes and pink to orange garnets up to 7mm across are >90% of the rock. Rutile (more common than opaques) form stringers up to 15 x 1mm.

Relict pale-greenpyroxenes, up to 1mm across, could be omphacite. Some garnets have cloudy cores and darker, Fe-rich rims. Rare blue amphibole (glaucophane) preserved within garnets.

gt - cpx - hb - ru - ab - ore - gl

ECLOGITE

Napp

WL362: Opaques up to 7mm across are main primary mafic phase, later overgrown by biotite and then by orthopyroxene. Opaques-orthopyroxene and biotite-orthopyroxene symplectites.

plag - ore - cpx - bi - opx - sp - gt - ol

ORE-BEARING GABBRO

WL363: Ophitic texture of orthopyroxene, opaques and plagioclase (some with exsolved antiperthitic potassium feldspar) with minor biotite; grain size up to 0.5mm. Euhedral plagioclase up to 10mm across. Carlsbad and albite twinning in both groundmass and phenocryst plagioclase.

plag - opx - ore - bi - ksp

PORPHYRITIC NORITE (Romey, 1971)

WL317: Fine-grained (about 0.5mm across) plagioclase is 60% of the rock. Minor antiperthite. Garnet rims between opaques and plagioclase, and clinopyroxene and plagioclase.

plag - cpx - ksp - ore - opx - bi - gt

GABBRO

Borge Gabbro, Vestvågøy

WL462: Very similar to WL323, except there is no foliation and it is not as fine-grained. Garnets, up to 3mm across, full of minute inclusions in the cores, most have clear rims. Fine-grained intergrowth of albite and hornblende (but the latter may be up to 3mm across) make up the bulk of the rock. Biotite-rich patches are abundant, calcite-rich patches less so. Accessory zoisite, opaques and rutile.

ab - hb - gt - bi - cc - zs - ore - ru

GARNET AMPHIBOLITE

WL465: Dusty plagioclase (An_{42}) up to 3mm, 50% of rock. Olivine and clinopyroxene, c.1mm across, latter overgrows former and is much more abundant; both contain opaques, as needles along cleavage planes in cpx, irregular in olivine. biotite overgrows cpx and fine-grained garnet rims are developed. Minor alteration of cpx to hornblende. Accessory calcite.

plag - cpx - bi - ore - ol - gt - hb - cc

GABBRO

WL467: Orthoclase (some perthitic) and quartz make up about 80% of the rock. Orthopyroxene (10mm) and a single opaque (7mm) are the main mafic phases. Opx is rimmed by hornblende and garnet. Shear zones contain a carbonate phase. Zircon (2mm) and metamict allanite (3mm) are quite abundant.

ksp - qz - plag - opx - ore - all - zr - hb - cc - gt

CHARNOCKITE PEGMATITE

WL468: Contact between gabbro (identical to WL465, except no olivine), and quartzo-feldspathic vein c.30cm wide. Quartz and very fine-scale perthites up to 10mm across with minor plagioclase. Other phases total 5% of rock. Opaques up to 4mm across with fine-grained garnet rims plus accessory zircon and rutile.

ksp - qz - ore - plag - gt - zr - ru

CHARNOCKITE VEIN

Vik-Våje, Vestvågøy

WL437: Plagioclase, Carlsbad and albite twinned, up to 5mm long. Orthopyroxene, up to 10mm long, pink-green pleochroic, abundant opaque needles along cleavage, hornblende rims. Garnet rims to biotite. Opx + plag → hb. Bi + plag → gt. Clinopyroxene less common than orthopyroxene and overgrown by it. Subophitic texture.

plag - opx - cpx - bi - ore - hb - gt

GABBRONORITE

WL438: Altered version of WL437. Relict ortho- (and some clino-) pyroxene with bi + hb + gt rims. Otherwise similar to WL462. Pink garnets, many euhedral, up to 1mm across, as rims to hornblende-albite-biotite-opaque pseudomorphs after pyroxenes. Also blue amphibole-glaucophane: "occurs during retrograde

metamorphism of eclogite" (Deer et al., 1966, p.183). Accessory apatite and rutile. Patches of white mica in albite.

ab - hb - gt - opx - bi - ore - cpx - gl - ap - ru - musc

GARNET AMPHIBOLITE

WL439: Similar to WL438 except no relict pyroxenes.

ab - hb - gt - bi - ore - sph - ru

GARNET AMPHIBOLITE

WL440: Very similar to WL437, except more clino- than orthopyroxene.

plag - cpx - opx - bi - ore - hb - gt

GABBRO

WL441: Alkali feldspar (some perthitic) >95%, up to 20mm across. Quartz up to 3mm, but most occurs in fine-grained patches along feldspar grain boundaries and in small shear zones with opx and garnet.

ksp - qz - opx - gt

CHARNOCKITE PEGMATITE

WL442: Four slides from single sample showing gabbro-eclogite transition:

D: gabbro akin to WL437 and WL440. Opx easily distinguished from cpx: the former is pink-green pleochroic and inclusion free, the latter has minute inclusions colouring it brown, plus ore needles following cleavage traces. Both opx and cpx are being altered to hb and bi. Garnet rims common around biotite only. Dusty plagioclase, up to 3mm long.

plag - cpx - opx - bi - hb - ore - gt - ru

GABBRO

C: further breakdown of pyroxenes to hb + bi + gt. Continued growth of ore neeldes in cpx. Plagioclase clouded with fine needles.

plag - opx - bi - hb - cpx - gt - ore - ru

B: most pyroxenes completely pseudomorphed to hb - ab - bi - gt symplectites. White mica in albite-rich areas.

hb - ab - gt - bi - opx - cpx - ru - ore - musc

A: similar to WL438. Garnet up to 2mm across. Rare relict cpx, no opx seen. White mica. Opaques altered to rutile.

ab - hb - gt - bi - ru - ore - opx - musc

ECLOGITIC AMPHIBOLITE

WL443: Similar to WL442A, except no relict pyroxene, white mica patches are commonly developed in albite-rich patches often associated with small needles of zoisite, biotite is rare, and rutile is present.

ab - hb - gt - ru - ore - musc - zs - bi

GARNET AMPHIBOLITE

WL444: As WL443, except more biotite, less rutile and garnet, and fine-grained diopside-albite aggregates.

ab - hb - gt - di - bi - ore - ru - musc - zs

GARNET AMPHIBOLITE

WL445: As WL443, except that zoisite is very abundant and up to 1mm long and garnet is less abundant.

ab - hb - zs - gt - bi - ore - ru - musc

GARNET AMPHIBOLITE

WL446: >95% of slide is a pale-green clinopyroxene with minute inclusions producing dusty appearance. The rest of the rock consists of olivine altered to iddingsite, plus opaques and very rare hornblende.

cpx - ol - ore - hb

PYROXENITE

WL447: Pyroxenite, as WL446 without olivine, in contact with garnet-biotite-rich rock, which is fine-grained and contains quartz and plagioclase. There is a garnet-hornblende contact zone between the pyroxenite.

WL448: Very rich in hornblende, in places c.90%. Most of the slide is fine-grained aggregate of hornblende, albite and garnet with abundant aggregates of rutile. Accessory sphene and opaques. Acicular actinolite.

gt - hb - ab - ru - sph - ore - act

GARNET AMPHIBOLITE

WL476: Strong foliation in garnet amphibolite, defined by biotite laths, zoisite needles and by lenses of very fine-grained diopside. Also lenses of quartz and hornblende.

ab - hb - di - gt - bi - ore - qz - ru

GARNET AMPHIBOLITE

WL477: Garnet, individual grains up to 3mm across, occurring in aggregates up to 10mm across. Some garnets have spongy cores. Pale green actinolite (?) is c.60% of the rock. Minor albite, quartz and biotite. Accessory opaques, rutile and sphene.

act - gt - ab - qz - bi - ore - ru - sph

GARNET AMPHIBOLITE

WL478: Garnet, up to 5mm across, with many inclusions; some are full of rutile, some full of plagioclase. Garnet is 50% of slide's area. Most of the rest is pale green actinolite (?), grains up to 3mm long. Opaques altered to rutile.

gt - act - ru - qz - ab - ore - sph

GARNET AMPHIBOLITE

WL479: Garnet up to 3mm across. Shear zones through rock are calcite-rich.

qz - gt - hb - ore - ru - ab - cc

GARNET AMPHIBOLITE

Gabbro, West of Ramberg

WL455: Relict ortho- and clinopyroxene up to 3mm across. Opx-opaque symplectites. Most pyroxenes altered to hornblende and biotite, both up to 5mm across. Garnets, many euhedral up to 0.5mm across, present as incomplete rims to hornblende. Plagioclase altered to fine-grained mass rich in white mica. Apatite up to 1mm.

hb - opx - cpx - ab - bi - gt - ore - ap - musc

HORNBLENDE GABBRONORITE

WL456: No pyroxenes. Hornblende up to 10mm across. More garnet than WL455. Late vein filled by plagioclase and acicular actinolite (?). Two types of hornblende - one green, less abundant yellow-green - occurring in separate domains.

hb - ab - gt - bi - ore - ru - act

GARNET AMPHIBOLITE

WL457: Foliation defined by biotite, zoisite and hornblende alignment Blue amphibole-glaucophane. Garnet rims around hornblende-rich areas are complete in many places. Single patch of relict pyroxenes.

hb - gt - ab - zs - bi - ore - gl - cpx - opx - ru

GARNET AMPHIBOLITE

WL458: Relict pyroxenes. Abundant zoisite. No foliation.

hb - ab - gt - bi - zs - cpx - ore - ap - gl - ru

GARNET AMPHIBOLITE

WL459: Quartz and perthites up to 19mm across, but also <1mm, especially at grain boundaries. Single clinopyroxene altering to hornblende with garnet rim, single 3mm long biotite grain, single paque with garnet rim. Accessory zircon.

ksp - qz - bi - cpx - ore - hb - zr - gt

CHARNOCKITE VEIN

WL460: Four slides of sheared garnet amphibolite patches in sheared charnockite:

A: clinopyroxene up to 4mm across, with well-developed opx exsolution lamellae. Opx grains less common. Pyroxenes altering to quartz and hornblende. Plagioclase saussuritized. Spongy garnet up to 1.5mm across.

cpx - hb - ab - gt - opx - bi - qz - musc - ore - sph - ru

AMPHIBOLITIZED GABBRO

B: amphibolitized gabbro in fine-grained quartzo-feldspathic matrix with <1mm size biotite laths defining foliation. Relict alkali feldspar up to 2mm across, and quartz and biotite aggregates behave as augen.

ksp - qz - bi - ore - sph - ru

AMPHIBOLITIZED GABBRO IN CHARNOCKITE VEIN

C: as above, sheared quartzo-feldspathic rock. Alkali feldspar, up to 10mm across, cut by fine-grained shears.

ksp - qz - bi - musc - ore - sph - gt - ru

CHARNOCKITE VEIN

D: as WL460C, except for 5mm orthopyroxene augen altering to hornblende and quartz. Biotite foliation wraps around pyroxene. Opaques altered to rutile.

CHARNOCKITE VEIN

WL461: Very fine-grained quartz - ksp - bi - musc rock with very strong foliation, defined by stringers of quartz up to 20 x 1 mm with individual grains up to 2mm across. Clinopyroxene augen around which foliation is wrapped.

ksp - qz 1 bi - cpx - musc - gt - sph

CHARNOCKITE VEIN WITH AUGEN

Hovden Gabbro, Langøy

WL708: Plagioclase, An_{55}, up to 5mm long, and olivine up to 3mm across are the primary phases. Symplectites of orthopyroxene and opaques associated with biotite and green spinel. Rare clinopyroxene,with opaque needles along cleavage traces, overgrown by opx. Reddish-brown biotite in aggregates up to 5mm across.

plag - ol - opx - ore - bi - sp - cpx

TROCTOLITE

WL709: Rounded olivine up to 2mm across, altering to serpentine and opaques, is about 80% of the rock. Minor clinopyroxene. Plagioclase, fractured, interstitial to olivines. Green spinel. Brown amphibole, probably ferrohastingsite as WL716.

ol - plag - cpx - hst - ore - sp

PERIDOTITE

WL716: Rounded olivines, c.2mm across, altering to serpentine and opaques; volume change causes cracks radial to olivine fracturing the surrounding plagioclase. The olivines may be enclosed by greenish-brown amphibole altering to brown amphibole along cracks - probably pargasite. Accessory spinel.

plag - ol - parg - bi - opx - ore - sp

TROCTOLITE

WL717: Well-developed foliation defined by segregations of quartz and alkali feldspar. Quartz, individual grain size up to 2mm across occurs as narrow bands or lenses with fine-grained feldspar between. Other phases total <5%.

ksp - qz - cpx - ore - opx - ep - sph - ru

CHARNOCKITE VEIN

WL718: Very coarse rock. Clinopyroxene, with abundant opaque inclusions, up to 20mm across, including olivine and rutile up to 2mm across, with plagioclase, An_{65}, up to 15mm long, between the euhedral clinopyroxene. Orthopyroxene associated with olivine and rutile in the cores of the large cpx.

cpx - plag - ol - ru - opx - ore

OLIVINE GABBRO

WL720: Granophyric intergrowth of quartz and alkali feldspar. Quartz up to 10mm across. Some alkali feldspar is perthitic. Green clinopyroxene up to 3mm long.

ksp - qz - cpx - ore - sph - ru

GRANOPHYRIC VEIN

Bø Gabbro, Langøy

WL711: Grain size c.0.5-1mm. Plagioclase c.60% of slide, some is antiperthitic. Pale-green clinopyroxene with less abundant pink-green pleochroic orthopyroxene.

plag - cpx - bi - opx - ore - ap

GABBRO

Eidsfjord Gabbro, Langøy

WL736: Coarse-grained, pink-green pleochroic orthopryoxene. One 20mm long opx broken by kink bands into pieces 0.2-4mm wide. Clinopyroxene in exsolution lamellae and recrystallized along kink fractures. Opx is c.95% of the slide.

opx - ore - plag - cpx - bi

ORTHOPYROXENITE

Petrography of the Mid Proterozoic Mangeritic Intrusions

The Sund-Ølkona Mangerite

Ølkona, Moskenesøy

W1202: Antiperthitic plagioclase up to 7mm across, >90% of the rock. Green clinopyroxene up to 5mm across, with rare garnet rims.

plag - ksp - cpx - ore - ap - gt

WL204: Mesoperthite (very fine-scale) grains with sutured margins, up to 5mm across comprise >90% of the rock. Green clinopyroxene with minor alteration to hornblende and garnet rims.

cpx + plag → gt + qz.

ksp - plag - cpx - ore - ap - gt - qz - hb - opx

WL210: Mesoperthites up to 20mm across. Green clinopyroxene with rare colourless orthopyroxene. Garnet rims to pyroxenes and opaques.

ksp - plag - cpx - ore - opx - ap - gt

WL211: Mesoperthite up to 5mm across. Opaques rimmed by garnet.

ksp - plag - opx - ore - cpx - ap - gt - zr

Sund, Flakstadøy

WL302: Plagioclase, quartz (c.20%) and orthoclase up to 2mm across, but most of the rock is granular texture of above minerals, <0.5mm across. Fine-grained aggregates of hornblende, quartz and biotite, and opaques.

ksp - qz - plag - ore - hb - bi - zr

RETROGRADED CHARNOCKITE

WL303: Intergrown mesoperthite and quartz c.1-3mm. across with fine-grained recrystallization along margins. Orthopyroxenes exsolution lamellae in clinopyroxene. 0.5mm zircon. Reactions present:

opx + plag → gt + qz

ore + plag → gt + qz

ksp - plag - qz - opx - ore - gt - zr - cpx - ru

WL332: Plagioclase grains <1mm across are c.60% of the rock. Relict orthopyroxene, altering to aggregates of hornblende. Most pyroxenes completely altered to hornblende, rimmed by garnet

and quartz:

$$cpx + plag + H_2O \rightarrow hb + gt + qz$$

plag - hb - bi - ore - opx - ksp - gt - qz

XENOLITH

WL333: Coarse perthites up to 15mm across and granular quartz aggregages are >95% of the rock. The remainder is aggregates of green clinopyroxene, hornblende, quartz, calcite, orthopyroxene, opaques, biotite, rutile and garnet.

ksp - plag - qz - cpx - hb - cc - opx - ore - bi - ru - gt

WL334: Granular quartz, plagioclase and K-feldspar grains, 0.1-1.5mm across. Hornblende-quartz-biotite pseudomorphs after pyroxenes. Pink, sieve-like garnets.

qz - plag - ksp - bi - gt - hb - ore - ep

RETROGRADED CHARNOCKITE

WL335: Similar to WL303. Complete and partial pseudomorphs after pyroxenes.

ksp - qz - plag - opx - bi - gt - ore - cpx - cc

WL336: Equigranular, c.0.5mm, plagioclase, K-feldspar and clinopyroxene. Opaques rimmed by biotite and subsequently by garnet. Mafics c.40%.

plag - ksp - cpx - ore - bi - gt

XENOLITH

WL337: Mesoperthite up to 5mm c.80% of the rock. Green clinopyroxene with orthopyroxene exsolution lamealle and quartz and garnet coronas.

ksp - plag - cpx - opx - ore - ap - gt - qz - cc

WL366: Fine-grained granular matrix of quartz, plagioclase and K-feldspar c.50% of the rock. Reddish-brown biotite grains in optical continuity define foliation. Pale-green clinopyroxene up to 2mm across but most is fine-grained.

qz - cpx - plag - ksp - bi - gt - ore

XENOLITH

The Flakstadøy Charnockite

WL306: Quartz (c.30%) and mesoperthite grains, up to 4mm across, with sutured margins are c.98% of the rock. Minor opaques overgrown by biotite and garnet. Rare relict orthopyroxene. Rock traversed by shear zones.

ksp - plag - qz - ore - bi - gt - opx - ap

The South-West Lofoten Mangerite

WL307: Mesoperthite and plagioclase up to 5mm across c.85% of the rock. Aggregates of pyroxenes, opaques, biotite (some primary) and apatite are common. Rare quartz.

opx + plag → bi

plag - ksp - opx - cpx - ore - bi - ap - qz - gt

WL309: Similar to WL307, except feldspars only c.80% of the rock. Apatite even more abundant, more quartz, orthopyroxene much more abundant than clinopyroxene.

plag - ksp - opx - ore - ap - cpx - bi - qz - gt

WL315: Pyroxene-opaque-biotite-apatite aggregates in plagioclase and perthite matrix. Perthites are much less abundant than plagioclase in the rock and the mafic aggregates are c.25% of the rock. 5mm wide ultramylonite. Primary biotite.

ore + plag → hb (+ bi + gt)

bi + plag → hb

opx + plag → hb

ore + plag → gt

bi + plag → gt

plag - opx - ksp - cpx - ore - ap - bi - gt - hb - qz

WL316: Plagioclase with minor mesoperthite and accessory quartz, up to 4mm across, totals c.80% of the rock. Orthopyroxene is the main mafic phase, overgrown by hornblende. Opaques may be rimmed by biotite-hornblende intergrowths. Primary biotite.

plag - ksp - opx - cpx - ore - bi - ap - hb - qz - gt

WL319: Mafic phases <5% of the rock. Mesoperthite, up to 10mm across, with fine-grained recrystallization and myrmekite development at grain margins. Opaques with garnet rims. Relict orthopyroxene.

ksp - plag - qz - ore - bi - opx - ap - gt

WL331: Mafic phases c.10% of the rock. Fine-scale perthites and plagioclase, up to 5mm across with sutured margins and abundant inclusions of minute opaque needles. Pyroxenes, opaques and apatite occurring in aggregates. No garnet coronas.

plag - ksp - opx - ore - cpx - ap - bi - ep - zr

WL401: Mafic phases, c.10% of the rock. Coarse perthites up to 10mm across, with exsolved patches of plagioclase. Orthopyroxene (and accessory clinopyroxene) altering to biotite and opaques. Apatite up to 1mm long. Biotite coronas to opaques - laths grow perpendicular to the margin and up to 0.1mm long. Rare quartz and myrmekite.

opx + plag → bi + hb + gt

plag - ksp - opx - ore - ap - bi - hb - qz - cpx - gt

WL402: Very similar to WL401, with the addition of a single sphene.

WL403: Very coarse bleb mesoperthites as well as the usual ribbon perthites. Similar to WL401. 2mm wide ultramylonite zone.

WL404: Hornblende-quartz-biotite pseudomorphs after pyroxenes, up to 10mm across, also contain apatite and opaques. Garnets up to 0.5mm across. Rare relict pyroxenes.

plag - ksp - bi - qz - ore - hb - ap - gt - cpx - ru

RETROGRADED MANGERITE

WL405: Mesoperthite, up to 10mm across, is c.95% of the rock. Orthopyroxene with clinopyroxene exsolution lamellae. Biotite up to 2mm long. Opaques with garnet coronas.

plag - ksp - opx - cpx - ore - bi - ap - qz - gt - zr

WL410: Mafic phases c.20% of the rock. Plagioclase is more abundant than perthites and up to 3mm across. Some pyroxenes completely pseudomorphed by hornblende, quartz and biotite, others with quartz and hornblende coronas. opx + plag → hb + qz

plag - ksp - opx - bi - ore - hb cpx - gt - ap - qz - sph

WL416: Mafic phases c.20% of the rock. Plagioclase is much more abundant than perthites and up to 2mm across. Biotite laths, some rimmed by garnet, are the dominant mafic phase. Opaques with garnet coronas (and also biotite). Rare relict pyroxenes and myrmekite.

plag - bi - ksp - ore - cpx - gt - opx - hb - ap - qz - sph

RETROGRADED MANGERITE

WL417: Plagioclase and vernicular mesoperthites up to 10mm across. Very fine-grained granular zones. Mafics c.10% of the rock. Pyroxenes altered to hornblende-quartz-biotite pseudomorphs, rimmed by garnet.

plag - ksp - cpx - bi - hb - opx - ore - ap - gt - qz

WL436: Perthites with myrmekite containing aggregates of hornblende, biotite, quartz, opaques, apatite, garnet and relict pyroxenes.

ksp - plag - hb - bi - cpx - ore - opx - ap - gt - qz - sph

RETROGRADED MANGERITE

Includes fine-grained area (<0.5mm)

ksp - plag - bi - ore - gt - cpx - hb - ap

XENOLITH

WL471: Mafic phases c.40% of the rock. Aggregates of pryoxenes with minor biotite and hornblende, up to 2mm grain size, in equigranular texture pyroxenes, plagioclase and K-feldspar. Weak foliation.

ksp - plag - opx - cpx - bi - hb - ore

XENOLITH

The Hopen Mangerite/Charnockite

WL508: Mesoperthite and quartz up to 1.5mm across ic c.80% of the rock. Green clinopyroxene, up to 0.5mm across, orthopyroxene, hornblende and opaques (the latter rimmed by biotite and garnet) comprise the rest of the rock, with accessory apatite.

$$cpx + plag + H_2O \rightarrow hb\ (+ qz)$$

$$opx + plag + H_2O \rightarrow hb\ (+ qz)$$

ksp - plag - qz - ore - cpx - hb - opx - bi - gt - ap - qz

WL510: Mesoperthite and quartz, up to 4mm across is c.80% of the rock. Hornblende - quartz-biotite pseudomorphs after pyroxenes, rimmed by garnet.

$$ore + plag \rightarrow hb + gt$$

$$opx + plag + H_2O \rightarrow hb + qz$$

ksp - plag 1 qz - opx - cpx - ore - bi - hb - ap - gt - ru - zr

WL514: Mesoperthite, up to 4mm across, is c.60% of the rock and has sutured margins with quartz which is c.40% of the rock. Aggregates of dark green-brown clinopyroxene and opaques with garnet and quartz coronas. Mesoperthite commonly exsolved to plagioclase and orthoclase. Some pyroxenes completely pseudomorphed.

plag - ksp - qz - cpx - ore - opx - gt - hb - ap - bi - zr

WL516: Mesoperthite, up to 4mm across, some exsolved to plagioclase and orthoclase, and minor quartz makes up c.90% of the rock. Rare myrmekite. ore + plag → gt (+ bi)

opx + plag → gt + qz

hb + plag → gt

Hornblende, up to 1.5mm across, enclose opaques.

plag - ksp - qz - opx - hb - ore - cpx - gt - bi - ap - zr

WL517: Foliation defined by pyroxene-rich and pryoxene-poor bands. Plagioclase, orthoclase and quartz in granular texture, grain size up to 2mm. No perthites. Bands of up to 70% green clinopyroxene.

plag - ksp - cpx - qz - ore - opx - gt - bi - ap - zr

WL518: Foliation defined by layers of mafic phases, which total c.15% of the rock. Mesoperthite up to 4mm across. Rare quartz and myrmekite. Green clinopyroxene up to 2mm, and apatite up to 0.5mm across.

ksp - plag - opx - cpx - ore - hb - ap - gt - qz - zr - ru

WL544: Layered rock.

Layer 1: clinopyroxene, up to 1mm across, pale green, c.95% of this layer. Minor opaques and altered plagioclase.

cpx - ore - plag CLINOPYROXENITE

Layer 2: 11mm wide. Equigranular, 0.5-1mm green hornblende is c.95% of this layer, the remainder is biotite laths with accessory opaques and altered plagioclase.

hb - bi - ore - plag - ap

IGNEOUS HORNBLENDITE

Layer 3: 5mm wide. Alternating bands of hornblendite as Layer 2, and finer-grained bands (0.1-0.5mm) of hornblende with c.30% clinopyroxene. No biotite or plagioclase.

hb - cpx - ore - ap CPX-BEARING HORNBLENDITE

Layer 4: 7mm wide. As Layer 2.

Layer 5: >15mm wide. Bands of pale-green clinopyroxene and biotite up to 0.5mm grain size; some monominerallic biotite bands, some bands c.80% plagioclase with minor biotite and clinopyroxene. Accessory opaques.

cpx - bi - plag - ore BIOTITE-BEARING CLINOPYROXENE

WL547: Mafics c.40% of the rock. Grain size up to 2mm. All minerals fractured in a single orientation, regardless of cleavage directions. Plagioclase c.60%. Opaques first-formed mafic. Clinopyroxene and minor orthopyroxene overgrows opaques. Accessory olivine. Primary biotite and hornblende are common. Hornblende coronas to biotite.

plag - cpx - ore - bi - opx - hb - ol - cc - ap

GABBRO

WL548: Apatite up to 15mm long. Plagioclase altering to calcite and epidote. Mafics c.50% of the rock. Clinopyroxene-opaque symplectites. Primary biotite and hornblende.

cpx - plag - ap - ore - bi - opx - hb - cc - ep

APATITE GABBRO

WL563: Foliation defined by layers of mafics, comprising c.5% of the rock. Some perthites, but most has exsolved to plagioclase and orthoclase. Orthopyroxene with exsolution lamellae of clinopyroxene. ore + plag → gt (+ bi)

opx + plag + H_2O → gt + qz

plag - ksp - opx - hb - cpx - qz - ore - gt - bi - ap

WL565: Quartz up to 20mm long and partially exsolved mesoperthites present in equal amounts. Hornblende-quartz-biotite pseudomorphs more common than relict pyroxenes.

qz - plag - ksp - hb - bi - ore - zr - ap - opx - cpx - gt - ru - musc

RETROGRADED CHARNOCKITE

WL567: Contact between foliated charnockite and amphiboite xenolith. At the contact is a green pyroxene-rich band. The amphibolite consists of hornblende, biotite and plagioclase grain size <1mm.

plag - hb - bi - ap XENOLITH

WL568: Mafic phases c.60%. Grain size <1mm, predominantly biotite and hornblende with euhedral garnet developed. Relict pyroxenes present. Feldspars are mainly plagioclase; some orthoclase. Quite similar to amphibolites at Festvåg, especially WL557.

plag - hb - bi - gt - ksp - ore - cpx - opx

XENOLITH

WL570: Coarse, flame mesoperthites and quartz up to 5mm across. Green clinopyroxene and an opaque are the main mafic phases which total

c.5% of the rock. Sphene and rutile up to 1mm across.

plag - ksp - qz - cpx - ore - sph - ru - opx - gt - hb - ap

WL572: Quartz and mesoperthite, often exsolving plagioclase and orthoclase, have sutured margins and are up to 3mm across. Mafics are c.5% of the rock. ore + plag → gt (+ qz)

cpx + plag + H_2O → hb + qz

Secondary hornblende is blue-green

plag - ksp - qz - cpx - hb - opx - ore - gt - ap

WL573: Quartz and mesoperthite, often exsolving plagioclase and orthoclase, have sutured margins and are up to 3mm across. Mafics are c.15% of the rock. Secondary hornblende after pyroxenes.

plag - ksp - qz - hb - ore - bi - ap - gt - zr - sph

RETROGRADED CHARNOCKITE

WL575: Foliation defined by layers of mafics which make up c.10% of the rock. Quartz, up to 3mm across, is c.50% of the rock; mesoperthite, c.40%. Abundant sphene up to 0.5mm long.

qz - plag - ksp - cpx - opx - hb - sph - ore - bi - ep

The Austvågøy Grey Mangerite

WL522: Mesoperthite, up to 4mm across, is c. 70% of the rock, quartz and mafic phases c.15% of the rock each. Perthites show exsolution to plagioclase and orthoclase. Pale-green clinopyroxene rimmed by dark-green hornblende and quartz. Biotite, hornblende and garnet all occur as rims to opaques.

plag - ksp - qz - cpx - ore - hb - opx - gt - ap - sph - bi - zr

WL525: Mesoperthite, up to 4mm across, is c.85% of the rock. Quartz is rare and mafic phases are c.15% of the rock. Hornblende rims to opaques and pyroxenes. Biotite and garnet rims the opaques. Hornblende grains up to 1.5mm across are primary.

plag - ksp - hb - cpx - ore - opx - ap - bi - qz - sph

HORNBLENDE MANGERITE

WL526: Quartz, up to 7mm long is c.30% of the rock, mesoperthite c.65% and mafics c.5%. Mafics are mainly orthopyroxene with clinopyroxene exsolution lamellae and hornblende and garnet rims, opaques and minor primary hornblende.

plag - ksp - qz - opx - cpx - ore - hb - gt - bi - zr sph - ap

WL529: Feldspars, up to 5mm across, are rarely perthitic, most has exsolved to plagioclase and orthoclase. Minor quartz and myrmekite is present. Rare orthopyroxene in the cores of primary magmatic hornblende up to 2mm across, with biotite rims. Mafics c.20% of the rock. Minor shear zones cut the rock. hb + plag → bi

plag - ksp - hb - ore - bi - qz - opx - gt - ap - sph

HORNBLENDE MANGERITE

WL530: Perthites, plagioclase and quartz, up to 5mm across are c.80% of the rock. Abundant Carlsbad twinned plagioclase. Pyroxenes have hornblende rims. Primary hornblende is full of dusty opaque inclusions, is up to 2mm across and occurs in aggregates with opaques, apatite and biotite.

plag - ksp - opx - hb - cpx - ore - qz - ap - sph - gt

WL533: Fine-scale perthites up to 10mm across in granular texture with quartz and mafics. Myrmekite developed between quartz and plagioclase. Colourless clinopyroxene overgrown by primary magmatic hornblende. Mafics total c.15% of the rock.

plag - ksp - opx - cpx - hb - ore - qz - ap - bi - zr - sph - musc

The Raftsund Mangerite

WL538: Very coarse 'graphic intergrowth' of plagioclase and orthoclase as perthite components exsolve. 1.5mm pale green clinopyroxene containing hornblende, biotite and euhedral opaques, overgrown by 0.2mm wide magmatic reaction rim of colourless orthopyroxene which itself contains opaques and apatite. Apatite up to 1.5mm across. Many magmatic overgrowths seen: one clinopyroxene, mantled by orthopyroxene, is subsequently overgrown by hornblende. Mafics c.10%. ore + plag → gt (+ bi)

plag - ksp - cpx - ore - opx - hb - ap 1 bi - gt - sph

CPX-OPX ZONE

WL539: Mesoperthite, up to 5mm across, c. 90% of the rock, some has exsolved to plagioclase and orthoclase. Minor quartz and myrmekite. Magmatic hornblende overgrows pyroxenes and opaques.

plag - ksp - opx - cpx - ore - hb - ap - gt - qz - bi

CPX-OPX ZONE

WL540: Mesoperthite, up to 5mm across, c.90% of the rock, much has exsolved separate feldspar phases. Minor plagioclase. Colourless orthopyroxene rims to pale green clinopyroxene. Minor magmatic hornblende. Some pryoxenes pseudomorphed by hornblende-quartz-biotite with garnet coronas. Rare patches of breakdown products of olivine surrounded by multi-layer coronas.

plag - ksp - opx - cpx - hb - ore - ap - bi - gt - ol - qz - sp

CPX-OPX-OL ZONE

WL101: Perthites and plagioclase, up to 7mm across, are c.75% of the rock. Mafics c.25% are largely retrograded to biotite-garnet-opaques, relict pyroxenes quite common. Feldspars <1mm across near retrograded spots. Apatite up to 1.5mm long.

plag 1 ksp - bi - cpx - gt - ap - opx - ore - ru

RETROGRADED MANGERITE

CPX-OPX ZONE

WL102: Perthites and plagioclase, up to 5mm long, c.80% of the rock. Pale green clinopyroxene, with opaque needle inclusions, grain size up to 3.5mm, with hornblende coronas. Otherwise similar to WL101.

plag - ksp - cpx - bi - gt - opx - ap - ore - hb

RETROGRADED MANGERITE

CPX-OPX ZONE

WL103: 10mm exsolved perthites and plagioclase partially saussuritized with epidote and white mica developed. Aggregates of biotite and quartz with garnet coronas - pseudomorphs after pyroxenes. Single pyroxene altering to hornblende. Apatite up to 1mm long.

plag - ksp - bi - hb - qz - ore - ap - cpx - ep - musc

RETROGRADED MANGERITE

CPX-OPX ZONE ?

WL104: Similar to WL103, except biotite-quartz-garnet aggregates are up to 6mm across. 5mm orthopyroxene grain with hornblende corona, but also breaking down to biotite and quartz. Sphene associated with opaques in pyroxene pseudomorphs.

plag - ksp - bi - qz - hb - ap - ore - gt - opx - cpx - sph

RETROGRADED MANGERITE

CPX-OPX ZONE

WL105: Perthites and plagioclase up to 5mm across, partially saussuritized cores, c.90% of the rock. 1.5mm clinopyroxene with inner corona of quartz and hornblende and complete outer rim of hornblende. Also complete hornblende-quartz pseudomorph surrounded by ring of sieve-like subhedral garnets. Sphene up to 1mm across, in cores of biotite aggregates. Hornblende may be blue-green.

plag - ksp - bi - hb - qz - gt - ap - sph - cpx - ore - cc

RETROGRADED MANGERITE

CPX-OPX ZONE

WL602: Mesoperthite and plagioclase grains up to 5mm across are c.90% of the rock. Mafics c.10% are largely pryoxenes and opaques. One 3mm orthopyroxene contains two pale green clinopyroxenes, 1mm across, which contain opx exsolution lamellae. The large orthopyroxene also contains opaques rimmed by primary hornblende, and apatite. Broad cpx lamellae present in opx grains, which may be overgrown hy clinopyroxene. Myrmekite present.

plag - ksp - opx - ore - cpx - hb - bi - ap - qz

CPX-OPX ZONE

WL603: Feldspars up to 10mm across. Plagioclase grains are more abundant than in WL602, also exsolution of perthites to plagioclase and orthoclase is more common. Garnet and hornblende very rare.

plag - ksp - opx - ore - bi - cpx - ap - gt - hb

CPX-OPX ZONE

WL604: Plagioclase, fine-scale perthites and exsolved plagioclase-oligoclase grains up to 10mm across is c.85% of the rock. Mafic phases include: 5mm primary biotite grain; clinopyroxene with euhedral opaque inclusions and overgrown by orthopyroxene; olivine, up to 2.5mm across and altered to an opaque and a bright red phase - iddingsite (?); and opaques rimmed by hornblende, biotite or garnet. Vernicular cpx-quartz intergrowth. Orthopyroxene occurs between cpx and olivine, suggesting it is a magmatic reaction product.

plag - ksp - cpx - opx - bi - ore - ol - ap - hb - gt - qz

CPX-OPX-OL ZONE

WL605: Plagioclase, ribbon perthites and exsolved perthites up to 10mm across. Mafics c.5% of the rock. Orthopyroxene, 1mm across with broad cpx exsolution lamellae, rimmed by magmatic hornblende.

Apatite up to 1.5mm long, zircon up to 1mm. Rare myrmekite.

plag - ksp - cpx - opx - ap - ore - hb - bi - zr - gt - qz

CPX-OPX ZONE

WL606: Plagioclase and perthites, most of which are exsolved to plagioclase and orthoclase domains, are c.90% of the rock. Rare myrmekite along grain boundaries. Grain size up to 6mm. Clinopyroxene, up to 2.5mm across, with euhedral opaque inclsuions, overgrown by orthopyroxene, which in turn is overgrown by primary hornblende, which may have garnet coronas. Biotite is also rimmed by garnet where it was in contact with feldspar, but it may also be primary-seen overgrown by orthopyroxene. Apatite up to 1.5mm across.

plag - ksp - opx - cpx - hb - bi - ap - ore - gt - qz

CPX-OPX ZONE

WL607: Unusually mafic-rich: mafics c.35%; grain size up to 4mm, most grains much smaller. The feldspars - plagioclase, orthoclase and minor perthites - are usually <1mm across. Primary biotite is up to 4mm long and overgrows opaques. No retrograde reactions.

plag - ksp - opx - bi - ore - cpx - ap - sph

CPX-OPX ZONE

WL609: Perthites, up to 8mm across, with exsolved plagioclase and orthoclase domains are c.80% of the rock. Rare myrmekite. Pleochroic, green clinopyroxene with abundant oriented opaque needles, intergrown with opaque-rich breakdown products of olivine. Abundant aggregates of fine-grained orthopyroxene and opaques overgrown by fibrous opx and an outer rim of garnet and fibrous garnet, taken to be pseudomorphs after olivine (Griffin and Heier, 1969; Fig. 9).

plag - ksp - cpx - bi - opx - ore - gt - ap - hb

CPX-OL ZONE

WL610: Complete breakdown of pyroxenes. Mafics tota- c.25% of the rock. Feldspars are all exsolved plagioclase-orthoclase from perthites, up to 5mm across. Hornblende-quartz calcite pseudomorphs after pyroxenes with outer, almost complete, hornblende rim. Sphene overgrows opaques - radially in one case. Blue-green amphibole up to 2mm long. Subhedral pink garnets associated with pseudomorphs. Apatite up to 1.5mm across.

ksp - plag - hb - qz - bi - gt - sph - cc - ap

RETROGRADED MANGERITE

WL614: Feldspars, plagioclase and exsolved perthites, up to 5mm across, 90% of the rock. Green clinopyroxene, some with opaque inclusions, overgrown by orthopyroxene. Cpx also overgrown by dark green hornblende up to 3mm along, which also overgrows apatite and opaques. Rare myrmekite.

plag - ksp - hb - cpx - opx - ore - ap - bi - gt - zr - qz

CPX-OPX ZONE

WL614A: Inclusion in WL614. Plagioclase and orthoclase in granular texture, grain size 0.2-1mm. Mafics c.35% of the rock. Orthopyroxene overgrown by primary hornblende, up to 2mm long.

plag - hb - bi - ksp - opx - zr

XENOLITH

WL617: Mesoperthite, up to 4mm across, many grains with exsolved plagioclase and orthoclase. Rare discrete plagioclase. Mafics c.15% of the rock. Pale green clinopyroxene overgrown by orthopyroxene and hornblende. Opaque-rich pseudomorphs after olivine, surrounded by orthopyroxene-garnet coronas, as described by Griffin and Heier (1973). Garnet overgrown by biotite. Apatite up to 1.5mm across.

plag - ksp - cpx - opx - ore - hb - ap - bi - gt

CPX-OPX-OL ZONE

WL618: Layered mangeritic sample

D: Layer 1 (lowest) is typical retrograded mangerite with relict pyroxenes. Grain size up to 3mm.

plag - ksp - bi - hb - qz - gt - cpx - ore - ap

Layer 2 has 10mm flame perthites, minor quartz and zircon and accessory biotite and garnet. 20mm wide.

ksp - plag - qz - zr - bi - gt

C: Layer 3 consists of plagioclase, perthites and quartz, up to 3mm across, c.75% of the layer. Hornblende is the main mafic with accessory biotite and opaques. Graded layer: hornblende-rich base. 28mm wide.

plag - ksp - hb - qz - bi - ore - ap

B: Layer 4 consists of perthites, quartz and minor plagioclase up to 2mm across, c.75% of the layer. Hornblende is the main mafic with pale green clinopyroxene and less biotite than

Layer 3. No opaques. 18mmwide.

ksp - plag - hb - qz - cpx - bi - gt - ap

Layer 5 has more mafics: c.35% of the layer, due to more pale green clinopyroxene. 24mm wide.

ksp - plag - hb - qz - cpx - bi - gt - ap

A: Layer 6 has mafics c.5% of the layer, which occur in spots - hornblende and biotite. 40mm wide.

ksp - plag - qz - hb - bi - ap

WL620: Perthites, exsolved to plagioclase and orthoclase, up to 5mm across c.85% of the rock. Pyroxenes pseudomorphed by hornblende-quartz-biotite and rimmed by garnet. Aggregates, 10mm across, of primary hornblende overgrowing ortho- and clinopyroxene, apatite and opaques.

ksp - plag - cpx - hb - opx - bi - gt - qz - ap

CPX-OPX ZONE

WL625: Perthites, with exsolved plagioclase and orthoclase domains, are c.85% of the rock and up to 4mm across. Primary olivine breaking down to opaques and granular orthopyroxene. Green clinopyroxene with opaque inclusions. Hornblende overgrows clinopyroxene. Accessory zircon, rutile and apatite.

ksp - plag - cpx - ore - ol - hb - bi - ap - zr - ru

CPX-OL ZONE

WL626: Perthites, with exsolved plagioclase and orthoclase domains, c.80% of the rock and up to 8mm across. Pale green clinopyroxene with opaque inclusions, overgrown by hornblende. Opaques contain apatite and are overgrown by hornblende. Olivine with double coronas against feldspars. Zircon up to 1.5mm across, apatite 0.5mm.

ksp - plag - cpx - ore - hb - ol - ap - zr - ru - bi

CPX-OL ZONE

WL628: Ribbon perthites up to 4mm across, few perthites showing exsolution to plagioclase and orthoclase, c.85% of the rock. Pale green clinopyroxene in aggregates with opaques and apatite, and pseudomorphs after olivine.

ksp - plag - cpx - ore - hb - ap - bi - zr

CPX-OL ZONE

WL649: Mafics c.50% of the rock. Similar to WL568. Plagioclase <1mm across. Biotite, hornblende, pyroxenes up to 1.5mm. Charnockite vein at one end, grains up to 5mm across.

plag - hb - bi - cpx - opx - ore - zr

XENOLITH

WL650: Cumulate peridotite layer. c.45% each olivine and clinopyroxene, about 3mm across. Abundant apatite, minor orthopyroxene, hornblende and biotite with interstitial opaques.

ol - cpx - ore - ap - opx - hb - bi

WEHRLITE LAYER

The Hovden Mangerite

WL703: Megacrysts of ribbon mesoperthites up to 10mm across, containing biotite, opaques and apatite, are c.50% of the rock. The matrix is mesoperthite, plagioclase and orthoclase up to 0.5mm across. Mafics total c.5% of the rock. Hornblende and biotite rims to pyroxenes and opaques.

ksp - plag - opx - ore - ap - cpx - hb - bi

WL707: Megacrysts of ribbon mesoperthite and finer-grained matrix as for WL703. Mafics total c.10% of the rock. Clinopyroxene overgrown by orthopyroxene. Biotite much more common than in WL703.

ksp - plag - opx - ore - bi - cpx - ap - hb

The Steine Mangerite

WL713: Fine scale perthites, up to 5mm across, plagioclase and rare myrmekites at grain boundaries at c.95% of the rock. Biotite and opaques are the main mafic phases.

plag - ksp - ore - bi - cpx - opx - ap - hb - qz - zr

WL714: As WL713 except biotite-quartz-opaque pseudomorphs after pyroxene are present. Zircon surrounded by pleochroic haloes in biotite.

plag - ksp - bi - ore - cpx - opx - ap - qz - zr

The Søberg Mangerite

WL715: Perthite megacrysts up to 15mm across. Matrix is plagioclase and perthites <2mm across with mafics c.15% of the rock.

plag - ksp - opx - cpx - ore - bi - ap - qz

The Eidsfjord Mangerite

WL732: Mesoperthite megacrysts up to 3mm across. Matrix is mesoperthite and plagioclase <1mm across. Mafics are c.20. of the rock.

plag - ksp - opx - cpx - ore - bi - ap

WL733: Fine-grained (<1mm), equigranular texture. Plagioclase and minor orthoclase over c.65% of the rock. The mafic phases are pyroxenes and opaques with minor biotite. Accessory apatite is also present.

plag - opx - ksp - cpx - ore - bi - ap

XENOLITH

The Engenykken Mangerite

WL737: Coarser and finer bands. Coarse bands have perthites up to 7mm across and c.15% mafic phases. Ortho- and clinopyroxene are the dominant mafics. Opaques, biotite and apatite are less abundant. The finer-grained bands have grain size <1.5mm and c.25% mafic phases. Minor and accessory phases are the same as in the coarser bands. It is not possible to say whether the 'layering' is igneous or due to the presence of xenoliths.

ksp - plag - opx - cpx - ore - bi - ap

Petrography of the Dolerite Dykes

WL313: Plagioclase laths, 3mm, in sub-ophitic texture with clinopyroxene and olivine, the latter overgrown by clinopyroxene-orthopyroxene-spinel-garnet coronas. Minor opaques and biotite.

plag - cpx - ol - ore - bi - gt - opx - sp

OLIVINE DOLERITE

WL613: Plagioclase phenocrysts, up to 15mm, euhedral, about 10% of the rock. Sericitization of plagioclase cores - indicates zonation: Na-rich rim not sericitized. Biotite-plagioclase groundmass <0.5mm grain size. Carbonate phase present in groundmass and plagioclase cores. Accessory apatite.

plag - bi - cc - musc - ep - ap

PLAGIOCLASE-PHYRIC LAMPROPHYRE

Petrography of the Lødingen Granite

WL629: Quartz, microcline and plagioclase comprise about 98% of the rock. Quartz grains typically 1mm, feldspars are generally more fine-grained, but one microcline is 3mm. Green biotite is the main mafic phase.

qz - ksp - plag - bi - ore - musc - gt - zr - ap

WL631: Mafic phases total 10%. Grain size typically 2mm. Green biotite in aggregates of up to 4mm. Abundant sphene associated with opaques. Strongly pleochroic blue-green to yellow hornblende.

ksp - plag - qz - bi - sph - pre - hb - ap - zr

WL632: Similar to WL631. Pleochroic haloes in green biotite caused by zircon. Abundant sphene. Accessory apatite and epidote.

ksp - plag - qz - bi - sph - ore - hb - ap - ep - zr

WL634: Similar to WL631. Epidote with allanite cores. No hornblende.

ksp - plag - qz - bi - sph - pre - ep - all - cc - musc

WL665: Grain size up to 6mm. Mafics occur in aggregates: green biotite, blue-green hornblende, sphene and opaques. Minor sericitization of feldspars.

qz - ksp - plag - bi - sph - ore - ap - musc - ep

WL666: Mafic phases about 20%, same as above but greater abundance. Sphene 1mm. long.

ksp - qz - plag - ore - hb - sph - bi - ep - zr

WL670: As WL665.

WL671: As WL666. Includes area 2cm across in which biotite, hornblende, sphene and opaques comprise >50% of the rock.

WL672: Mafic phases about 1%. Grain size up to 2mm.

qz - ksp - plag - bi - ore - hb - musc - chl - zr

WL673: Plagioclase up to 5mm across, single opaque grain and single aggregate of biotite grains.

plag - qz - ksp - cc - bi - ore - ep

WL678: Equigranular quartz, plagioclase, microcline, green biotite about 0.5mm. Minor hornblende. Accessory zircon.

bi - qz - plag - ksp - hb - zr

AMPHIBOLITE

WL687: Brown biotite plates and plagioclase laths define foliation. Grain size 0.5mm. Opx- and cpx-rich layers.

plag - bi - cpx - opx - hb - zr - ru

BIOTITE GABBRO

WL690: Quartz and microcline with minor plagioclase, up to 5mm. Mafics c.2%: opaques. Sericitized feldspars.

qz - ksp - plag - chl - ep - musc - ore

APPENDIX 3. Constants Used in Isotope Diagrams

Rb-Sr

$\lambda(^{87}Rb) = 1.42 \times 10^{-11}. yr.^{-1}$ — Steiger and Jäger (1977)

$BABI = (^{87}Sr/^{86}Sr)_o = 0.69898$ — Papanastassiou and Wasserburg (1969)

U-Pb

$\lambda(^{238}U) = 0.155125 \times 10^{-9}. yr.^{-1}$

$\lambda(^{235}U) = 0.98485 \times 10^{-9}. yr.^{-1}$

Steiger and Jäger (1977)

$a_o = (^{206}Pb/^{204}Pb)_o = 9.307$

$b_o = (^{207}Pb/^{204}Pb)_o = 10.294$

T = Age of Earth = 4.57×10^{9}yr

Tatsumoto et al. (1973)

Sm-Nd

$\lambda(^{147}Sm) = 6.54 \times 10^{-12}. yr.^{-1}$ — DePaolo and Wasserburg (1976)

Chondritic uniform reservoir (CHUR) values at the present day:

$^{143}Nd/^{144}Nd = 0.512637$ — Goldstein et al. (1984)

$^{147}Sm/^{144}Nd = 0.1976$ — N.W. Jones pers. comm.

$(^{143}Nd/^{144}Nd)_{CHUR}$ values in the past are calculated from these figures. A rock with the same $^{143}Nd/^{144}Nd$ value as that for CHUR at any given time has $\varepsilon_{Nd} = 0$ by definition. The depleted mantle model used in the present study is that of DePaolo (1981) in which ε_{Nd} in the depleted mantle is 0 at 4.57Ga and +8.5 at 0Ga. ε_{Nd} in the depleted mantle between these times is calculated from an algorithm representing even removal through time of light rare earth enriched continental crust-forming material from the upper mantle.

APPENDIX 4. Analytical Techniques

(i) X-Ray Fluorescence Spectroscopy

A Philips PW1400 automatic spectrometer at the University of Nottingham was used by P.K. Harvey and B.P. Atkin to determine the whole-rock concentrations of SiO_2, TiO_2, Al_2O_3, Fe_2O_3, MgO, CaO, Na_2O, K_2O, MnO and P_2O_5. The fusion beads used for these determinations were prepared at the University of Oxford according to the method of Norrish and Hutton (1969) as modified by Harvey et al. (1973).

A Philips PW1410 semi-automatic spectrometer at the University of Oxford was used to determine the whole-rock concentrations of Rb and Sr. Analyses were performed on 4cm diameter pressed powder pellets, made by compressing 9g of rock powder with 0.5 - 0.8 ml of Moviol PVA under a pressure of 10 tons/in^2. Analyses were made using a modification of the published method (Pankhurst and O'Nions, 1973) using a Mo tube.

(ii) Sr Isotope Analysis

Sr for isotopic analysis was separated by ion exchange using the method described by Pankhurst and O'Nions (1973). Sr isotope ratios were determined on a Vg-Micromass 30 mass spectrometer and a VG-Isomass 54E mass spectrometer using a modification of published procedures (O'Nions and Pankhurst, 1973). Measurements of the Eimer and Amend Sr standard solution, on the Micromass 30, gave average $^{87}Sr/^{86}Sr$ ratios of 0.70780 ± 4 during an 18-month period. $^{87}Sr/^{86}Sr$ ratios measured on the Micromass 30 were normalised to a value of 0.70800 for the Eimer and Amend standard by addition of 0.00020; ratios were also normalised to $^{88}Sr/^{86}Sr$ of 8.37521. Errors on $^{87}Sr/^{86}Sr$ ratios are better than 0.01%.

(iii) Pb Isotope Analysis

Pb for isotopic analysis was separated by the electro-deposition method of Arden and Gale (1974) and loaded onto single Re filaments with silica gel and phosphoric acid. Pb isotope ratios were determined on a VG-Isomass 54E mass spectrometer. The mass fractionation correction on Pb isotope analyses, determined by repeated analysis of the NBS981 standard solution, was -0.1% during the 12-month period of analysis. Errors on Pb isotope ratios are 0.1%.

(iv) Nd Isotope Analysis

a) Sample Dissolution

Nd for isotopic analysis was separated using a method developed by J.W. Arden and this author from that of Richard et al. (1976).

Powdered sample (ca. 200mg) and a spike solution (ca. 250mg) are weighed precisely into a teflon bomb. Add ca.5ml HF and 100 µl 16MHNO_3 and heat in an oven overnight at about 150°C. Allow the bomb to cool and evaporate the sample in a polytetrafluoroethylene (PTFE) evaporator under infra-red (K.R.) lamps with a supply of filtered air.

Add about 5ml 6MHCl and heat in an oven for a few hours/overnight at about 150°C. Evaporate to dryness.

Add 2-2.5ml 3.0M (precisely determined by titration) HCl, and leave the closed bomb standing overnight.

Transfer the sample solution to a quartz centrifuge tube and centrifuge it at about 2000 rpm for 10 minutes. Load 2ml of the centrifuged solution onto prepared columns.

b) Separation of LREE as a Group

Cation exchange columns : quartz, 14cm long, 7mm internal diameter, ca. 30ml reservoir

Resin : AG 50W - X12 200-400 mesh
7mm x 8.5cm (in H_2O)
7mm x 7.8cm (in 3.0MHCl)

(Resin in columns changed after about ten separations)

Column cleaning: 5ml H_2O
5ml 3M HNO_3
20ml 6M HNO_3
5ml 3M HNO_3
20ml H_2O
5ml 3M HCl
20ml 6M HCl
5ml 3M HCl
20ml H_2O

(Resin should not be left HNO_3 stage overnight)

After cleaning equilibrate columns with 5ml 3.0MHCl.

Load sample solution: 2ml
Wash : 2ml, 1ml, 1ml
Pass : 14ml
Collect : 1ml

Evaporate the collected REE fraction to dryness

Add 100µl of 0.15M (precisely determined by titration) HCl and allow the closed beaker to stand for a few hours or overnight.

c) Separation of Sm and Nd

Column (PTFE-HDEHP) cleaning: 20ml 6M HCl
equilibration: 2x5ml 0.15 MHCl
flow-rate: ca. 0.2ml. min.$^{-1}$

Load sample solution in	0.1ml			
Wash	3x0.1ml			
Pass	18ml			0.15M HCl
Collect	6ml	→	Nd	
Pass	2ml			
Pass	7ml			0.30M HCl
Collect	10ml	→	Sm	

Evaporate Sm and Nd fractions to incipient dryness (not baked).

d) Mass Spectrometry

Samples loaded onto Ta side filaments with a Re centre filament were analysed on a VG-Isomass 54E mass spectrometer. Measurement of the BCR-1 standard rock powder gave a $^{143}Nd/^{144}Nd$ ratio of 0.512632 ± 15, well within error of the CHUR value used of 0.512637.