Geology and geochemistry of mafic-ultramafic rocks (Köli) in the Handöl area, central Scandinavian Caledonides

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The Rödberget ultramafic-mafic Complex in the Handöl area in Jämtland, central Swedish Caledonides, consists of various ultramafic rocks, metacumulate rocks, metagabbros, mafic dykes, plagiogranites and a melange. It is interpreted as a fragmented ophiolite. The overlying Bunnran Formation is a metasedimentary-meta-igneous unit usually consisting of finely laminated calcareous psammitic and pelitic rocks with probable tuffitic components, intruded by and interlayered with mafic intrusive and extrusive rocks. Combined geological and geochemical evidence indicates a close spatial and temporal genetic relationship between the Rödberget ultramafic-mafic Complex and the Bunnran Formation. It is suggested that they originated in a basin where oceanic crust was formed during deposition of material derived from continental and volcanic source regions, an environment comparable to the Andaman Sea in the Indian Ocean. As a general model, such basins with their high frequency of fracture zones/transform faults could be the original settings of the solitary and detrital ultramafic rocks also in correlateable units in the central Scandinavian Caledonides.

A wide variety of igneous and sedimentary settings are represented in the allochthonous units in the Scandinavian Caledonides. Rocks formed in various early Palaeozoic oceanic environments are found in the eugeoclinal terranes (Köli Nappes) which structurally overlie the late Precambrian–early Palaeozoic continental margin of Baltica (Seve Nappes). In recent tectonic models of the central Scandinavian Caledonides (Stephens & Gee 1985; Stephens 1988; Stephens & Gee 1989), three eugeoclinal terranes have been distinguished: the Virisen, Gjersvik and Hølonda terranes, which correspond tectonostratigraphically to the earlier defined (Stephens 1980) Lower, Middle and Upper Köli Nappes, respectively. The metasedimentary rocks in the Virisen terrane record evidence for proximity to the Baltoscandian margin from the Middle Ordovician; the Gjersvik and Hølonda terranes have been inferred to have accreted to Laurentia during the Early Ordovician (Stephens & Gee 1989) prior to the closure of Iapetus. The Köli unit of the present study area was initially included in the Virisen terrane (Stephens & Gee 1985), but recently (Stephens & Gee 1989) it was assigned to a separate isolated terrane (metamorphic complex of less certain affinity).

The brief presentations of the evidence for an ophiolite along the boundary between the Köli and Seve Nappes at Handöl, central Scandinavian Caledonides (Gee & Sjöström 1984; Sjöström 1986), promoted this study and preliminary field results have been published previously (Bergman 1987). This paper presents geological and geochemical data from the Rödberget ultramafic-mafic Complex and the Bunnran Formation and discusses models for their original setting and subsequent evolution in terms of present and Caledonian environments.

Field relations and petrography

Previous work

Brief descriptions of the geology in the Handöl area (Fig. 1) are included in publications based on early regional work in the central Caledonides (Högboom 1894; Törnebohm 1896; Frödin 1922). More recent contributions are based on detailed studies of the Handöl–Storlien (Sjöström 1983, 1986) and Tännforsen areas (Beckholmen 1984). A small number of chemical analyses of rocks from the study area have been presented in regional studies of the ultramafic rocks in the Scandinavian Caledonides (DuRietz 1935; Stigh 1979). In addition, some petrographic, textural and geochemical work in the area has previously been carried out (Hardenby 1974; van der Kamp 1974; Westlund 1983).

The geology of the study area is shown in Fig. 1. The Seve Nappes in the area have been subdivided into three parts (Sjöström 1983): the lowermost Blåhammarfjället Nappe (greenschist–amphibolite facies) consisting of amphibolite, mica schist and calc-silicate rocks over lain by the paragneiss-dominated Snasahögarna Nappe (lower granulite facies), in turn overlain by the Täljstensvalen Complex composed of imbricated thrust slices dominated by mica schist and amphibolite metamorphosed in amphibolite facies.
Fig. 1. A. Location of map B in Scandinavia. T = Trondheim, O = Östersund. B. Map of the Tannforsen area and surroundings, simplified after Gee & Sturt (1985) and Beckholmen (1984). C. Geological map of the study area with sample locations. Most of the boundaries from the western part of the area are from Sjostrom (1983).
The Köl units (upper greenschist–lower amphibolite facies) tectonically overlying the Seve in the southern part of the Tännforsen region (Tännforsfället; Törnbohm 1896) have previously been described as the Bunnerviken lens (Beckholmen 1984) and its western part as the Handol Formation (Sjöström 1983). The subdivision presented here differs slightly regarding the Bunnerviken lens as defined by Beckholmen (1984, see Fig. 2). Based on field relationships, two units are distinguished: the Rödberget ultramafic-mafic Complex overlain by the Bunnran Formation. A schematic composite section (Fig. 2) shows the relations between the different compo-

ments in these two units. There are few exposures of the contact between the Rödberget ultramafic-mafic Complex and the Bunnran Formation and its nature is not obvious. However, in the Bunnerviken quarry (Fig. 1) the impression is a primary transition, from the calcareous melange in the former to the calcareous metasedimentary rocks of the latter, rather than a tectonic contact.

Rödberget ultramafic-mafic Complex

The Rödberget ultramafic-mafic Complex consists of tectonic lenses of mainly ultramafic-mafic rocks dispersed over a distance of ca. 20 km along the tectonic boundary between the Köl and Seve units. Many lenses have an entirely ultramafic composition but several (at least six) are composite, displaying two or more rock types. The exposed components are ultramafic rocks, various cumulate rocks, gabbros, mafic dykes and plagiogranites (Fig. 2), mostly containing metamorphic mineral assemblages. A lens containing all these components is exposed at Mount Rödberget and River Järpån (Fig. 1). A melange (see definition in Raymond 1984) contains diverse fragments (see below) in a dominantly ultramafic matrix. The ultramafic rocks of the present area are found at two structural levels. They are separated by a zone of strongly sheared amphibolite, mica schist and more or less calcareous Cr-rich quartzite (with Cr-mica and locally uvarovite and wollastonite). The latter is an excellent marker horizon throughout most of the area.

Lower ultramafic level. – This level is dominated by foliated, more or less serpentinized dunite in well-exposed tectonic lenses (maximum recorded thickness is ca. 250 m) bounded by strongly deformed serpentinite. It is underlain by mylonitized mica schist, amphibolite and psammitic of the Täljstärvallen Complex (Seve). Primary features within the better preserved dunitic bodies have been recorded locally (Fig. 3A, B). These include compositional- and grain-size layering with alternating orthopyroxene-bearing porphyritic dunite and fine-grained dunite and rare mm-thick chromite trails. Irregular 0.3–1 m thick pods of fine-grained dunite have gradational contacts to a more coarse-grained variety. A grain-shape foliation defined by coarse olivines or pyroxenes is developed in both rock types. Gabbroic rocks of the cumulate section have been found in tectonic contact with serpentinized dunite in several places indicating structural breaks in the lithological column.

In thin section, large olivines with undulose extinction have recrystallized to small strain-free crystals with straight grain boundaries meeting in triple points. Several generations of serpentine and late talc veins replace olivine. Reddish brown chrome spinel is partly altered to magnetite and is associated with secondary Al-rich chlorite. Relict orthopyroxene is found in some samples.
Fig. 3. A. Layering in slightly serpentinized dunite. The hammer is 54 cm long. 400 m southeast of the Bunnerviken quarry. B. Fine-grained dunite pod (upper part) with gradational contact to porphyritic orthopyroxene-bearing dunite. The foliation defined by oriented grains and aggregates is interpreted as primary (see text). The pen is 16 cm long. 400 m southeast of the Bunnerviken quarry. C. Grain-size layering in meta-cumulate rock with actinolite pseudomorphs after pyroxene. The pen is 14 cm long. Mount Rödberget. D. Layered metagabbro with mafic dyke transecting the layering. The hammer is 54 cm long. Mount Rödberget. E. Multiple sheets of metadolerite (shaded) and plagiogranite (crosses). The contacts and fine-grained margins (dotted) are clarified on the line drawing. Loose debris is white. The hammer is 54 cm long. River Järpdån. F. The melange: fragments of ultramafic to felsic igneous rocks, jasper, quartzite, quartz, schist, marble and manganiferous rock in a dominantly ultramafic matrix. The hammer-head is 15 cm long. Bunnerviken quarry. G. Psammite beds and laminae with sharp bases grading upwards into psammitic. Note incipient formation of compositional layering transecting bedding due to metamorphic differentiation along crenulation cleavage. The pen is 14 cm long. River Handölan. H. Amygdaloidal metabasalt overlain by graded (right-way up) metasedimentary sequence. The pen is 14 cm long. River Handölan. I. Psammite with ripple-drift cross-lamination. The pen is 14 cm long. River Handölan. J. Metaconglomerate with clasts of psammite, pelite, marble, granite and Cr-bearing mica in a quartz-white mica-calcite matrix. The pen is 14 cm long. River Handölan. K. Intrusions of metagabbro and metadolerite, now concordant and lens-shaped, in laminated pelite and marble to the left and psammite and pelite in the centre. Note thin metadolerite intrusion within metagabbro in upper right corner (see arrows). The thickest part of the dark metagabbro lens is ca. 3 m. Tjufloarma quarry.
Mafic-ultramafic rocks, Handö l
Upper ultramafic level. - This consists of up to 50 m thick strongly deformed tale schist with antigorite, chlorite, carbonate, tremolite and magnetite in addition to talc as principal constituents. Small (dm-sized) retrogressed mafic pods interpreted as disrupted dykes and more coherent but isoclinically folded retrogressed amphibolite dykes have been recorded. In close association with the tale schist there is a ca. 10 m thick melange (see below). A tectonic lens of massive or slightly foliated serpentine is situated within the cumulate rocks. The serpentine has an entirely metamorphic mineral assemblage consisting of antigorite, magnetite, talc and carbonate but pseudomorphs after olivine are outlined by a network of fine-grained magnetite.

Cumulate rocks and amphibolites. - The cumulate section is best exposed in the easternmost part of the area (Mount Rödberget and River Järpån, Fig. 2). It is at least 400 m thick (the upper contact is not exposed) and comprises layered metapyroxenite, metagabbro and meta-anorthosite, massive leucogabbro and Fe–Ti-amphibolite.

In the layered cumulate rocks, layers range in thickness from less than a centimetre to several decimetres and can be followed along strike for a few tens of metres. The layering in the lower parts is defined by size variations of crystal aggregates that consist mainly of actinolite and minor antigorite and carbonate (Fig. 3C). Such aggregates often have rectangular outlines up to 5 × 3 cm in size. They are interpreted as pseudomorphs after pyroxene (uralite). A few crystals of relict clinopyroxene have also been found locally.

The massive leucogabbro and the layered metagabbro (Fig. 3D) consist of varying proportions of actinolite and more or less altered plagioclase (labradorite). The degree of alteration is extremely variable over short distances. Some meta-anorthosite layers consist almost entirely of plagioclase, while other leucocratic layers are almost completely saussuritized. Pockets of pegmatitic gabbro have been recorded within the leucogabbro, which is most abundant in the upper part of the cumulate section. The upper contact of this leucogabbro is intrusive, as demonstrated by its fine-grained amphibolite xenoliths.

The amphibolites overlying the layered cumulates are inconspicuous in the field but display a distinctive chemistry. They are depleted in most incompatible elements and show signs of accumulation of hornblende/Fe–Ti-oxides (see below). These rocks are mineralogically similar to the mafic dykes although their abundance of opaque phases is higher. Sheets of leucogabbro are present within this unit.

Plagiogranite. - The felsic rocks (plagiogranite) are found within the layered cumulates and most commonly in the mafic dykes and the Fe–Ti-amphibolites as cross-cutting dykes (Figs. 2 and 3E) (0.1–0.4 m wide) or lenses or sheets with unclear contact relations. Their fine-grained groundmass consists of quartz, plagioclase, hornblende, chlorite and opaque minerals. Well-preserved zoned euhedral plagioclase phenocrysts are common.

Mafic dykes. - Mafic dykes have intruded the layered cumulate rocks (Fig. 3D), often at a high angle to the layering, and the Fe–Ti-amphibolites and the plagiograniites. Fine-grained margins are common and dyke-intruding-dyke relationships (Fig. 3E) have been found at several places, although never in sufficient numbers to be recognized as a sheeted dyke complex. Dykes are present in all of the composite ultramafic-mafic lenses. Porphyritic, sub-ophitic and massive fine-grained varieties have been recorded and their thicknesses vary between 0.1 and 1 m. Mineralogically the mafic dykes are amphibolites consisting of hornblende, plagioclase (oligoclase) and epidote with minor amounts of opaque minerals, biotite, chlorite and Ti-oxides. Apatite and actinolite are rare. Secondary calcite has been observed in some samples. Variably altered plagioclase phenocrysts are the only primary magmatic phases present.

Melange. - The melange (Raymond 1984) contains fragments of various rock types in a dominantly ultramafic matrix (Fig. 3F). At the type locality in the Bunnerviken quarry its thickness is ca. 10 m. It is matrix-supported and the strongly foliated matrix is dominated by antigorite and talc but chlorite- and carbonate-rich zones are also present in the upper part. Local concentration of carbonate in patches and irregular veins provide evidence for secondary carbonate introduction. The fragments range in size from a few millimetres to several decimetres (the largest observed is 0.5 × 0.3 m). Their aspect ratios vary from 1:1 to less than 1:10, in part depending on their competence relative to the matrix. Some are angular but most are rounded and oblate in shape with both long and intermediate axes subparallel to the matrix foliation. At one locality (north of Lake Bunnersjöarna), the foliation in an angular serpentinite fragment is at a high angle to the matrix foliation, indicating the possibility of deformation prior to formation of the melange. A strongly retrogressed and boudinaged mafic dyke has been found in the melange.

Many rock types are represented by fragments: serpentinite, white and green tale schist, metagabbro, amphibolite, felsic plutonic rock, biotite schist, quartzite, vein quartz, marble and manganiferous rock. Rocks of ultramafic composition dominate the fragment assemblage. Serpentinite fragments are dark green or black and fine-to medium-grained. Apart from some chromites rimmed by magnetite, no relict igneous phases have been found. Metagabbroic fragments often have a well-preserved igneous texture although the primary mineralogy is totally replaced. Many of the quartzitic fragments are red and fine-grained, suggesting that they originated as jaspers. Grey and white varieties have also been recorded. The marble fragments consist of either white calcite or, more rarely, pink rhodochrosite. This Mn-rich rock is found as massive ellipsoidal aggregates, disrupted layers or as rims around white carbonate fragments. An important exotic felsic plutonic fragment consists of 1–2 mm large rounded bluish quartz, altered feldspar, white
mica and opaque minerals. Discontinuous layers, ca. 1 cm thick, composed of biotite and chlorite may represent pelitic fragments. A fragment of uncertain origin consists of abundant carbonate and opaque phases with minor talc and serpentineite and well-shaped plagioclase porphyroblasts.

Bunnran Formation

Overlying the Rödberget ultramafic-mafic Complex is the Bunnran Formation, a heterogeneous metasedimentary and meta-igneous unit. Most common are calcareous psammitic and pelitic rocks with some marble and intercalations of mafic metavolcanic rocks. Intrusions of metagabbro and metadolerite are abundant. Multiple deformation and metamorphic mineral growth in upper greenschist–lower amphibolite facies have obscured most primary structures. However, the heterogeneous nature of the deformation has led to local preservation of primary sedimentary and igneous structures and textures. Younging directions, mostly recorded from graded beds, are indicated in Fig. 1. It has not been possible to establish a meaningful lithostratigraphy because of internal shear zones and isoclinal folds, the lack of persistent marker horizons and limited exposure. Similar lithological units are commonly repeated throughout the entire exposed thickness of the Bunnran Formation (Fig. 2), which is at least 1000 m. The relative importance of tectonic repetition versus cyclicity related to the depositional environment is unknown, but both are probably significant.

Metasedimentary rocks. – A characteristic feature of the metasedimentary rocks is the compositional layering (Fig. 3G–I, K) with individual layers of widely varying thickness, but commonly 0.5–5 cm. The layering can in places be shown to represent sedimentary bedding or lamination by associated sedimentary structures (Fig. 3G–I, but elsewhere an origin by metamorphic differentiation during formation of crenulation cleavage can be demonstrated (Fig. 3G). The rocks consist of varying proportions of quartz, mica, plagioclase and calcite, with the addition of hornblende, garnet, chlorite or clinozoisite in particular layers. Scattered grains of green Cr-bearing mica in quartz-rich layers have been found in several places. These are probably reaction products after detrital chromite.

Compositionally graded beds, generally 2–5 cm thick, are characterized by a quartz-rich base and a continuously increasing amount of mica towards the top of the bed (Fig. 3G, H). Rare cross-lamination (ripple drift) is shown by asymmetric lenses of quartz-rich metasedimentary material enveloped by mm-thick mica-rich laminae (Fig. 3I). Laminations probably disturbed by soft sediment deformation have also been noted.

Calcite is a common matrix constituent, especially in psammitic rocks. Thin carbonate layers are frequent in the finely laminated metasedimentary rocks and a calcite marble unit close to the base of the Bunnran Formation at River Handölän is several metres thick. Although the marble is intensely deformed, recognizable but poorly preserved Pelmatozoan stem segments have been observed here. These were discovered by A. G. Högblom and mentioned in Strand & Kulling (1972) but have escaped attention elsewhere in the literature, since they are of limited use as age or environmental indicators.

In the stream bed of Handölän, there is a matrix-supported polymict metaconglomerate (Fig. 3J, cf. Fig. 2). It is ca. 6 m thick, laterally impersistent and is in contact with pelitic and mafic igneous rocks. The matrix consists of quartz, white mica and calcite. The more or less strongly flattened clasts are composed of psammites-pelites, brown weathering marble, granite and abundant aggregates (≤0.5 cm) of green Cr-bearing mica (0.5% Cr₂O₃). The clast population is critical for the interpretation of the provenance of the metasedimentary rocks in the Bunnran Formation and is further discussed below.

At a few localities, knots (≤1.5 cm) of quartz, calcite or quartz-cored calcite are present in a quartz-white mica-dominated matrix. Hexagonal outlines of quartz knots have been observed. The origin of this rock is unknown.

Euhedral plagioclase crystals embedded in a fine-grained quartz-mica matrix are not uncommon and are regarded as igneous in origin. Alternating thin layers of quartzo-feldspathic metasedimentary material and amphibolite have been interpreted as tuffite (Sjöström 1983).

Meta-igneous rocks. – Mafic igneous rocks (metagabbros, metadolerites and metabasalts) make up a significant proportion (ca. 30–50%) of the rocks in the Bunnran Formation. In the field, they stand as topographic highs, while the surrounding metasedimentary rocks are mostly covered with Quaternary deposits. The mafic rocks are generally concordant with the layering in the metasedimentary rocks and have tectonic lensoidal shapes (Fig. 3K) up to 130 m thick. They are composed of blue-green hornblende, plagioclase, oxides, ± epidote, ± biotite, ± chlorite, ± carbonate. Some mafic rocks carry garnet. Plagioclase is the only primary magmatic phase recorded, and is locally very abundant as euhedral–subhedral phenocrysts. Relict clinopyroxene was reported by DuRietz (1935). Selective growth of metamorphic hornblende and garnet in pelitic metasedimentary rocks along the contact to the mafic rocks is not uncommon. The microstructure of these porphyroblasts is similar to undoubted regional metamorphic porphyroblasts further away from mafic bodies. This implies that the contact phenomena are mainly due to material transfer by magma-sediment interaction rather than contact metamorphism.

The concordant relationships and locally preserved chilled margins suggest that the medium-grained metagabbroic bodies intruded as sills in the metasedi-
Discussion of field relations and petrography

The transition from calcareous melange in the Rödberget ultramafic-mafic Complex to the calcareous metasedimentary rocks of the Bunnran Formation and the presence of metadolerite dykes in both units suggest that they have a close spatial genetic relationship. The presence of felsic plutonic and pelitic detrital material together with the Cr-rich material in both units supports such an interpretation.

There is no general agreement in previous work concerning the exact location of the boundary between the Seve and the Kölí rocks in the area. In this study, the spatial association of dunitic ultramafic rock and cumulate metagabbro (Fig. 2), and the presence of cumulate metagabbro at the same structural level as the talc-schist (Kölí, Fig. 2), are considered to be good reasons for including all ultramafic rocks in the Kölí Nappes. This is also consistent with textural and metamorphic data from associated pelitic rocks (Bergman 1992).

It is suggested that the association of ultramafic rocks, layered cumulate rocks, gabbros, plagiogranite, mafic dykes and melange in the Handöl area represents a dismembered ophiolite, in agreement with Gee & Sjöström (1984). The two basal layers of an ophiolite as defined by the Penrose Conference (Conference Participants 1972) appear to be represented in the area. An extensive mafic sheeted dyke complex (100% dykes) and mafic volcanic rocks in contact with other ophiolitic components have not been found, however. This may be due to erosion, poor exposure and/or high strain obliterating primary features (see also below). An alternative model for the origin of the Rödberget ultramafic-mafic Complex as a layered intrusion is considered less likely since the dunitic rocks have characteristics (foliation, layering, fine-grained dunite pods) which resemble those in the transition zone between the metamorphic peridotite and the overlying cumulates in ophiolites (Coleman 1977). Although this transition zone has commonly been interpreted as cumulate in origin (e.g. Coleman 1977), evidence for a dominantly residual origin by partial melting of orthopyroxene out of harzburgite has been presented (Nicolas & Prinzhofer 1983).

The foliation in the better preserved dunitic rocks may have originated in the mantle. Support for this is given by van der Kamp (1974), who reported slip along \{0kt\}[100] in the older large olivines at Handöl indicating upper mantle, high-temperature deformation (Rayleigh 1968; Carter & Ave'Lallement 1970). This also argues against an origin as a layered intrusion for the Rödberget ultramafic-mafic Complex.

The layered cumulate rocks may have formed by accumulation of Fe-Mg-silicates and plagioclase. Repeated supply of magma is indicated by the dykes transecting the layering. The cumulate rocks seem to be in a normal upward facing position in view of the upwards increasing leucocratic component and typical high-level massive gabbro, pegmatitic gabbro and plagiogranite in the upper parts.

Melanges are widespread in the modern and ancient geological record and may be formed in a variety of environments by tectonic or sedimentary processes (e.g. olistostromes, serpentinite diapirism, tectonic disruption) or they may be polygenetic (Raymond 1984). In general, distinctive criteria indicating their origin are difficult to establish and later tectonic overprinting, such as in the Handöl area, further complicates interpretation. Apparently the igneous fragments in the melange come from various levels of the ophiolite. The seemingly transitional nature of the contact between the melange and the Bunnran Formation may indicate a sedimentary origin for the melange.

Mn-rich rocks similar to those found as clasts in the melange may be formed in a variety of settings. Mn-rich nodules and coatings are well known from marine and lacustrine environments where sedimentation rates are very low (0.5–5 mm/year). Mn-mineralizations have also been recognized in aragonite veins deposited from cold sub-seafloor solutions, in ultramafic rocks from the Romanché Fracture Zone in the Atlantic Ocean (Bonatti et al. 1980) and Zabargad Island in the Red Sea (Bonatti et al. 1983). Manganese carbonate hands (Sugisaki et al. 1991) concentrated at specific horizons associated with chert or shale have been suggested as indicative of a hemipelagic-neritic environment in isolated basins such as fore-arc and back-arc basins or rifted continental margins with narrow basins. Since any of these modes of formation may be applicable to the Mn-rich fragments observed in the Handöl area, their depositional environment and mode of incorporation into the melange cannot be inferred.
In the Bunnran Formation, the intercalation of mafic extrusive rocks with water-lain metasedimentary rocks demonstrate eruption in a subaerial environment, although unequivocal pillow lavas have not been found. The water depth during eruption is constrained to less than ca. 2000 m by the presence of amygdales in basaltic rocks (Cas & Wright 1988; cf. Moore 1965).

The clasts in the local conglomerate (channel deposit?) in the Bunnran Formation indicate possible sediment sources for this unit. The presence of granite and quartzite clasts strongly suggests a clastic input from a continental source. The green Cr-mica aggregates, probably derived from detrital chromite, also indicate ultramafic or mafic rocks in the source region.

Geochemistry
The chemical composition of an igneous rock representing a liquid is a function of the composition of the source, amount of and conditions during partial melting and differentiation, interaction with wallrock and other magma, etc. Assignment of one or more of these processes to a given suite of rocks, and quantification of the variables involved, requires extensive trace element and isotope studies of fresh rocks with preserved phenocryst assemblages. The ambition here is to obtain geochemical information from the ultramafic to felsic igneous rocks in the area, emphasizing those mafic rocks which represent liquid compositions, to complement field and petrographic data so as to create a basis for further discussion of the environment of formation of the Koli rocks in the Handol area.

During sampling, an effort was made to obtain large, fresh, non-porphyritic and non-amygdaloidal samples that were not strongly sheared, weathered or affected by other secondary alterations. In order to maximize their representativity, samples weighing several kilograms were taken in most cases. Largest possible geographical spreading of sample sites was aimed at but a concentration in some areas was inevitable due to the distribution of appropriate outcrops.

Samples for analysis were crushed in steel jaws and rock chips were powdered in a tungsten carbide shatterbox. W and Co analyses are not presented due to the possibility of contamination from the shatterbox. Whole-rock chemical analyses of 54 samples were made by ICP-AES (Inductively Coupled Plasma-Atomic Emission Spectroscopy) at the Swedish Geological Co., Luleå. All calibrations were based on certified international standards. The stated precision is better than 1.5% (relative standard deviation) for major elements and for trace elements at concentrations ten times greater than the detection limit (Table 1), 10% at trace elements at concentrations five times greater than the detection limit and 50% at the detection limit.

The precision was tested by duplicating the analysis of two samples (numbers 298 and 126). The results (Table 1) are generally within the precision stated by the laboratory for sample 298 but deviations in Se, V, Nb and Yb are clearly higher for number 126. In order to evaluate the accuracy of the analyses, the international standard BCR-1 was included twice in the sample series. The results of this test are also presented in Table 1. With respect to the detection limits and reliability of published standard values (Flanagan 1973), the results obtained

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Table 1. Comparison between duplicate analyses of samples 126, 298 and the international standard BCR-1, together with recommended values for BCR-1 (Flanagan 1973), and the detection limits for ICP-AES during the analytical procedure, for evaluation of precision and accuracy.

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During sampling, an effort was made to obtain large, fresh, non-porphyritic and non-amygdaloidal samples that were not strongly sheared, weathered or affected by other secondary alterations. In order to maximize their representativity, samples weighing several kilograms were taken in most cases. Largest possible geographical spreading of sample sites was aimed at but a concentration in some areas was inevitable due to the distribution of appropriate outcrops.

Samples for analysis were crushed in steel jaws and rock chips were powdered in a tungsten carbide shatterbox. W and Co analyses are not presented due to the possibility of contamination from the shatterbox. Whole-rock chemical analyses of 54 samples were made by ICP-AES (Inductively Coupled Plasma-Atomic Emission Spectroscopy) at the Swedish Geological Co., Luleå. All calibrations were based on certified international standards. The stated precision is better than 1.5% (relative standard deviation) for major elements and for trace elements at concentrations ten times greater than the detection limit (Table 1), 10% at trace elements at concentrations five times greater than the detection limit and 50% at the detection limit.

The precision was tested by duplicating the analysis of two samples (numbers 298 and 126). The results (Table 1) are generally within the precision stated by the laboratory for sample 298 but deviations in Se, V, Nb and Yb are clearly higher for number 126. In order to evaluate the accuracy of the analyses, the international standard BCR-1 was included twice in the sample series. The results of this test are also presented in Table 1. With respect to the detection limits and reliability of published standard values (Flanagan 1973), the results obtained

<table>
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<th>Ni</th>
<th>Sr</th>
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Table 1. Comparison between duplicate analyses of samples 126, 298 and the international standard BCR-1, together with recommended values for BCR-1 (Flanagan 1973), and the detection limits for ICP-AES during the analytical procedure, for evaluation of precision and accuracy.
were less accurate for Ba, Cr and Yb (one analysis) and for Nb (both analyses). The conclusion is that Yb and Nb cannot be used in the geochemical discussion, Sc, V and Cr will be used with the above results in mind.

The results of the chemical analyses of ultramafic rocks (n = 8) and metacumulate rocks (n = 5) are presented in Table 2A and the results for amphibolites (n = 9) and plagiogranites (n = 2) in Table 2B. The fine-grained mafic dykes (Table 2C), which are spatially related to cumulate rocks and ultramafic rocks, are labelled depending on which of the main outcrop areas they are sampled; Handol (n = 6) and Bunnarsjöarna (n = 5) plus Järpån (n = 1). The mafic rocks from the Bunnarn Formation (Table 2D) have been divided into three groups; intrusive rocks (n = 10), intrusive fine-grained (n = 4) and gabbroic rocks (n = 4), respectively.

Table 2A. Whole-rock chemical analyses of amphibolites (A) and plagiogranites (PG) from the Rödberget ultramafic-mafic Complex.

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</table>
The uncertainty in some cases, whether a fine-grained mafic rock is intrusive or extrusive, is indicated by parentheses in Table 2D. The results for the mafic intrusive and intrusive rocks from both units will be discussed first, thereafter the remaining rocks from the Rödberget ultramafic-mafic Complex.

**Mafic intrusive and extrusive rocks**

**Element mobility and classification.** – Many elements, including Ca, Sr, Ba, K and Na, are susceptible to mobility during hydrothermal alteration, low- to medium-grade metamorphism and weathering. However, some high field strength and transition elements, which are of petrogenetic interest (e.g. Ti, P, Zr, Y, Nb, V, Cr, Ni, Sc), are considered to be relatively little affected by metamorphism below amphibolite facies conditions (e.g. Cann 1970; Coish 1977; Winchester & Floyd 1977; Humphris & Thompson 1978; Seyfried & Mottl 1982; Shervais 1982). The results of this study indicate that these considerations are also valid for the rocks analysed here; the elements Ca, Sr, Ba, K and Na show significant scatter on variation diagrams (not shown), while the more stable elements correlate well with each other (e.g. Figs. 5D, 7F, H), which shows that their concentrations are not severely altered by post-magmatic processes.

Of the 30 analysed mafic intrusive and extrusive rocks (Tables 2C, D), 18 are classified as basalts on the total alkali-silica plot (Le Bas et al. 1986; diagram not shown here), and four are basaltic andesites. Higher total alkali content classify the remaining eight samples in the range trachybasalts-basaltic trachyandesites. This apparent alkaliinity is most probably due to the addition of secondary sodium. This is also the case for the apparent calc-alkaline signature on the FeO1-MgO-(Na2O + K2O) diagram (Fig. 4A). Nearly all samples plotting in the calc-alkaline field have a Na2O/CaO ratio higher than normal unaltered basalts. More significant is the Zr-P2O5 plot (Fig. 5F; Floyd & Winchester 1975) involving less mobile elements on which most samples have subalkaline characteristics. The Zr/TiO2 ratio indicates that nearly all samples are subalkaline basalt or andesite/basalt (Winchester & Floyd 1977).

**Variation diagrams.** – The chemical variation of igneous rocks suites is conveniently displayed on variation diagrams where oxides, elements or ratios are plotted against an index of differentiation. The variation of FeO, TiO2, P2O5, MgO, Cr, Ni, Al2O3, MnO, Sc, Zr, Y and V with FeO/Sc and Zr is shown in Figs. 5–7 for the mafic intrusive and extrusive rocks in the Rödberget ultramafic-mafic Complex and the Bunnran Formation. By using several variation indices, the trends can be compared and evaluated. Diagrams with Y as variation index (not shown here) are very similar to those with Zr.

An important feature that is immediately apparent on the variation diagrams is the almost complete overlap of rocks from the Rödberget ultramafic-mafic Complex and the Bunnran Formation. The weak tendency for higher Zr/Y-ratios (Figs. 7F, 8B) in the latter is not significant within the analytical uncertainties. An effort aimed to disprove a cogenetic relationship between the two groups...
Table 2D. Whole-rock chemical analyses of mafic igneous rocks (E = extrusive, I = intrusive, G = metagabbro) from the Bunram Formation.

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using the variation diagrams (Russell & Stanley 1990) has failed. The significance of the chemical similarity will be discussed later.

A differentiation trend similar to that of abyssal tholeiites is indicated on the TiO$_2$-FeO$_1$/MgO diagram (Fig. 5A). A general decrease in Al$_2$O$_3$, MgO, Cr and Ni (Figs. 6A–H), also more or less pronounced in the more mobile CaO and Sr (not shown), indicates that olivine, plagioclase, Cr-spinel and/or clinopyroxene were the major crystallizing phases during differentiation. The vague curvatures of the trends show that the separating phases were changing composition during the crystallization process. A break in slope at FeO$_1$/MgO = 1.3 and Zr = 130 ppm for MnO and in particular Sc (Figs. 7A–D) is thought to reflect the beginning of extensive crystallization of clinopyroxene. This pattern of olivine and plagioclase crystallization prior to pyroxene is typical for low pressure fractionation in ocean-floor basalts (e.g. Bender et al. 1978) and has been verified both experimentally and from natural samples. The SiO$_2$-content has been shown to be constant or to slightly increase during early differentiation.

Relatively tight positive correlations between Zr, Y and TiO$_2$ are consistent with their behaviour as incompatible elements. The most evolved extrusive rock falls clearly off the trend for TiO$_2$, and also for FeO$_1$ and V on the FeO$_1$–FeO$_1$/MgO diagram (Fig. 5A). A general decrease in Al$_2$O$_3$, MgO, Cr and Ni (Figs. 6A–H), also more or less pronounced in the more mobile CaO and Sr (not shown), indicates that olivine, plagioclase, Cr-spinel and/or clinopyroxene were the major crystallizing phases during differentiation. The vague curvatures of the trends show that the separating phases were changing composition during the crystallization process. A break in slope at FeO$_1$/MgO = 1.3 and Zr = 130 ppm for MnO and in particular Sc (Figs. 7A–D) is thought to reflect the beginning of extensive crystallization of clinopyroxene. This pattern of olivine and plagioclase crystallization prior to pyroxene is typical for low pressure fractionation in ocean-floor basalts (e.g. Bender et al. 1978) and has been verified both experimentally and from natural samples. The SiO$_2$-content has been shown to be constant or to slightly increase during early differentiation.

Relatively tight positive correlations between Zr, Y and TiO$_2$ are consistent with their behaviour as incompatible elements. The most evolved extrusive rock falls clearly off the trend for TiO$_2$, and also for FeO$_1$ and V
Fig. 6. Variation diagrams showing the variation of MgO, Cr, Ni and Al₂O₃ with FeO/MgO and Zr. Symbols as in Fig. 5.

(Figs. 5A-D, G-H, 7H), suggesting saturation of the melt with respect to an Fe-Ti-bearing phase at minimum concentrations of Zr = 200 ppm and TiO₂ = 2.6% (cf. DeLong & Chatelain 1990). These considerations are based on only one sample and must therefore be viewed with caution. There is no evidence for apatite removal in the P₂O₅-diagrams (Figs. 5E-F).

**Discrimination diagrams.** – The use of discrimination diagrams is a convenient way to characterize mafic igneous rocks and to compare them with recently formed rocks in well known tectonic environments. Several diagrams have been presented (e.g. Pearce & Cann 1973; Floyd & Winchester 1975; Miyashiro 1975; Pearce 1975, 1980; Shervais 1982) using both incompatible (e.g. Ti, Zr, Y, Nb, V) and compatible (e.g. Cr, Ni) elements. The diagrams are empirical, and some caution should be taken in applying them, although explanations for the element variations in different tectonic environments are given in terms of different source characteristics, melting conditions, etc. (e.g. Pearce & Norry 1979). Furthermore, the fields in many diagrams overlap and may not give an unequivocal result.

In the discrimination diagrams (Figs. 5H, 8A-F) the rocks in both the Röderberget ultramafic-mafic Complex and the Bunnran Formation give a consistent indication of affinity to mid-ocean ridge basalt (MORB). Most or all samples plot within the appropriate MORB field, although the different fields overlap in many diagrams. Some of the more evolved samples tend towards having an apparent within-plate basalt-signature (Figs. 8A, B, D). The analytical uncertainties for V and Cr (see above) do not affect the conclusions based on the discrimination diagrams (Figs. 5H, 8D-F).

Some MORBs deviate from normal (N-type) MORB by being enriched (E- or P-type) in for example the incompatible elements K, Rb, Ba, Nb and light rare earth elements and by having distinct isotopic signatures (e.g. Schilling 1975; Sun et al. 1979; Wood 1979; LeRoex et al. 1983). These features are indicative of heterogeneities in the suboceanic mantle. The slopes on the Zr/Y- and Y/TiO₂-diagrams (Figs. 7F, H) suggest closest affinity to N- or T-(transitional-) type MORB (Perfit et al. 1980; LeRoex et al. 1983, 1985).

**Ultramafic rocks**

The distribution of all analysed ultramafic-felsic samples is shown on the triangular diagrams in Figs. 4A–B. The best preserved ultramafic rocks are characterized by high amounts of MgO, Cr and Ni and low Al₂O₃ contents.
and plot within the field of metamorphic peridotites (Coleman 1977). Their chemical characteristics are more typical of tectonites than of cumulates following the classification of ultramafic rocks in the Scandinavian Caledonides of Qvale & Stigh (1985). The talc-schist samples have higher contents of SiO₂, TiO₂, Al₂O₃, CaO, Sr, Zr, Y, V and are poorer in MgO than the best preserved ultramafic rocks.

Microprobe analyses of olivines have been carried out using a Cameca SX 50 electron microprobe at the Department of Mineralogy and Petrology, Uppsala. Large, old, strained and small, young, strain-free olivines fall in the compositional range forsterite = 92.2–92.6% and NiO = 0.34–0.40% in good agreement with previous analyses (Filippidis & Annersten 1981). Such high NiO contents are characteristic of mantle-derived olivines (Sato 1977), which supports a previous study indicating that early olivines were deformed at upper mantle conditions (van der Kamp 1974). This, together with the almost constant composition of the different olivine generations, argues against the suggestion by Stigh (1979) that the olivine-rich rocks are metamorphosed serpentinite protrusions, since olivines having that origin would have been more forsterite rich (Filippidis & Annersten 1981; cf. Sjöström 1986). The olivine data also argue against an origin as a layered intrusion for the Rödberget ultramafic-mafic Complex.

Recent microprobe data yield different compositions of chromite from dunite compared to chromite from the Cr-rich quartzite and the melange (P. Palmblad, Uppsala, unpublished results). If this is a result of different origins for the chromites, more complex models than those presented in the following section will have to be considered.

**Cumulate rocks**

The leucocratic metagabbros have high MgO, CaO and Al₂O₃ contents and low amounts of FeO, TiO₂, P₂O₅ and other incompatible elements (Table 2, Figs. 4A–B). They likely formed by accumulation of plagioclase and Fe–Mg-silicates during fractional crystallization of mafic magma. This is consistent with available field information. They plot close to the backward projection of the early part of the inferred liquid line of descent (shaded in Figs. 9A–D), supporting the interpretation that they are the plutonic complements to the high-level metadolerites, metagabbros and metabasalts.

**Fe–Ti-amphibolites**

It is evident on many diagrams (for example Figs. 9A–D) that this group of rocks differs significantly in chemical composition from the metadolerites of the Rödberget ultramafic-mafic Complex. At similar Zr-contents, FeOᵢ, TiO₂, MnO, V and Y are significantly higher in most Fe–Ti-amphibolites and MgO, Al₂O₃, Cr, Ni are lower than in the metadolerites (Table 2B, C). High FeOᵢ/MgO ratios and low Cr and Ni contents suggest a higher degree of differentiation, but this explanation is not in accordance with the generally low abundance of Zr. A higher degree of partial melting of the same source as the other groups would dilute the Zr contents but this fails to explain the high FeOᵢ, V and TiO₂. A mantle source of different bulk chemical and/or mineralogical composition or a higher degree of partial melting including a mantle phase rich in high field strength elements could explain the observed pattern for these rocks. However, the preferred interpretation of the origin of this group of rocks is by accumulation of crystals from an evolving basaltic liquid. Addition of variable amounts of hornblende and Fe-Ti-oxide to a differentiated basaltic liquid could explain the relatively high contents or steep trends of Fe, Ti, V and Y, the diluted concentration of the incompatible element Zr and the small amounts of the compatible elements Cr and Ni. This is supported by Figs. 9A–D, where the Fe–Ti-amphibolites may be interpreted as complementary cumulate rocks to more evolved liquids along the inferred liquid line of descent. The more evolved part of the latter is controlled by only one sample, however.

Hornblende–ilmenite/magnetite-bearing gabbros or microgabbros chemically similar to the Fe-Ti-amphibolites have been described from for example the Mid-Atlantic Ridge near 24°N (Thompson 1973), the Oceanographer Fracture Zone in the north Atlantic Ocean (DeLong & Chatelain 1989, 1990), the Sarimiento Ophiolite Complex in southern Chile (Saunders et al. 1979) and from the Løkken Ophiolite in central Norway (Grenne 1989).

**Felsic rocks**

The felsic rocks have the chemical characteristics of differentiates of subalkaline basalts, i.e. plagiogranites (Coleman & Peterman 1975). These typically display low K₂O (<1%) and Al₂O₃ (<14.5%) contents and generally follow the later stage of the Skaergaard liquid trend on the AFM-diagram (Fig. 4A). Field appearance and chemistry (major and trace elements) compare well with similar rocks on Newfoundland in the Bay of Islands Ophiolite (oceanic lithosphere) as opposed to those of the Little Port Complex, which represents the basement of an island arc assemblage and whose felsic rocks were formed by partial fusion of amphibolite (Malpas 1979).

**Summary of geochemistry**

The results of the geochemical investigation suggest that the vast majority of the analysed mafic intrusive and extrusive rocks are tholeiitic basalts or basaltic andesites with modified contents of mobile elements. Such modifications could have occurred shortly after solidification by sub-seafloor hydrothermal alteration and metamorphism and/or during later regional syndeformational metamorphism.
The analysed igneous rocks cover a range of compositions that are similar to those commonly found in Recent ocean floor and in many ophiolites. The chemical variations can be simply explained by partial fusion of sub-oceanic mantle, fractional crystallization in shallow magma chambers resulting in formation of ultramafic-mafic cumulates and variously differentiated liquids.

No significant chemical differences have been detected between the metadolerite dykes of the inferred ophiolite and the mafic meta-igneous rocks of the Bunnran Formation. The original distinction between these major groups was based on their associated rocks, ultramafic-mafic or metasedimentary. In the light of their chemical similarities, a cogenetic hypothesis can be considered, although this is by no means a unique explanation of the data. If the most evolved extrusive sample in the Bunnran Formation is representative of the later part of the liquid evolution, then support is given to a complementary magmatic relationship with the Fe–Ti-amphibolites in the Rödberget ultramafic-mafic Complex.

Models for the formation of the Köli rocks in the Handöld area

A genetic model must account for the presence of: (1) a fragmented ophiolite with a MORB signature for the dolerite dykes, (2) a melange with mainly ophiolitic debris and minor continental components, (3) mafic igneous sheets and flows interlayered with metasedimentary rocks, formed at <2000 m depth and with a MORB chemistry, and (4) a varied metasedimentary succession with common quartz-rich (continent-derived) rocks and probable volcanic components.

Two models will be considered. One assumes a close genetic relationship between the Rödberget ultramafic-mafic Complex and the Bunnran Formation, the other assumes separate rock-forming events.

The first model accepts that the dykes in the ophiolite and the dykes in the Bunnran Formation are equivalent and that they fed the high-level intrusions and extrusions. Proximity to both a continental and a volcanic source permitted high sedimentation rates during mafic magma formation at a spreading centre. Most of the magma formed intrusions and extrusions in the overlying metasedimentary rocks rather than sheeted dykes. Geological and geochemical support for this interpretation has been presented above.

The second model involves the formation and early accretion of the Rödberget ultramafic-mafic Complex to a continental margin. This was followed by erosion, formation of a melange and deposition of the sedimentary part of the Bunnran Formation during a later phase of rift magmatism. This model would favour earlier
The general environment of formation was characterized by three input sources: an igneous source supplying magma of MORB composition, a source for quartz-rich sediments and a source for tuffaceous material. A plate tectonic scenario with these ingredients is envisaged where a spreading centre was situated close to a continental margin and within the reach of pyroclastic material from island arc volcanoes.

A well-known site of present ocean crust formation close to a continental margin is found in the Gulf of California, where mostly MORB magma (e.g. Saunders et al. 1982a, b) is intruded into rapidly accumulating continent-derived sediments, leading to formation of basaltic sill-sediment complexes (Einsle 1985). The spreading ridges in the Gulf of California consist of short segments separated by long transform faults. A similar situation is found off Burma in the Andaman Sea (Fig. 10; Rodolfo 1969; Curray et al. 1982; Hja Maung 1987), where oblique subduction of the Indian plate has lead to formation of a pull-apart basin with spreading centres and transform faults behind an island arc system and close to the continental slope off Burma. Unfortunately, the volcano-sedimentary succession in this basin is not well known. However, it is assumed here that the rock types and their relations in parts of the Andaman Sea are similar to the conditions in the better known Gulf of California (with the addition of volcanic activity) and to those prevailing when the rocks of the Handölf area were formed. The Andaman Sea environment is especially appealing as a model because of the high density of transform faults and fracture zones where diapirc and/or sedimentary melanges are likely to be formed (e.g. Saleeby 1984). Sea-water circulation and hydration in these zones will promote formation of serpentine protrusions, and uplifted oceanic crust along fault scarps will be eroded to form sedimentary deposits (cf. Karson & Dewey 1978; Simonian & Gass 1978). These zones of weakness may have been used to accommodate strain during later orogenic events and are likely to have been obscured by overprinting. Some of them could presently be expressed as ultramafic belts, deformation zones and terrane boundaries.

The Andaman Sea has already been used as an analogue for the Solund–Stavfjord Ophiolite Complex (Skjerlie & Furnes 1990; Furnes et al. 1990) in the southwestern Norwegian Caledonides (Fig. 10). Other areas which have been compared with the Andaman Sea are the Sierra Nevada ophiolite belt (Saleeby 1982), where multiple ensimatic rifting events and sedimentation patterns record the production of a marginal or inter-arc basin system adjacent to both primitive island arc and continental source regions, and the Devil's Elbow ophiolite remnant and cover in the Klamath Mountains of California (Wyld & Wright 1988).

Regional implications
Several authors have correlated the tectonostratigraphically lowest Köli units in the Tännforsen region with
units at a comparable tectonostratigraphic level further north in the Swedish counties of Jämtland and Västerbotten (Sjöstrand 1978; Beckholmen 1978, 1984; Gee & Kumpulainen 1980; Sandwall 1981; Sjöström 1983; Stephens & Gee 1985) and with units to the south at Feragen and Otta in Norway (Stephens & Gee 1985). They have previously been included in the Virisen terrane (Fig. 10; Stephens & Gee 1985) and the Lower Köli Terrane (Roberts 1988). However, the problems concerning tectonostratigraphic correlation of the Tännforsen Köli rocks along the mountain belt were discussed briefly by Stephens & Gee (1989). As a consequence, these authors proposed a more cautious approach and placed the Tännforsen rocks in a separate isolated terrane.

Previously published geochronological results that may be of relevance for the present study are presented in Fig. 11. No definite age of formation of the Rödbergget ultramafic-mafic Complex and the Bunnnan Formation can be suggested based on this compilation, but it will serve as a basis for discussion. Caledonian rifting-related mafic magmatism (in a convergent plate setting) is concentrated into two main periods at ca. 490 Ma and ca. 435 Ma (data summarized by Pedersen et al. 1991 and Tucker et al. 1991). The older phase of magmatism (and tectonic activity) occurred along both the Baltoscandian and Laurentian margins that were separated at this time by the Iapetus Ocean. At the time of the younger magmatic phase, most terranes were close to each other as the Iapetus Ocean was in its waning stages. Correlation can be made with: (a) terranes which were close to Baltica in the Early Ordovician or with (b) successions in the Mid-Late Ordovician—Early Silurian. These are: (a) the older rocks in the Virisen terrane, maybe including Otta and (b) the younger part of the Virisen terrane and the rocks associated with for example the Solund—Stavfjord Ophiolite Complex and the igneous complex at Sulitjelma (Boyle 1980; Pedersen et al. 1991).

The Virisen terrane as defined by Stephens & Gee (1985) has its type area in the Bjørkvatnet—Virisen area (Fig. 10; Kulling 1933) in the county of Västerbotten, Sweden. It is a volcano-sedimentary complex with a wide range in age (Early Ordovician to Early Silurian). The major components are, from its base and upwards: (a) solitary and detrital serpentinites (with Ordovician gastropods; Holmqvist 1980; Bruton & Harper 1981), (b) transitional within-plate basalts (Stephens et al. 1985), (c) subduction-related igneous rocks (U-Pb zircon age 488 ± 5 Ma; Claesson et al. 1983), (d) clastic rocks (Mid-Ordovician) derived from a platformal/cratonic source with minor mafic-ultramafic detritus, and (e) Ashgill limestones and Llandovery phyllites with rift-related basalts (Stephens et al. 1985). The clastic rocks (turbidites) were presumably easterly-derived and have been interpreted to reflect uplift of the outer margin of Baltica due to an early orogenic event (Stephens & Gee 1985). Deposition of turbidites is also recorded in the Lower Allochthon in Jämtland where there is good stratigraphic control (Karis; in Gee & Kumpulainen 1980).

In the Scandinavian Caledonides, ultramafic bodies are mainly concentrated in the lower and upper parts of the Seve Nappes and in the Lower Köli (e.g. Zachrisson 1969; Stigh 1979). Most of these bodies are solitary, primitive ultramafic rocks and, although it was not excluded that some are basal parts of ophiolites, an origin as serpentinite protrusions was considered most likely (Stigh 1979; Qvale & Stigh 1985). In the Lower Köli, they are often associated with monomict detrital serpentinites. According to Stigh (1979), these are located at various stratigraphic levels within the Lower Köli, from the lowest part (Early Ordovician or older) to the upper part of the Gilliks Formation (pre-Ashgill) below the Voitja conglomerate.

The partly well-preserved ultramafic rocks at Feragen (Røros area in Norway) have been interpreted to be cumulates formed in a back-arc environment (Moore & Hultin 1980). In Västerbotten, mafic-ultramafic igneous rocks associated with clastic metasedimentary rocks have been interpreted as having formed in relation to initial island-arc development or back-arc spreading, associated with the formation of a relatively small ocean basin close to the margin of the Baltoscandian craton (Stephens 1977). However, Stephens & Gee (1985) found the evidence of proximity of the Virisen terrane (Lower Köli) to the Baltoscandian miogeocline difficult to reconcile with the character of its detrital serpentinites. Therefore, they proposed that the ultramafic rocks may have originated in the Early Ordovician in a truly ensimatic environment, remote from the Baltoscandian margin. The ultramafic rocks were subsequently transported towards the margin in the Middle Ordovician, and were tectonically emplaced into the depositional basin of the Virisen terrane immediately outboard of the Baltoscandian margin. However, no unconformity reflecting this emplacement has yet been recognized within the type areas of the Virisen terrane.

An Andaman Sea-type setting (Fig. 10) will have large facies variations with a dominance in different areas of
for example arc volcanic rocks, continent-derived sedimentary rocks or back-arc basin oceanic crust, and different facies may overlap in time and space. The variations in this setting may explain and integrate into a coherent model the following components of the Virisen terrane: (1) the large amount of serpentinite protrusions and detrital serpentinites, (2) the presence of fragments of oceanic/marginal basin crust (Handöll, Feragen; possibly Raudfjellet in the Snåsa area in Norway, cf. Stigl 1979), (3) the mafic intrusions and extrusions among the metasedimentary rocks, (4) the continent-derived metasedimentary rocks, and (5) the rifted arc intrusive and extrusive rocks (e.g. Anka rede in Västerbotten in Sweden) and distal tuffs.

Recent tectonic models of the Scandinavian Caledonides are not specific concerning the polarity of the Early Ordovician Virisen arc subduction zone (Stephens & Gee 1985, 1989), but an easterly dip was preferred by Gee (1975) and Stephens (1988). The model presented in Fig. 10 is merely an expanded version of the latter adding the complexities arising from oblique convergence leading to a large concentration of fracture zones in the basin. If the Handöll rocks are Early Ordovician in age or older, then correlation with the Virisen terrane is supported.

Another possibility is to compare with the Caledonian events in the Late Ordovician–Early Silurian. The Solund–Stavfjord Ophiolite Complex (Skjerlie & Furnes 1990; Furnes et al. 1990), which has been dated to 443 ± 5 Ma (Dunning & Pedersen 1988), has important characteristics in common with the rocks of the Rödbergert ultramafic-mafic Complex and the Bunnran Formation. These characteristics include an ophiolite complex overlain by a clastic sequence which contains extrusive and intrusive rocks that are cognetic with the ophiolitic rocks. Detrital ultramafic rocks are also known from this area (Andersen et al. 1990). The coeval rocks at Sullitjelma, dated to 437 ± 2 Ma (Pedersen et al. 1991), have also been interpreted to contain an ophiolite (Boyle 1980).

Conclusions
The combined geological and geochemical evidence favours a close genetic relationship between the Rödbergert ultramafic-mafic Complex and the Bunnran Formation. This implies formation of oceanic crust in a basin infilled with material from both continental and volcanic sources. Fracture zones in the oceanic crust are likely sites for formation of serpentinite diapirs and melange. The complex environment in the Andaman Sea can serve as a modern analogue for the site of formation of the Kölí rocks of the Handöll area, and can be expanded to be a model for the entire complex igneous-sedimentary Lower Kölí unit (Virisen terrane) in the Scandinavian Caledonides. However, correlation with other units in the late Ordovician–Early Silurian is still possible. Until age determinations of rocks and minerals have been carried out, the questions of timing and correlation remain open.

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References


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