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Along the northern margin of the Central Lapland greenstone belt, Archean basement gneisses are exposed in a number of structural culminations. One such basement window, at Möykkelmä, is a domal structure flanked mostly by Lapponian quartzite, but also preserves a sequence of subaerial mafic and ultramafic volcanites about 250 m thick, that were extruded directly on the gneisses. Within the volcanic sequence, five stratigraphical units have been identified – the lower ultramafic, lower mafic, upper ultramafic, middle mafic and upper mafic units respectively. At the base of the lower ultramafic unit is a volcanic breccia, containing gneiss fragments in an ultramafic matrix, consistent with volcanism having taken place in a sialic crustal setting. Both ultramafic sequences consist for the most part of agglomerates and lapilli tuffs, with only subordinate lavas. The lowest mafic unit contains some amygdaloidal, in addition to the predominant massive lavas and tuffaceous deposits, whereas the middle mafic unit consists principally of amygdaloidal lava flows, and the upper unit contains both pyroclastic lithologies and massive lavas.

On the basis of their MgO (13 – 21 %) contents and TiO₂ (<0.60 %) values the ultramafic units can be classified principally as basaltic komatiites, with only sporadic true komatiite compositions. Nevertheless, they differ from true komatiites in a number of respects, in part due to intense hydrothermal alteration, resulting in a wide variation in SiO₂ contents (43 – 62 %) and typically low CaO contents (1 – 11 %), while enrichment in K, Rb, and Cs is common. Less mobile elements also exhibit anomalous characteristics. LREE enrichment is typical (La_N/Yb_N = 4.4 – 5.7) and a clearly negative Ta anomaly is present. In addition, Ti/Zr is exceptionally low (< 60) and Al₂O₃/TiO₂ equals or exceeds the chondritic value. All of these anomalous immobile element characteristics suggest a considerable amount of Archean upper crustal contamination.

Contamination has also affected the composition of mafic volcanics which, along with the mobilization of alkali- and alkali-earth elements, in turn influences their geochemical classification. Thus, according to their present composition, the lowermost unit belongs to the calc-alkali series, while the middle unit falls between transitional calc-alkaline and tholeiitic trends and the uppermost unit exhibits tholeiitic characteristics. Classification on the basis of SiO₂ contents shows a so-called antidromic trend – that is, a stratigraphical progression from andesites through basaltic andesites to basaltic compositions. This is consistent with the concomitantly observed increase in compatible trace element contents and conversely, decrease in amounts of many incompatible elements (such as LREE, Zr). In contrast, the contents of the moderately incompatible elements (Fe, Ti, V) are higher for basalt than for the basaltic andesites, thus eliminating a simple genetic relationship between these two magma types, although it is plausible that the basaltic andesites and andesites are differentiates of a single parent magma. The unusually high Cr content (av. 890 ppm) in basalt remains difficult to explain. The generally low TiO₂ content (< 0.80 %) and high LIL/HFS element ratios are characteristic of island arc mafic rocks, in spite of the Möykkelmä volcanites having been erupted in a continental setting. Their composition is nevertheless explicable in terms of derivation of parental magma from a depleted upper mantle, and later contamination and fractional crystallization in the upper crust, as well as subsequent hydrothermal alteration.

Key words: komatiite, tholeiite, gneisses, pyroclastics, geochemistry, contamination, Archean, Proterozoic, Möykkelmä, Sodankylä, Finland.

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INTRODUCTION

The Lapland Volcanite Project was initiated in 1984 and was intended, during the ensuing five year period to involve amongst its principal objectives, "effective sampling and other related studies of the widespread, but generally poorly exposed volcanic successions in northern Finland, in order to develop genetic models for their mode and environment of eruption, as well as be of general benefit to regional mineral exploration".

The volcanics of the Möykkelmä area were selected as a specific target for investigation in the 1984 project outline, as a result of their differing from other volcanic units in Central Lapland. According to Tyrväinen (1983, p. 45) "At Möykkelmä, volcanics were evidently erupted directly onto basement gneisses and apparently represent the initial stages of

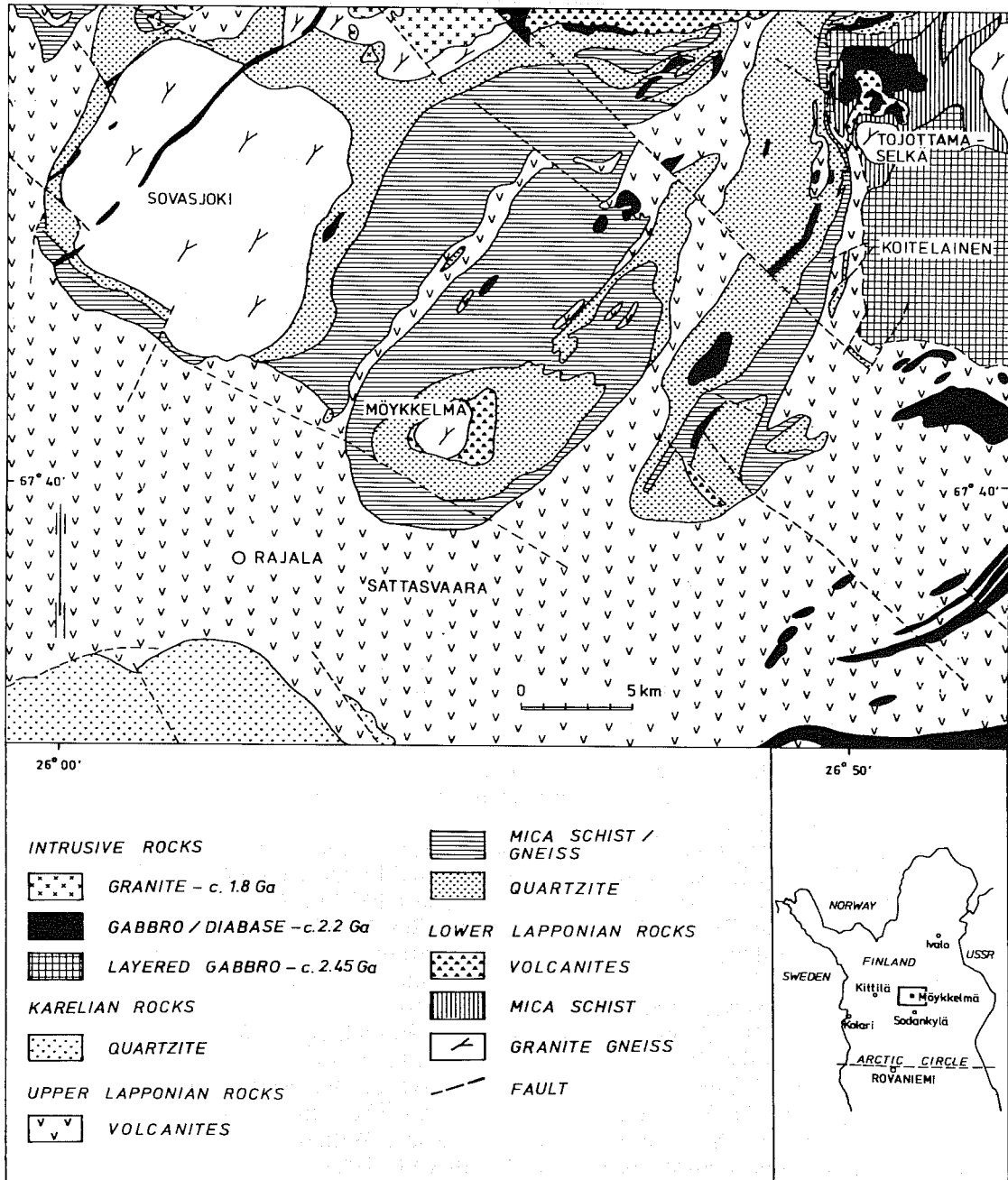


Fig. 1. Major geological units surrounding the study area.

development of the greenstone belt, or at least volcanic activity predating the main phase of volcanism”.

Studies at Möykkelmä were intended to elucidate the nature, origin, internal stratigraphy and petrographical and geochemical characteristics of the volcanics, as well as to ascertain their relationship to other volcanic sequences in Central Lapland. Likewise the nature of the contact with the associated underlying basement gneisses was to be examined.

The Möykkelmä study area is located in the municipality of Sodankylä in Central Lapland, about 10 km NE of the village of Rajala (Fig. 1), comprising a somewhat elevated terrain about 10 km² in extent. Previous geological investigations of the area are confined to the 1 : 100 000 scale Sattanen (Sheet 3714) map and explanatory notes (Tyrväinen 1980, 1983), according to which gneissose granodiorite (considered by Tyrväinen to be Archean) and mafic volcanics form a roughly 8 km² window within the surrounding quartzite.

Prior to the commencement of field work for the Lapland Volcanite Project a regional map of the lithological association occurring in Central Lapland was compiled, based on all available data. The Möykkelmä gneissose granodiorites were thus assigned to the late Archean basement complex while the volcanic sequence was correlated with the early Proterozoic Lower Lapponian Series (Lehtonen *et al.* 1984).

The neighbouring Madetkoski volcanites, in the vicinity of the Porttipahta area were likewise considered to be Lower Lapponian and, in common with the Möykkelmä greenstones, are both non-magnetic and devoid of intensely electrically conductive horizons. An extensive quartzite succession, often greenish due to its distinctive fuchsite content, overlies the Lower Lapponian volcanites and has been assigned to the Upper Lapponian.

Preliminary results of the Möykkelmä studies were presented at the 17th Nordic Countries Geological Meeting in Helsinki May 1986 (Räsänen *et al.* 1986).

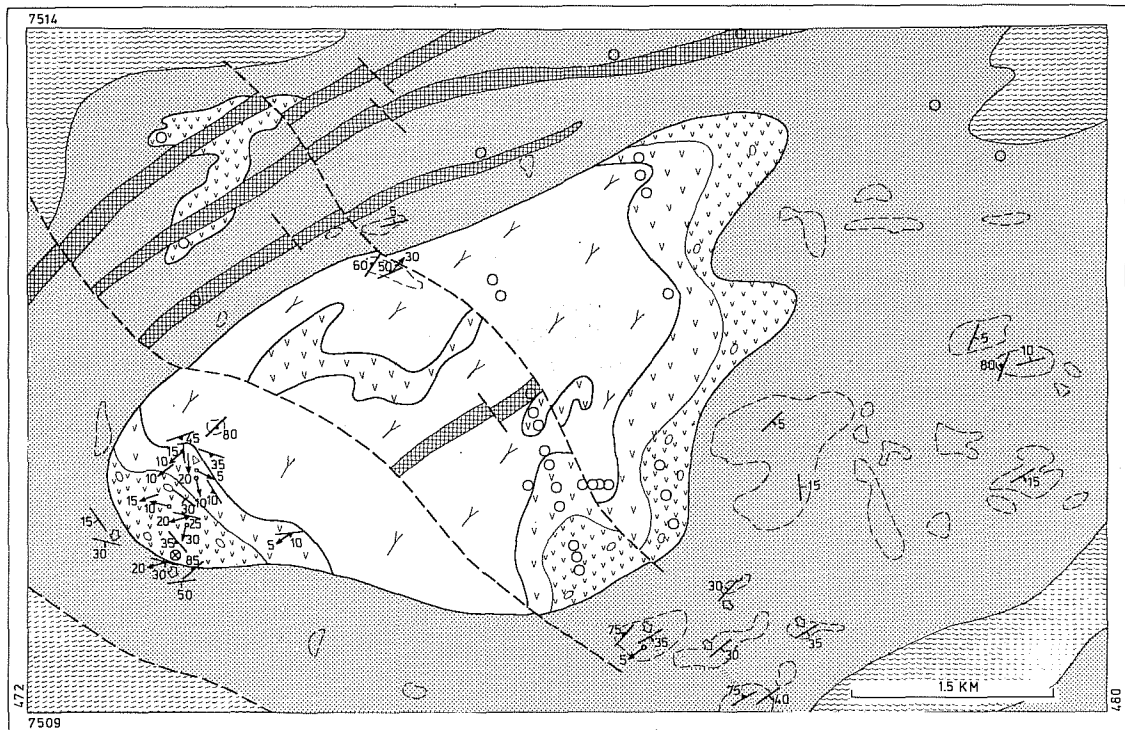
Jorma Räsänen was responsible for field investigations, structural and stratigraphical interpretations, and lithological descriptions, while geochemical studies of the volcanites were undertaken by Eero Hanski, and the geochemistry of the basement gneisses, as well as regional geological relationships were examined by Matti Lehtonen.

NATURE OF INVESTIGATIONS

Investigations were concentrated towards the southwestern part of the Möykkelmä dome, where it was attempted to examine in detail the volcanites exposed between underlying basement gneiss and the overlying supracrustal rocks (cf. Tyrväinen 1980, 1983). The area was mapped at a scale of 1 : 5 000 and a special effort was made to locate even small exposures in areas where ground plants and soil cover were sufficiently thin. Four exploration trenches, with a combined length of 400 m were also excavated to expose critical lithological boundaries and were mapped in detail at a scale of 1 : 500.

The volcanic sequence was also drilled to a depth of about 250 m in order to obtain representative samples from the whole formation. Magnetic and electrical resistivity properties were measured in the exploration trenches and their surroundings and U radiometric values were recorded from the drill hole.

Chemical analyses were performed on a total of 126 samples. All major and some trace elements (Cr, Ni, V, Sr, Zr, S, Cu, Ba, Zn) were analyzed by the XRF method using a Philips 1400 spectrometer at the Raahe laboratories of Rautaruukki Ltd. (Analysis package A5). In addition, Fe, Na, As, Sb, Zn, Sn, Ag, W, Mo, Au, La, Sm, Lu, Cr, Ni, Co, Sc, Ba, Cs, Rb, Ta, U, Th, Br were analyzed from each sample using the instrumental neutron activation facilities of the Technical Research Centre at Otaniemi (Analysis package E5). Lanthanide analyses (La, Ce, Nd, Sm, Eu, Tb, Yb, Lu) were made on fifteen of the least altered drill core samples at the same laboratory, using the INAA methods described by Rosenberg *et al.* (1982). In general, acceptable results using the XRF method were obtained for V, Zr, Cu, S and Rr, in addition to the major elements, while precision of the INAA (E analytical package) appears sufficient for the elements Cr, Ni, Co, Sc, La, Sm, Ba, Rb, U, Th, Ta and Cs. The oxidation state of Fe and volatile contents are missing from the XRF major element analyses. With the exception of the granitoid analyses all the volcanite analyses have been normalized to 100% volatile free. In general, Fe is taken as Fe²⁺ (ferrous) except for the purpose of norm calculations, where part of the total Fe is counted as ferric, according to the ratio Fe³⁺/Fe²⁺ = 0.15.



LEGEND

	Granitic gneiss		Amygdaloidal lava		Drill hole
	Ultramafic volcanic rocks		15 Bedding		Percussion drill hole
	Mafic volcanic rocks		25 Foliation		Contact
	Quartzite		15 Fold axis		Unconformity
	Mica schist		10 Lineation		Fault
	Gabbro / diabase		Top of the beds		
	Volcanic breccia		Outcrop		

Fig. 2. Geological map of the Möykkelmä area.

GEOLOGICAL SETTING

In the surroundings of the study area supracrustal rocks of the Lapponian Supergroup are predominant consisting principally quartzites, mica schists and gneisses and volcanic rocks (Figs. 1 and 2, see also Lehtonen *et al.* 1984). Quartzites were deposited at least in part directly on gneissose basement (Tyrväinen 1983; Räsänen 1986). The Lapponian volcanites are considered to have erupted principally after deposition of the quartzites and mica schists (e.g. Lehtonen *et al.* 1985); and generally comprise komatitic and tholeiitic lavas and pyroclastic deposits, with associated tuffites and chemical sediments. The youngest supracrustal formations in Central Lapland occur as discrete outliers of commonly coarse-clastic arkosic quartzite and conglomerates, formerly known as "Kumpu quartzite" (Mikkola 1941), but currently assigned to the Karelia Supergroup (Silvennoinen 1985). The contact between the Lapponian and Karelian Supergroups is an erosional surface (Lehtonen *et al.* 1985; Räsänen & Mäkelä 1988).

Intrusive rocks in the vicinity of the study area are represented by Presvecokarelian gabbros and dolerites as well as Svecokarelian granitoids. The Koitelainen layered gabbro complex is situated some 20 km ENE of the study area (Fig. 1) and has yielded a U-Pb zircon age of 2.45 Ga (Puustinen 1977). Throughout the Central Lapland, numerous differentiated dolerite dykes intruding the greenstone terrain give U-Pb zircon and sphene ages of 2.0 – 2.2 Ga, constraining the minimum age for eruption of the Lapponian Supergroup volcanites (e.g. Rastas 1980).

Structurally, Möykkelmä belongs to the extensive Pomokaira basement complex area, which comprises antiformal culminations of gneissose granitoids with intervening depressions occupied by Lapponian quartzite and mica schist formations (Lehtonen *et al.* 1984). At the southern margin of the basement gneiss complex, the Möykkelmä study area forms a distinct domal structure, and there exist several analogous domal structures (e.g. Sovasjoki and Tojottamaselkä, Fig. 1). A U-Pb age of 3.1 Ga for zircons from Tojottamaselkä (Kröner *et al.* 1981) confirms the inferred Archean age for the Pomokaira gneisses, while paleoregoliths observed at the basement-cover boundary further indicate a substantial age difference between the gneisses and the overlying supracrustal formations (cf. Tyrväinen 1983).

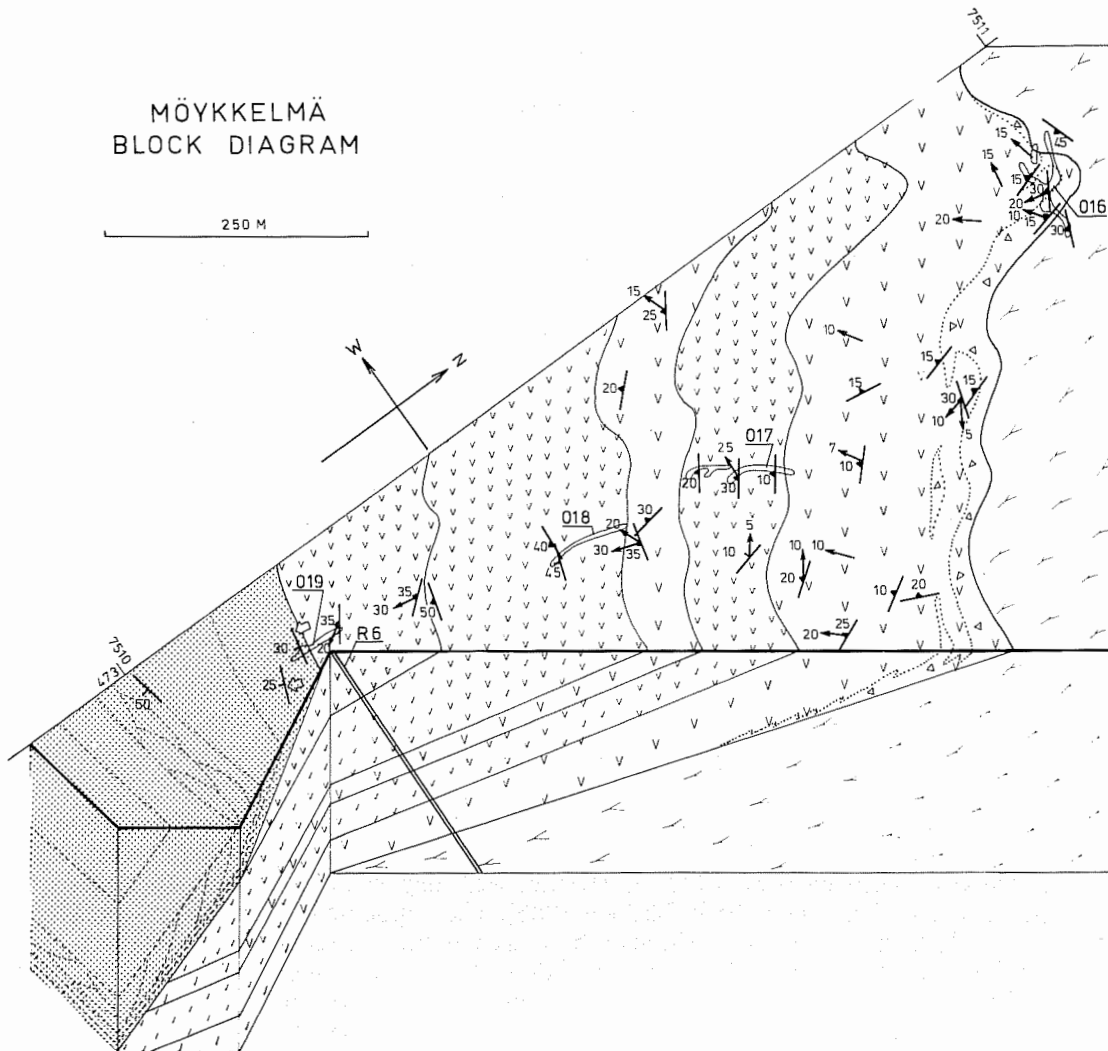


Fig. 3. Block diagram showing the volcanic units and their relation to the overlying quartzite formation and the underlying granite gneiss.

STRUCTURE AND STRATIGRAPHY

A detailed structural analysis of the study area was not feasible, due to the poor exposure and in addition small size and generally frost-shattered nature of most outcrops. Gneissose banding is commonly seen in the basement rocks but the majority of lithologies (with the exception of some tuffites and quartzites) tend to be massive. Nevertheless, on the basis of structural observations, the Möykkelmä gneisses and volcanites are located at a culmination formed by antiformal fold axes of differing trend, and are surrounded by gently-dipping quartzites. That is, the quartzites form a stratigraphic window through which older units can be seen.

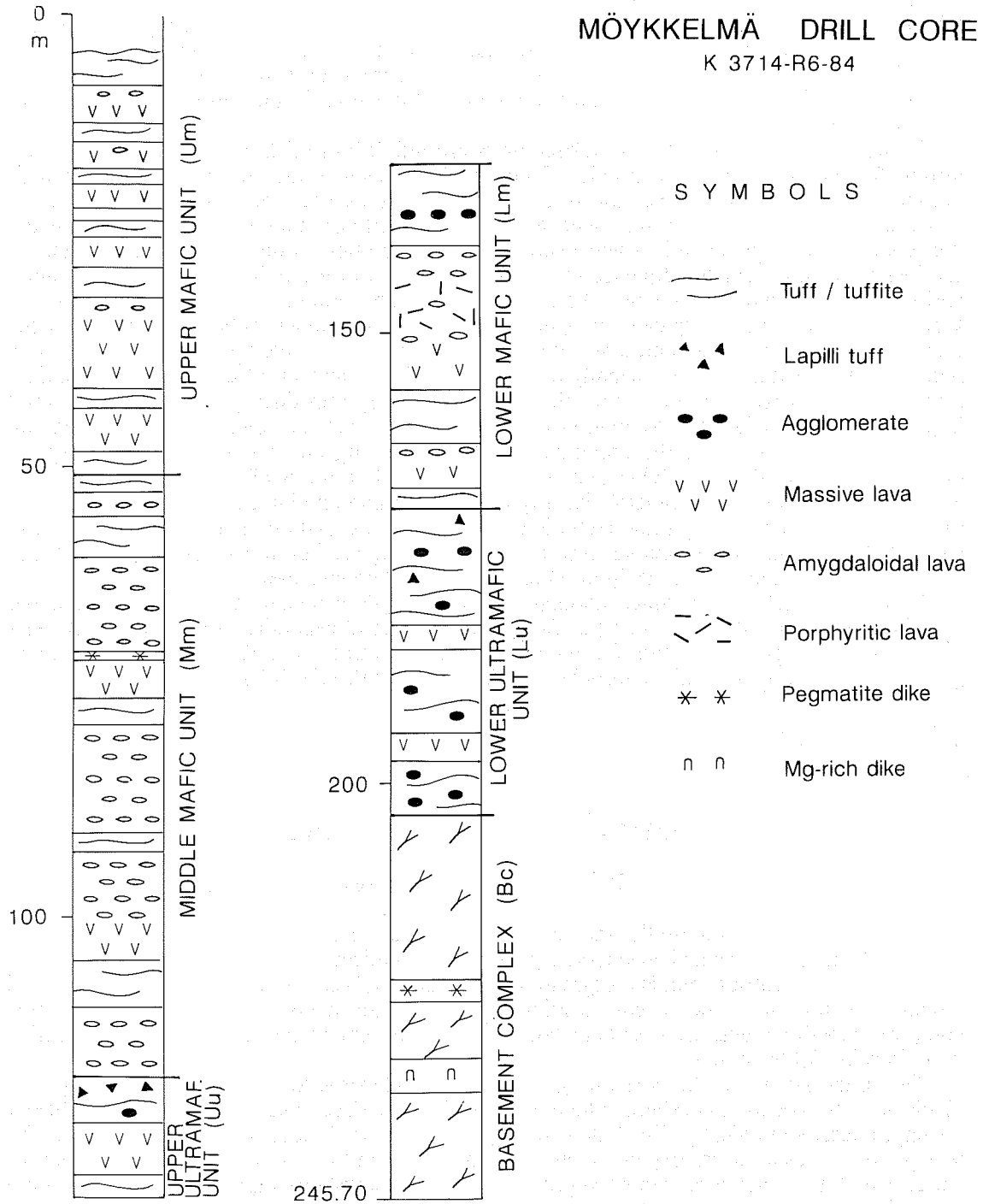


Fig. 4. Diamond drill section through the Möykkelmä volcanic formation. Location of the drill hole is shown in Figs. 2 and 3.

Although competence contrasts between the quartzites and volcanites influenced their behaviour during deformation, there also appear to be differences in their deformational history. The earliest folding phase to affect the quartzites produce low-angle structures with gently plunging fold axes consistently of NE–SW trend. In the underlying volcanites, however, these structures appear to have deformed an earlier, roughly NNW–SSE trending, intensely banded fabric. In addition, the volcanites are deformed into generally E–W trending chevron fold which not only deforms the above mentioned banding but also appears to predate the folding observed in the quartzites. Nevertheless, there is insufficient evidence from the study area to ascertain whether or not the quartzite had already experienced tectonic deformation at this stage.

Granite-gneiss fragments in the volcanic breccia at the base of the Möykkelmä sequence indicate that the magmas ascended through and were erupted onto the gneissic basement. Observations from exploration trenches confirm an unconformable relationship between the granitic gneisses and volcanites, where the contact is seen to be discordant to the gneissic banding in the former (Fig. 5). Furthermore, the occurrence of a compositionally distinctive biotite schist immediately below the contact indicates weathering of the granitic gneisses prior to eruption of the volcanites.

In the study area, the volcanite formation is around 250 m thick and is divisible into two ultramafic and three mafic units (Figs. 3 and 4). The lower ultramafic unit consists of both pyroclastic and homogeneous, massive, but strongly schistose volcanites, with a breccia containing gneissose basement fragments forming the basalmost member. The lowermost mafic unit contains both amygdaloidal and massive lavas in addition to tuffaceous horizons. This is succeeded by an ultramafic sequence that differs from the lower one in being generally thinner and lacking in gneissic fragments. The middle mafic unit comprises successive lava flows of variable thickness, and is typically amygdaloidal, whereas the upper mafic unit consists for the most part of massive lavas and associated tuffaceous horizons. Overlying the volcanites is a sequence of metasediments, principally quartzites which, on account of their wide distribution, form a useful marker horizon within the Lapponia Supergroup throughout much of Central Lapland. The quartzites are commonly cross-bedded and may be found deposited either on volcanites or directly upon the granitic-gneiss basement, suggesting that regionally, they are disconformable on the underlying units (Figs. 1 and 2). Exposures of the basal contact in an exploration trench revealed general skarn-type alteration and an abundance of vein quartz and quartz fragments suggesting some tectonic dislocation or brecciation at the contact, so that in spite of the differing deformational history between the quartzites and volcanites, no obvious angular unconformable relationship in outcrop could be demonstrated.

Pegmatitic dykes intrude other lithologies at Möykkelmä. Albitic dolerites were encountered only in percussion drill samples and as float, and correspond generally to those of the Karelia Supergroup throughout Lapland. On the other hand, an ultramafic dyke seen to cut basement gneisses in drill core material, is probably genetically related to the Möykkelmä volcanites.

LITHOLOGICAL DESCRIPTIONS

Basement granitic gneisses

These form both structurally and lithologically a rather heterogeneous unit consisting of massive, banded and veined leucocratic gneisses, migmatized to varying degrees, as well as discordant, cross-cutting pinkish pegmatitic veins and dykes. Diffuse to well differentiated banding of the order of several centimeters occurs throughout the gneisses, but some compositional differences are found at a larger scale up to some meters thick, generally reflecting variations in biotite content.

The mineralogy of the gneissic granites is straightforward, being principally quartz, plagioclase (oligoclase-andesine), which is strongly sericitized, and biotite, with K-feldspar porphyroblasts sporadically distributed throughout the more leucocratic bands. However, K-feldspar occurs as an accessory phase, along with apatite and zircon. Grain size of the gneisses varies from fine to medium, with however some coarsely recrystallized portions, and textures are granoblastic, or less commonly lepidoblastic.

Pegmatitic granite is in contrast only weakly oriented and contains biotite and muscovite in addition to the major phases quartz, plagioclase and K-feldspar.

The granitic gneisses do not contain any features that might indicate their possible protolith, the banded nature reflecting metamorphic segregation during penetrative deformation. Nevertheless, compositional differences at a larger scale could conceivably represent the kind of variation observed between sandy graywacke units.

Volcanite formation

Lower ultramafic unit

Drill core data indicate a present thickness of about 40 m for this unit, which consists principally of tuffaceous pyroclastic rocks containing variable amounts of ultramafic clastic fragments and two more massive intervals. The biotite-rich schists marking the (weathered ?) contact with the basement gneisses are overlain by a 3 – 4 m thick pyroclastic breccia containing ultramafic ejecta as well as gneiss fragments 5 – 25 cm in diameter and having been assimilated to varying degrees (Fig. 5). The smallest of these are rounded whereas large ones tend to be angular (Fig. 6) and the unit is interpreted as an explosion breccia formed during the initial stages of volcanism. The matrix is ultramafic, consisting of chlorite, tremolite-actinolite and, in the

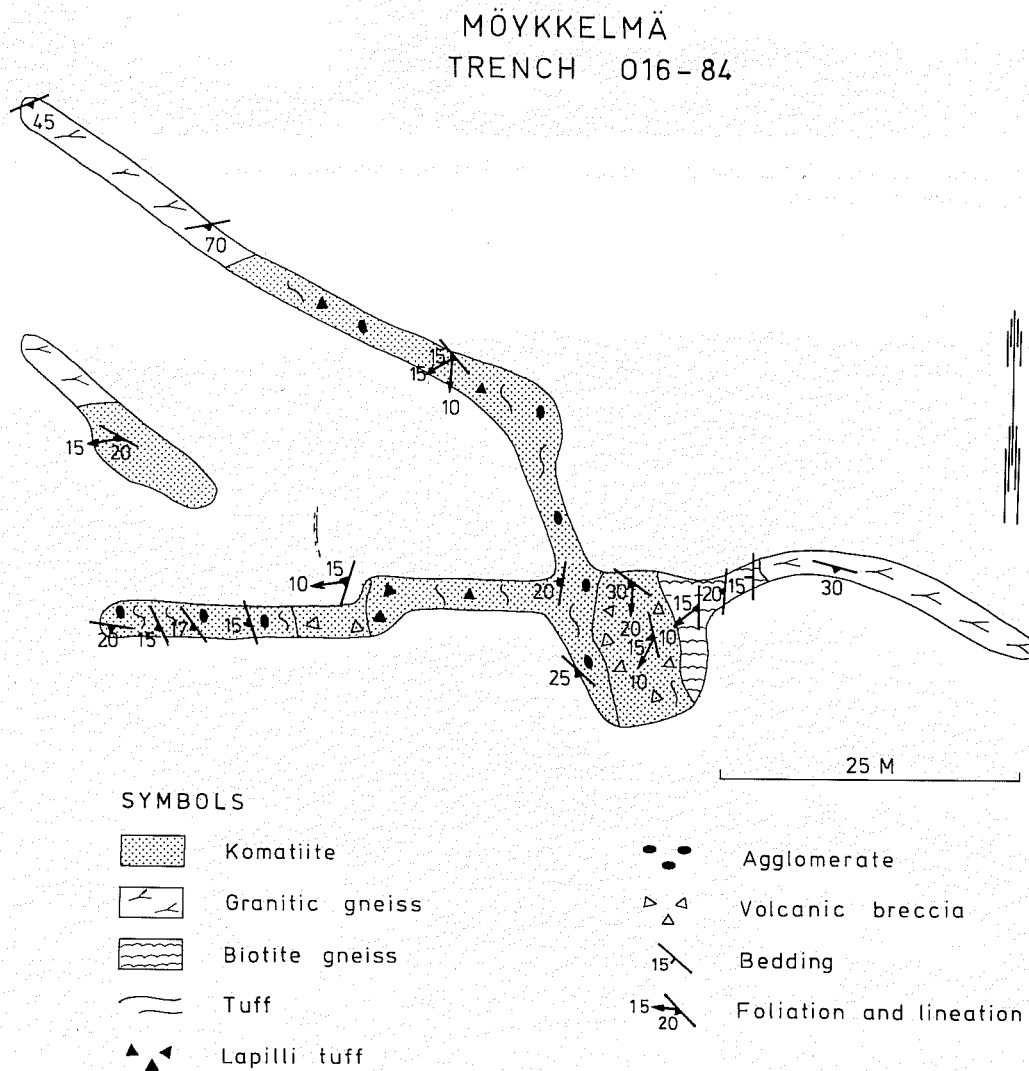


Fig. 5. Lithology of and tectonic observations from exploration trench 016, Möykkelmä. Location of the trench is shown in Fig. 3.

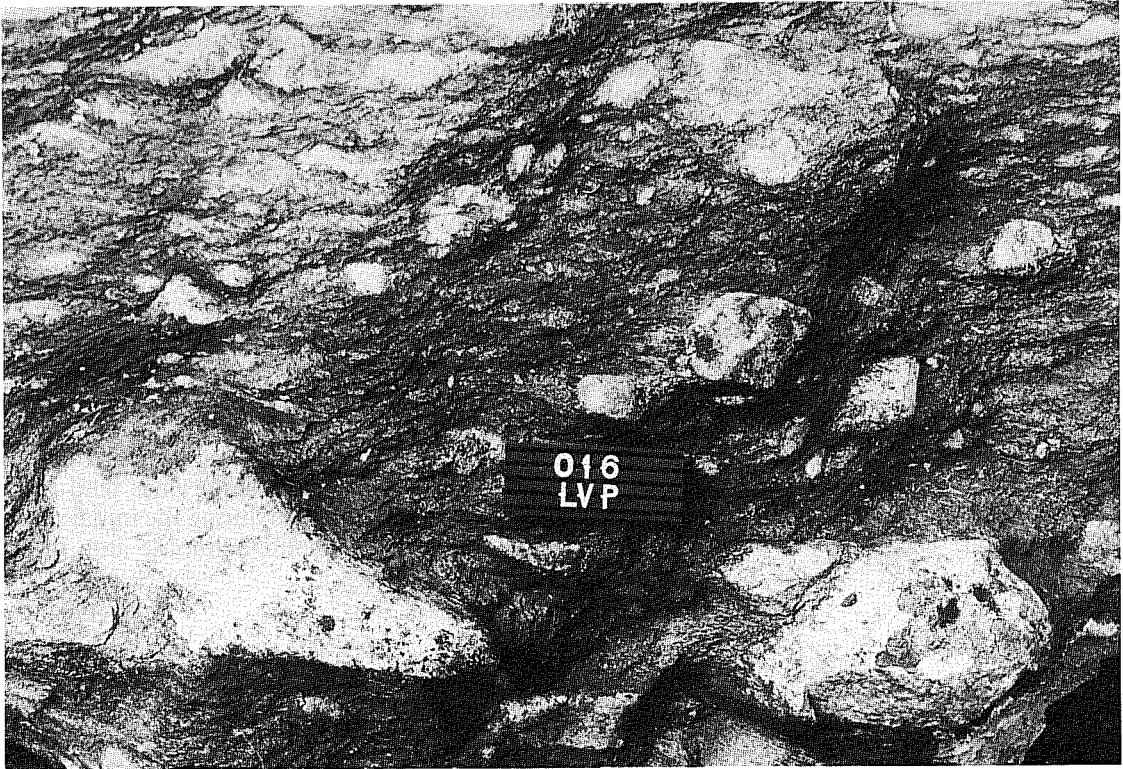


Fig. 6. Angular and subrounded granite gneiss fragments in a komatiitic tuff matrix at the base of the Möykkelmä volcanic formation. Tag = 10 cm.



Fig. 7. Coarse agglomeratic lapilli tuff from the lower ultramafic unit. Note zoned bomb in the centre of the picture. Match 5 cm.



Fig. 8. Amygdaloidal lava from the lower mafic unit. Match = 5 cm.



Fig. 9. A partially melted granitic gneiss fragment in amygdaloidal lava of the lower mafic unit. Note the secondary rim around the fragment caused by partial melting. Match = 5 cm.

lowermost parts, also of biotite and quartz. The matrix is strongly foliated, and coarser fragments are elongate. The proportion of fragments decreases rapidly upwards as the rock becomes a very fine grained biotite-bearing chlorite-amphibole schist.

At about 8 – 10 m above the basal contact, the volcanic breccia gradually changes to a tuff containing lapilli fragments up to 5 – 25 mm in size (Fig. 5). The tuff is characterized by porphyroblasts, 1 – 3 mm in size, which generally contain a chlorite core surrounded by a plagioclase rim and are evident as light-colored blebs on the weathered surface. Gradually the rock becomes somewhat agglomeritic, with 2 – 10 cm sized oval-shaped fragments consisting of tremolite-actinolite and chlorite (Fig. 7). These are typically lighter in color than the enclosing matrix and are surrounded by a 0.2 – 0.5 cm thick, darker and fine grained chlorite-amphibole rim. The originally tuffaceous matrix consists of variable proportions of chlorite and amphibole, with some biotite and as accessories talc, calcite and in places pyrite and porphyroblastic olivine. The occurrence of olivine and talc in the same paragenesis is taken to indicate a moderate degree of metamorphism (cf. Winkler 1979, p. 161).

Diamond drilling also intersected, amongst the pyroclastic layers, strongly foliated but homogeneous amphibole-chlorite rocks as two intercalations. Both of them are less than 3 m thick and they are interpreted as lava flows, although no primary flow structures are preserved. However, their lower contacts are sharp and well defined but upper contacts are gradual over a short (10 – 30 cm) interval. They consist of colorless amphibole, which occurs as a finely distributed groundmass as well as acicular prisms, and pale green chlorite. In addition to accessory talc and biotite, quartz and strongly zoned plagioclase are present.

Lower mafic unit

According to estimates from drill core material, this unit is around 35 m thick and consists of alternating lavas and tuffaceous horizons. The former are typically gray, with a faintly greenish tinge, and generally massive and very fine grained, although both porphyritic and amygdaloidal variants occur. At the base of the sequence is a dark gray tuffaceous unit abruptly overlying the lower ultramafic sequence and, like the remainder of the mafic tuffs, consists of hornblende, plagioclase, biotite and quartz. A sharp, diopside-rich contact occurs with the succeeding massive, less deformed, pale gray unit which, on the basis of sporadic green chloritid amygdules, is interpreted as lava flow. The groundmass is very fine-grained and is composed of biotite, plagioclase, amphibole and quartz.

The next lava flow begins in a similar massive fashion, but becomes progressively porphyritic, and then amygdaloidal in the very uppermost part (Fig. 8). Phenocrysts are of euhedral and zoned plagioclase (1 mm in size), while amygdules are typically 1 – 5 mm across and consist mainly of plagioclase and amphibole, as well as quartz and biotite. Partially melted granitic gneiss inclusions up to 10 cm in size are found in a trench (Fig. 9).

The highest part of the lower mafic unit consists of tuffs with slightly agglomeritic horizons. The tuffs are very fine grained and commonly dark greenish gray in color, but some deep gray layers rich in biotite are found. The very uppermost part, up to two meters, of this sequence is pale gray and it contains a high proportion of quartz.

Upper ultramafic unit

On the basis of drill-hole intersections, this unit appears to be 13 m thick and is both mineralogically and texturally very similar to the lower ultramafic unit. It comprises a massive, featureless unit about 6 m thick, interpreted as a lava flow, (in spite of the absence of such typical ultramafic flow features as spinifex texture or flow top breccias), both underlain and overlain by ultramafic pyroclastics in which depositional features are still discernible (Fig. 10). These pyroclastics differ, however, from their counterparts in the lower ultramafic unit, in that while ejecta are in general more abundant, granite gneiss fragments have not been observed.

Middle mafic unit

This unit, totaling 65 m in the drill core, comprises five amygdaloidal lava flows with relatively thin tuffaceous intercalations. All of the lava flows tend to be fine grained and homogeneous, and consist of hornblende, oligoclase, biotite and quartz. Chlorite is the most

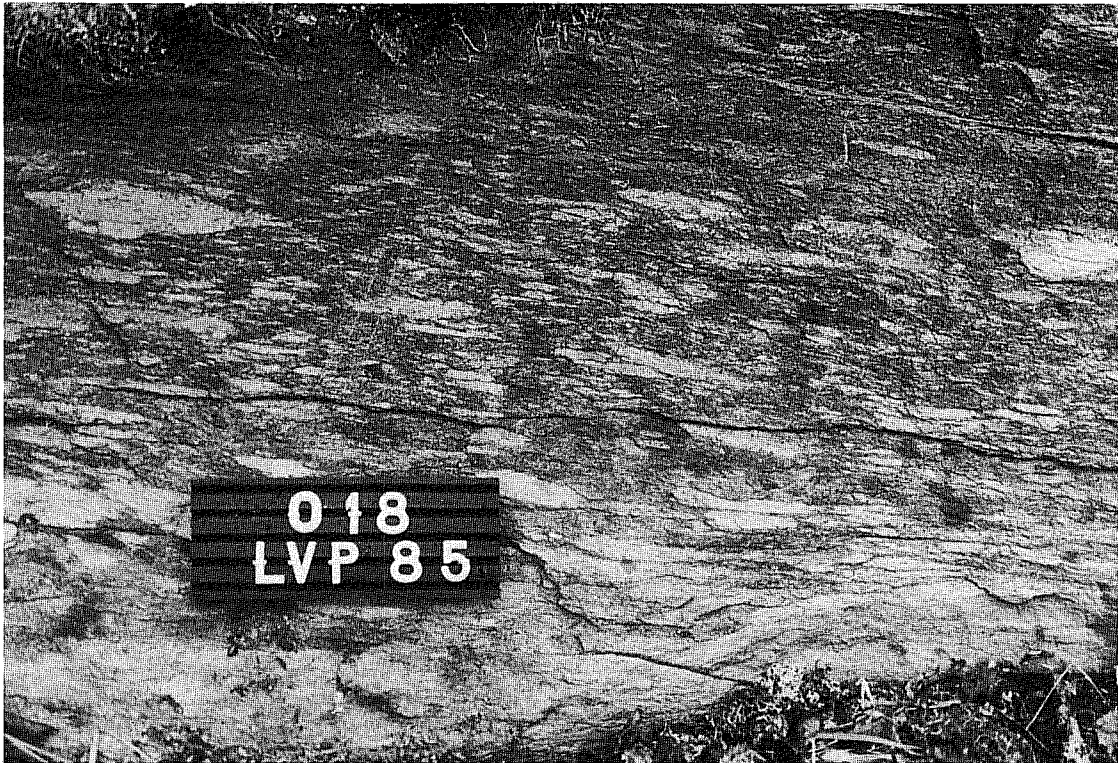


Fig. 10. A deformed pyroclastic deposit in the upper ultramafic unit. Tag = 10 cm.



Fig. 11. Amygdule-rich bands in a metalava of the middle mafic unit probably indicating primary flow structures. Tag = 10 cm.

widespread accessory phase, and carbonate minerals are common in places. Amygdules are general feature of these lavas but vary considerably in size, abundance and distribution, and usually have plagioclase-amphibole fillings but may contain also quartz. They range from 0.2 – 3 cm in diameter and form layer-like horizons or trails of variable thickness, interpreted as primary magma flow features (Fig. 11).

Upper mafic unit

This is a sequence nearly 50 m thick consisting of fine grained tuffs and massive lavas, in which relict structures resembling amygdules can be discerned. Mineralogically, the lavas consist of hornblende, oligoclase and quartz with some biotite. The intervening tuffaceous units are biotite-rich and contain, in addition to amphibole and quartz, patches of carbonate and chlorite-rich intercalations. Thin calc-silicate bands containing garnet porphyroblasts are widespread. The uppermost part of the unit consists of banded amphibolite containing diopside in addition to hornblende and plagioclase. The mafic unit is overlain by a coarse-grained skarn rock about 3 m thick consisting mainly of diopside and fibrous, radiating amphibole intergrowths.

Quartzite formation

The quartzites overlying the volcanite formation are arkosic in composition, and blastoclastic in texture. Lowermost, immediately overlying the amphibole-diopside skarn rock, is a fine grained sequence of recrystallized laminated arenites which contain diopside in addition to quartz and plagioclase, and may include further biotite-bearing amphibole-diopside skarn interlayers. These latter become less frequent upwards, and quartzite comprises 0.5 – 2 cm thick beds alternating from arkose to occasionally almost orthoquartzites in composition. Higher up in the sequence, coarser grained arkosic quartzites with well-preserved cross-bedding occur, and once more contain skarn mineral assemblages. Commonly, the quartzites are pinkish in color, but diamond drilling in the western part of Möykkelmä revealed that the quartzite sequence also contains greenish sericite quartzite and orthoquartzite as well as graded mica schist intercalations.

GEOCHEMISTRY

Granitic gneiss basement

Major and trace elements were analyzed from nine samples (Table 1, analyses 1 – 9), of which six represent fine grained granitic gneisses of variable composition (Table 1, analyses 1 – 6) and three somewhat coarser-grained, massive quartz-plagioclase dominated granite (trondjemite) (Table 1, analyses 7 – 9). Lanthanides were also analyzed from two samples (Table 4).

Most major components of the granitic gneisses typically show a rather broad variation, e.g. SiO₂ (63.18 – 76.42%), Al₂O₃ (13.60 – 18.31%), Fe₂O₃ (1.21 – 4.85%), CaO (0.91 – 2.89%) and MgO (0.35 – 2.36%). The ratio Na₂O/K₂O is generally below 2 and fairly uniform. The trondjemites are characterized by high and constant SiO₂ (73.21 – 75.71%) and Na₂O (4.93 – 5.54%) contents and a high Na₂O/K₂O ratio (2 – 3.5). The values of Fe₂O₃, MgO, CaO, TiO₂ and P₂O₅ are constant and lower than those in the gneisses.

Trace elements abundances for both gneisses and trondjemites are given in Table 1 (analyses 1 – 9) while the mantle-normalized abundances of Rb, Ba, Th, U, Ta, La, Sr, Zr, Sm and K₂O, P₂O₅ and TiO₂ are presented in Fig. 12. Granitic gneisses show a large variation in trace element contents, with the exception of a constant Ta level. They generally have high Ba contents (707 – 2760 ppm) while Rb and Sr are relatively low (55.6 – 149 ppm and 110 – 330 ppm respectively). Trondjemites display a different trace element pattern with respect to the granitic gneisses (Fig. 12). In particular, Ba and P₂O₅, but also Sr, Zr, Sm, K₂O and TiO₂ values are lower than in the granitic gneisses; Ta is conversely higher in the trondjemite. La contents and La/Sm indicate a similar REE distribution for both trondjemites and gneisses (Fig. 13), showing clear LREE fractionation ((La/Sm)_N = 3 – 7) with variable enrichment ((La)_N = 20 –

Table 1. Major and trace element compositions of selected granitoid samples from the Möykkelmä area.

	1	2	3	4	5	6	7	8	9
SiO ₂	66.36	71.22	74.79	68.84	62.26	72.48	72.45	74.18	75.37
TiO ₂	0.32	0.23	0.13	0.23	0.43	0.11	0.12	0.11	0.16
Al ₂ O ₃	16.44	14.13	13.74	14.10	18.04	14.49	14.98	13.99	13.41
FeO _{tot}	3.77	2.45	1.07	3.65	4.30	1.52	1.42	1.22	1.46
MnO	0.04	0.03	0.02	0.05	0.06	0.06	0.05	0.03	0.02
MgO	1.92	1.52	0.35	2.31	1.96	0.72	0.90	0.43	0.74
CaO	2.26	1.27	1.12	1.13	2.85	1.51	1.54	1.06	0.90
Na ₂ O	5.39	3.81	4.09	3.38	4.35	4.84	5.48	5.41	4.11
K ₂ O	1.90	3.26	3.31	3.95	3.93	2.43	2.00	1.53	2.39
P ₂ O ₅	0.31	0.11	0.05	0.10	0.32	0.02	0.02	0.02	0.05
Total	98.70	98.02	98.67	97.75	98.50	98.19	98.95	97.97	98.61
Cr	137	54.1	28.7	29.2	29.1	24.8	22.3	23.5	25.6
Ni	40.6	35.2	19.1	21.2	21.5	19.3	17.7	18	18.5
Co	16.7	8.19	1.67	11.9	12.4	1.7	1.59	5.5	6.51
Sc	14.6	2.87	2.09	10.2	7.83	2.42	2.58	1.83	2.26
V	60	40	20	50	70	30	20	20	20
Zr	90	220	140	90	160	60	60	120	390
La	43.3	35.1	8.51	38.1	32.5	34.8	15.1	32.3	97.3
Sm	5.68	4.37	1.06	5.54	6.21	3.96	1.89	3.2	8.56
Ba	724	1370	2760	1220	1800	603	561	683	707
Sr	280	210	220	120	330	120	120	140	110
Rb	74.9	82.5	76.3	149	101	113	95.2	38.9	55.6
Ta	0.267	0.375	0.268	0.349	0.367	1.86	1.08	0.678	0.471
Th	5.09	17.6	7.28	3.19	6.57	6.66	10.7	11.1	60.4
U	0.477	2.24	0.684	1.19	1.52	3.09	2.99	2.91	4.13
Cs	4.42	2.51	1.85	7.89	4.33	1.54	1.16	0.742	1.16
Cu	30	0	10	0	150	10	10	130	40
S	400	30	120	20	3390	120	350	1600	460
O	10.85	18.38	26.20	30.42	34.49	38.79	27.47	30.03	33.77
An	12.24	9.30	5.05	5.72	5.33	4.22	7.60	7.51	5.26
Ab	37.35	46.19	29.25	32.88	35.07	35.26	46.86	41.71	46.72
Or	24.54	11.87	24.49	20.36	21.11	14.70	12.28	14.98	9.61
C	2.06	2.09	2.40	2.19	1.29	2.50	0.99	1.15	1.51

1 = 5157, granite gneiss
 2 = 5602 "
 3 = 5603 "

4 = 5604 "
 5 = 5605 "
 6 = 5606 "

7 = 5609, trondhjemite
 8 = 5610 "
 9 = 5611 "

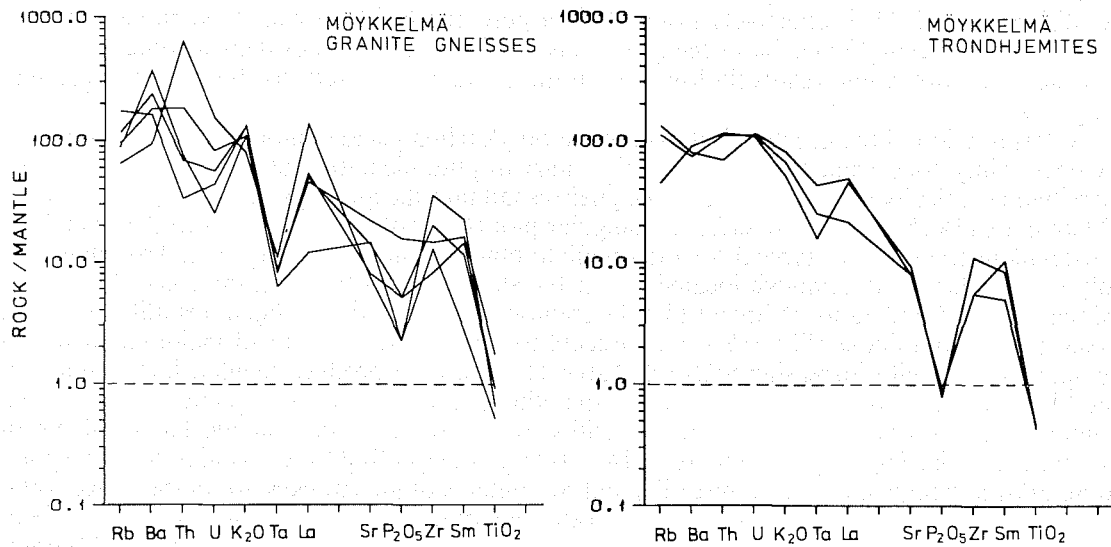


Fig. 12. Spidergrams of granite gneisses and trondhjemites from Möykkelmä normalized with respect to primitive mantle (Wood, 1979). Data from Table 1.

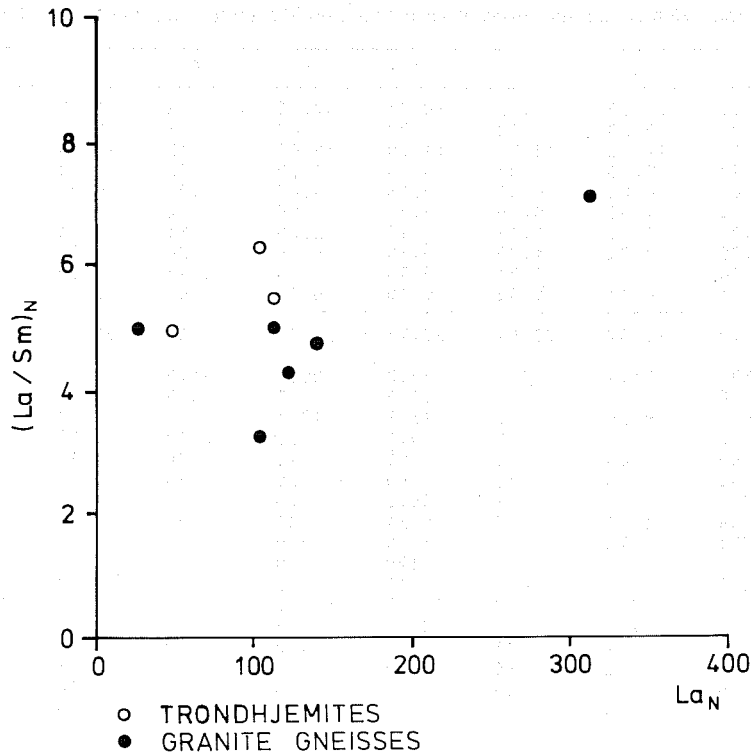


Fig. 13. Chondrite-normalized La/Sm and La of Möykkelmä trondhjemites and granite gneisses.

300). Results from the two gneiss samples analyzed more comprehensively are consistent with the above La/Sm data, also indicating strong lanthanide fractionation ($(La/Yb)_N = 10 - 14$). Analysis 5603 (granitic gneiss; Table 4, Fig. 22a) exhibits a weak positive Eu anomaly.

The commonly gradual transition from granitic gneisses to trondhjemitic lithologies at Möykkelmä suggests that the latter developed by partial melting and migmatization of the former, hornblende- and biotite-rich gneisses having remained below the solidus as restite. Partial melting of granitic gneiss is also supported by the major and trace element data, since the high SiO_2 contents (73 - 75%) of the trondhjemites indicate melting and recrystallization at conditions close to the granite eutectic. The low P_2O_5 contents of the trondhjemites with respect to the gneisses (Table 1, Fig. 12) are also considered a good indicator of minimum melt conditions since P_2O_5 is effectively insoluble in a pure quartz-feldspar melt (White & Chappel 1983). The higher K_2O and Ba in the gneisses is explicable in terms of their tending to remain in residual biotite, while relatively lower Sr in the trondhjemites reflects the effects of residual plagioclase.

Interpretation of the geochemical data of the Möykkelmä gneisses more precisely is hindered by uncertainty concerning the possible sedimentary or plutonic nature of their protoliths. When classified chemically as plutonic rocks, the gneisses fall into the granite and trondhjemite field of O'Connor's (1965) normative feldspar triangular plot (Fig. 14), and on the AFM diagram form a calc-alkali trend rather typical for intermediate plutonic intrusives (Fig. 15). Trondhjemites follow a similar trend, although located nearer the alkali apex of the diagram. According to the $Al/(Na + K + Ca/2)$ ratio of Shand (1951) granitic gneisses and trondhjemites alike are both meta- and peraluminous ($Si = 0.9 - 1.1$). Except for the alkalis, contents of major elements fall sympathetically with increasing SiO_2 (cf. Table 1) but do not produce regular, linear trends. If the Möykkelmä gneisses are considered, on the other hand, to be metasediments whose present composition represents that of their protoliths, they would be the chemical equivalents of graywackes (Fig. 16); Al_2O_3 contents of 14 - 18% and Na_2O/K_2O are too high to correspond to an arkosic composition. The mean Rb and Sr contents of graywackes are in the ranges 40 - 100 ppm and 220 - 520 ppm, respectively (Peterman *et al.* 1967); the former is comparable to the Möykkelmä gneisses but Sr is clearly too low, while the high Ba content (up to 2700 ppm) observed at Möykkelmä is atypical for graywackes, it being to a large extent adsorbed onto clay minerals in shallow water (Sahama 1947).

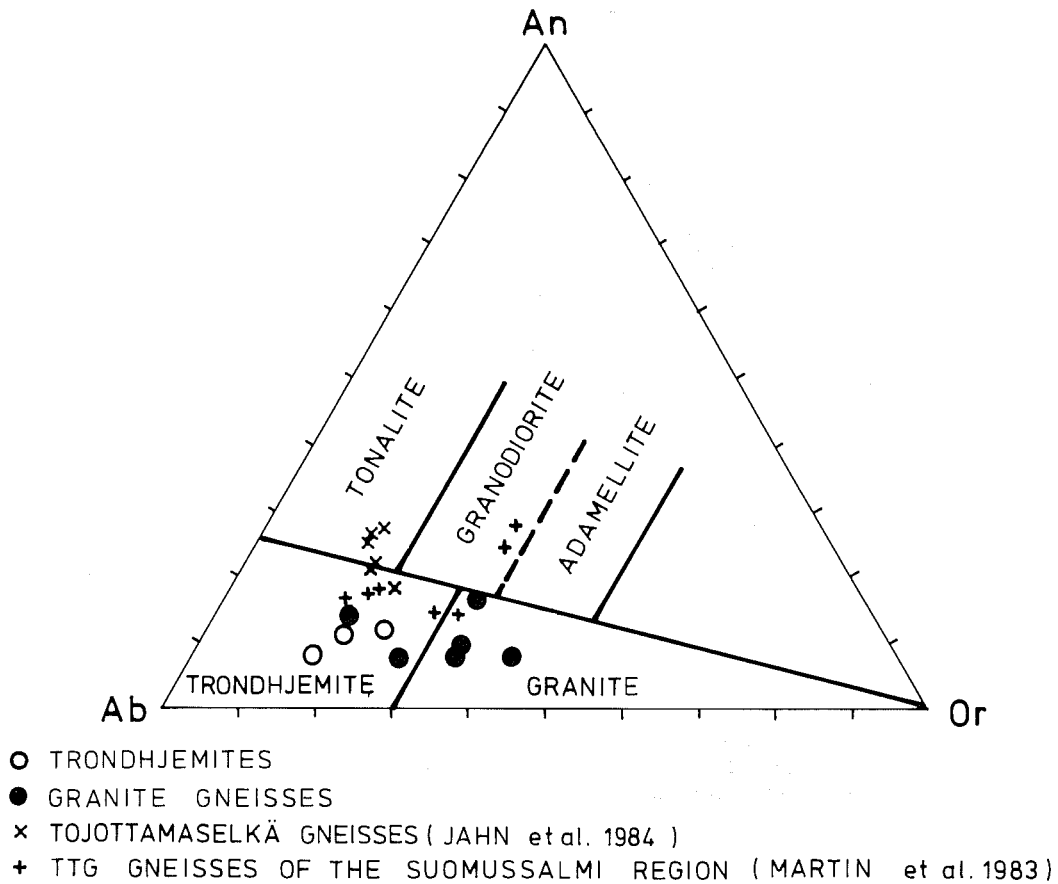


Fig. 14. Ab - An - Or diagram for Möykkelmä trondhjemites and granite gneisses.

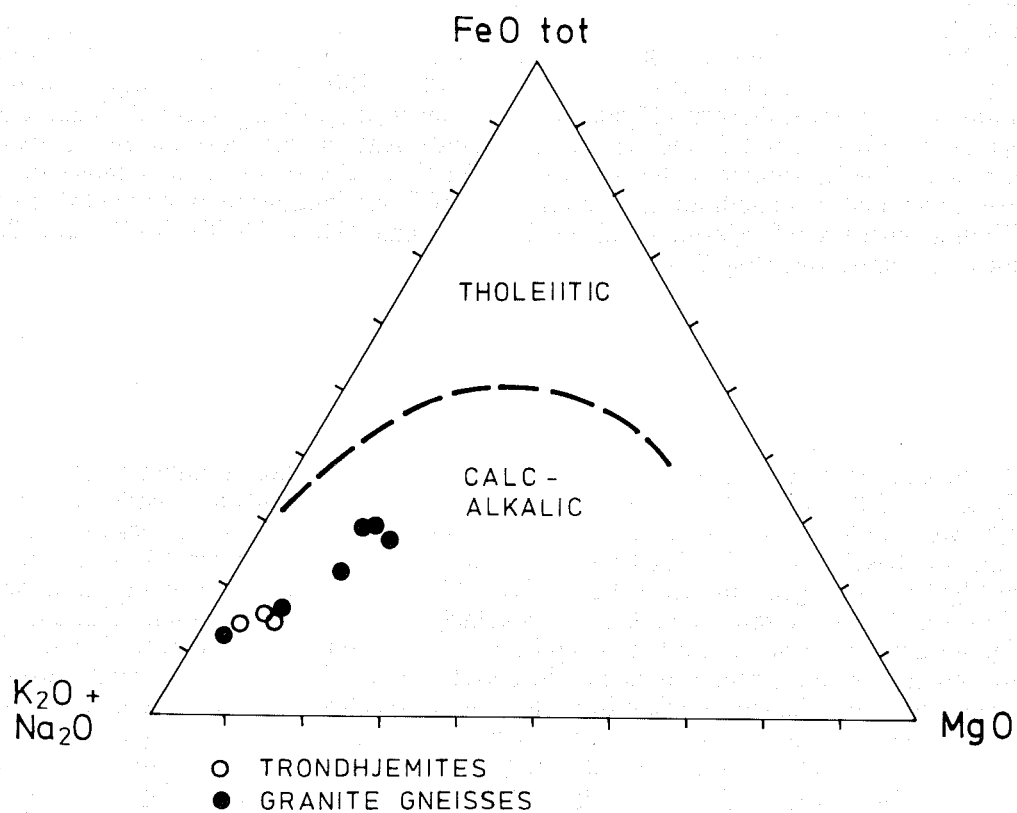


Fig. 15. AFM diagram for Möykkelmä trondhjemites and granite gneisses.

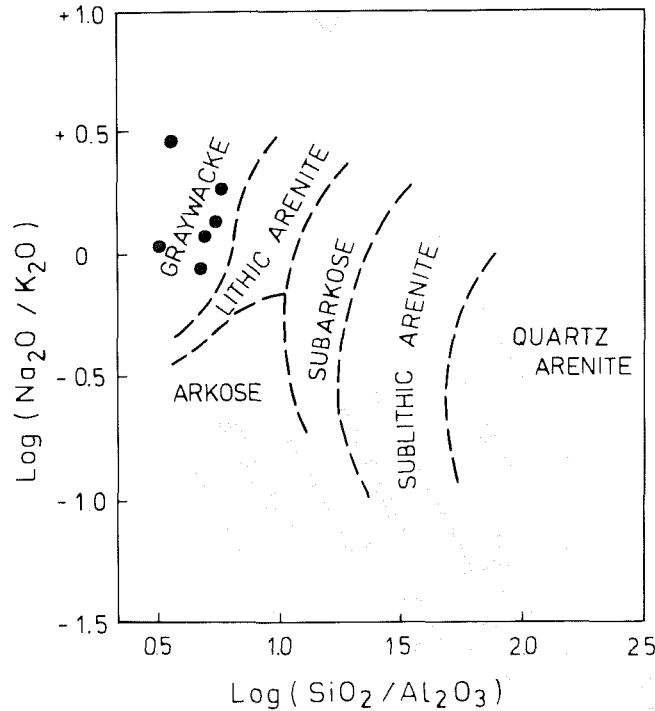


Fig. 16. Analyses of Möykkelmä granite gneisses plotted on a diagram of Pettijohn *et al.* (1973) for comparison with compositions of various sedimentary rocks.

The Möykkelmä granitic gneisses approach a true granitic composition more closely than do the Suomussalmi gray gneisses, and especially the Tojottamaselkä TTG gneisses (Fig. 14), and differ chemically from typical Archean gneisses (e.g. Glikson 1979) in their generally higher SiO_2 (>70 vs. 60–70%) and Ba (700–2700 ppm vs. 20–1000 ppm) contents and lower FeO/MgO (<2 vs. >2) and Sr (110–330 ppm vs. 300–1000 ppm). In comparison to the Tojottamaselkä gneisses, those at Möykkelmä have higher SiO_2 , Fe_2O_3 , MgO, K_2O and Rb (cf. Kröner *et al.* 1981, tables 2 and 4), but lanthanide patterns are nevertheless comparable (cf. Jahn *et al.* 1984), being in both cases strongly fractionated and with some samples showing positive Eu anomalies. In addition to similarities of REE distributions between the Möykkelmä and Kuhmo gneisses (cf. Martin *et al.* 1983; Horneman 1985), Th/Ta, Ta/La and Zr/Ti relations are comparable (Fig. 22a).

Komatiites

On the commonly used discrimination diagrams, the Möykkelmä ultramafic volcanites are clearly seen to fall within the komatiite field (Fig. 17). MgO contents in volatile-free analyses range between 11.7–21.4% with a mean of 15.5%. With a few exceptions, the rocks are classified as komatiitic basalts, thus differing from the Sattasvaara komatiites some 10 km to the S, in which MgO contents generally exceed 20% (Fig. 18). Apart from certain mobile components, both lower and upper ultramafic units at Möykkelmä are mutually comparable (Fig. 25).

The ultramafic volcanites are both chemically and mineralogically strongly altered. This is readily apparent from the chemical data in the correlation matrix (Fig. 19), in which correlation coefficients for most elements deviate from the values expected for purely magmatic processes. SiO_2 shows for example significant negative correlation with TiO_2 , Al_2O_3 , K_2O , P_2O_5 , Zr, Ba, Rb and Cs whereas the opposite should be expected and conversely, the anticipated positive correlation of SiO_2 with $\text{FeO}/(\text{FeO} + \text{MgO})$, La, Sm, U, Th and Ta is lacking. The behaviour of CaO is exceptional in that it shows a negative correlation with two groups of elements having converse behavior during komatiite fractionation, namely MgO, Ni, Cr, Co and TiO_2 , Al_2O_3 ,

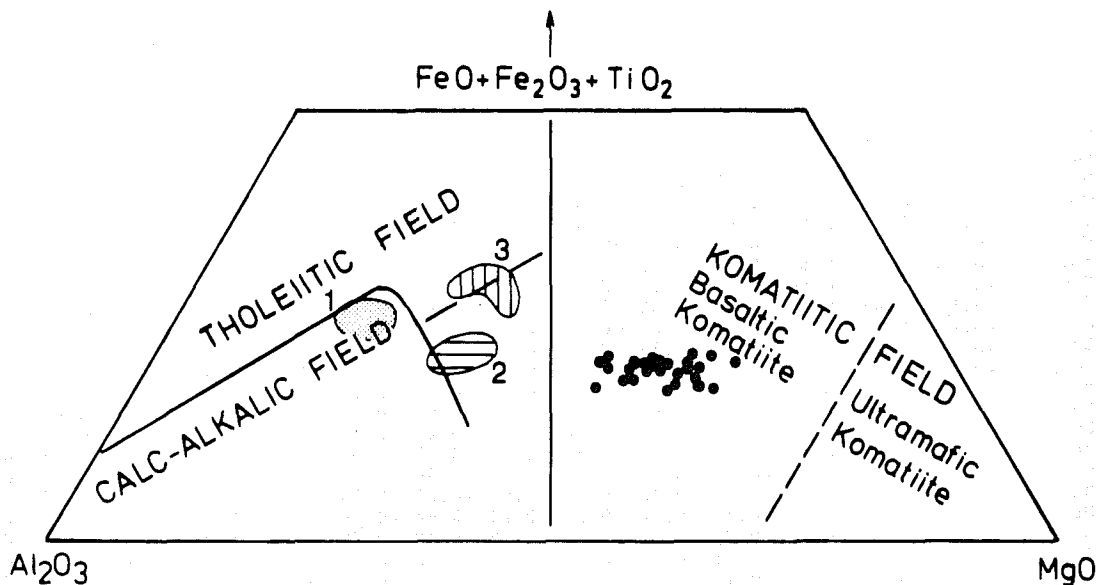


Fig. 17. Jensen plot of ultramafic and mafic volcanites from Möykkelmä. Fields: 1 = lower mafic unit, 2 = Middle mafic unit, 3 = upper mafic unit.

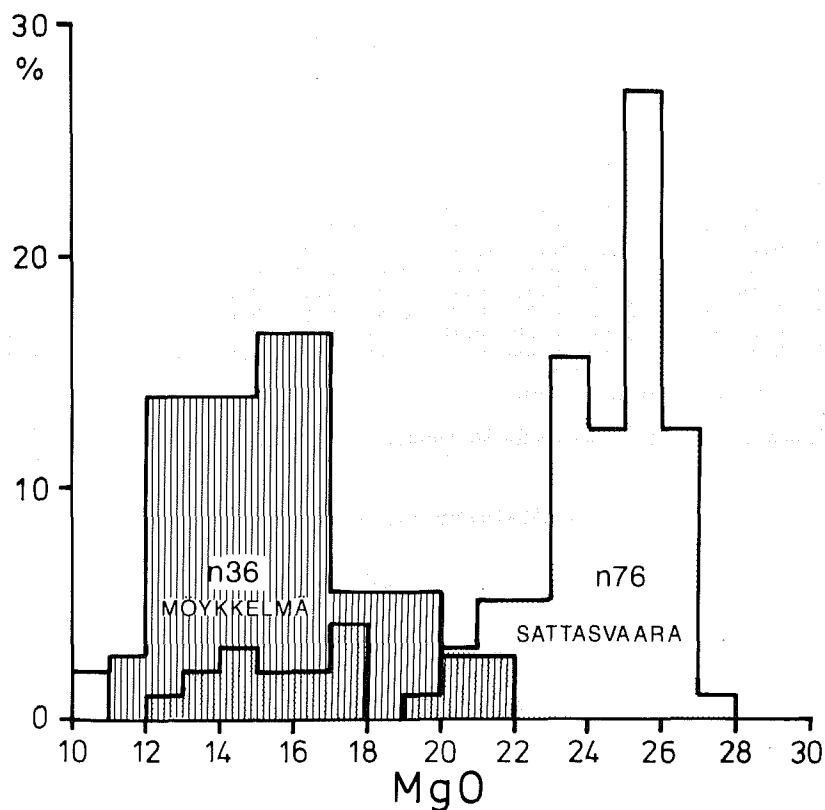


Fig. 18. Histogram showing the distribution of MgO contents in komatiites from the Möykkelmä and Sattasvaara areas.

P_2O_5 . The correlation matrix also shows total FeO to decrease sympathetically with falling MgO values, although it should normally be expected to increase somewhat during differentiation of a komatiitic melt.

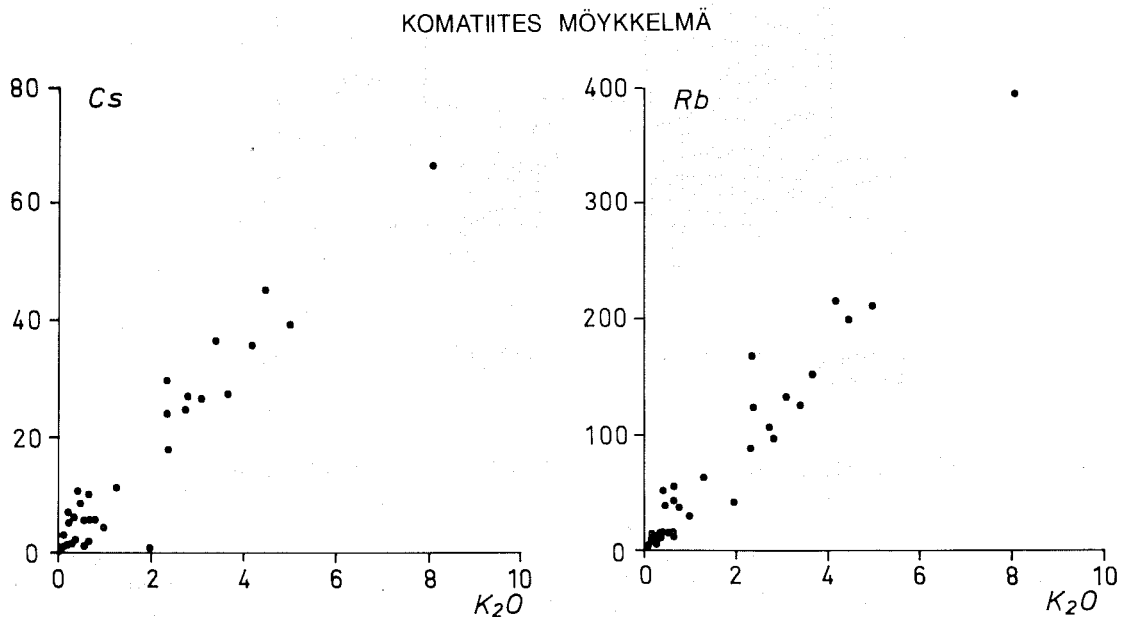
Al_2O_3 and TiO_2 show a highly positive mutual correlation but with respect to MgO for example, their coefficients are 0.36 and 0.27, in contrast to the clear negative correlation which would be expected after magmatic processes. MgO and elements having similar behaviour notably Cr, Ni and Co do not in general display a significant negative correlation with respect

	SiO ₂	TiO ₂	Al ₂ O ₃	FeOtot	Fe*	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Cr	Ni	Co
SiO ₂	1.00													
TiO ₂	-0.70	1.00												
Al ₂ O ₃	-0.77	0.79	1.00											
FeOtot	-0.84	0.59	0.47	1.00										
Fe*	-0.07	0.06	0.11	0.04	1.00									
MnO	0.06	-0.39	-0.44	0.20	0.22	1.00								
MgO	-0.54	0.36	0.27	0.68	-0.70	-0.01	1.00							
CaO	0.45	-0.60	-0.54	-0.48	0.51	0.34	-0.72	1.00						
Na ₂ O	0.32	-0.29	0.00	-0.49	0.59	0.09	-0.75	0.53	1.00					
K ₂ O	-0.70	0.64	0.61	0.46	0.12	-0.27	0.21	-0.51	-0.33	1.00				
P ₂ O ₅	-0.51	0.60	0.64	0.38	-0.13	-0.33	0.38	-0.43	-0.03	0.19	1.00			
Cr	-0.59	0.42	0.38	0.58	-0.09	0.07	0.49	-0.20	-0.24	0.14	0.34	1.00		
Ni	-0.60	0.41	0.29	0.59	-0.22	-0.07	0.55	-0.23	-0.50	0.29	0.21	0.73	1.00	
Co	-0.71	0.46	0.35	0.76	-0.17	0.09	0.66	-0.35	-0.56	0.37	0.22	0.73	0.83	1.00
Sc	-0.60	0.16	0.22	0.58	-0.20	0.24	0.56	-0.23	-0.40	0.31	0.12	0.65	0.60	0.72
V	-0.71	0.60	0.60	0.56	0.04	-0.01	0.36	-0.47	-0.28	0.70	0.22	0.47	0.45	0.58
La	-0.01	-0.01	0.17	-0.22	0.06	-0.15	-0.20	0.02	-0.03	0.24	-0.21	-0.12	0.01	0.01
Sm	-0.02	-0.36	-0.03	0.02	0.16	0.34	-0.06	0.23	0.31	-0.24	-0.11	0.16	-0.03	0.04
Zr	-0.53	0.69	0.73	0.38	0.01	-0.37	0.28	-0.45	-0.01	0.32	0.75	0.16	0.12	0.08
Sr	0.20	-0.20	0.13	-0.47	0.49	-0.02	-0.68	0.44	0.89	-0.16	0.75	0.16	-0.39	-0.50
Ba	-0.42	0.42	0.35	0.40	0.42	0.01	-0.05	-0.21	-0.06	0.61	0.10	-0.08	0.12	0.20
Rb	-0.69	0.68	0.63	0.45	0.08	-0.33	0.24	-0.54	-0.36	0.98	0.22	0.14	0.27	0.34
U	-0.34	0.54	0.55	0.28	0.01	-0.39	0.22	-0.40	-0.20	0.26	0.41	0.00	-0.01	0.06
Th	-0.24	0.41	0.45	0.15	-0.04	-0.33	0.15	-0.42	-0.11	0.31	0.30	-0.01	0.08	0.09
Ta	-0.14	0.42	0.29	0.17	0.38	-0.17	-0.15	-0.11	0.18	0.17	0.06	0.13	0.05	0.02
Cs	-0.68	0.61	0.58	0.45	0.04	-0.23	0.26	-0.54	-0.35	0.97	0.17	0.18	0.32	0.36
Cu	-0.29	0.45	0.23	0.18	0.26	-0.35	-0.09	-0.01	-0.15	0.43	-0.01	0.09	0.09	0.08
S	-0.49	0.72	0.54	0.42	0.23	-0.38	0.10	-0.44	-0.26	0.69	0.22	0.02	0.10	0.12

	Sc	V	La	Sm	Zr	Sr	Ba	Rb	U	Th	Ta	Cs	Cu	S
Sc	1.00													
V	0.56	1.00												
La	0.19	0.33	1.00											
Sm	0.43	0.02	0.29	1.00										
Zr	0.04	0.28	-0.05	-0.09	1.00									
Sr	-0.36	-0.19	0.07	0.20	0.09	1.00								
Ba	0.10	0.42	0.21	-0.07	0.39	0.06	1.00							
Rb	0.29	0.70	0.22	-0.28	0.34	-0.19	0.53	1.00						
U	-0.14	0.25	0.07	-0.14	0.64	-0.17	0.20	0.30	1.00					
Th	0.04	0.38	0.33	0.04	0.53	-0.06	0.49	0.29	0.41	1.00				
Ta	-0.21	0.18	-0.06	-0.06	0.36	0.11	0.35	0.17	0.37	0.44	1.00			
Cs	0.35	0.72	0.23	-0.22	0.26	-0.20	0.51	0.97	0.20	0.28	0.12	1.00		
Cu	-0.11	0.21	-0.12	-0.42	0.20	-0.18	0.31	0.46	0.33	0.18	0.39	0.40	1.00	
S	-0.10	0.37	-0.02	-0.50	0.46	-0.21	0.48	0.71	0.56	0.23	0.39	0.66	0.74	1.00

Fe* = FeOtot / (FeOtot + MgO)

Fig. 19. Interelement correlation coefficients for 36 samples of Mykkelm komatiites.

Fig. 20. Covariation of Rb and Cs with K₂O in Mykkelm komatiites.

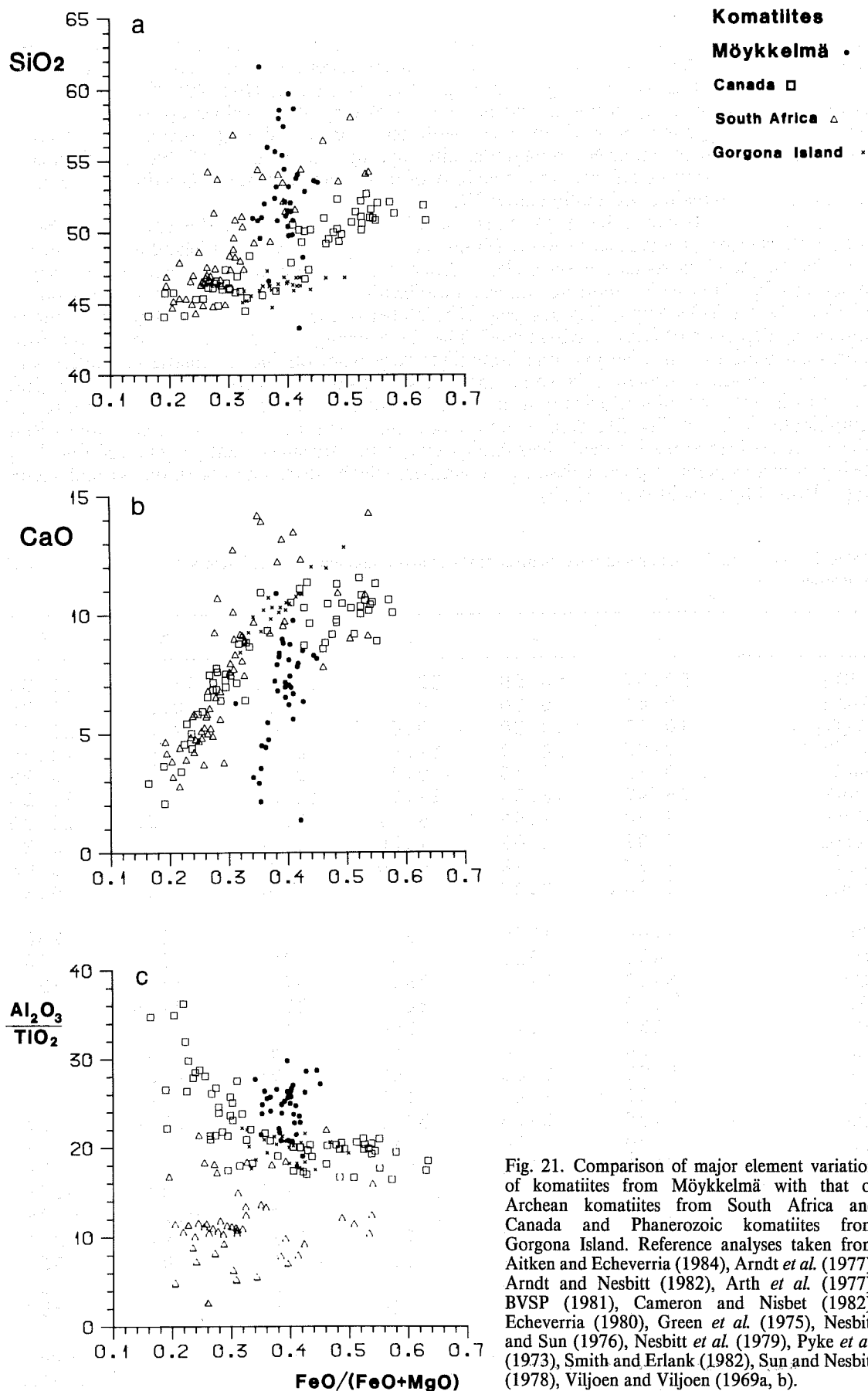


Fig. 21. Comparison of major element variation of komatiites from Möykkelmä with that of Archean komatiites from South Africa and Canada and Phanerozoic komatiites from Gorgona Island. Reference analyses taken from Aitken and Echeverria (1984), Arndt *et al.* (1977), Arndt and Nesbitt (1982), Arth *et al.* (1977), BVSP (1981), Cameron and Nisbet (1982), Echeverria (1980), Green *et al.* (1975), Nesbitt and Sun (1976), Nesbitt *et al.* (1979), Pyke *et al.* (1973), Smith and Erlank (1982), Sun and Nesbitt (1978), Viljoen and Viljoen (1969a, b).

to other incompatible elements. Incompatible elements show on the whole widely varying mutual correlation coefficients, with a virtual absence of significant positive values. For example, coefficients for La fall within the range 0.12 - 0.33. An exception is, however, provided by Rb, Cs and K_2O which show very high mutual positive correlations (Fig. 20). K_2O contents are several per cent in many instances and in some cases may attain values up to 8%, obviously indicating introduction of K_2O together with Rb and Cs during secondary alteration processes, which is reflected mineralogically by the abundance of biotite. Potassium metasomatism was more effective in the vicinity of the basement, since K_2O values of the upper ultramafic unit remain consistently below 2%.

A comparison between the Möykkelmä komatiites and well preserved examples from South Africa (Barberton), Canada (principally Munro Township) and Gorgona Island is shown in Fig. 21. The Möykkelmä rocks present a clear vertical trend on the SiO_2 vs. $FeO/(FeO + MgO)$ diagram (Fig. 21a) which indicates intense mobilization of SiO_2 , while most analyses on the CaO vs. $FeO/(FeO + MgO)$ diagram (Fig. 21b) fall below the values for the reference examples, indicating depletion of CaO at Möykkelmä. The above changes are also readily discernible from norm calculations, since Möykkelmä komatiites vary from olivine- to quartz-normative, and even to corundum-normative in some cases. The considerable variation in SiO_2 and CaO contents is reflected mineralogically in the variation of the relative proportions of chlorite and amphibole. Although Fig. 21a depicts the Möykkelmä komatiites as occurring generally within the South African komatiite field, their Al_2O_3/TiO_2 ratios are substantially higher (cf. Fig. 21c), being in fact as great or greater than for chondrites, which corresponds with the Al-undepleted type defined by Nesbitt *et al.* (1979).

Table 2. Representative major and trace element analyses and minimum and maximum values of the Möykkelmä komatiites.

	1	2	3	4	5	6
SiO_2	52.37	50.99	52.07	59.70	43.33	61.59
TiO_2	0.41	0.50	0.46	0.44	0.33	0.86
Al_2O_3	10.93	13.30	9.61	9.09	7.64	16.35
FeO_{tot}	9.98	10.19	11.68	8.71	7.65	12.26
MnO	0.23	0.15	0.27	0.21	0.11	0.29
MgO	16.42	18.35	17.78	12.85	11.59	21.38
CaO	7.20	4.46	6.52	7.40	1.31	10.90
Na_2O	1.01	1.09	1.20	1.24	0.30	2.83
K_2O	1.27	0.78	0.20	0.21	0.10	8.08
P_2O_5	0.06	0.08	0.09	0.06	0.05	0.11
Cr	1700	1970	1950	1730	1190	2360
Ni	450	513	483	386	263	587
Co	67.3	79.2	78.5	57.1	46.2	79.2
Sc	31.9	33	30.9	27.1	21.8	35.9
V	150	170	160	160	120	220
Zr	60	70	50	50	50	100
La	5.61	7.63	.68	5.76	2.18	10.4
Sm	1.97	1.88	1.8	1.75	0.923	2.22
Ba	135	36.8	37.9	36.3	31.4	574
Sr	20	30	20	30	10	140
Rb	63.5	35.4	14.3	6.6	5.9	395
Ta	0.145	0.104	0.191	0.156	0.098	0.355
Th	0.426	0.355	0.178	0.26	0.129	0.754
U	1.39	1.59	1.2	1.12	0.411	2.48
Cs	11.4	5.89	6.93	5.19	0.868	66.8
Cu	10	0	10	30	0	220
S	30	20	40	40	10	3570

1 = 5588 } komatiite
2 = 5594 } (lower)

3 = 5569 } komatiite
4 = 5570 } (upper)

5 = minimum values
6 = maximum values

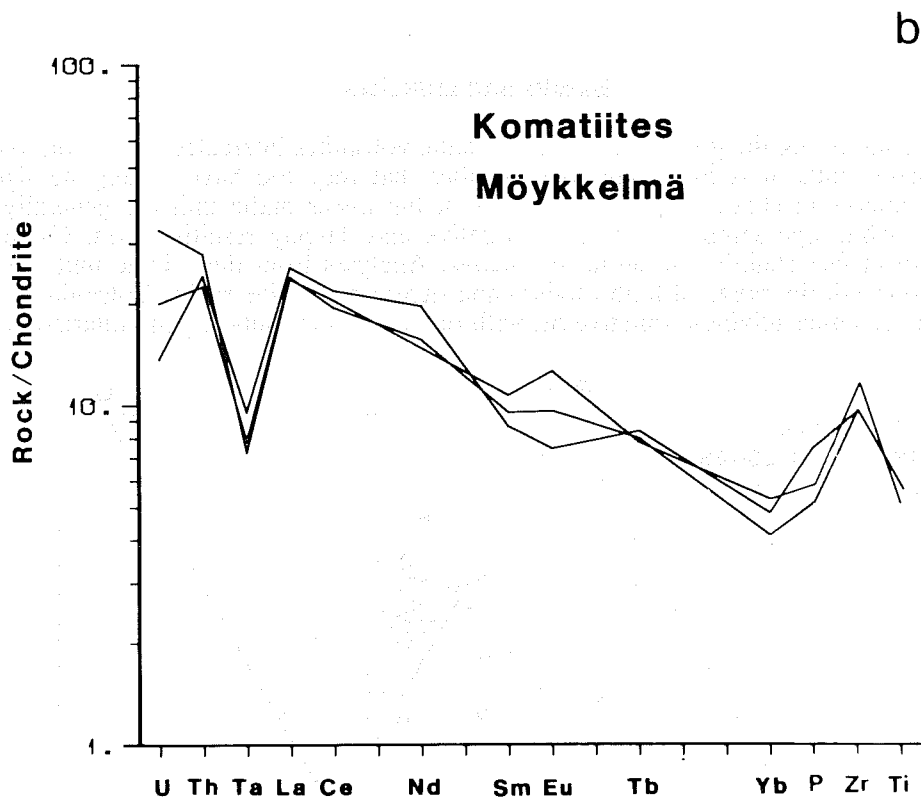
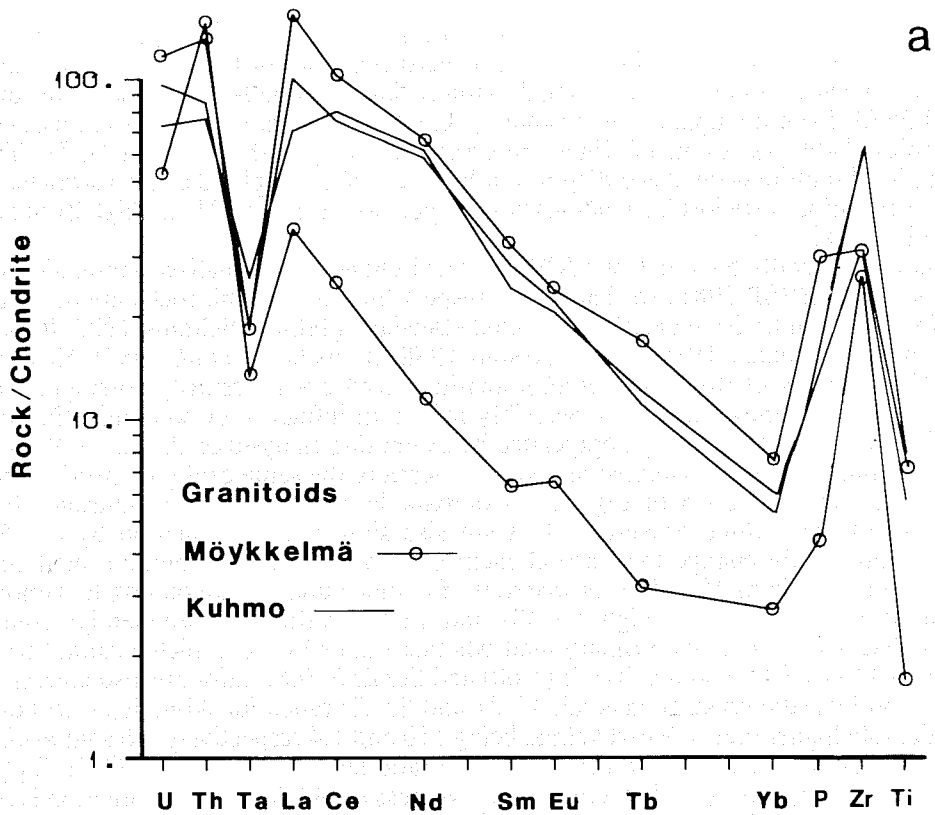


Fig. 22. Chondrite-normalized abundance diagrams of REE and other trace elements for granitoids from Möykkelmä and Kuhmo (a) and for komatiites from Möykkelmä (b). Kuhmo analyses are from Horneman (1985) and unpublished data of the University of Oulu (The Archean Research Project).

Owing to this pervasive alteration, most chemical analyses do not reveal much concerning the primary magmatic nature of the rocks, but nevertheless four of the best preserved drill-core samples have been examined in more detail, namely 5588 and 5594 (from the lower unit) and 5569 and 5570 (from the upper unit) (Table 2). Lanthanide elements were also analyzed from these samples (Table 4), results of which are presented, along with values for U, Th, Ta, P, Zr and Ti, in the chondrite normalized diagram in Fig. 22b. The samples display a straight, LREE-enriched distribution, very low Ta contents with respect to La, Th and U and high Zr values with respect to P and Ti.

Komatiites generally have flat or LREE-depleted chondrite-normalized lanthanide patterns (Arth *et al.* 1977; BVSP 1981) so that in this respect the Möykkelmä rocks are atypical, even though similar examples have been documented elsewhere (Jahn & Schrank 1983; Rajamani *et al.* 1985; Arndt & Jenner 1986; Sun & Nesbitt 1978; Ryabchikov *et al.* 1985). However, the present REE patterns of the Möykkelmä komatiites need not necessarily correspond to their primary values since they could very plausibly reflect an influence of contamination by sialic crust. This possibility can be easily appreciated by examining analyses of U, Th, Ta, P, Zr, Ti and REE from granitoid gneisses sampled at a greater depth in the same drill core profile, presented in chondrite-normalized form in Fig. 22a. Comparable analyses of two granitoids from the surroundings of the Kuhmo greenstone belt are also shown. Contamination by the Archean upper crust can subtly change the trace element compositions of a komatiitic melt such that assimilation of as little as 10 – 12% of material of composition corresponding to sample 6505 would result in the observed high La/Yb and La/Ta values. Consequently, conclusions concerning REE-patterns of the primary melt whether flat or LREE depleted cannot be drawn.

The high $\text{Al}_2\text{O}_3/\text{TiO}_2$ and low Ti/Zr (<60) and Sc/Zr (<0.6) ratios are also consistent with contamination by sialic crust. In general, Ti/Zr and Sc/Zr ratios for Al-undepleted komatiites are considerably higher than at Möykkelmä, being 110 and 1.4 respectively (Nesbitt *et al.* 1979). Moreover, komatiites having a high MREE/HREE ratio tend to have low $\text{Al}_2\text{O}_3/\text{TiO}_2$ (Jahn *et al.* 1982). This is in fact observed at Sattasvaara, whereas at Möykkelmä the situation is reversed (Fig. 21c). The possibility of contamination is clearly further supported by the occurrence of granitoid fragments in the lower komatiite unit.

Basalts and andesites

When considering the geochemistry of the mafic volcanites intercalated with and overlying the ultramafic units, it is important to remember that they too have undergone significant secondary alteration (Figs. 23 and 24). Rocks of the lower mafic unit are generally quartz normative, while one sample is olivine normative and, largely resulting from Ca loss, two samples are in fact slightly corundum normative. Analyses from the middle unit are widely dispersed through the range of both olivine- and quartz-normative rocks. Volcanites from the upper unit are either olivine-normative or, with one exception, only slightly quartz-normative.

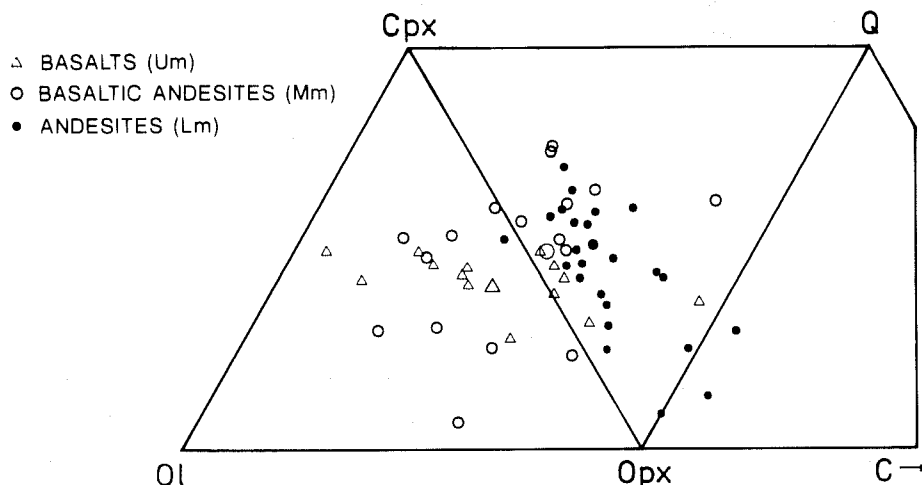


Fig. 23. CIPW normative compositions of basalts, basaltic andesites and andesites. Average compositions shown by larger symbols.

Within the unit, however, considerable variation in normative compositions exists. On the basis of changes in contents of immobile incompatible elements, such as Zr, it seems clear that this variation in normative compositions is not a product of magmatic processes but rather secondary alteration has been important. In some cases, quartz-filled amygdules are so abundant that it proved impossible to avoid them during sampling.

A large proportion of analyses fall outside Hughes' (1973) "igneous spectrum" (Fig. 24). For both the lower and upper units, the total of alkalis remains rather constant, but K_2O/Na_2O ratio varies considerably, in the lower unit commonly being lower than in "pristine, unaltered" rocks, while in the upper unit deviating equally from the mean in both directions. Analyses from the middle unit are also dispersed vertically and commonly deviate from the "igneous spectrum" field towards the left, in the direction of small K_2O/Na_2O ratios.

All of the Möykkelmä mafic volcanites are subalkaline in character with the upper unit falling into the tholeiitic field in all tholeiite-calc alkali major element discriminant diagrams (viz. AFM, Jensen cation plot (Fig. 17), FeO vs. FeO/MgO, SiO_2 vs. FeO/MgO, Al_2O_3 vs. normative plagioclase; Irvine & Baragar 1971; Jensen 1976; Miyashiro 1974). The lower unit falls within the calc-alkali field on all diagrams except for the FeO vs. FeO/MgO plot, while the rocks of the middle unit plot on either or both calc-alkali and tholeiitic fields in all the above mentioned diagrams. However, three of these plots include the elements Si, Ca, Na and K which at Möykkelmä were in at least some cases demonstrably mobile, so that results should be treated

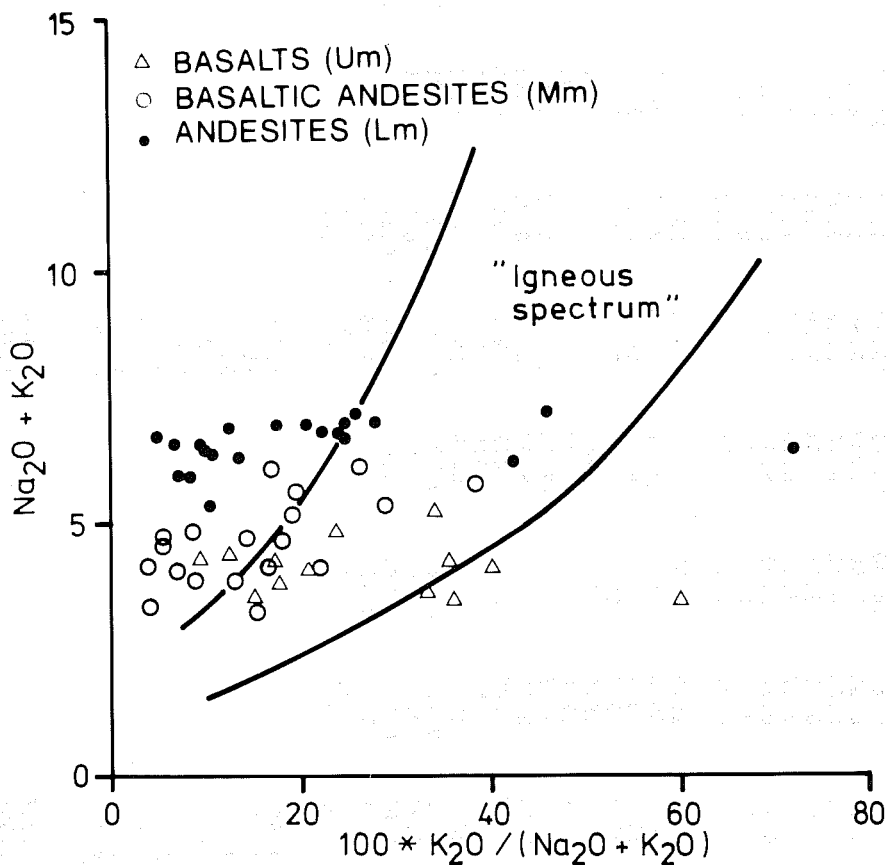


Fig. 24. Sum and ratio of alkali elements for volcanites from the andesite, basaltic andesite and basalt units. "Igneous spectrum" after Hughes (1973).

with some reservation. The effects of secondary alteration are less pronounced in the Jensen diagram (Fig. 17) where each of the mafic units plots as a discrete group, with rocks of the lower unit being located in the calc-alkali field and the other two units being tholeiitic. In spite of the fact that in an overall sense the three units all share broadly similar trace element characteristics (Fig. 27), they cannot be simply classified as belonging to the same series. It must also be remembered that in addition to secondary alteration, the effects of silic contamination (to be considered below) may have changed the primary compositions such that classification becomes very difficult.

Table 3. Average, minimum and maximum values and standard deviation of major and trace elements for volcanites from the three mafic units. CIPW norm calculations based on average compositions.

	1	2	3	4	5	6	7	8	9	10	11	12
SiO ₂	60.01	1.09	58.65	61.83	56.47	2.42	52.51	60.50	52.44	1.53	49.83	54.48
TiO ₂	0.64	0.04	0.53	0.73	0.52	0.04	0.39	0.55	0.63	0.02	0.61	0.66
Al ₂ O ₃	14.64	0.40	14.00	15.41	13.48	1.01	11.42	15.13	13.14	0.48	12.13	13.92
FeO _{tot}	8.69	0.82	7.50	10.03	8.84	1.25	6.83	10.81	12.80	1.28	11.30	15.82
MnO	0.16	0.03	0.12	0.20	0.18	0.02	0.15	0.23	0.23	0.03	0.20	0.29
MgO	3.83	0.33	3.16	4.44	7.38	0.81	5.96	8.52	8.01	0.97	6.31	9.73
CaO	5.19	0.82	3.57	6.75	8.24	1.72	5.57	10.84	8.57	0.84	7.36	10.11
Na ₂ O	5.45	0.60	3.55	6.33	3.95	0.53	3.17	5.03	2.89	0.70	1.35	3.91
K ₂ O	1.13	0.64	0.33	2.61	0.74	0.45	0.14	1.59	1.07	0.52	0.39	2.00
P ₂ O ₅	0.14	0.01	0.13	0.16	0.09	0.01	0.07	0.10	0.07	0.01	0.05	0.11
Cr	57.7	15.8	32	85.9	426.6	86.1	236	579	884.5	112.3	732	1110
Ni	34.9	10.3	18.2	54.2	96.4	21.4	66.9	130	143.0	43.5	47.1	247
Co	35.3	5.1	27.3	43.3	46.4	6.49	34.4	54.7	63.7	6.36	55.8	80.6
Sc	27.4	3.0	20.0	31.5	38.2	5.8	24.1	47.9	55.5	7.9	39.0	68.1
V	181	19.8	160	240	182	16	140	210	283	62	230	460
Zr	136	12.6	110	170	78	6	70	90	60	4	50	70
La	14.16	6.28	4.41	27.20	12.31	6.05	4.57	27.0	5.51	2.12	2.04	10.80
Sm	3.37	0.92	1.64	5.97	2.93	0.99	1.43	5.91	2.06	0.23	1.79	2.71
Ba	322	209	34.2	829	242	161	45.8	560	363	177	120	739
Sr	174	50	100	310	122	42	40	190	125	40	30	170
Rb	26.3	19.7	5.29	64.4	18.4	14.5	5.27	60.5	41.4	29.7	6.28	118
Ta	0.425	0.130	0.267	0.798	0.293	0.080	0.195	0.516	0.284	0.104	0.214	0.610
Th	5.31	0.52	4.08	6.54	2.46	0.24	1.90	2.82	2.66	0.15	2.45	2.95
U	1.00	0.226	0.754	1.63	0.544	0.143	0.376	0.973	0.568	0.124	0.333	0.763
Cs	2.18	1.68	0.439	5.77	1.19	0.84	0.404	3.57	4.09	2.94	0.508	11.5
Cu	64	13	10	160	78	65	10	250	90	78	20	310
S	176	260	10	1110	730	1055	10	3810	555	950	20	3470
Q	4.805				1.698							
An	12.15				16.87				19.72			
Ab	46.12				33.43				24.46			
Or	6.678				4.373				6.324			
Wo-Di	5.298				9.781				4.773			
En-Di	2.338				5.597				4.322			
Fs-Di	2.945				3.754				11.26			
En	7.200				12.78				10.20			
Fs	9.071				8.572				2.742			
Fo	—				—				2.736			
Fa	—				—				1.197			
Il	1.216				0.988				1.197			
Mt	1.869				1.900				2.756			
Ap	0.332				0.213				0.166			
Chr	0.012				0.092				0.190			

1 = andesite, average (n=19)
 2 = " standard deviation
 3 = " minimum
 4 = " maximum

5 = basaltic andesite, average (n=16)
 6 = " standard deviation
 7 = " minimum
 8 = " maximum

9 = basalt, average (n=13)
 10 = " standard deviation
 11 = " minimum
 12 = " maximum

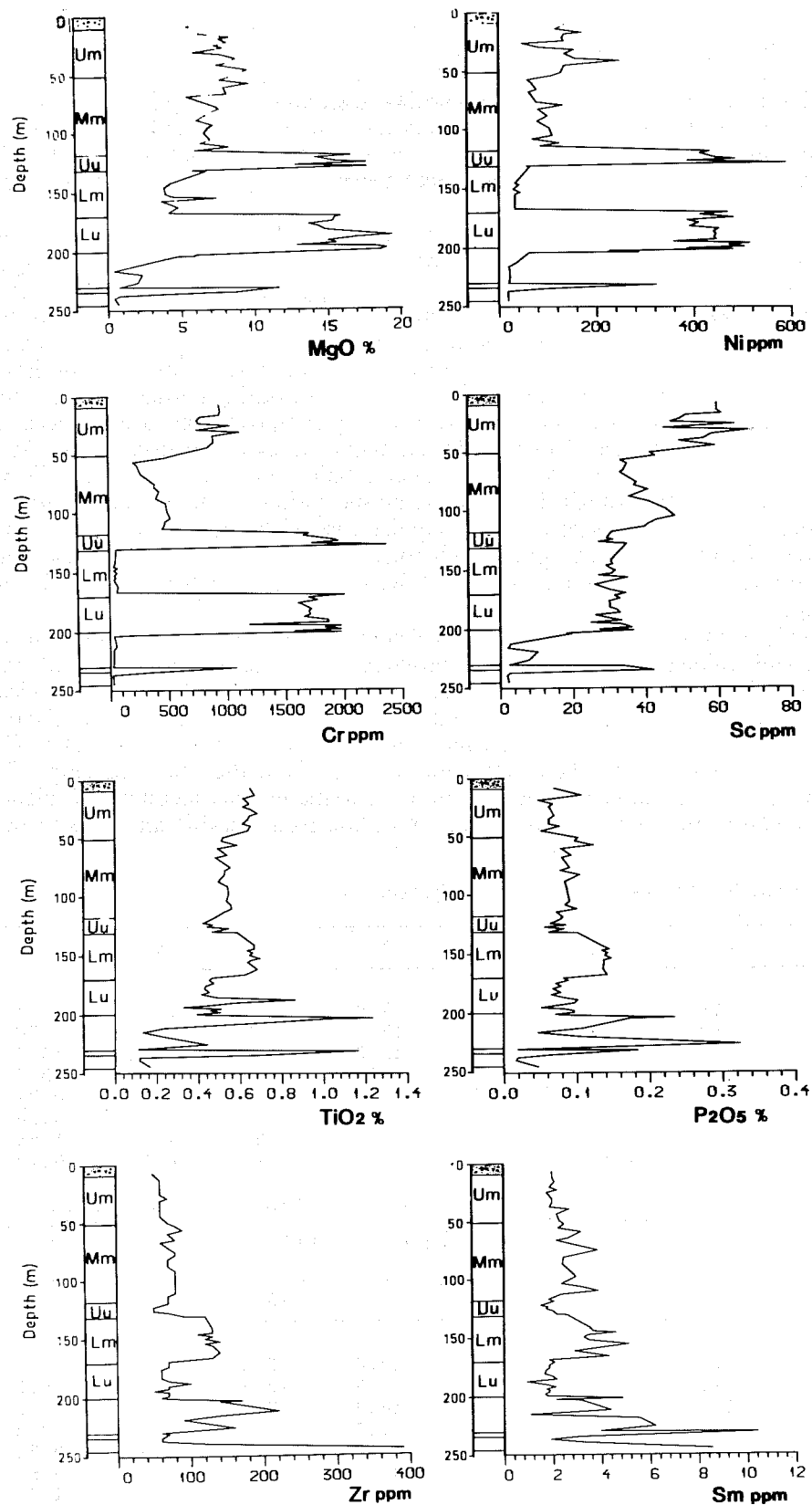


Fig. 25. Major and trace element profiles of a drill core through the Møykkelmä volcanic sequence. (Symbols shown in Fig. 4.)

Table 3 presents mean, absolute maximum and minimum values and standard deviations for each of the three mafic units, after excluding the most intensely altered samples. Normative compositions based on the mean values are also given, and are distinguished from other analyses in Fig. 23 by the use of larger symbols. According to its SiO₂ content (58.7 – 61.8%), the lower unit can be considered to be predominantly andesitic; MgO values fall within the range 3.2 – 4.4%. Drill core material from the very uppermost part of the unit, immediately below the second komatiitic unit, shows a transition from andesite to a more felsic, thin dacitic pyroclastic unit.

SiO₂ contents of the middle unit vary from 52.5 – 60.5% with a mean of 56.5% so that the rocks are of appropriate composition for basaltic andesites (Table 3), with MgO varying between 6.0 – 8.5% and having a mean of 7.4%. The upper unit, with SiO₂ values in the range 49.8 – 54.5% and MgO between 7.3 – 9.7% is accordingly classified as basaltic. Thus, from oldest to youngest, the mafic units show a trend of decreasing SiO₂ with a converse increase in MgO content. This so called antidromic transition from andesitic basalts through andesites to basalts contrasts with the more typical homodromic trend expected during normal magma fractionation.

Chemical profiles from the drill core data, showing compositional variation as a function of stratigraphic level, are presented in Fig. 25. As might be expected from the existence of the above-mentioned antidromic trend, contents of compatible elements such as Ni, Cr, Co and Sc clearly increase upwards through the stratigraphic succession, while the incompatible elements P₂O₅, Zr and light lanthanides correspondingly decrease. Other incompatible element amounts also tend to similarly decrease except that differences between the andesitic and basaltic units are only slight for Ta, U, Th and Sr, and in general, Ba and Rb have higher values in the uppermost, basaltic unit.

REE analyses for the Möykkelmä mafic volcanites are presented in Table 4. The chondrite-normalized lanthanide data clearly shows a progressive decrease in LREE upwards through the stratigraphy (Fig. 26). The two lower units – andesites and basaltic andesites clearly show sloping REE patterns ((La/Sm)_N = 3.3 and 3.1, (La/Yb)_N = 7.8 and 7.1 respectively) whereas the upper, basaltic unit shows a markedly flatter distribution ((La/Sm)_N = 1.6, (La/Yb)_N = 2.5).

Low TiO₂ contents (below 0.8%) are common to all the mafic volcanites (Fig. 25) but both TiO₂ and V contents are anomalous when compared to many other incompatible elements (such as Zr and P₂O₅) in that they rise during the transition from basaltic andesites to basalts. The

Table 4. REE contents of various samples from the Möykkelmä area.

	1	2	3	4	5	6	7	8
La	48	11.2	7.3	7.9	7.4	19.3	18.4	17.9
Ce	83	20	16.5	17.6	15.7	42	30	36
Nd	40	6.9	—	11.8	9.4	16.4	15.2	23
Sm	6.3	1.23	2.1	1.70	1.87	3.7	3.6	3.4
Eu	1.74	0.48	0.97	0.55	0.71	1.08	0.92	1.00
Tb	0.80	0.15	0.37	0.40	0.38	0.75	0.38	0.57
Yb	1.59	0.57	1.11	1.01	0.87	1.79	1.56	1.45
Lu	0.22	0.05	0.12	0.13	0.10	0.16	0.22	0.19
	1 = 5605 } 2 = 5603 }	granite gneiss	3 = 5588 4 = 5569 5 = 5570 }	komatiite (lower) komatiite (upper)		6 = 5572 7 = 5575 8 = 5577 }	andesite	
	9	10	11	12	13	14	15	
La	13.9	9.7	14.9	13.8	6.0	4.9	5.3	
Ce	21	19.5	27	29	12.2	8.8	9.1	
Nd	13.8	—	15.7	16.7	7.1	—	8.2	
Sm	2.6	2.6	2.8	2.7	2.2	1.98	2.2	
Eu	0.76	0.72	0.76	0.76	0.73	0.59	0.73	
Tb	0.48	0.45	0.84	0.49	0.53	0.52	0.59	
Yb	1.3	1.17	1.26	1.23	1.54	1.42	1.46	
Lu	0.16	0.12	0.14	0.15	0.17	0.17	0.14	
	9 = 5550 10 = 5562 11 = 5558 12 = 5561 }	basaltic andesite	13 = 5546 14 = 5545 15 = 5547 }	basalt				

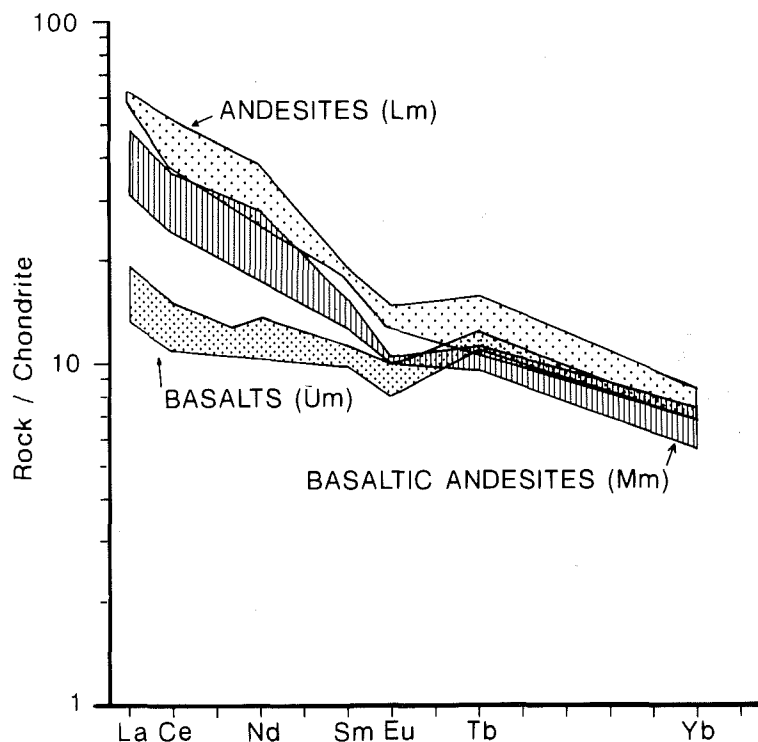


Fig. 26. Ranges of chondrite-normalized REE patterns for the andesite, basaltic andesite and basalt units.

FeO/(FeO + MgO) ratio also clearly rises between the middle and upper units, the reverse of what would be expected for an antidromic trend (Table 3), as a result of the high total Fe content (mean = 12.8%) of the basalts relative to the basaltic andesites (mean = 8.8%).

The above mentioned differences in Fe and Ti values, as well as in REE patterns militate against the basalts and the underlying basaltic andesites being the products of fractionation of the same kind of parental magma, or being different temperature melts derived from the same source material. Neither does contamination by silicic crust explain all the observed differences between the basalts and basaltic andesites – the effect on the FeO/(FeO + MgO) ratio would be in the opposite direction. Therefore it is difficult to argue for a straightforward genetic connection, even though they possess broadly similar incompatible element distributions (Fig. 27) and were presumably both erupted within a relatively short interval, since there is no evidence for intervening sedimentation or erosion.

Incompatible element patterns for the basaltic andesites and andesites are very similar except that the latter are in general higher (Fig. 27) and the behaviour of other elements is also appropriate for the two units being derived by fractional crystallization of the same parental magma. Nevertheless, contamination by silicic crust should still be considered, as indicated by high Th/Ta ratios and relatively low P/Zr- and Ti/Zr ratios (Fig. 27). Contamination has been particularly marked in the first two units, in which exceptionally low Ti/Zr ratios (mean values 28 and 40 respectively in Table 3) are associated with distinct LREE enrichment (Fig. 26). Partially assimilated leucocratic enclaves within the lower unit further attest to the significant role of contamination.

Cr-contents and FeO/(FeO + MgO) ratios for the Möykkelmä mafic volcanites are compared with corresponding data taken from the literature for a wide range of mafic and ultramafic rocks including, amongst others, komatiites, boninites, various tholeiites, alkali basalts, calc-alkali basalts, and andesites (Fig. 28). It is evident that analyses from the first and second units fall broadly within the field of the reference data, even though Cr contents of the second unit are relatively high (av. 430 ppm). In contrast, the third, uppermost unit contains anomalously high Cr and is thus situated distinctly above the reference data. The mean Cr content for the unit, at 890 ppm, is a value normally encountered in rocks with MgO between 10 – 18% and FeO/(FeO + MgO) between 0.4 – 0.5.

As Fig. 28 shows, only two examples from the available literature have Cr contents as high as those in the upper Möykkelmä unit. Rajamani *et al.* (1985) refer to an exceptionally high Cr

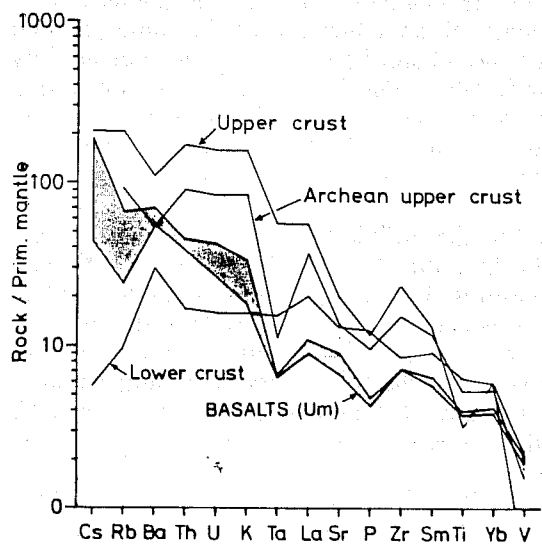
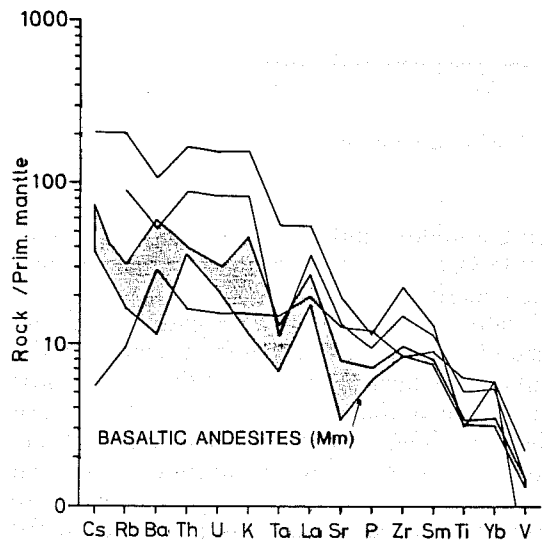
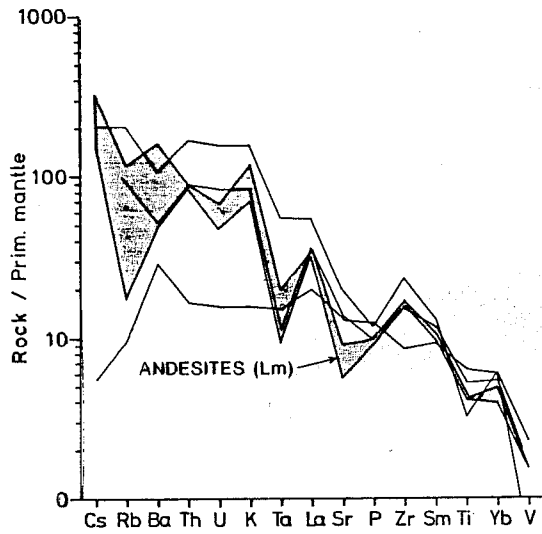


Fig. 27. Spidergrams for volcanites from the andesite, basaltic andesite and basalt units compared with those for present-day upper and lower crust and Archean upper crust (Taylor & McLennan 1981, 1985; Weaver & Tarney 1984). Normalizing factors are the same as used by Thompson *et al.* (1984).

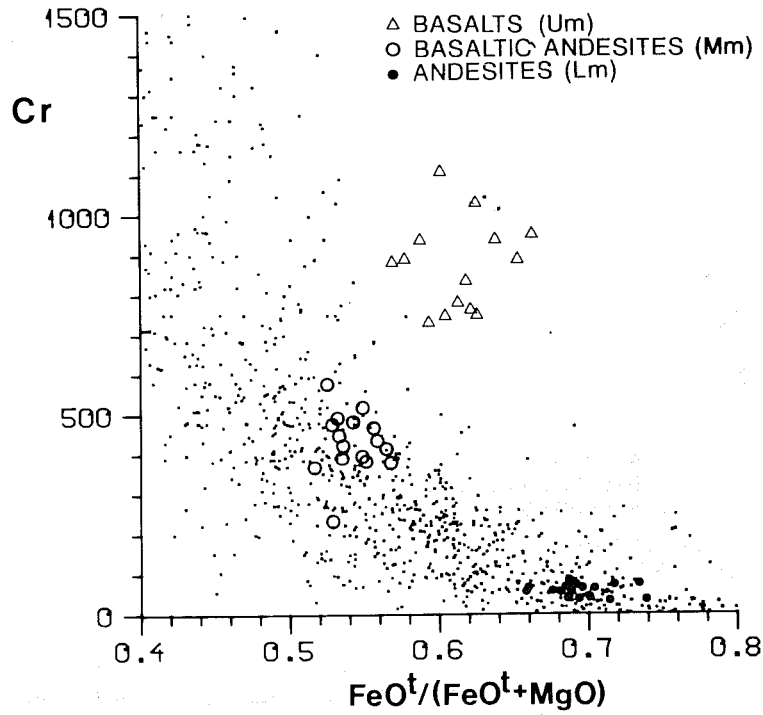


Fig. 28. Cr contents as a function of $FeO/(FeO + MgO)$ ratio for volcanites from the andesite, basaltic andesite and basalt units, compared with those of ultrabasic to andesitic volcanites (small dots) from the literature (see text).

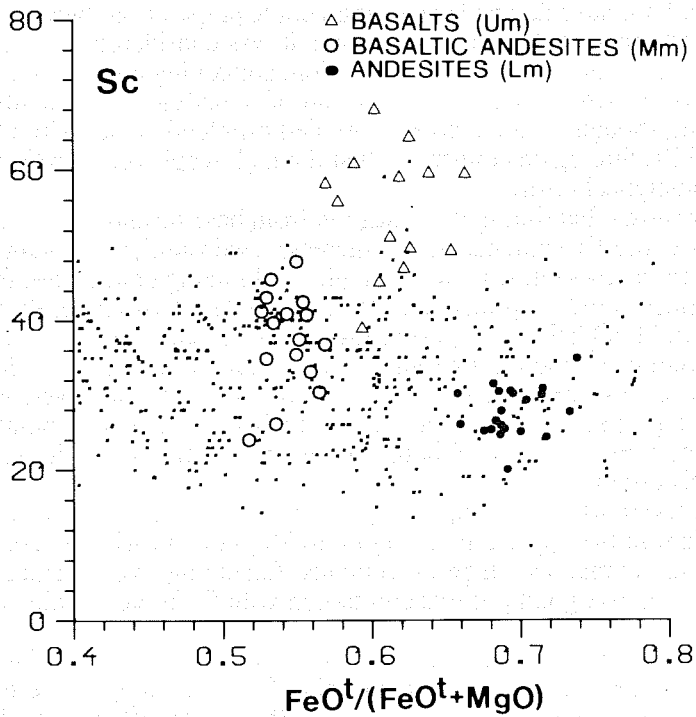


Fig. 29. Sc contents as a function of $FeO/(FeO + MgO)$ ratio for volcanites from the andesite, basaltic andesite and basalt units compared with those of ultrabasic to andesitic volcanites (small dots) from the literature (see text).

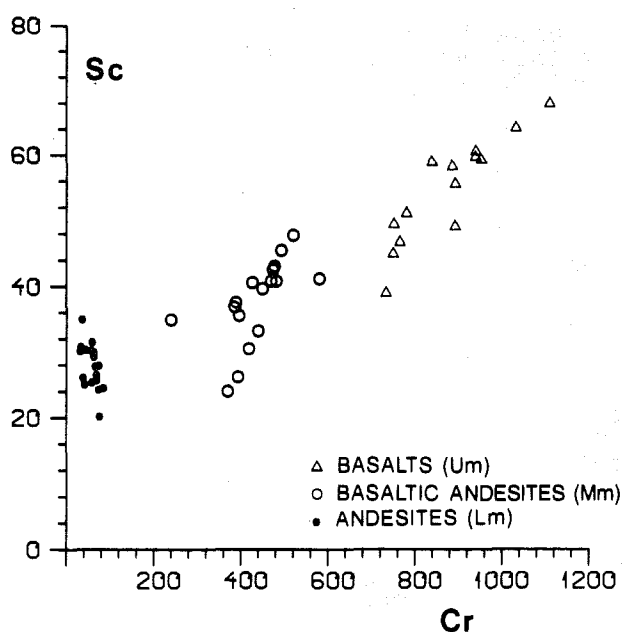


Fig. 30. Sc and Cr contents of volcanites from the andesite, basaltic andesite and basalt units.

value of 1019 ppm for a tholeiitic metabasalt from the Archean Kolar schist belt in India, but make no attempt to explain it. Nisbet *et al.* (1977) likewise reported a Cr value from basalt exceeding 1000 ppm, again without further comment. It is possible that both these examples, being isolated, could result from the presence of sporadic chromite phenocrysts (or even analytical error), but it is more difficult to invoke such a mechanism at Möykkelmä, where anomalously high Cr values occur consistently through a distinct volcanic unit many tens of meters thick.

Since similar Cr results were obtained by both XRF and INA analysis, it seems improbable that the high values can be attributed to calibration or measurement errors. The crushing equipment was tested with pure quartz samples, without adverse results, demonstrating that Cr contamination could not have taken place during sample preparation. Because chromite does not appear modally, and other oxide phases are scarce, it was considered that Cr must therefore be bound within amphibole, and hence one thin section, containing approximately 65% amphibole was chemically investigated by microprobe, a total of 6 analyses being made. The amphiboles are Mg-hornblendes, though situated close to the Fe-hornblende field, with Cr contents varying between 0.13 – 0.21%, thus again confirming that the high whole rock Cr_2O_3 values are real and not an artefact of analytical error.

Fig. 28 further shows that during the transition from basaltic andesites to basalts, both $\text{FeO}/(\text{FeO} + \text{MgO})$ ratios and Cr contents clearly increase, again militating against a simple genetic relationship between the two units. However, a physical mixing of andesites from the lower unit and Cr-rich komatiites could, in principle at least, generate a Cr-rich basalt. Calculations using mean Cr contents and $\text{FeO}/(\text{FeO} + \text{MgO})$ ratios for Möykkelmä komatiites (1700 ppm and 0.4 respectively) and andesites produce a basaltic composition falling between the basaltic andesite and basaltic fields on Fig. 28. The Cr content of this mixture, however, is too low for this mechanism to be a propable explanation for the high Cr contents of basalts. Moreover, the markedly lower total Fe contents of both end members eliminates it from further serious consideration in this instance.

The above observations concerning Cr apply also to Sc, as is evident from Fig. 29. Sc is particularly abundant in the upper unit, rising up to 70 ppm and while rocks of the middle unit do not have quite so anomalously high Sc contents. Excluding two aberrant analyses from the middle unit Sc shows a very good positive correlation with Cr in the middle and upper unit (Fig. 30).

Internal compositional variations within each of the three mafic units will now be considered. Two analyses are available for the more felsic lithology at the top of the first unit which, according to the scheme of Le Bas *et al.* (1986), are classified as andesites and dacites, having SiO_2 contents of 61.1 and 70.1%. The high MgO contents (6.8% and 5.8% respectively) are noteworthy, considering that recent or relatively young volcanites at corresponding SiO_2 levels seldom have MgO contents in excess of 3.5% or 1.5% respectively (e.g. Ewart 1979). MgO contents are also greater than those of the underlying andesites (maximum = 4.4%), while K_2O

values are substantially lower (0.06% and 0.07%). The above mentioned features, and additionally, rather low incompatible element contents in the felsic member indicate that its present composition is not a product of mixing of felsic and andesitic melts, even though drill core material record a progressive transition between a pale and a greenish pyroclastic unit. Rather, alteration, such as silicification of the upper part of the andesitic unit is deemed a more probable explanation, in which case primary felsic volcanites are in fact completely lacking from the Möykkelmä sequence.

Excluding this felsic member from consideration reveals that there is no systematic internal variation within either the first or second mafic units (Fig. 25). In contrast, rocks of the third unit show a progressive upwards decrease in Cr, Sc and Ni contents suggesting, as do converse but less pronounced increases in Al, Ti and P, that fractional crystallization was an operative process. Similar increases in incompatible elements (e.g. La, Th) should be expected, although no clear trend was observed in the present instance.

DISCUSSION

Although radiometric ages are not available for the Möykkelmä granitic gneisses, a late Archean age seems probable since the domal tectonic setting, geochemistry and geological environment are reminiscent of those at Tojottamaselkä some 20 km away, for which an age of 3.1 Ga has been obtained (Kröner *et al.* 1981). The Möykkelmä mafic and ultramafic rocks have not been dated, but they are older than the bulk of the volcanites in Central Lapland, which were erupted after deposition of the Lapponian quartzites. Likewise, at Tojottamaselkä, a sequence of mafic to intermediate volcanites occurs beneath these quartzites and a dacitic member has yielded a U-Pb age of 2.526 Ga (Pihlaja & Manninen 1988). This would indicate that the age of the Möykkelmä volcanites is close to the Archean-Proterozoic boundary.

In general, depositional basement to Archean greenstone belts is difficult to identify or demonstrate since contacts are commonly either intrusive or tectonic in nature, and intense metamorphism, anatexis and deformation have almost invariably obliterated all evidence of primary textural features. Nevertheless, eruption of greenstones onto a preexisting sialic basement has been documented from Zimbabwe and Canada (e.g. Bickle *et al.* 1975; Beakhouse 1985; Easton 1985). The granitic gneiss fragments occurring in the volcanic breccia at the base of the lower ultramafic unit are convincing evidence for eruption of the Möykkelmä komatiites through a sialic crust.

Pyroclastic units are particularly abundant at Möykkelmä with quantitatively even more than half of the komatiites being either agglomerates or lapilli tuffs. Pyroclastics are products of explosive activity in general attributed to high H₂O contents of magma in combination with high viscosity. Ultramafic magmas, however, are high temperature, low viscosity melts (Huppert & Sparks 1985b) so that explosive eruptions should be subordinate to extrusion as lava flows. However, retrograde boiling of magma in contact with ground or surface water, or with sediments can cause explosive phreatomagmatic activity. Since these are common in association with basaltic and andesitic volcanism (e.g. McBirney 1973; Schmincke 1977) it may be inferred that similar processes occurred during komatiitic eruptions.

In both ultramafic units, the more subordinate massive horizons are interpreted as lava flows, though neither display the features normally considered diagnostic of komatiite flows. The mafic lavas are predominantly amygdaloidal, which is typical for flows on land or in shallow water, whereas low vesicularity suggests their emplacement in deep water (e.g. Moore 1965, Jones 1969). Pillow lavas, which are generally regarded as indicators of a submarine environment of eruption (e.g. Dimroth *et al.* 1978) are lacking from the Möykkelmä sequence, as also are intercalated chemical, pelagic or other sediments that might point to deposition in a submarine setting. On the contrary, the generally weakly defined bedding, and gradual transitions across boundaries of tuffaceous units suggest subaerial deposition, while sharp lithological boundaries are typical for tuffs deposited and reworked in submarine conditions (e.g. Norin 1958).

No suitable chemical criteria were found to satisfactorily discriminate between a metasedimentary or plutonic origin for the Möykkelmä gneisses. The dominance of plagioclase in, and leucocratic nature of the Möykkelmä granitic gneisses is typical for the Archean tonalite-trondhjemite-granite (TTG) series, although the wide chemical variation and alternation between biotite-rich and quartz-rich units could also indicate a metasedimentary origin: overall compositions also correspond broadly to those of average graywacke. The sample population is too small to define distinct trends on, for example, Harker diagrams, in spite of major element contents (excluding alkalis) tending to decrease with increasing SiO₂. On the other hand, the

high alumina index value ($Al/Na + K + Ca/2 = 0.9 - 1.1$), presence of normative corundum ($>1\%$) and relatively low Ca and Sr are, using the classification and criteria of Chappell and White (1974), more consistent with a metasedimentary than igneous origin. If the rocks are interpreted as metasediments, the good correlation of the Möykkelmä gneisses in REE distributions with the Tojottamaselkä and Kuhmo TTG-gneisses is consistent with the model that TTG-gneisses represent a source material for the Möykkelmä gneisses. This is supported by the fact that weathering of relatively stable mineral assemblages (such as granite and granodiorite) is not known to cause selective depletion of lanthanides (Nesbitt 1979). The influence of metamorphism upon ultimate major and trace element compositions nevertheless remains speculative.

The geochemistry does not favour a simple model of geological development for the Möykkelmä gneisses. A polyphase history was indeed proposed for the Tojottamaselkä gneisses (Jahn *et al.* 1984) which have very similar REE patterns to the Möykkelmä gneisses. The Möykkelmä granitic gneisses have also been subjected to anatexis, subsequent to the gneiss forming event. Trondhjemite magma formed under near minimum melt P/T conditions crystallized in the gneisses as small scale. Biotite and to some extent plagioclase represent residual refractory restite with the former retaining much of the total rock K and Ba, while Sr remained preferentially in andesine.

The high Ba-content in the Möykkelmä gneisses, higher than for either graywackes or TTG gneisses, could be a function of later metasomatic effects, since the contact zone between the granitic gneisses and overlying volcanites at Möykkelmä appears to have been permitted extensive metasomatic fluid flow. This is evidenced by the fact that the lowermost komatiites are more extensively altered than those higher in the sequence and that some samples close to the contact show relatively high U, Mo, W and As contents.

Because komatiitic magmas are generally considered to result from high-temperature melting, many elemental ratios in the melt may closely represent corresponding compositions in the original source material, an argument which has been widely used in deducing mantle composition during the early history of the earth. Recent theoretical studies have indicated, however, that a hot komatiitic melt of low viscosity will experience turbulent flow, efficiently transferring heat into adjacent rock units. This may result in up to 30 - 40% contamination by sialic material during the ascent of the magma through crust of granitic composition (Huppert & Sparks 1985a). Thermal erosion during eruption and extrusion of lava can also cause assimilation of less refractory country rocks up to 10%. Hence, attempts to infer mantle compositions based solely upon data from komatiites should consider the possible influence of sialic contamination (cf. Arndt 1986a). Crustal contamination may also lead to erroneous interpretation of isotopic age determinations (see Chauvel *et al.* 1985).

Convincing arguments for contamination by older crust have been presented for komatiites and basalts at Kambalda and Negri in Western Australia (Barley 1986; Arndt & Jenner 1986) as well as from Newton Township komatiites in Canada (Cattell 1987; Cattell & Arndt 1987). The Möykkelmä komatiites and tholeiites also furnish evidence for crustal contamination and assimilation having influenced the composition of komatiitic magmas. Even though the rocks in question have clearly also experienced intense secondary alteration, primary contamination is not thereby excluded, since most immobile incompatible element relations deviate significantly from chondritic values.

Komatiites of the Möykkelmä type occur in the Kolari district in western Lapland, where they also occupy a similar setting, directly overlying felsic gneisses (Väänänen 1989, in press). On the contrary, the Möykkelmä komatiites differ substantially from the stratigraphically higher Lapponian Sattasvaara komatiites near Sodankylä, and also the Kummitsoiva and Kiimasselkä komatiites in the vicinity of Savukoski (Juopperi & Räsänen, in prep.; Lehtonen *et al.* 1986; Manninen *et al.* 1986; Peltoniemi 1984; Saverikko 1983, 1985). The latter resemble the Al-depleted type defined by Nesbitt *et al.* (1979) whereas the Möykkelmä rocks are Al-undepleted types. Further contrasts with respect to the Möykkelmä komatiites include LREE depletion in the Sattasvaara komatiites, and absence of a distinct Ta anomaly, while their Ti/Zr ratio is close to that for chondrites.

Chemical studies of mafic and ultramafic units with Archean greenstone belts have revealed that more than one conclusion can be drawn concerning the relationship between komatiites and spatially associated subalkalic basalts (Arndt *et al.* 1977; Sun & Nesbitt 1978; Nesbitt *et al.* 1984). At Möykkelmä, the mafic and ultramafic volcanites form distinct units both chemically and stratigraphically, unlike at Kolari, where a continuous progression from komatiites to basalts has been documented, accompanied by a steady decrease in MgO down to a level of 5% (Väänänen 1989, in press). These basalts are relatively Cr-rich and TiO₂-poor (0.60 - 0.70%);

the latter feature corresponds well with the upper mafic volcanite at Möykkelmä but on the MgO vs Cr diagram, Möykkelmä basalt lie clearly above the Kolari trend. On the other hand, the basaltic andesites of the middle Möykkelmä unit extrapolate well onto the Kolari trend, although their TiO₂ and FeO total contents are far too small for them to represent differentiation from a contaminated komatiitic magma. Therefore it is not possible to invoke a direct genetic connection between the komatiites and associated mafic rocks at Möykkelmä.

Low Ta/La ratios and high Zr/Ti ratios indicate that upper crustal contamination was not restricted to the komatiites but has also affected the final composition of Möykkelmä basaltic melts. Since SiO₂ contents, the LREE/HREE ratio and negative Ta-anomaly all decrease upwards through the sequence, the importance of this contamination would seem to have decreased with time. Such trends have been documented from continental basalts elsewhere and has been explained by proposing that basaltic magma becomes progressively isolated from adjacent sialic country rock by crystallization of melt against the margins and walls of magma conduits and chambers (Dostal *et al.* 1986; Jolly 1987). Contamination alone does not adequately explain the differences in compatible trace element compositions between the basalts and basaltic andesites, in which case the andesites must represent the product of considerable more evolved fractional crystallization.

One feature common to all the Möykkelmä mafic volcanites is their low TiO₂ contents. Although assimilation of sialic crust can indeed reduce TiO₂ values, average Archean crustal TiO₂ compositions of 0.6 – 0.8% TiO₂ (Weaver & Tarney 1984; Taylor & McLennan 1985), being almost the same as that in the Möykkelmä basalts, would seem inappropriate as a contaminant in this case. However, the basement gneisses analyzed here (taken from near the basement-cover contact) have TiO₂ contents between 0.11 – 0.44%, markedly lower than the above mentioned mean values. Unfortunately, relevant data from samples at the core of the gneiss window, away from the contact, are not available, but similar tonalitic gneisses at Tojottamaselkä have TiO₂ contents in the range 0.44 – 0.58% (Kröner *et al.* 1981). Even were the TiO₂ values of the assimilated material as low as the Möykkelmä analyses suggest, the primary TiO₂ content of the basalts and basaltic andesites could scarcely have exceeded 1%, since the high Cr contents impose constraints upon the degree of contamination.

Since neither contamination nor subsequent alteration can adequately account for the low TiO₂ in the basalts, the ultimate source magma presumably contained only a small amount of TiO₂. This could in turn be attributed to low Ti contents in the source region or that during melting a residual Ti-rich phase remained, such as rutile, ilmenite or sphene, in which Ti and other high-field-strength cations were preferentially retained (Jolly 1987). According to recent experimental data, the solubility of Ti in mafic melts at both high and low pressures should result in TiO₂ values considerably in excess of those observed at Möykkelmä (cf. Rutherford *et al.* 1980; Green & Pearson 1986; Medvedev *et al.* 1987; Ryerson & Watson 1987). Hence, the present TiO₂ contents would seem to be a function of low source contents, rather than retention in a residual phase. The most simple explanation for incompatible element depletion of mantle material is to have extracted them preferentially in connection with some previous melting event. It may be noted that enrichment of volcanites in LIL elements and LREE is not at variance with derivation from a depleted source since such enrichment can be explained by either sialic contamination, or mantle metasomatism caused by circulation of H₂O-rich fluids. Many investigations of mantle-derived xenoliths have demonstrated, on the basis of major element and isotopic characteristics, that metasomatically altered, LIL- and LREE-enriched peridotites represent refractory residues depleted in basaltic constituents during an earlier melting episode (e.g. Erlank *et al.* 1987). Owing to the altered nature of the rocks at Möykkelmä, it is not possible to establish whether the high LIL element contents of both komatiitic and basaltic rocks is a result of mantle metasomatism, in addition to contamination and subsequent alteration processes.

Low amounts of Ti and other high-field strength elements (Ta, Nb, Zr, Hf) and high LIL element contents are characteristic features of volcanic rocks from active continental margins and oceanic island arcs (Pearce 1982). Many discrimination schemes are based on ratios of elements from these two groups including high Ba/Ta (>450) which, according to Gill (1981), is the single most diagnostic geochemical feature of arc magmas. Many analyses of the Möykkelmä basic and intermediate volcanics fulfill this and other criteria for island arc magmas. However, the lack of dacitic and rhyolitic derivatives among the Möykkelmä volcanics and their subaerial eruption unconformably on the Archean granitic gneiss basement preferably suggest a continental environment. The same problem was also encountered by Jolly (1987) when studying Huronian bimodal volcanism in central Ontario where low-Ti tholeiites and calc-alkaline rhyolites, dated about 2450 Ma, are associated with quartzose metasediments deposited

on an Archean craton. Jolly explains the island arc characteristics of the basic volcanism by proposing subcontinental mantle metasomatism during a previous, Archean subduction. Another analogous case is reported by Crow and Condie (1987) from western Orange Free State and southwestern Transvaal (South Africa), where late Archean (2.8 Ga) quartzites and metavolcanic rocks of the Dominion Group rest unconformably on the granitic basement of the Kaapvaal Craton. The 1100 m thick volcanic sequence is composed of calc-alkaline andesites, basaltic andesites and minor felsic porphyries. Crow and Condie (1987) also consider the possibility that the island arc geochemical signature of the Dominion group is inherited from older mantle lithosphere retaining such a component from an earlier subduction regime. Alternatively, the succession may have formed adjacent to a continental-margin arc system in an incipient foreland basin on the Kaapvaal Craton.

A third, much younger example of low-Ti basalts not directly associated with any subduction zone, is provided by the Kirkpatrick Basalt in Antarctica (Siders & Elliot 1985). This Mesozoic continental flood basalt sequence containing both low-Ti and high-Ti flows is related to the breakup of Gondwana. The TiO_2 content (0.59 – 0.82%) of the low-Ti basalts is unusually low for continental flood basalts and the incompatible element ratios such as Ta/La, Ta/Th, P/Zr, Ti/Zr are low, indicating a distinct crustal signature.

The high Cr content of the basalts deserves further consideration. According to recent experimental investigations, the solubility of chromite in silicate melts is dependent upon melt composition, temperature and oxygen fugacity (Murck & Campbell 1986; Barnes 1986). Under the QFM-buffer conditions, a melt compositionally equivalent to the Möykkelmä basalts would have a chromite solubility threshold corresponding to 400 ppm Cr in the melt, approximately half of the amount actually measured at Möykkelmä. This indicates that chromite should have been a liquidus phase but because the basalts also exhibit fractionated Fe/Mg ratios, concomitant fractionation of chromite would be expected to lower the Cr content of the melt to considerably less than the presently observed values. It is possible that the high Cr contents correlate with an abundance of chromite phenocrysts, but this is difficult to verify since the primary magmatic mineral assemblage has not been preserved. However, an original absence of chromite phenocrysts is indicated by the strongly positive correlation between Sc and Cr (Fig. 30), since Sc is not readily incorporated into the spinel lattice (Leeman 1976; Nagasawa *et al.* 1980). The solubility of chromite increases when oxygen fugacity decreases but in order that the solubility limit approach 1000 ppm, O_2 fugacity should fall to a level approximately midway between the QFM and IW buffers (calibrated logarithmically). Under these conditions a significant amount of Eu would be divalent (Drake 1975), so that separation of Eu from other REE might be anticipated, with ensuing anomalies showing in REE patterns. No distinct Eu-anomaly has, however, been observed.

A positive correlation between Cr and Sc is qualitatively consistent with the enrichment of basalt in clinopyroxene phenocrysts since both Sc and Cr have distribution coefficients significantly in excess of unity, when considering partition between clinopyroxene and a mafic silicate melt (e.g. Irving & Frey 1984). Nevertheless, the proportion of phenocrysts required would be high, being many tens of per cent, in which case the Cr content would accordingly rise from the typical amount of less than 400 ppm to the observed values (Fig. 29). This should also be reflected in major element geochemistry since the abundance of pyroxene phenocrysts would mean higher CaO; such however has not been observed and in fact the basalts have on average less CaO than the relatively Cr-poor basaltic andesites. This phenomenon of lower than expected CaO in the basalts could perhaps be explained by secondary alteration. However, it is known from studies of Na-metasomatism in differentiated intrusions that, in contrast to plagioclase dominated rocks, pyroxene-rich lithologies retain the bulk of their CaO during alteration to Ca-amphibole (Hanski 1984). Since petrographical evidence for abundant pyroxene phenocrysts or their pseudomorphs is also lacking, the high Cr-content of the Möykkelmä basalts remains thus problematical.

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