Petrology and isotope geology of the Hunnedalen Monzonoritic Dyke Swarm, SW Norway: a possible late expression of Egersund Anorthosite magmatism

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The widest and northernmost dyke of the post-tectonic Late Proterozoic Hunnedalen Monzonoritic Dyke Swarm has been studied petrographically, chemically (major and trace elements) and isotopically (Rb-Sr; Sm-Nd; K-Ar) together with its country rocks. Samples from the core of the dyke appear fresh or nearly fresh and have a monzonoritic composition. Samples from within 1 m of the contacts appear severely altered, petrographically, chemically and isotopically, most probably during a very low-grade regional phase of metamorphism, M4, related to Early Palaeozoic burial and to Caledonian orogenesis. The unaltered central part the dyke is slightly inhomogeneous with some concentration of earlier formed minerals in the centre of the dyke (higher plagioclase content, higher Mg-ratio) due to flow differentiation. Trace element discrimination diagrams indicate a continental, within-plate setting. In altered samples the change in chemistry involved variable shifts in the contents of Rb and Sr, as well as in Sr-isotope ratios.

Fresh samples produce concordant mineral-whole isochrons with both the Rb-Sr (834 ± 9 Ma; $Sr_{(i)}$ 0.70442 ± 0.00004; MSWD 0.594) and Sm-Nd (835 ± 47 Ma; $Nd_{(i)}$ 0.51166 ± 0.00004; MSWD 1.57) methods, most probably representing the age of intrusion. Whole-rock samples give Rb-Sr and Sm-Nd ages with large uncertainties, due to restricted spread of data points. K-Ar whole-rock ages of fresh samples exhibit some spread, but the average (715 ± 30 Ma) is considerably lower than the probable intrusion age. On the other hand, biotite K-Ar ages are higher (average 893 Ma). This apparent discrepancy is explained by the incorporation of excess Ar during emplacement, which still is present in minerals (e.g. biotite) with closure temperatures above the temperature of the M4 phase of metamorphism, but lost from other minerals (e.g. feldspars) with a closure temperature below the M4-temperature.

A remarkable similarity exists between the Hunnedalen monzonorites (c. 835 Ma) and other monzonorites in Rogaland-Vest Agder, especially the chilled margins of the Hidra monzonorite and of the Bjerkreim-Sokndal layered intrusion (c. 930 Ma). This similarity in both mineralogy and chemistry (major and trace elements, element ratios, REE patterns, spidergrams, Eu/Eu* anomalies and $Sr_{(i)}$ values) supports a genetic relationship.

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Introduction

The Proterozoic basement of Southwest Norway contains two distinct, post-tectonic, mafic dyke swarms, the Hunnedalen and Egersund Dyke Swarms (Versteeve 1975, Falkum 1982, Verschure 1985). The Egersund Dyke Swarm strikes ESE-WNW and crops out, in general, close to the North Sea coast, roughly between Undheim in the northwest and Flekkefjord in the southeast. It consists of olivine dolerites, dolerites and trachydolerites, emplaced at upper crustal levels (P_{tot} 1 atm-1 kb) in anorthositic intrusion and granulite-facies gneisses (Venhuis & Barton 1986). The Egersund dykes are dated at 630-650 Ma (whole-rock Rb-Sr and Sm-Nd; Sundvoll 1987) and 616 ± 3 Ma (baddeleyite U-Pb, Bingen et al. 1997, 1998a). Their emplacement seems to be related to a Late Precambrian extensional tectonic event, i.e. the opening of the Iapetus Ocean (Sundvoll 1990), as is supported by their OIB to E-MORB-like chemistry (Miller & Barton 1992, 1993).

The Hunnedalen Dyke Swarm strikes ENE-WSW and is found inland, about 40 km southeast to 90 km east of Stavanger (Fig. 1). The petrology and geochemistry of 9 dykes were studied by Drent (1982). An accurate age has hitherto not been established. Versteeve (1975) reported 4 widely differing whole-rock K-Ar ages of 940, 848, 655 and 570 Ma (recalculated with the I.U.G.S recommended constants). Sundvoll (1990) mentioned «a preliminary Rb-Sr date on one dyke indicates an age about 820 Ma» but without giving any details. The age of the Hunnedalen Dyke Swarm was thought to be older than that of the Egersund dykes on both geological [e.g. coarser grain size, less finer grained chilled margins, lacking 'glass' that occurs in the Egersund dykes (Antun 1956)] and paleomagnetic grounds (Poorter 1972, 1975, 1981).

In this paper, new results are presented on the Hunnedalen Dyke Swarm, with the following objectives:

1 - to study mineralogical and chemical variations across and along the dyke and the influence of alteration/retrograde metamorphism on the mineralogy, chemistry and isotope systematics;
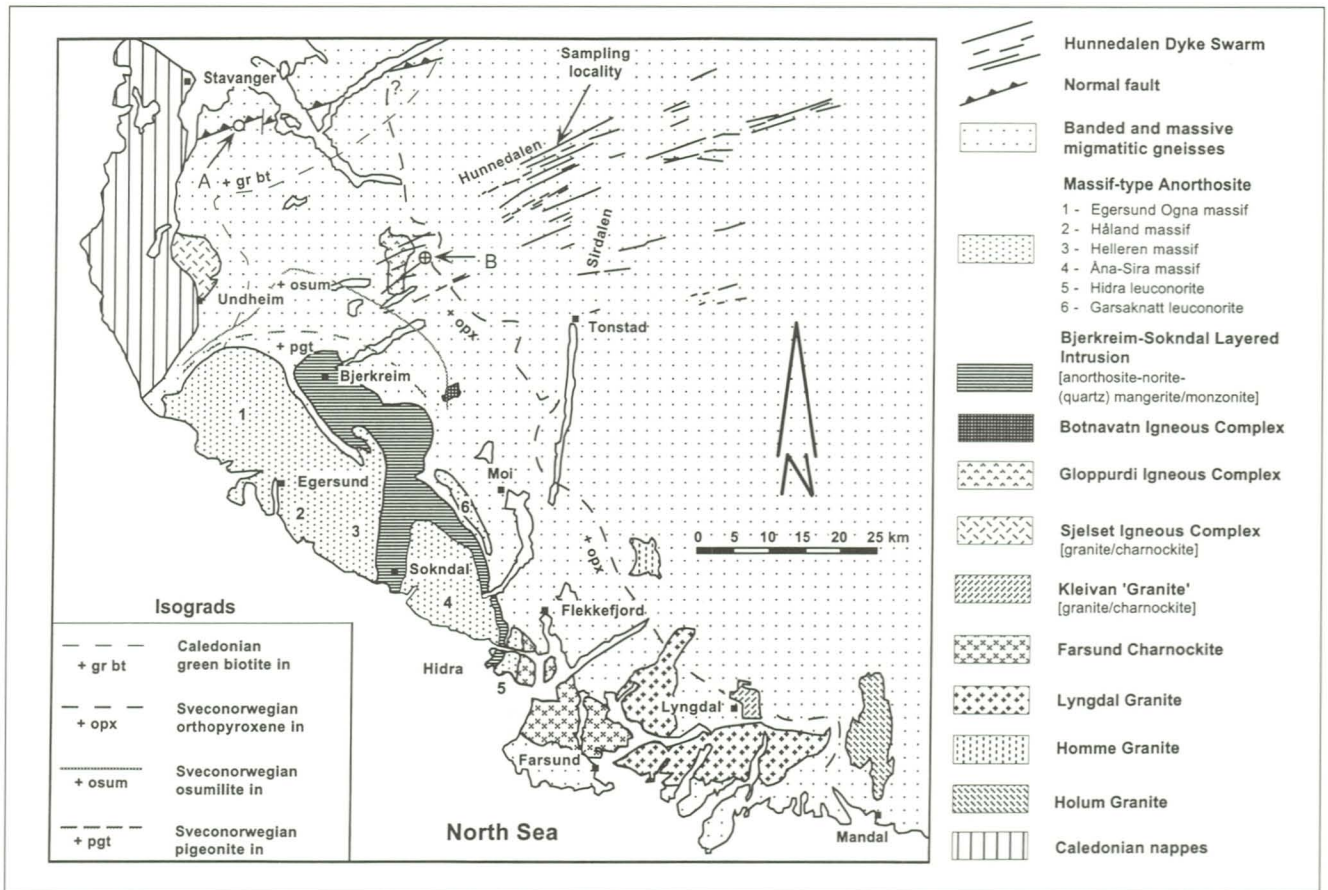
Fig. 1. Geological sketch map of SW Norway with the Hunnedalen Dyke Swarm. Locality A - Caledonian green biotite from ENE-WSW striking normal fault zone.. Locality B - Caledonian pseudotachylite from ENE-WSW striking joint/fault zone.

2 - to date the Hunnedalen Dyke Swarm and to evaluate various dating methods (Rb-Sr, Sm-Nd and K-Ar);

3 - to evaluate a possible relationship between the Hunnedalen Dyke Swarm and the voluminous older Anorthosite/Monzonorite/Charnockite/Granite (AMCG) suite magmatism of Rogaland/Vest Agder.

Geological Setting

The Proterozoic basement of the southwestern part of Norway, is dominated by 2 main units (Fig. 1): (i) a large igneous complex, and (ii) its metamorphic envelope.

The igneous complex comprises several massif-type anorthosites (e.g., the Egersund-Ogna, the Håland-Helleren and the Åna Sira anorthosites), leuconorites (e.g., the Hidra and the Garsaknatt leuconorites), the folded anorthositic to mangeritic Bjerkreim-Sokndal layered intrusion and several charnockitic to granitic bodies (e.g., the Farsund charnockite, the Kleivan charnockite/granite and the Sjelset charnockite/granite (Duchesne & Michot 1987, Maijer et al. 1994).

Two different basic parental magmas are recognised to account for the Rogaland massifs (Duchesne et al. 1985, Duchesne & Maquil 1987). A basaltic magma produced the voluminous anorthosite massifs, while a jotunitic (= monzonoritic) one formed large dykes and intrusions (Bjerkreim-Sokndal layered intrusion and the Hidra leuconorite; Duchesne et al.

1989). The monzonorites are intimately associated and coeval, but not comagmatic, with the massif-type anorthosites (Duchesne et al. 1989, Duchesne 1990). The basaltic magma is considered to have been mantle derived. According to Vander Auwera et al. (1998) the origin of jotunites «remains a subject of considerable debate, despite their textural and geochemical characteristics from one anorthosite complex to another». Six proposed models have been listed. One model involves the partial melting of basic to intermediate rocks in the lower crust (Duchesne et al. 1989) possibly triggered by hot anorthositic diapirs rising slowly to their final level of emplacement (Duchesne et al. 1985).

The high-grade polymetamorphic envelope consists of charnockitic and granitic migmatites with intercalations of mafic rocks [(pyroxene-) amphibolites, pyroxene gneisses], augen gneisses and pelitic or quartzitic/calcareous metasediments (Hermans et al. 1975, Tobi et al. 1985, Maijer 1987). They underwent 4 or 5 phases of metamorphism. The earliest well-defined regional phase of metamorphism, M1, reached upper amphibolite facies, including anatexis and possibly locally granulite facies. Relict mineral assemblages indicate temperatures of 600-700°C and pressures of 6-8 kb (Jansen et al. 1985). M1 probably was preceded by an ill-defined Gothian metamorphic phase, M0 (Maijer 1987).

The peak of metamorphism, M2, is found in a 20-40 km wide zone around the igneous complex. A sequence of more

or less concentric isograds: orthopyroxene-in (in granitic rocks), hornblende + quartz-out, garnet-out, osumilite-in, pigeonite-in (Maijer et al. 1981, Tobi et al. 1985) reflects increasing temperatures from about 750°C at the Opx-in isograd to about 1000°C close to the igneous complex (Jacques de Dixmude 1978) at relatively low pressures of about 4 kb (Jansen et al. 1985). The isograd pattern, the high temperatures reached and the static character of M2 recrystallisations are reminiscent of deep-seated contact metamorphism (Tobi et al. 1985) related to the emplacement of the igneous complex.

The ages of M2 metamorphism and anorthosite magmatism in Rogaland/Vest Agder have long been, and still are, a matter of debate. Recently, the massif-type anorthosites have been dated at 931 ± 2 Ma by U-Pb on zircons and baddeleyites (Schärer et al. 1996). This age is interpreted as the emplacement age and is much younger than the best estimate of the M2 peak of metamorphism of 1000-1020 Ma (Bingen & van Breemen 1998). The anorthosite intrusion ages of 931 ± 2 Ma (Schärer et al. 1996) would also imply a short duration of anorthosite magmatism, contrary to earlier speculations of the emplacement as a long-lasting, multiphase intrusion history (Duchesne 1984, Duchesne et al. 1985, Duchesne & Michot 1987, Duchesne & Maquil 1987). To overcome the age controversy, Bingen et al. (1998b) - accepting the genetic relationship of the thermal metamorphism M2 to the intrusion of the magmatic complex - attribute an age of 930-920 Ma to M2, while zircon U-Pb ages of around 1050 Ma (Wielens et al. 1981) and an oldest generation of monazites 1024-997 Ma (Bingen et al. 1998b) should be regarded to be related to the main phase of the Sveconorwegian orogeny and the M1 phase of metamorphism. The M3 metamorphic phase, though omnipresent, is of much lesser importance in volume and reflects slow, near isobaric cooling after the peak of metamorphism. It is expressed by well developed exsolution phenomena and partial decomposition of M2 minerals and the formation of the hydrous minerals hornblende and biotite around M2 minerals. Geothermobarometry indicates PT conditions of 550-750°C at 3-5 kb (Jansen et al. 1985).

After the M3 event, southern Norway underwent further cooling and a general uplift and peneplanation. Further cooling is demonstrated by regional hornblende $^{40}Ar/^{39}Ar$ ages (916 +12/-14 Ma in the Rogaland granulite-facies area; 871 +8/-10 Ma in the amphibolite-facies area; Bingen et al. 1998b) and by regional biotite Rb-Sr and K-Ar ages (895-850 Ma, averaging c. 870 Ma; Verschure et al. 1980, Bingen et al. 1998b). The present surface is probably close to, or not far below, the sub-Cambrian peneplain, i.e., above the sub-Cambrian 120°C isotherm (Hansen et al. 1996). The area was, on account of regional fission track ages, subsequently buried to depths with temperatures in excess of 120°C, i.e. an overburden of >4 km, assuming a normal geothermal gradient of 30°C/km (Hansen et al. 1996). The overburden of Lower Palaeozoic sediments and overthrusting produced slightly elevated temperatures resulting in a retrograde metamorphic phase, M4, in prehnite-pumpellyite facies to lower greenschist facies in the west, separated by a green biotite-in isograd at 5-15 km from the present Caledonian thrust front (Fig. 1). East of the green biotite-in isograd, the M4 event produced minerals such as prehnite, pumpellyite, stilpnomelane, epidote, chlorite, serpentinite, talc, white mica, diaspore, tremolite, albite and microcline. These minerals are found locally in varying but usually small amounts throughout the area. They are supposed to have resulted from very low-grade burial metamorphism, M4a, of Early Palaeozoic age and low-grade metamorphism, M4b, related to Caledonian overthrusting (Verschure et al. 1980, Sauter et al. 1983). Temperatures of burial metamorphism did not exceed 200-250°C in the Evje area, where Late Proterozoic titanite fission track ages (590-790 Ma) have not been annealed by the M4 phase of metamorphism. On the other hand, the M4 event reached temperatures of c. 400°C at the Caledonian green biotite-in isograd (Fig. 1; Verschure et al. 1980). Interpolation between these values suggests a temperature of c. 300-350°C during the M4 event in the Hunnedalen area.

The low to very low-grade retrogradation is found most abundantly along fault zones. A Caledonian (reactivation) age of these faults is supported by:

(i) A K-Ar age of 469 Ma on green biotite formed from the hornblende of a mylonitic biotite amphibolite at Horve-Bratthetland, (A on Fig. 1; Verschure et al. 1980). This mylonite manifests a major ENE-WSW striking normal fault, which can be followed for over 30 km with an estimated net slip in the order of 1-2 km.

(ii) A whole-rock K-Ar age of 516 ± 15 Ma on a pseudotachylite (unpublished result) found at the foot of a steep cliff on the south side of the Maudal glacial valley (B on Fig. 1; UTM coordinates $^3452^{65}159$) This valley is one of a series of parallel ENE-WSW striking glacial valleys following a major joint/fault zone, forming the most prominent lineament in the Rogaland topography. This lineament is parallel to the Hunnedalen Monzonoritic Dyke Swarm.

Post-Caledonian uplift and exhumation removed the Lower Palaeozoic overburden in most parts of southern Norway. Regional apatite fission track age patterns support a dome style of late-stage uplift, with two phases of increased exhumation in Triassic-Jurassic times and since the Neogene, i.e. from 30 Ma onward (Rohrman 1995, Rohrman et al. 1995). Andriessen (1990) presented a tentative reconstruction of the post-Caledonian thermo-tectonic history of the Hunnedalen area, based on apatite fission track data, which do not indicate an undisturbed, continuous, post-Caledonian cooling history of the Hunnedalen rocks but rather suggest a period of rapid uplift and a mild temperature increase.

Previous paleomagnetic, petrological and geochemical studies

Paleomagnetic investigations (Poorter 1972, 1975, 1981,

Pesonen et al. 1989, 1991) revealed that all basement rocks, magmatic as well as metamorphic, have similar magnetisation directions, obviously due to a simultaneous magnetisation of the Gothian-Sveconorwegian crustal block during the post-orogenic uplift and cooling at about 1.0-0.8 Ga ago. The magnetisation of the Hunnedalen Dyke Swarm differs only slightly from that of the overall magnetisation directions of the basement. The same applies to the Egersund dykes, but these dykes have a reversed polarity (polarity defined according to Pesonen & Neuvonen, 1981). Therefore, Poorter concluded that the Hunnedalen dykes intruded shortly after the magnetisation of the basement, while the intrusion of the Egersund dykes took place later.

Drent (1982) studied the petrology and geochemistry of 21 samples from 9 different Hunnedalen dykes. Fresh rocks are biotite mononorites composed of plagioclase + orthopyroxene + clinopyroxene + biotite + Fe-Ti oxides with minor orthoclase + apatite ± quartz. One of his samples, taken c. 20 km south-southeast of our sampling site, is olivine-bearing. Two of his samples stem from the same dyke that we investigated, collected about 800 m east-northeast of our sampling site. Thin dykes and the margins of thicker dykes are locally altered to amphibolites. Subsequent low-grade chloritisation, correlated with the regional phase of (very) low-grade metamorphism M4, severely affected both the mineralogy and the chemical composition.

Based on Harker diagrams using the Mg-ratio as differentiation index, Drent (1982) showed that the variation in major and trace elements in the least altered Hunnedalen dykes resulted from fractionation of plagioclase, orthopyroxene and Fe-Ti oxide. One of his samples, from the same dyke that we sampled, slightly deviates from the observed fractionation trends, which is attributed to flow differentiation in this relatively thick dyke.

Drent (1982) calculated a temperature of c. 1200°C and a pressure P_{tot} of c. 6.25 kb for equilibrium between first formed plagioclase and melt, using the models of Kudo & Weill (1970) and Ghiorso & Charmichael (1980) in the case of dry conditions. This pressure is higher than the regional post-M3 pressure of 3-5 kb (or lower), indicating that plagioclase started crystallising in, or during ascent from a deeper reservoir. Low water contents would imply slightly lower temperatures and higher pressures.

Field relations and sampling

The Hunnedalen Dyke Swarm consists of several tens of parallel to sub-parallel, ENE-WSW striking, sub-vertical dykes in an area measuring c. 25 by 80 km, to the east of Stavanger (Falkum 1982). The ENE-WSW orientation of the Hunnedalen dykes is very conspicuous in the field from a number of parallel valleys and gullies, easily seen on topographic maps and aerial photographs. Surprisingly, no dykes have been observed in the most pronounced ENE-WSW striking glacial valleys in this area, e.g. Hunnedalen, Maudalen,
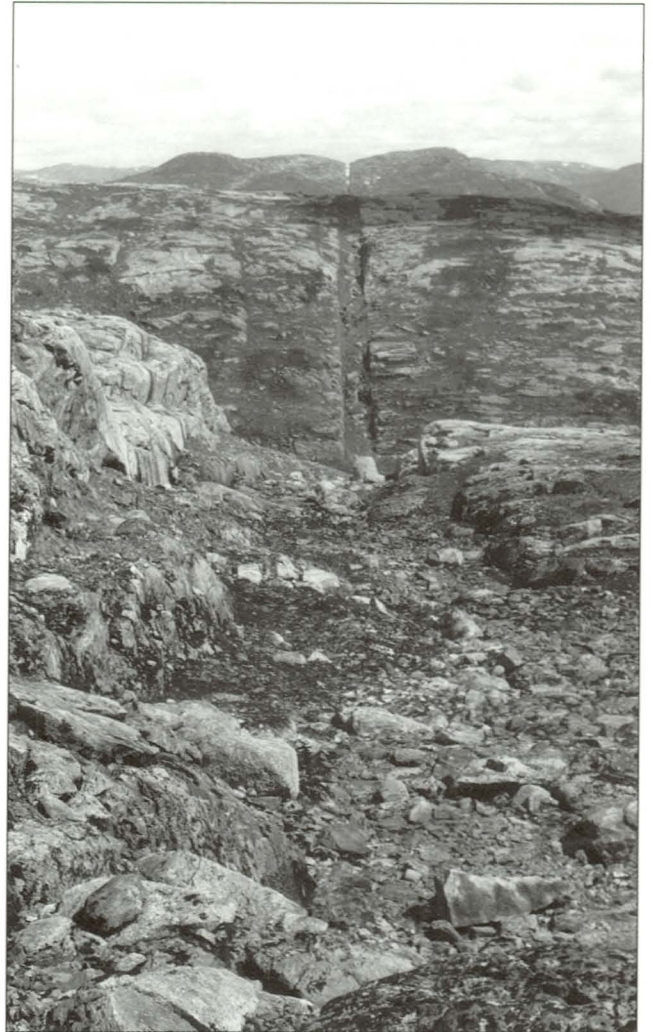
Fig. 2. Photograph taken at the Hunnedalen West site showing the sampled Hunnedalen dyke. Note the straight continuation of this dyke for at least 5 km towards ENE and the excellent degree of exposure of the country rocks.

Austrumsdalen and Ørsdalen.

The dyke sampled for dating is the widest and northernmost one of the Hunnedalen Dyke Swarm west of Sirdalen. This dyke is generally c. 10 to 20 m wide with sharp contacts and is remarkably straight over about 10 km (Fig. 2). Irregularities and offshoots are rare. Xenoliths of country rocks are very rare and have only been found c. 25 km further southwest of our sampling sites. The dyke rock is fine- to medium-grained and appears to be homogeneous. Chilled borders are found, but they are usually inconspicuous in view of the alteration of the marginal parts of the dyke. To check for variations across and along the dyke, two profiles of 10 samples each have been sampled across the dyke (Fig. 3a, b), further referred to as: Hunnedalen West (UTM coordinates. $^3671^{65}301$, height 920 m: Rog 389-399) and Hunnedalen East (UTM coordinates. $^3678^{65}304$, height 870 m: Rog 475-484), c. 750 m apart.

The country rocks of the investigated Hunnedalen dyke consist of migmatitic granitic or charnockitic gneisses, inter-
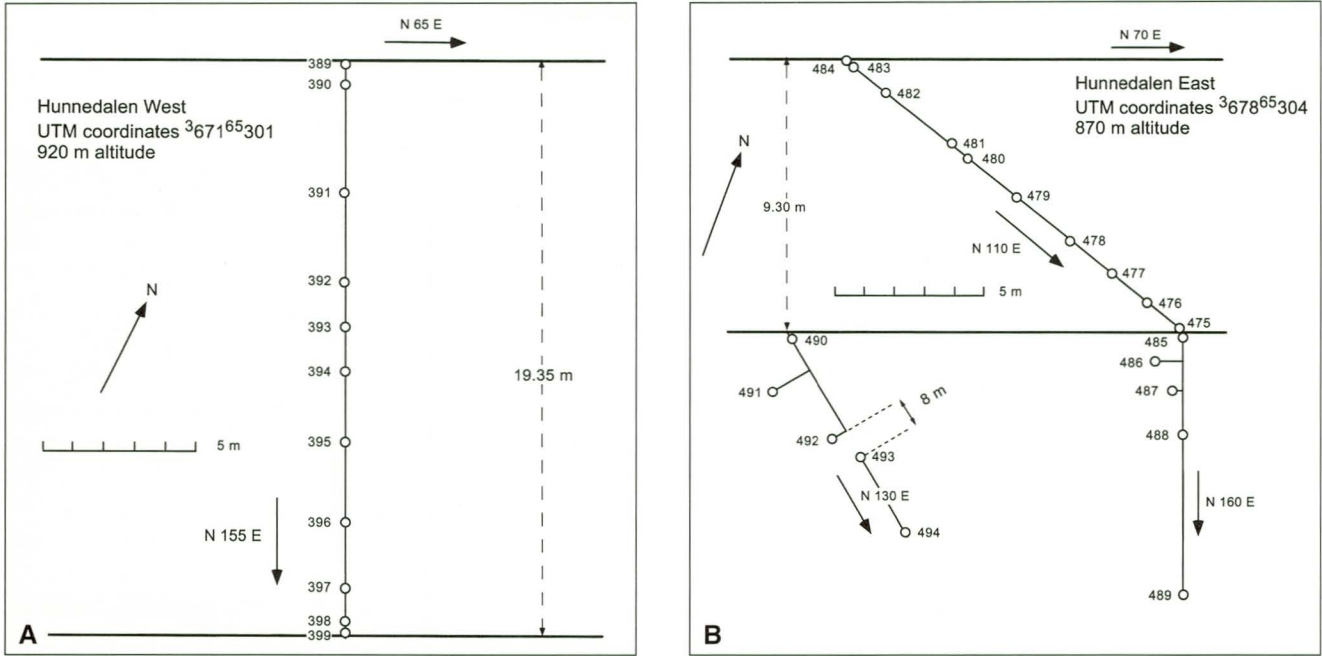
Fig. 3. Sample profiles of the Hunnedalen West (3a) and Hunnedalen East (3b) locations.

bedded with biotite-garnet ± sillimanite gneisses and pyroxene amphibolites. At Hunnedalen East, biotite-garnet gneisses [Rog 485-489] and amphibolites [Rog 490-494] have been collected to check for possible interaction between the country rocks and the intruding dyke magma (Fig. 3b).

Petrography

Dyke rocks

The dyke rocks are fine- to medium-grained and appear fresh, or almost fresh, in the central parts of the dyke, but are

		Sample	Quality	Qtz	Or	Pl	An %	Cpx	Opx	Bt	Hbl	Grt	Ox	Ap
Monzonorite Hunnedalen West		Rog 389	—	Ø	Ø	[XXX]		[ ]	[ ]	[ ]			[ ]	Ø
		390	-	Ø	[Ø]?	[XXX]		[ ]	[ ]	[ ]			[X]	Ø
		391	+	a	X	XXX	35-40	X	XX	X			X	Ø
		392	+	a	X	XXX	35-40	X	XX	X			X	Ø
		393	++	Ø	X	XXX	35-40	X	XX	X			X	Ø
		394	++	Ø	Ø	XXX	35-40	X	XX	X			X	Ø
		395	++	Ø	X	XXX	35-40	X	XX	X			X	Ø
		396	+	Ø	X	XXX	35-40	X	XX	X			X	Ø
		398	—	Ø	[a]	[XXX]		[ ]	[ ]	[X]			[X]	Ø
		399	—	Ø	Ø	[XXX]		[ ]	[ ]	[X]			[X]	Ø
Monzonorite Hunnedalen East		Rog 475	-	Ø	X	[XXX]	35-40	[X]	[ ]	[X]			[X]	Ø
		476	+	Ø	X	XXX	35-40	X	XX	X			X	Ø
		477	++	Ø	X	XXX	35-40	X	XX	X			X	Ø
		478	++	Ø	X	XXX	35-40	X	XX	X			X	Ø
		479	++	Ø	Ø	XXX	35-40	X	XX	X			X	Ø
		480	++	Ø	Ø	XXX		X	XX	X			X	Ø
		481	++	Ø	X	XXX		X	XX	X			X	Ø
		482	++	Ø	X	XXX		X	XX	X			X	Ø
		483	—	Ø	?	[XXX]		[ ]	[ ]	[ ]			[X]	Ø
		484	—	Ø	?	[ ]		[ ]	[ ]	[ ]			[ ]	
Country rock Hunnedalen East	Biotite garnet gneiss	Rog 485	—	XX	XX	[XXX]				[ ]		[ ]	[ ]	a
		486	+	XX	XX	XXX				Ø		X	a	a
		487	++	XX	XX	XXX	± 20			X		X	a	a
		488	++	XX	XX	XXX	± 20			X		X	a	a
		489	+	XX	XX	XXX	± 20			[Ø]		X	a	a
Country rock Hunnedalen East	Pyroxene amphibolite	Rog 490	-			[ ]		XX	XX	[Ø]	X		Ø	a
		491	-			[ ]		XX	[ ]	[a]	XX		Ø	a
		492	±			XXX	65-80	X	XX	Ø	XX		Ø	a
		493	++			XXX	65-80	XX	XX	a	XX		Ø	a
		494	++			XXX	65-80	XX	XX	a	XX		Ø	a

a	=	< 1 % (accessory)	Quality		
Ø	=	1 - 5 %			
X	=	5 - 15 %	++	=	fresh
XX	=	15 -30 %	+	=	nearly fresh
XXX	=	30 - 55 %	±	=	slightly altered
[X]	=	partly altered mineral	-	=	altered
[ ]	=	completely altered mineral	—	=	severely or completely altered

Table 1. Modal compositions of the primary minerals of the Hunnedalen dyke and country rocks.

commonly severely altered at the c. 1-m wide margins. This holds for both profiles (Table 1). The fresh samples [Rog 391-396 and Rog 476-482] are mineralogically homogeneous, with 50-55 % andesine and 20-30 % pyroxene. Minor minerals are 7-12 % brown biotite, 6-7 % Fe-Ti oxide, 1-1.5 % apatite, 5-10 % orthoclase and 0.5-2 % quartz. Subhedral plagioclase laths are 0.5-1.5 mm in length and slightly zoned [An 35-40; optically determined]. Drent (1982) reported microprobe analyses of plagioclase from other Hunnedalen dykes, varying from An_{60} (in a more primitive olivine-bearing dyke or in phenocrysts) down to c. An_{30} (rims of phenocrysts and matrix). Plagioclase is twinned according to albite and Carlsbad laws. Plagioclase does not show any preferred orientation. Orthopyroxene is the main mafic mineral. Textures suggest two different types; (i) larger subhedral prismatic crystals, or glomerophyric aggregates, usually crowded with exsolved ilmenite platelets and associated with subordinate clinopyroxene as tiny exsolution lamellæ, (ii) smaller anhedral to subhedral interstitial grains and aggregates, which give the rock its intergranular texture. These orthopyroxenes are irregularly intergrown with variable amounts of clinopyroxene, occurring as lamellæ and grains or rims around the orthopyroxene host. The amount and orientation of clinopyroxene inclusions indicate that some of these intergrowths are reminiscent of inverted pigeonite. Mg-ratios (Mg# = MgO/MgO + FeO) of orthopyroxenes from samples of the same dyke (Drent 1982) vary from 0.39 to 0.38, i.e. falling in the range of pigeonite composition. Clinopyroxene occurs also as dispersed, small, interstitial grains. Brown biotite forms small flakes commonly associated with Fe-Ti oxide and pyroxenes. Biotite occurs dispersed throughout the dyke without enrichment towards the contacts. It does not show any preferred orientation and is regarded as a late magmatic product. Drent (1982) also considered (most of) the biotite in the Hunnedalen dykes as late-magmatic crystallisation products, a conclusion mainly based on their high Al^{IV} and TiO_2 contents (4.0-7.2 wt % TiO_2 in biotites of the least altered dykes, indicating high temperatures of formation). Fe-Ti oxides, both ilmenite and magnetite (Drent 1982), occur predominantly as elongated grains.

Interstitial quartz and anhedral perthitic, in places mesoperthitic, orthoclase occur in between the plagioclase laths. Slender, occasionally hollow prisms of apatite form the main accessory constituent and are most abundant in interstitial perthite-quartz aggregates. Zircon has not been identified, although sporadic pleochroic haloes in biotite around tiny grains (< 10 μm) attest to the presence of mineral matter containing radioactive elements.

Chilled border samples [Rog 10, 13] from the same dyke are porphyritic (Versteeve 1975), with phenocrysts of plagioclase (c. 5 %, c. An_{40}, up to 1 mm in length) occurring in a fine-grained matrix of plagioclase, pyroxene, biotite and Fe-Ti oxides. Matrix plagioclase shows some preferred orientation parallel to dyke contacts, whereas biotite does not show any preferred orientation.

Petrographically, the Hunnedalen dyke rock may be classified as biotite monzonorite to norite. Apparently crystallisation started with plagioclase and was successively accompanied by orthopyroxene, pigeonite + Ca-rich clinopyroxene, biotite and orthoclase + quartz.

In the altered samples of the dyke margins [Rog 389, 390, 398, 399, 475, 483, 484] the primary minerals have been variably replaced by (very) low-grade minerals. Obviously, orthopyroxene is the mineral most sensitive to alteration, and is replaced by aggregates of fibrous actinolitic amphibole and/or chlorite. Plagioclase is altered into turbid crystals of oligoclase-albite, with abundant sericite and saussurite. Biotite is replaced by chlorite with abundant fine lenses, veinlets and rims of titanite. Fe-Ti oxides are replaced by titanite. The quartz content seems slightly higher than in fresh samples. Some samples near the contact with the country rocks are slightly brecciated [Rog 483] or strongly sheared [Rog 484]. Titanite-chlorite ± albite ± epidote ± carbonate veinlets [Rog 389, 399, 475, 483] or prehnite-rich veinlets [Rog 389] are found locally. The cores of plagioclase crystals in Rog 389 are replaced by irregular blebs of K-feldspar.

The alteration minerals demonstrate metamorphic recrystallisation in prehnite-pumpellyite or lowest greenschist facies, indicative of metamorphic temperatures of 300-350°C. Such a (very) low-grade metamorphism corresponds with the regional M4 phase of metamorphism, thus supporting an Early Palaeozoic/Caledonian age for this recrystallisation. Most likely the alteration resulted from an influx of fluids (mainly water) along the dyke contacts, combined with minor displacements and shearing. Both the dyke and the country rocks were infiltrated for about one metre from the dyke contact.

Country rocks

The country rocks appear as severely retrograded to (very) low-grade assemblages close to the contacts with the mafic dyke [Rog 485; 490-492], but are practically unaltered at 1-4 m distance from the dyke [Rog 486-489; 493 and 494]. The fresh metapelitic rocks are biotite-garnet gneisses, which contain 20-30 % quartz, 20-30 % microperthite, 35-40 % oligoclase [c. An_{20}], 1-3 % parallel-oriented brown biotite and 8-14 % garnet porphyroblasts; accessories are zircon, opaque (ilmenite?), monazite(?) and pyrite(?). Alteration in the samples Rog 485 and 486 produced chlorite (after biotite and garnet), sericite and saussurite (after plagioclase). Sample Rog 485, collected close to the mafic dyke, is severely sheared. White mica veinlets are found in Rog 486.

The fresh pyroxene amphibolites contain 35-45 % calcic plagioclase, showing reverse zoning (core c. An_{65}, rim c. An_{80}, with fine Huttenlocher intergrowths in between), 12-22 % anhedral, pale brown pleochroic, pargasitic hornblende, 35-40 % poikiloblastic ortho- and clinopyroxene, minor brown biotite (0.5-1 %) and Fe-Ti oxide (1-1.5 %; ilmenite?); apatite is the main accessory. Pyroxenes overgrow hornblende and plagioclase, demonstrating prograde granulite-facies meta-

Major elements by XRF (in wt %)

	Rog 389	Rog 390	Rog 391	Rog 392	Rog 393	Rog 394	Rog 395	Rog 396	Rog 398	Rog 399
SiO_2	48.27	50.34	51.65	51.29	51.43	51.14	51.09	51.24	51.07	50.09
TiO_2	3.78	3.58	3.60	3.69	3.69	3.62	3.61	3.62	3.61	3.78
Al_2O_3	15.44	15.66	15.43	15.67	15.66	16.25	15.96	15.59	15.57	16.13
FeO	12.42	12.45	12.17	12.05	12.25	11.85	11.94	12.23	12.18	12.57
MgO	4.07	4.18	3.74	3.81	3.92	4.10	4.02	3.89	4.10	4.31
CaO	5.36	6.01	6.40	6.46	6.38	6.55	6.55	6.38	6.40	5.56
MnO	0.13	0.15	0.13	0.13	0.14	0.13	0.13	0.14	0.14	0.14
Na_2O	3.01	3.62	3.71	3.58	3.69	3.78	3.72	3.71	3.69	3.40
K_2O	2.30	0.94	1.83	1.91	1.95	1.80	1.81	1.92	0.94	0.86
P_2O_5	0.68	0.65	0.65	0.62	0.63	0.56	0.59	0.65	0.64	0.67
BaO	0.10	0.050	0.08	0.08	0.08	0.07	0.07	0.08	0.06	0.06
LOI	2.78	1.84	0.52	0.68	0.24	0.36	0.44	0.54	1.55	2.32
Total	98.35	99.47	99.91	99.97	100.06	100.21	99.93	99.99	99.95	99.89

Trace elements by XRF (in ppm)

	Rog 389	Rog 390	Rog 391	Rog 392	Rog 393	Rog 394	Rog 395	Rog 396	Rog 398	Rog 399
Rb	60.1	38.5	46.8	49.1	47.5	44.0	45.1	49.5	35.6	34.7
Sr	389.3	478.1	484.7	493.0	515.6	558.7	529.4	510.3	477.1	388.8
La	24	14	17	13	15	10	5	13	16	12
Ce	77	73	65	79	69	57	67	69	70	71
Nd	48	43	44	43	36	36	42	43	42	46
Dy	6	5	6	3	4	5	1	4	5	3
Yb	4.0	3.5	3.0	3.1	3.5	2.6	2.3	3.5	3.9	2.7
Y	35.8	28.4	31.4	29.2	28.4	24.7	26.4	28.9	29.0	30.9
Th	1.5	2.2	2.5	1.8	1.8	1.3	1.6	0.6	2.3	0.9
Pb	5.1	5.7	5.7	4.5	5.6	5.4	3.9	5.2	5.3	4.6
Zr	291.5	270.1	278.8	269.3	266.1	233.2	245.6	271.4	260.1	279.3
Hf	8.2	7.5	7.8	8.0	8.1	7.2	7.5	8.0	7.1	7.9
V	176	162	155	156	159	159	154	160	154	167
Cr	51	44	38	33	35	37	37	35	32	30
Ni	53.9	56.9	53.3	54.8	57.8	61.6	60.5	54.9	55.9	57.5
Co	47	50	45	46	50	46	48	45	46	49
Zn	177.0	144.4	142.0	138.6	142.8	126.6	136.5	141.4	145.9	141.9
Cu	23	29	33	34	36	33	34	36	32	39
Ga	23.6	23.2	23.8	23.7	23.6	23.1	23.2	23.8	22.5	23.8

Element ratios

	Rog 389	Rog 390	Rog 391	Rog 392	Rog 393	Rog 394	Rog 395	Rog 396	Rog 398	Rog 399
R1	1172	1411	1265	1269	1226	1225	1235	1214	1442	1481
R2	1078	1158	1173	1188	1184	1223	1214	1182	1193	1125
$\frac{100(K_2O)}{K_2O+Na_2O}$	43.31	20.61	33.03	34.79	34.57	32.26	32.73	34.10	20.30	20.19
$K_2O + Na_2O$	5.31	4.56	5.54	5.49	5.64	5.58	5.53	5.63	4.63	4.26
K/Rb	318	203	324	323	340	338	333	322	218	205
Rb/Sr	0.1543	0.0804	0.0968	0.0996	0.0851	0.0787	0.0853	0.0970	0.0746	0.0892
$Mg\#\frac{MgO}{MgO+FeO}$	0.3687	0.3743	0.3539	0.3604	0.3632	0.3814	0.3750	0.3618	0.3749	0.4400

Table 2. (a) Major and trace element chemistry and relevant element ratios of Hunnedalen West dyke rocks.

Major elements by XRF (in wt %)

	Rog 475	Rog 476	Rog 477	Rog 478	Rog 479	Rog 480	Rog 481	Rog 482	Rog 483	Rog 484
SiO_2	50.01	50.70	50.10	50.66	50.91	50.71	50.85	50.92	49.76	48.03
TiO_2	3.73	3.58	3.60	3.63	3.61	3.63	3.60	3.65	3.56	3.64
Al_2O_3	15.72	15.14	15.19	15.38	15.80	15.60	15.48	15.39	15.05	15.72
FeO	11.55	12.16	12.30	12.39	12.10	12.11	12.26	12.42	12.41	11.66
MgO	3.84	3.88	3.88	3.99	4.09	3.99	4.00	3.93	3.94	4.19
CaO	6.39	6.45	6.36	6.43	6.57	6.53	6.48	6.51	6.93	6.30
MnO	0.14	0.14	0.13	0.13	0.13	0.13	0.13	0.16	0.20	0.12
Na_2O	2.90	3.63	3.71	3.72	3.78	3.78	3.76	3.70	3.39	4.29
K_2O	1.26	1.76	1.73	1.73	1.68	1.76	1.75	1.79	0.73	1.13
P_2O_5	0.67	0.62	0.62	0.61	0.59	0.62	0.62	0.63	0.63	0.66
BaO	0.05	0.07	0.07	0.07	0.06	0.07	0.07	0.07	0.04	0.05
LOI	2.20	0.21	0.09	0.01	0.04	0.05	0.00	0.03	1.63	2.81
Total	98.46	98.34	97.78	98.75	99.36	98.98	99.00	99.20	98.27	98.60

Trace elements by INA (*) and XRF (in ppm)

	Rog 475	Rog 476	Rog 477	Rog 478	Rog 479	Rog 480	Rog 481	Rog 482	Rog 483	Rog 484
Cs*	2.59	0.772	0.921	0.947	0.952	1.22		1.12	0.761	1.15
Ba*	426	501	445	445	379	440	433	475	241	374
Rb	28.7	35.9	36.1	35.6	34.4	36.6	36.0	38.5	20.5	24.4
Sr	433.3	484.1	492.3	498.4	519.0	506.9	505.6	491.9	452.4	357.9
La*	28.7	29	30.6	30.8	27.3	28.7	28.9	28.7	28.1	31.1
Ce*	69	67.6	69.9	71.6	59.6	69.8	64.2	66.1	69.3	73
Nd*	41.1	42.7	45.6	41.8	34.3	32.7	41.5	40.6	39	45.6
Sm*	9.25	9.19	9.5	9.08	8.6	8.98	9.11	8.94	9.31	9.56
Eu*	3.01	3	2.93	2.92	2.86	2.98	2.99	2.95	3.01	3.24
Tb*	1.24	1.09	1.15	1.23	1.1	1.21	1.17	1.21	1.17	1.33
Dy	6	6	6	6	5	8	4	3	3	8
Yb*	2.26	2.16	2.06	2.09	1.87	2.01	1.91	2	2.17	2.37
Lu*	0.353	0.299	0.29	0.319	0.246	0.311	0.299	0.305	0.34	0.385
Y	35.0	33.8	34.0	32.6	30.7	32.1	32.9	34.2	34.1	36.1
Th*	1.98	1.87	2.06	2.17	1.8	1.9	1.86	2.01	1.84	2.02
Pb	7.5	6.9	7.0	5.9	5.0	6.9	7.3	7.2	8.2	4.9
Zr	322.4	306.7	307.0	303.3	286.8	301.4	303.2	312.5	306.3	317.4
Hf*	8.58	7.93	7.93	8.19	7.34	7.83	7.9	7.88	7.86	9.28
Ta*	1.23	1.18	1.19	1.15	1.11	1.19	1.01	1.11	1.14	1.17
Nb	20	18	19	22	18	18	18	19	19	19
V	168	161	157	158	157	155	157	163	156	169
Cr*	56.4	41.7	45.3	41.3	43.7	38.9	31.4	48.9	48.6	32.5
Sc*	15.1	14.7	14.3	13.7	13.5	14.1	14.3	14	14.6	16.4
Ni	66.8	68.2	70.1	73.1	75.8	70.5	72.0	69.8	74.4	59.0
Co*	43.3	42.9	43.4	43.1	43.5	43.5	44.4	42.6	42.6	42.6
Zn	174.9	162.7	169.5	167.6	152.9	155.8	156.7	158.7	164.8	166.3
Cu	35	46	47	47	45	48	47	47	51	30
Ga	29.2	27.2	28.3	28.7	27.8	28.3	28.9	27.6	26.7	23.6

Element ratios

	Rog 475	Rog 476	Rog 477	Rog 478	Rog 479	Rog 480	Rog 481	Rog 482	Rog 483	Rog 484
R1	1590	1247	1182	1212	1228	1195	1210	1221	1504	0995
R2	1183	1180	1171	1188	1216	1203	1195	1193	1232	1190
$\frac{100(K_2O)}{K_2O+Na_2O}$	30.29	32.65	31.8	31.74	30.77	31.77	31.76	32.60	17.72	20.85
$K_2O + Na_2O$	4.16	5.39	5.44	5.45	5.46	5.54	5.51	5.49	4.12	5.42
K/Rb	364	407	398	403	405	399	403	386	295	384
Rb/Sr	0.0662	0.0741	0.0732	0.0715	0.0715	0.0663	0.0723	0.0711	0.0783	0.0453
$Mg\#\frac{MgO}{MgO+FeO}$	0.3720	0.3625	0.3598	0.3646	0.3759	0.3699	0.3676	0.3606	0.3613	0.3904

Table 2. (b) Major and trace element chemistry and relevant element ratios of Hunnedalen East dyke rocks.

morphism. Low-grade alteration produced sericite and saussurite (after plagioclase), actinolite or serpentine material (after hornblende, pyroxene) and chlorite (after biotite) in samples close to the mafic dyke.

The presence of orthopyroxene in the unaltered mafic country rocks and its absence in the unaltered metapelitic country rocks indicates that the metamorphic grade of the country rocks prior to dyke emplacement was transitional between the amphibolite and granulite facies. This is in agreement with the rarity of orthopyroxene in gneissic granitic to charnockitic country rocks (Bos 1984, Seepers 1988).

Chemistry

Major and trace elements of mafic dyke rocks of Hunnedalen West [Rog 389-399], Hunnedalen East [Rog 475- 484] and country rock samples of Hunnedalen East [Rog 485-494] were analysed by XRF. Trace elements of Hunnedalen East dyke rocks [Rog 475-484] were also analysed by INA. Analyses are presented in Tables 2a,b,c. Analytical methods are presented in the Appendix.

Small systematic differences between unaltered samples of Hunnedalen West and East (Table 2a and 2b; Fig. 4a and 5a) are observed for some major elements (SiO_2, K_2O), trace elements (e.g. Rb, Zr) and element ratios (e.g. K/Rb), while other major elements (e.g. TiO_2, CaO), minor elements (e.g.

Major elements by XRF (in wt %)

	Rog 485	Rog 486	Rog 487	Rog 488	Rog 489	Rog 490	Rog 491	Rog 492	Rog 493	Rog 494
SiO_2	72.59	68.05	69.76	65.71	68.68	49.08	48.32	48.74	48.09	49.40
TiO_2	0.29	0.44	0.54	0.56	0.43	0.66	0.72	0.81	0.76	0.78
Al_2O_3	13.81	16.14	14.78	16.43	15.33	15.96	16.20	16.26	16.45	15.57
FeO	2.61	3.35	3.68	5.13	3.48	7.77	7.95	8.40	7.86	8.09
MgO	0.79	1.08	1.24	1.44	1.07	8.69	9.12	9.19	9.16	9.30
CaO	0.69	1.47	1.71	1.50	1.63	11.71	11.84	11.58	12.66	12.46
MnO	0.03	0.04	0.03	0.08	0.04	0.15	0.15	0.15	0.14	0.15
Na_2O	4.09	3.76	3.60	3.59	3.45	1.22	1.65	1.98	1.98	2.08
K_2O	3.93	4.84	3.57	4.52	4.42	1.86	1.37	0.75	0.54	0.52
P_2O_5	0.04	0.05	0.05	0.07	0.05	0.06	0.06	0.06	0.06	0.06
BaO	0.13	0.09	0.08	0.10	0.11	0.02	0.02	0.01	0.00	0.00
LOI	0.73	0.87	0.13	0.24	0.20	1.91	1.79	1.10	0.97	0.86
Total	99.73	100.18	99.17	99.37	98.89	99.09	99.19	99.03	98.68	99.27

Trace elements by XRF (in ppm)

Rb	28.7	35.9	36.1	35.6	34.4	36.6	36.0	38.5	20.5	24.4
Sr	433.3	484.1	492.3	498.4	519.0	506.9	505.6	491.9	452.4	357.9
La	25	50	70	58	51	3	4	0	5	12
Ce	56	108	130	140	109	10	4	10	5	4
Nd	28	45	57	64	49	11	11	7	8	8
Dy	4	6	3	6	4	3	0	2	3	0
Yb	4.3	5.6	6.1	7.8	6.3	2.4	2.3	2.5	3.5	3.2
Y	37.0	46.3	44.9	68.0	54.8	15.1	16.3	18.0	17.2	17.2
Th	10.1	14.5	17.1	19.4	15.4	1.8	1.4	1.0	1.2	0.3
U	1.6	1.7	3.0	2.8	2.5					
Pb	23.2	29.1	26.6	34.9	31.6	8.4	6.1	6.9	8.0	5.9
Zr	151.9	183.6	360.5	284.5	290.2	45.4	42.7	46.7	42.1	49.8
Hf	4.3	4.8	9.7	7.8	7.7	1.6	1.2	1.2	1.4	1.8
V	35	40	48	57	40	213	225	232	228	232
Cr	117	121	106	131	103	220	234	225	234	218
Ni	15.8	24.1	22.3	29.9	22.6	34.8	34.3	36.7	34.5	31.9
Co	7	7	9	15	8	41	44	46	46	44
Zn	41.0	72.7	79.4	103.3	80.1	60.5	83.9	71.4	87.4	63.1
Cu	2	4	5	18	4	5	6	4	4	5
Ga	13.4	22.5	19.0	21.0	19.9	18.6	16.4	16.3	16.5	15.5

Element ratios

R1	2383	1958	2416	1887	2207	2167	2071	2113	2135	2184
R2	384	527	534	554	528	1997	2037	2014	2132	2100
$\frac{100\,(K_2O)}{K_2O + Na_2O}$	49.00	56.33	49.79	55.73	56.16	60.39	45.36	27.47	21.43	20.00
$K_2O + Na_2O$	8.02	8.61	7.17	8.11	7.87	3.08	3.02	2.73	2.52	2.60
K/Rb	380	246	213	227	234	195	246	336	352	389
Rb/Sr	0.0662	0.0741	0.07324	0.0715	0.06630	0.0723	0.07110	0.07826	0.0453	0.0682
$Mg\#\frac{MgO}{MgO+FeO}$	0.2324	0.2438	0.2520	0.2192	0.2352	0.5279	0.5343	0.5225	0.5382	0.5348

Table 2 (c) Major and trace element chemistry and relevant element ratios of Hunnedalen East country rocks.

P_2O_5) and element ratios (e.g. Mg#) are identical. With only two sampled profiles the significance of these lateral differences and similarities is difficult to evaluate.

Normative composition

Calculated with total Fe as FeO, the normative compositions of all fresh dyke samples [Rog 391-396; Rog 476-483] are quartz-free and olivine-bearing (1-6 %), which is not consistent with the actual mineralogical composition. The mineralogical composition can be visualised from a calculation of biotite + quartz instead of olivine + orthoclase [3 Ol + 1 Or = 1 Bt + 3 Qtz] (Fig. 4b and 5b). As mentioned before, olivine was reported (Drent 1982) from a Hunnedalen dyke 20 km to the south-southeast. However, the unaltered dyke rocks are olivine-free and quartz-bearing if calculated with 2.0 wt % Fe_2O_3, as recommended by Thompson et al. (1972), for
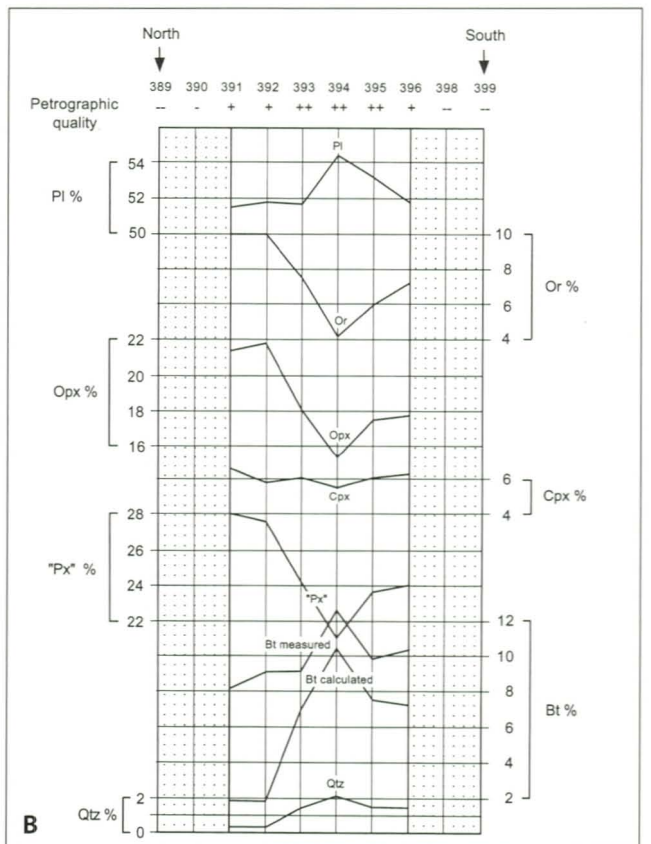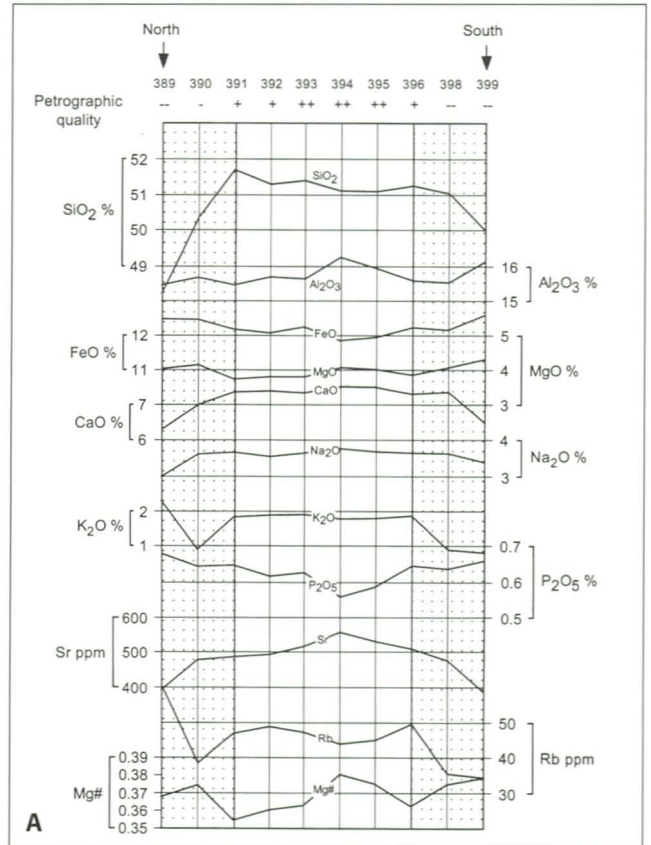
Fig. 4. Chemical (4a) and normative variations (4b, calculated with Bt + Qtz instead of Ol + Or) along sample profile Hunnedalen West. The stippled area represents the zone of altered rocks. Normative compositions calculated with total iron as FeO.

A

B
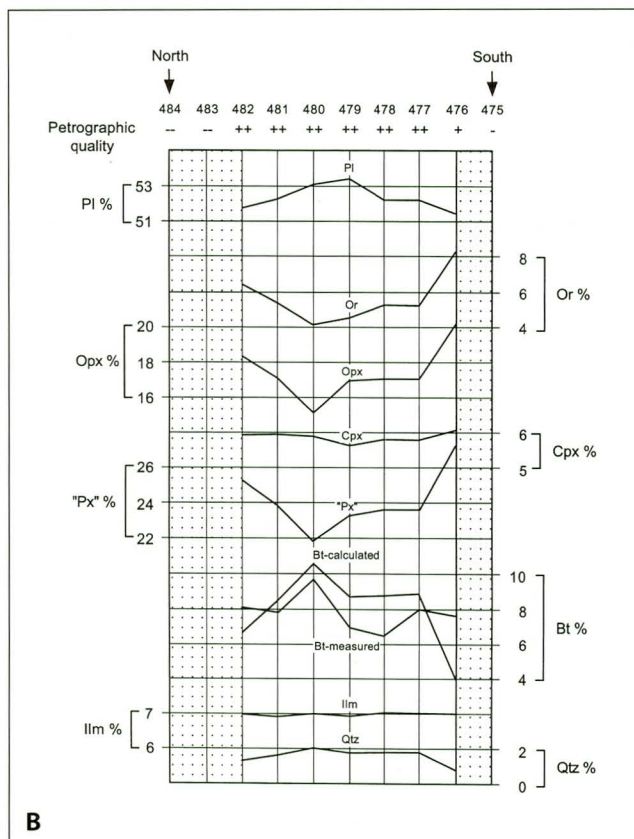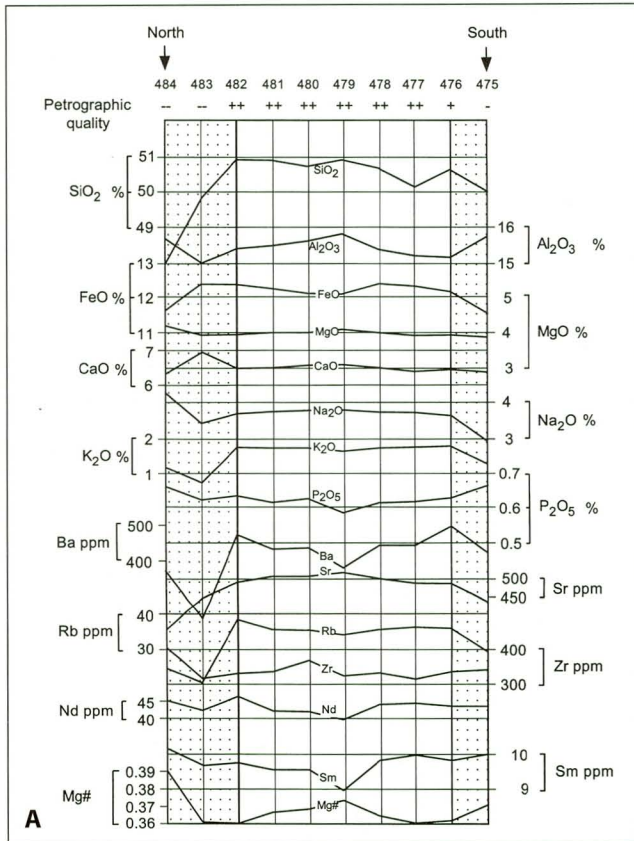
Fig. 5. Chemical (5a) and normative variations (5b, calculated with Bt + Qtz instead of Ol + Or) along sample profile Hunnedalen East. The stippled area represents the zone of altered rocks. Normative compositions calculated with total iron as FeO.
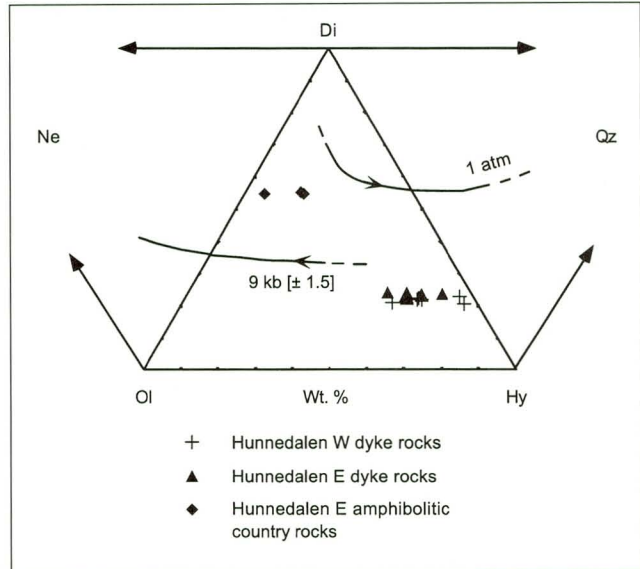
Fig. 6. Normative diopside, olivine, hypersthene and nepheline or quartz (CIPW, wt %) of Hunnedalen basic rocks (Thompson 1982). Major elements of unaltered rocks were recalculated to a volatile-free total of 100 % before plotting and with total iron as FeO. If Fe_2O_3 is present, the data points shift to the right. With Fe_2O_3 taken as 1.5 wt % (Thompson 1982) the Hunnedalen monzonorites cluster around the Di-Hy join, being either slightly Ol-normative or slightly Qtz-normative. With Fe_2O_3 taken as 2 wt %, as recommended by Thompson et al. (1972) for rocks with Na_2O + K_2O = 4 - 7 wt %, all Hunnedalen monzonorites are Qtz-normative and plot in the Di-Hy-Qtz field close to the Di-Hy join.

rocks with a Na_2O + K_2O content of 4-7 wt % (compare Fig. 6).

In Thompson's (1982) normative (nepheline-) diopside-hypersthene-olivine (-quartz) diagram (Fig. 6) the Hunnedalen dyke samples plot below the extrapolation of the experimentally determined 9 (± 1.5) kb anhydrous cotectic for basic magmas in equilibrium with olivine, plagioclase and clinopyroxene and far away from the 1 atmosphere cotectic (Thompson 1982). This would indicate that fractionation of the Hunnedalen dyke magma took place at great depth, probably close to the mantle-crust boundary. However, an anhydrous character of the Hunnedalen magma is doubtful because of the presence of 7-12 % biotite, considered to be of late-magmatic origin. This may imply that fractionation of the Hunnedalen magma took place at lower pressure (Gibson et al. 1997).

Chemical classification

On the Total Alkali-Silica [TAS] chemical classification diagram of Middlemost (1994), the analysed dyke rocks of Hunnedalen East and West plot in the monzogabbro to alkalic gabbro (= trachybasalt to alkalic basalt) field (Fig. 7). The unaltered dyke rocks of Hunnedalen East cluster in the monzogabbro field near the boundary line separating the monzogabbro field from the monzodiorite (= basaltic trachyandesite) field. Unaltered dyke rocks of Hunnedalen West cluster on this boundary line. Unaltered dyke rocks of both Hunnedalen East and West appear slightly alkaline accor-
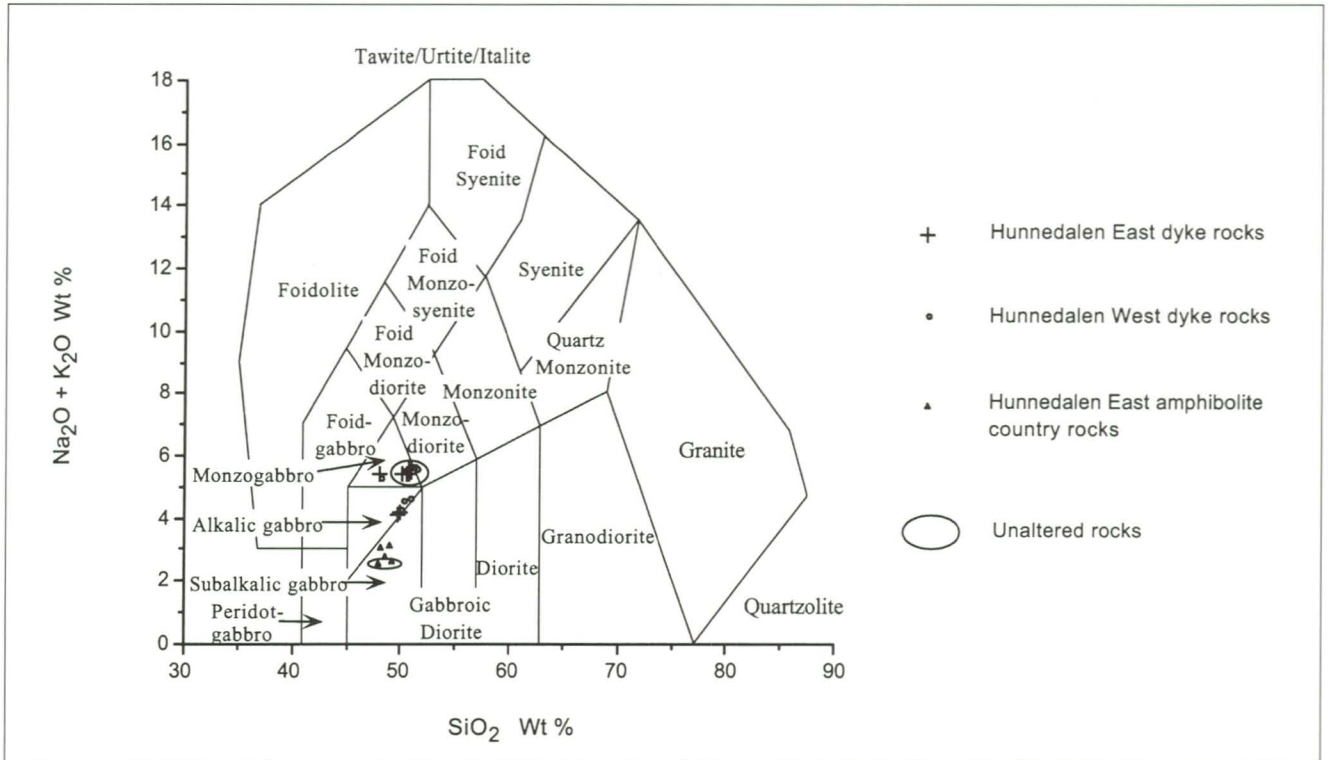
Fig. 7. The Hunnedalen rocks plotted in the TAS diagram of Middlemost (1994).

ding to the subdivision into alkaline and subalkaline (tholeii-te) series on the TAS diagram made by Irvine & Baragar (1971). Their slightly alkaline character is also evident in their plotting in the P_2O_5-Zr and the TiO_2-Zr/P_2O_5 x 10^4 discrimination diagrams (Winchester & Floyd 1976).

In the R_1R_2 diagram of De la Roche et al. (1980), the unaltered dyke rocks cluster in the latite (monzonite) field near the trachyandesite (syenodiorite) and the latibasalt (monzogabbro) fields. The cluster of Hunnedalen East straddles the critical line of silica saturation whereas that of Hunnedalen West lies on the silica-saturated side.

Alteration

Unaltered dyke rocks appear chemically rather homogeneous, but altered marginal samples [Rog 389, 390, 398, 399; Rog 475, 483, 484] are clearly modified (Figs. 4a and 5a). The most obvious modification is a strong increase of LOI values in marginal samples, in both dyke and country rocks. Moreover, in altered dyke rock samples;
- SiO_2 is lower in both Hunnedalen East and West,
- TiO_2 is equal [Rog 390, 398, 484] or higher in samples containing titanite-rich veinlets [Rog 389, 399, 475, 483],
- FeO is equal, or somewhat lower in Hunnedalen East and equal, or slightly higher in Hunnedalen West,
- MgO is equal, or somewhat higher [Rog 399],
- CaO is equal to clearly lower [Rog 389, 390, 399, 484] but higher in one sample [Rog 483],
- Na_2O is equal to clearly lower [Rog 389, 399, 475, 483] but higher in Rog 484, containing albite-rich veinlets,

- K_2O is distinctly lower in all marginal samples, except Rog 389, where it is higher,
- Rb values follow K_2O: they are also lower in all marginal samples, but higher in Rog 389. K/Rb ratios of altered samples are in general lower than those of unaltered samples,
- Sr values are lower or equal.

The lower K_2O and Rb contents of the altered marginal samples of Hunnedalen East go in stride with higher K_2O values of the altered marginal amphibolite samples [Rog 490, 491]. However, exchange of elements mobilised during alteration between different lithologies of the country rocks (e.g. gneisses and amphibolites) seems stronger than exchange between country rocks and dyke rocks.

The Hughes Igneous Spectrum (Fig. 8; Hughes 1973) clearly demonstrates the modification of K_2O (and Na_2O) values. All unaltered dyke rocks plot closely together in the central part of the igneous spectrum, while altered samples scatter in the margins of the spectrum. The K_2O-enriched country rock amphibolites plot far outside the igneous spectrum, contrary to relatively fresh amphibolites.

Zonation

The fresh or nearly fresh central samples (+ or ++ in Table 1 and Figs. 4 & 5) show some chemical zonation in both sampled sections, which are no weak extensions of the alteration in the marginal samples. These variations involve in the central samples Rog 394 and Rog 479 (and to a lesser extent also Rog 395 and Rog 480) a clear increase in Al_2O_3, a minor
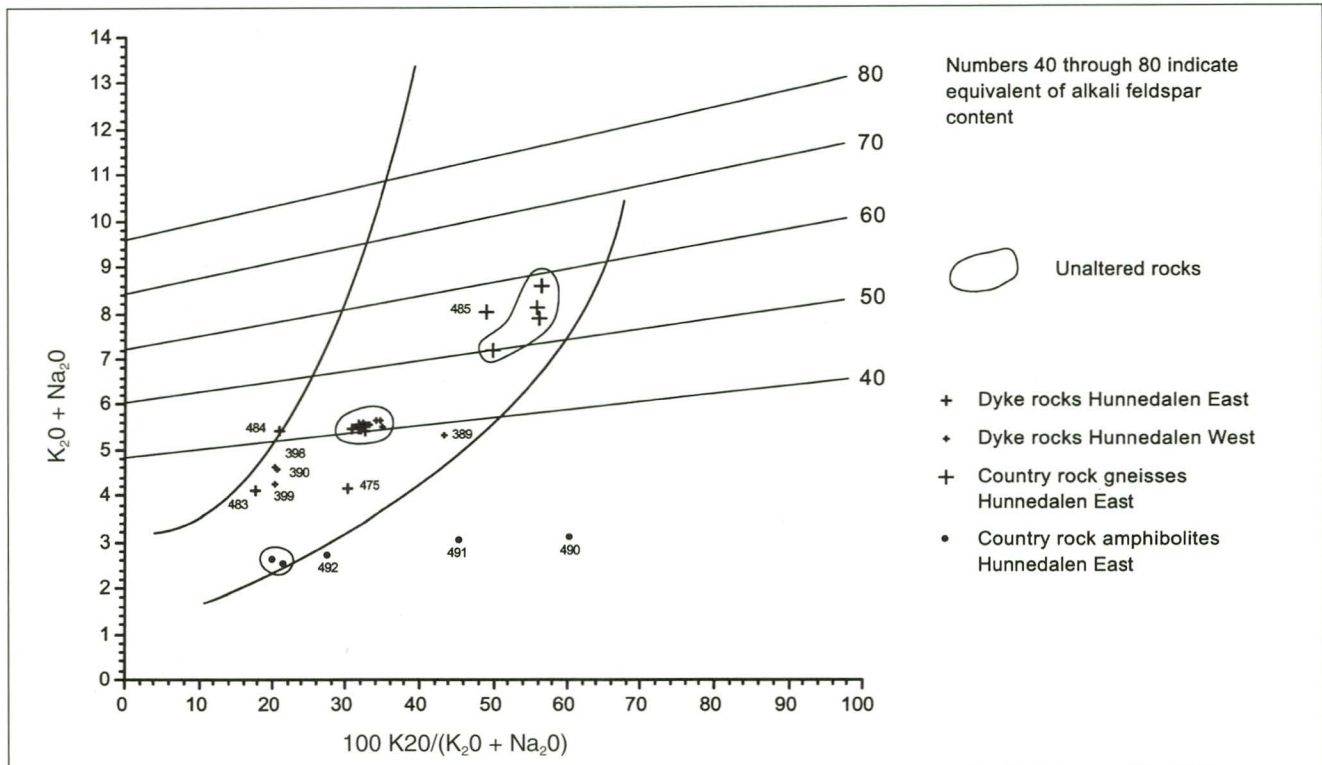
Fig. 8. Plot of the Hunnedalen rocks in relation to the Hughes Igneous Spectrum (Hughes 1973). The igneous spectrum is bounded by the two curved lines.

increase in MgO, CaO, Na_2O and Sr and a minor decrease in FeO, K_2O, P_2O_5, Ba, Rb and REE (Fig. 5a shows the plot of their Sm and Nd isotope dilution values). These variations in chemistry imply an increase in normative plagioclase, olivine and Mg# and a decrease in normative orthoclase and orthopyroxene in the central samples Rog 394 and Rog 479, while ilmenite and clinopyroxene remain approximately constant. Recalculated with biotite + quartz instead of olivine + orthoclase (Figs. 4b and 5b), the central samples exhibit an increase in normative plagioclase, biotite and quartz and a decrease in normative orthopyroxene and orthoclase.

Modal biotite contents (Figs. 4b and 5b) agree with calculated normative biotite contents in the most central samples, but deviate towards the marginal samples, suggesting that biotite (+ quartz) formed not only instead of olivine + orthoclase but also in place of orthopyroxene + orthoclase. The lower P_2O_5 values of these central samples show a correlation with lower values of REE, i.e., elements that are usually incorporated in apatite. These central samples also have slightly lower Zr values.

The zonation across the dyke, especially the higher contents of early formed plagioclase and higher Mg# in the centre of the dyke, indicates that the centre of the Hunnedalen dyke consists of more 'primitive' material than the more outer parts. Kalsbeek & Taylor (1986) reported a similar zonation from dykes in Greenland and explained it by a combination of: «Enrichment of early formed crystals in the central part of the dyke, and inhomogeneity of (the liquid fractionation of) the intruding magma, such that more fractionated magma was first intruded to form the dyke margins, follo-

wed by more primitive magma to form the dyke centers».

The relative low content of MgO (about 4 wt %), high content of FeO (about 12 wt %) — and therefore low Mg# (average 0.366) — indicate that the primary liquid from which the dyke magma formed underwent extensive fractionation before emplacement.

Trace elements

The dyke rocks of Hunnedalen East exhibit a consistent trace element signature. All plot in the field of within-plate basalt of classical geotectonic setting discrimination diagrams, for example in:

(a) the Ti-Zr-Y diagram (Pearce & Cann 1973, Fig. 9) and the Ti-Zr diagram (Pearce 1982),

(b) the Ti/Y-Zr/Y (Pearce & Gale 1977),

(c) the Zr/Y-Zr diagram (Pearce & Norry 1979),

(d) the Ti/Y-Nb/Y diagram (although in the tholeiitic subfield, rather than in the alkalic subfield of the within-plate field), the Cr-Y diagram and the Cr-Ce/Sr diagram (Pearce 1982),

(e) the Zr-Nb-Y diagram (Meschede 1986).

Moreover, in the La-Y-Nb diagram (Cabanis & Lecolle 1989) the samples of the Hunnedalen mafic dyke all plot in the field of continental basalt. Such a position in within-plate or continental fields is hardly surprising for dykes in the cratonised basement of southern Norway.

The fractionated character of the Hunnedalen monzonorites is also demonstrated by the enrichment of REE and especially of the LREE. The rather uniform contents of REE
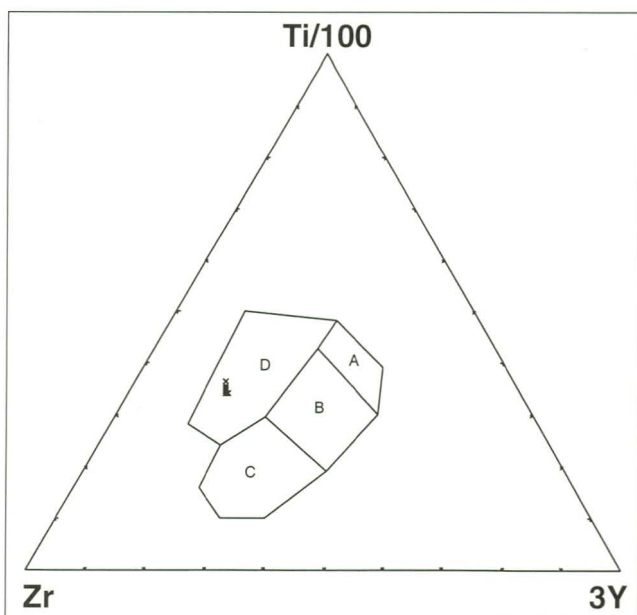
Fig. 9. Hunnedalen East dyke rocks in the Ti-Zr-Y discrimination diagram of Pearce & Cann (1973). (A) Island-arc tholeiites. (B) MORB, island-arc tholeiites and calc-alkali basalts. (C) Calc-alkali basalts. (D) Within-plate basalts.

Fig. 11. MORB-normalised spidergram of Hunnedalen East dyke rocks. Normalisation with average tholeiitic MORB (Pearce 1983). Primitive jotunites (shaded area) of the Egersund Igneous Complex (Vander Auwera et al. 1998) are shown for comparison.

(Table 2b, for Sm and Nd also Table 5) reflect the homogeneous character of the Hunnedalen monzonorites across the dyke, with slightly lower values for sample Rog 479 in the central part of the dyke. Σ REE (INA-values only) vary from 136 for the unaltered central dyke sample Rog 479 to 167 for the altered and strongly sheared marginal sample Rog 484. The other altered samples Rog 475 and 483 have Σ REE-values falling in the range of Σ REE-values for unaltered samples. Apparently, alteration did not significantly change the REE contents.

Chondrite-normalised (McDonough & Sun 1995) REE patterns are shown in Fig. 10. REE enrichments vary from 115-131 times chondritic values for La and from 11.6-14.7 times chondritic values for Yb. However, the LREE enrichment in altered dolerites differs slightly from those of unaltered rocks: $(La/Yb)_N$ is 9.1-10.2 (average 9.8) for unaltered rock

and 8.6-8.9, thus <9.0 for altered rocks, which may indicate a minor change in REE contents with alteration.

REE patterns show weak positive Eu anomalies. Eu/Eu* ratios (i.e., the ratio between the measured concentrations of Eu and the interpolated value between Sm and Tb) vary from 1.03 to 1.09, and average 1.07. A relationship with alteration is not shown. There is no clear correlation between the Eu/Eu* ratio and normative plagioclase content. In the chondrite-normalised REE pattern diagram for the Hunnedalen dykes (Fig. 10) the REE patterns of primitive and evolved jotunites of the Egersund Igneous Complex (according to data from Vander Auwera et al. 1998) are shown for comparison. The Hunnedalen REE patterns largely overlap with those of the primitive jotunites and are clearly lower than those of the evolved jotunites.

The distribution of incompatible elements is also homogeneous in all Hunnedalen dyke rocks (except for lower values of the LIL elements K, Rb and Ba in altered marginal samples Rog 475, 483 and 489), as is shown by an average MORB-normalised spidergram (Fig. 11; Pearce 1983). It shows - in unaltered rocks - markedly enriched values of the LILE (K, Rb, Ba, Th), lesser enriched values of LREE and moderately enriched values for Sr, Nd, P, Zr, Hf and Ti. Except for somewhat lower values for Th, Ta and Nb, this spidergram is most similar to the pattern of alkalic type of within-plate basalts (Pearce 1982) and distinctly different from spidergrams of basic rocks of other geotectonic settings. The spidergram of the unaltered Hunnedalen dyke rocks normalised to an estimated primordial mantle composition (Holm 1985) distinctly differs from those of 'initial rifting regimes'. Spidergram Fig. 11 shows a remarkable overlap of the Hunnedalen monzonorites with the primitive jotunites of the Egersund Igneous Complex (data from Vander Auwera et al. 1998).

Fig. 10. Chondrite-normalised (McDonough & Sun 1995) REE plot of the Hunnedalen East dyke rocks. Primitive and evolved jotunites (shaded areas) of the Egersund Igneous Complex (Vander Auwera et al. 1998) are shown for comparison.

Rb-Sr ages: results and discussion

Table 3 presents the results of Rb and Sr whole-rock analyses of the Hunnedalen East and West dyke rocks, the minerals of

Sample Nr.	Whole-rock /mineral	Rb [ppm wt]	Sr [ppm wt]	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$
Hunnedalen West: Dyke rocks					
Rog 389	whole-rock	60.1	389.3	0.4467 ± 22	0.711152 ± 19
Rog 390	whole-rock	38.5	478.1	0.2328 ± 12	0.708409 ± 23
Rog 391	whole-rock	46.8	484.7	0.2795 ± 14	0.707619 ± 12
Rog 392	whole-rock	49.1	493.0	0.2883 ± 14	0.707845 ± 14
Rog 393	whole-rock	47.5	515.6	0.2668 ± 13	0.707732 ± 16
Rog 394	whole-rock	44.0	558.7	0.2279 ± 12	0.707259 ± 25
Rog 395	whole-rock	45.1	529.4	0.2468 ± 12	0.707481 ± 22
Rog 396	whole-rock	49.5	510.3	0.2807 ± 14	0.708133 ± 24
Rog 398	whole-rock	35.6	477.1	0.2159 ± 11	0.708212 ± 20
Rog 399	whole-rock	34.7	388.8	0.2584 ± 13	0.709039 ± 35
Hunnedalen East: Dyke rocks					
Rog 475	whole-rock	28.7	433.3	0.1914 ± 10	0.707248 ± 40
Rog 476	whole-rock	35.9	484.1	0.2142 ± 11	0.706969 ± 10
Rog 477	whole-rock	36.1	492.3	0.2119 ± 11	0.706908 ± 65
Rog 478	whole-rock	35.6	498.4	0.2068 ± 10	0.706899 ± 15
Rog 479	whole-rock	34.4	519.0	0.1918 ± 10	0.706720 ± 40
Rog 480	whole-rock	36.6	506.9	0.2090 ± 10	0.706952 ± 14
Rog 481	whole-rock	36.0	505.6	0.2057 ± 10	0.706897 ± 07
Rog 482	whole-rock	38.5	491.9	0.2264 ± 11	0.707182 ± 14
Rog 483	whole-rock	20.5	452.4	0.1310 ± 10	0.707053 ± 21
Rog 484	whole-rock	24.4	357.9	0.1973 ± 10	0.709050 ± 13
Hunnedalen East: Country rocks					
Rog 485	whole-rock	85.7	221.5	1.1207 ± 56	0.725481 ± 16
Rog 486	whole-rock	163.3	209.3	2.2674 ± 113	0.751309 ± 20
Rog 487	whole-rock	139.1	211.1	1.9135 ± 96	0.744993 ± 21
Rog 488	whole-rock	165.2	193.2	2.4853 ± 124	0.756189 ± 16
Rog 489	whole-rock	156.7	243.1	1.8708 ± 94	0.743370 ± 11
Rog 490	whole-rock	79.3	332.9	0.6900 ± 35	0.716975 ± 15
Rog 491	whole-rock	46.3	378.9	0.3534 ± 18	0.712804 ± 10
Rog 492	whole-rock	18.5	241.3	0.2219 ± 11	0.708973 ± 11
Rog 493	whole-rock	13.8	237.8	0.1675 ± 8	0.708622 ± 24
Rog 494	whole-rock	11.2	227.6	0.1417 ± 7	0.708663 ± 28
Hunnedalen East: Minerals					
Rog 478	Bt	365.4	16	71.64 ± 36	1.55818 ± 72
Rog 479	Bt	360.9	13.3	86.16 ± 43	1.72644 ± 54
Rog 479	Pl	5.6	806.9	0.01991 ± 10	0.704651 ± 13
Rog 479	Opx	4.8	10.7	1.3110 ± 66	0.716682 ± 11
Rog 479	Cpx	12.5	41.5	0.8722 ± 44	0.713495 ± 10

Table 3. Whole-rock and mineral Rb-Sr data of Hunnedalen West and East dyke and country rocks.

Fig. 12. R-Sr mineral - whole-rock isochron plot of Hunnedalen East dyke rocks Rog 478 and Rog 479.

Hunnedalen East [Rog 478 Bt; Rog 479 Bt, Pl, Opx, Cpx] and Hunnedalen East country rocks. Analytical details are given in the Appendix.

The most significant Rb-Sr results are obtained from the mineral data. A Rb-Sr isochron plot of all minerals + their whole-rock data results in an errorchron age of 778 ± 93 Ma (n = 7), with $Sr_{(i)}$ = 0.70441 ± 0.00051 and MSWD = 329. The two biotites plot above and the two pyroxenes plot below the errorchron lineament. Apparently, not all minerals behaved as closed systems. The data points of the two pyroxenes (clinopyroxene and especially orthopyroxene) are most subject to small changes in Rb and/or Sr contents. Most probably this is the result of exsolution and/or minor alteration. Of the minerals present in the Hunnedalen dyke, orthopyroxe-ne is the mineral most sensitive to alteration.

An age calculation of the minerals [478 Bt, 479 Bt, Pl] + their whole-rocks - omitting the two pyroxenes - gives an isochron age of 834 ± 9 Ma, with $Sr_{(i)}$ = 0.70442 ± 0.00002 and MSWD = 0.594 (Fig. 12). The alteration of the pyroxenes will have influenced the whole-rock data in the case of an open system alteration. However, the results are not significantly changed if the two whole-rock data points are also omitted from the calculation. The remaining 478 Bt, 479 Bt and Pl define an isochron age of 833 ± 37 Ma, with Sr(i) = 0.70441 ± 0.00017 and MSWD = 0.671. It indicates that the system was closed for Rb and Sr on the scale of the rock samples.

Calculation from other combinations of minerals and their whole-rocks gives similar results. Most of these age results lean heavily on the two biotite data points, but Rog 479 whole-rock + plagioclase gives a comparable result (843 ± 18 Ma; Sr(i) = 0.70441 ± 0.00002).

Rb-Sr whole-rock ages show a considerable spread (Fig. 13). Due to the severely modified Rb and Sr contents of the altered marginal samples of the Hunnedalen dyke rocks (Table 2a,b, Figs. 4a and 5a), the large spread of Rb-Sr whole-rock ages is not at all surprising.

Unaltered samples of the central part of the dyke [Rog 391-396, Rog 476-488], on the other hand, show a reasonably linear arrangement, representing an errorchron age of 886 Ma, with a large uncertainty of ± 143 Ma. This large uncertainty is due to the restricted spread in $^{87}Rb/^{86}Sr$ ratios of these samples, 0.228-0.288 for Hunnedalen West and 0.192-0.226 for Hunnedalen East. The initial $^{87}Rb/^{86}Sr$ is 0.70429 ± 0.00051 with a MSWD of 21.4. Restrictions of the Rb-Sr whole-rock dating method for basic dykes was already discussed by Hanes (1987) in view of their low Rb contents and chemical homogeneity.

Different ages can be calculated from the slightly scatte-
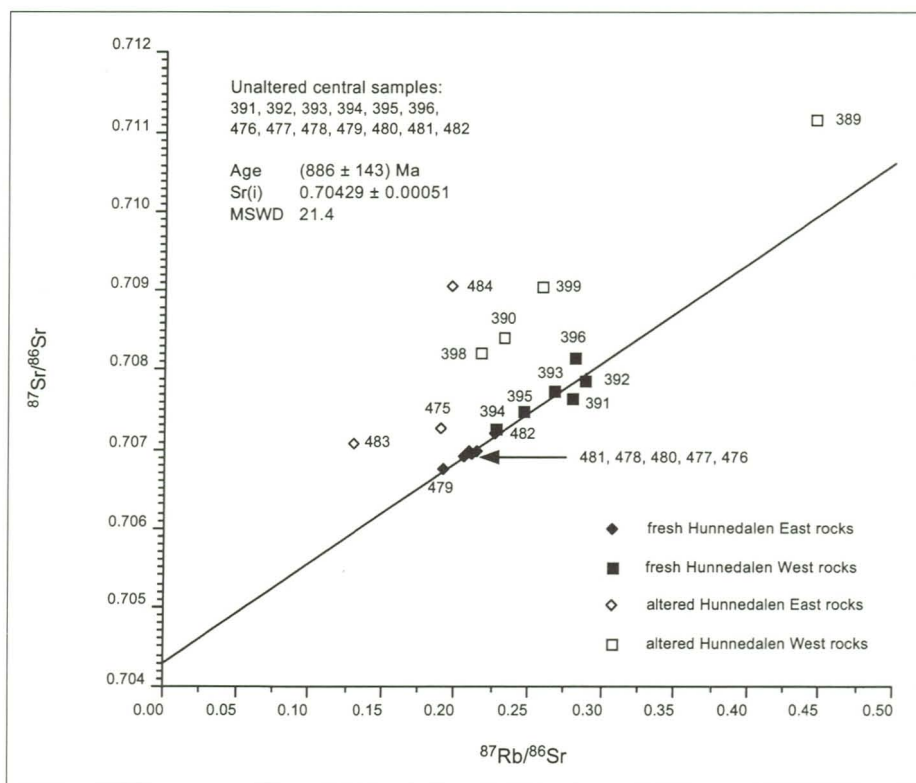
Fig. 13. Rb-Sr whole-rock errorchron plot of the Hunnedalen East and West dyke rocks.

red Rb-Sr data points of unaltered Hunnedalen dyke rocks, all with large errors and often unsatisfactory MSWD values. By omitting one sample, Rog 482, petrographically fresh but sampled close to the severely altered marginal samples Rog 483 and 484 - therefore possibly not strictly unaltered isotopically - the calculated whole-rock age for the Hunnedalen East rocks, drops from 941 ± 210 Ma to 739 ± 320 Ma.

Altered marginal dyke samples [Rog 389, 390, 398, 399; Rog 475, 483, 484] all plot above and - with the exception of the K and Rb enriched sample Rog 389 - to the left of the unaltered samples from the centre of the dyke (Fig. 13), indicating an increase of the $^{87}Sr/^{86}Sr$ ratio and/or a loss of Rb; or better, a loss of Rb greater than the loss of Sr, i.e. a reduction

Fig. 14.'Isochron cartoon' showing a hypothetical course of events for altered Hunnedalen dyke rocks, changing bulk Rb and/or Sr contents at time t = 1 between the initial emplacement t=0 and present situation t = 2 without changing the $^{87}Sr/^{86}Sr$ ratio.
Situation A with decreasing Rb/Sr ratio; Situation B with equal Rb/Sr ratio; Situation C with increasing Rb/Sr ratio.

in the Rb/Sr ratio. It should be remembered that Sr is reduced in all altered Hunnedalen dyke rocks, while Rb is reduced in all altered dyke rocks except Rog 389 (Table 2a,2b, Figs. 4a and 5a).

An 'isochron cartoon' (Fig. 14) visualises the behaviour of rocks that underwent changes of Rb and Sr at some time t = 1 between the initial emplacement time, t = 0, and the present situation, t = 2, without changing their $^{87}Sr/^{86}Sr$ ratio. There are 3 different situations possible:

– Situation A shows the effects of a decrease of the Rb/Sr ratio at time t = 1 exemplified by Rog 390, 398 and 475 and especially Rog 483, with a 1.5 [Rog 475] to 4.5 times [Rog 483] greater loss of Rb than of Sr (Table 4). The positions of data points of such rocks move to the left in an isochron plot at time t = 1, and will be found above the isochron at time t = 2.

– In situation B the changes of Rb and Sr are approximately equal [exemplified by Rog 399 and 484 (Table 4) at time t = 1]. The positions of data points of such rocks will not change much in an isochron plot at time t = 1 and they will perch on the isochron at time t = 2.

– Situation C shows the effects of an increase in the Rb/Sr ratio at time t = 1 [exemplified by Rog 389, with an increased Rb content accompanied by loss of Sr]. Data points of such rocks move to the right in an isochron plot at time t = 1 and occupy a place below the isochron at time t = 2.

The actual position of the altered samples exemplified by the situations B and C, i.e., Rog 389, 399 and 484, above the isochron is contrary to what was expected. It implies that their present position cannot be explained by the analysed changes in Rb and Sr contents, but that the $^{87}Sr/^{86}Sr$ ratio

Sample Nr.	Petro-graphic quality	Measured $^{87}Rb/^{86}Sr$	Rb ppm	average	Gain/loss %	Sr ppm	average	Gain/loss %	Recalculated $^{87}Rb/^{86}Sr$ for gain or loss of Rb and Sr
Rog 389	—	0.4467	60.1		+ 27.9	389.3		- 24.5	0.2587
Rog 390	-	0.2328	38.5		- 18.1	478.1		-7.2	0.2565
Rog 391	+	0.2795	46.8			484.7			
Rog 392	+	0.2883	49.1			493.0			
Rog 393	++	0.2668	47.5	47.0		515.6	515.3		
Rog 394	++	0.2279	44.0	(n=6)		558.7	(n=6)		
Rog 395	++	0.2468	45.1			529.4			
Rog 396	+	0.2807	49.5			510.3			
Rog 398	—	0.2159	35.6		- 24.2	477.1		- 7.4	0.2497
Rog 399	—	0.2584	34.7		- 26.2	388.8		-24.5	0.2619
Rog 475	-	0.1914	28.7		- 20.7	433.3		- 13.3	0.2039
Rog 476	+	0.2142	35.9			484.1			
Rog 477	++	0.2119	36.1			492.3			
Rog 478	++	0.2068	35.6	36.2		498.4	499.7		
Rog 479	++	0.1918	34.4	(n=7)		519.0	(n=7)		
Rog 480	++	0.2090	36.6			506.9			
Rog 481	++	0.2057	36.0			505.6			
Rog 482	++	0.2264	38.5			491.9			
Rog 483	—	0.1310	20.5		- 43.3	452.4		- 9.5	0.1714
Rog 484	—	0.1973	24.4		- 32.5	357.9		- 28.4	0.2036

Table 4. Calculated loss or gain of Rb and Sr (in %) in altered dyke rocks of Hunnedalen West and East compared to the average of unaltered dyke rocks.
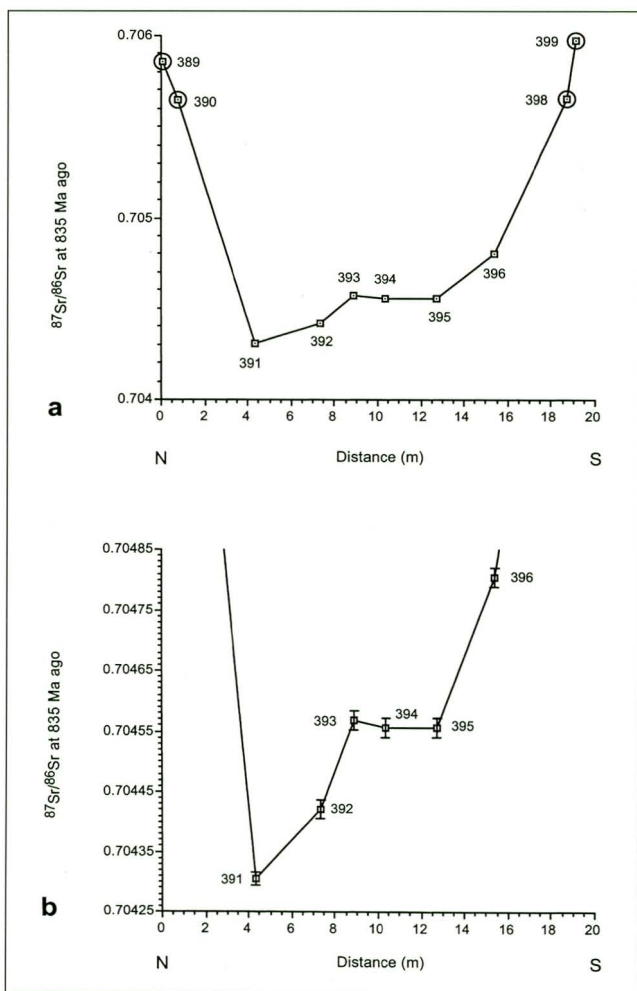
Fig. 15. (a) Plot of $^{87}Sr/^{86}Sr$ values (at 830 Ma ago) of the Hunnedalen West monzonorites versus their positions within the dyke. Altered rocks are circled. (b) Plot of $^{87}Sr/^{86}Sr$ values (at 830 Ma ago) of the Hunnedalen West monzonorites versus their positions within the dyke, with an 'extended' $^{87}Sr/^{86}Sr$ scale of 0.70425-0.70485.
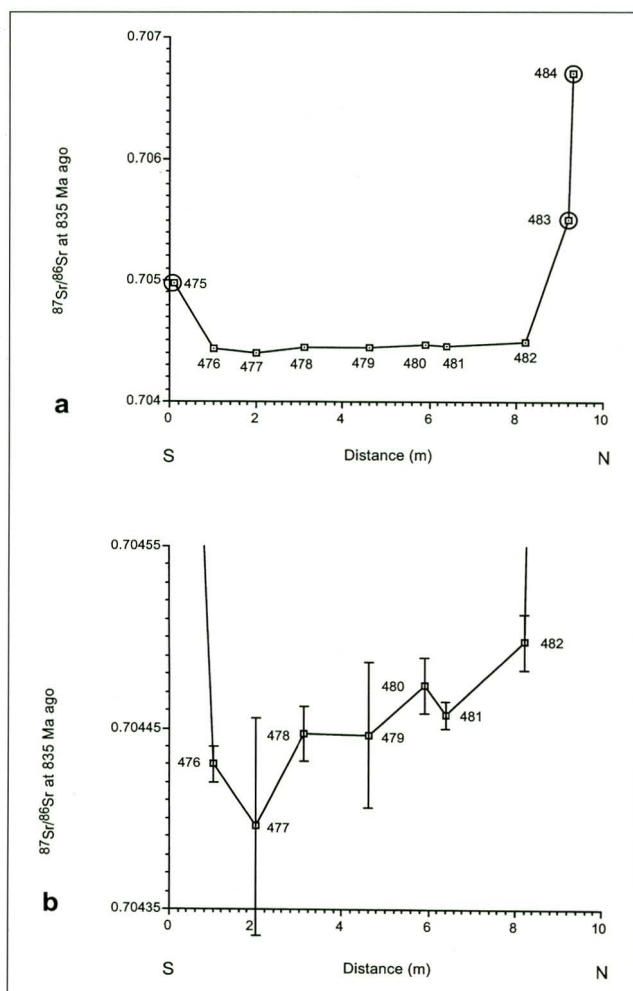
Fig. 16. (a) Plot of $^{87}Sr/^{86}Sr$ values (at 835 Ma ago) of the Hunnedalen East monzonorites versus their positions within the dyke. Altered rocks are circled. (b) Plot of $^{87}Sr/^{86}Sr$ values (at 835 Ma ago) of the Hunnedalen East monzonorites versus their positions within the dyke, with an 'extended' $^{87}Sr/^{86}Sr$ scale of 0.70435-0.70455.
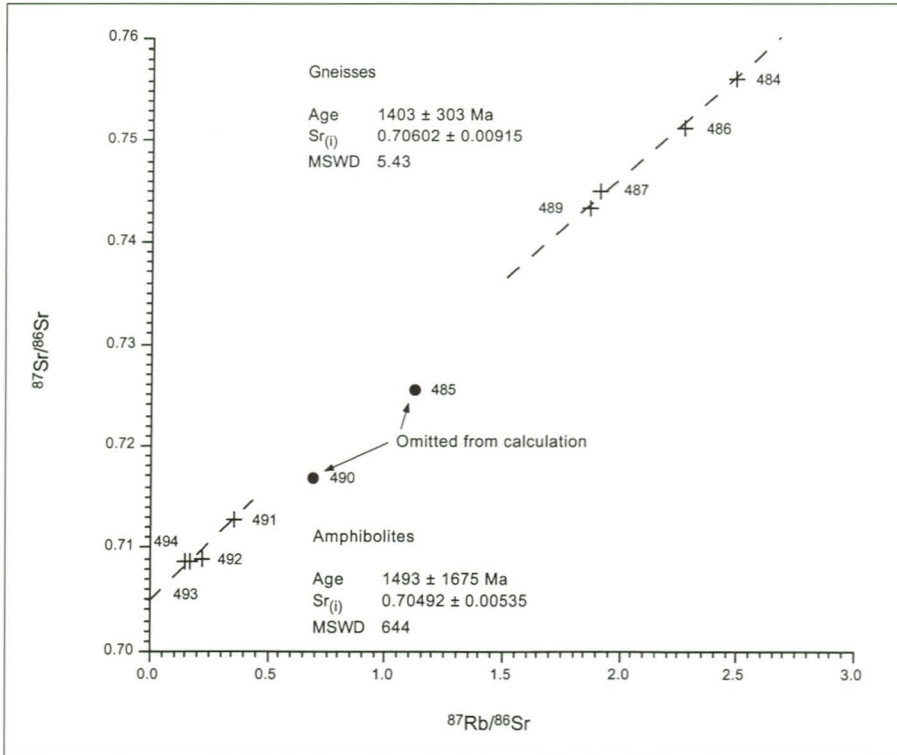
Fig. 17. Rb-Sr errorchron plot of Hunnedalen East country rocks.

must have been changed also, i.e., increased during alteration. This means that radiogenic ^{87}Sr must have operated differently than non-radiogenic ^{86}Sr. A possible explanation for such contrasting behaviour of Sr isotopes is the unequal behaviour during alteration of minerals containing predominantly non-radiogenic ^{86}Sr and those containing predominantly radiogenic ^{87}Sr.

Ca-bearing minerals, plagioclase and clinopyroxene, mainly incorporate non-radiogenic ^{86}Sr. During alteration, i.e., the replacement of clinopyroxene by actinolitic amphibole plus chlorite and the partial albitisation of plagioclase, these minerals have lost - simultaneously with Ca (Table 2a,b; Figs. 4a and 5a) - a considerable amount of their ^{86}Sr. On the other hand, radiogenic ^{87}Sr is enriched in the Rb (and K)-bearing minerals orthoclase and biotite. Some minor, interstitial orthoclase is still present in the altered samples, but the amount of orthoclase is less than in the unaltered samples. Chlorite, with abundant tiny lenses, veinlets or rims of titanite, partly or completely replaces biotite. If the radiogenic ^{87}Sr, freed by chloritisation, is incorporated in the Ca-mineral titanite, it is not (completely) lost during alteration. In such a scenario, a smaller loss of ^{87}Sr compared to that of ^{86}Sr is imaginable. Such an explanation seems more plausible than the assumption of an enrichment of ^{87}Sr liberated from the neighbouring country rock gneisses during alteration.

An increase in the $^{87}Sr/^{86}Sr$ values of altered samples is also shown when these values, recalculated to their original ratios at the assumed 'best age' of 835 Ma (see Sm—Nd ages), are plotted against their positions within the dyke (Figs. 15a and 16a). All the altered samples have clearly higher $Sr_{(i)}$ values than the unaltered samples; the latter look rat-

her uniform, especially those of Hunnedalen East (Fig. 16a). However, from an extended $Sr_{(i)}$ scale (Figs. 15b and 16b) it is clear that there is a variation in the $Sr_{(i)}$ values of unaltered samples which is larger than the (present) analytical uncertainty. Apparently, also the samples looking 'fresh' underwent some minor changes in their Sr-isotope systematics during alteration.

The Rb and K enrichment of the altered marginal sample Rog 389 seems related to the replacement of the plagioclase crystal cores by K-feldspar (see Petrography), probably due to K freed by alteration of country rock gneisses.

The country rocks at Hunnedalen East, 5 biotite-garnet gneisses [Rog 485-489], 5 pyroxene amphibolites [Rog 490-494], altered and unaltered, have been analysed for Rb and Sr isotopes (Table 3). By omitting the strongly altered gneiss Rog 485 and amphibolite Rog 490, age calculations produce a 'Gothian signal', 1403 ± 303 Ma and 1493 ± 1675, respectively (Fig. 17).

Sm-Nd ages: results and discussion

Table 5 presents the results of Sm and Nd analyses of Hunnedalen East whole-rocks [Rog 475-484] and minerals [Rog 478: Bt; Rog 479: Bt, Pl, Opx, Cpx, Ilm, Ap]. Analytical details are given in the Appendix. Apparently, Sm and Nd data (contents and isotope ratios) have not significantly been changed by the process of rock alteration.

The best results are produced by mineral isochron calculations. They are consistent with the best Rb-Sr mineral isochron calculation. By omitting the most deviating plagioclase data point from the calculation (Fig. 18), a good isochron with an

Sample Nr.	Whole-rock /Mineral	Sm (ppm Wt)	Nd (ppm Wt)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	ε_{Nd} (835 Ma)
Rog 475	Whole-rock	9.980	43.766	0.13786 ± 13	0.512395 ± 7	1.51
Rog 476	Whole-rock	9.807	43.747	0.13552 ± 11	0.512391 ± 2	1.68
Rog 477	Whole-rock	9.979	44.561	0.13539 ± 32	0.512382 ± 6	1.52
Rog 478	Whole-rock	9.809	44.124	0.13439 ± 10	0.512388 ± 5	1.74
Rog 479	Whole-rock	8.927	39.516	0.13658 ± 12	0.512404 ± 5	1.82
Rog 480	Whole-rock	9.521	42.298	0.13608 ± 14	0.512397 ± 7	1.74
Rog 481	Whole-rock	9.546	42.372	0.13621 ± 21	0.512385 ± 7	1.49
Rog 482	Whole-rock	9.747	43.355	0.13592 ± 15	0.512390 ± 6	1.62
Rog 483	Whole-rock	9.702	42.670	0.13746 ± 27	0.512396 ± 5	1.57
Rog 484	Whole-rock	10.213	45.202	0.13660 ± 12	0.512387 ± 9	1.48
Rog 478	Bt	0.652	3.095	0.12735 ± 10	0.512347 ± 10	1.69
Rog 479	Bt	0.316	1.405	0.13615 ± 25	0.512385 ± 11	1.50
Rog 479	Ilm	2.055	14.431	0.08611 ± 22	0.512125 ± 6	1.74
Rog 479	Opx	5.989	39.658	0.09130 ± 8	0.512167 ± 8	2.01
Rog 479	Pl	1.574	9.251	0.10285 ± 10	0.512252 ± 6	2.44
Rog 479	Cpx	8.167	40.280	0.12257 ± 12	0.512342 ± 9	2.10
Rog 479	Ap	306.8	998.0	0.185859 ± 35	0.512687 ± 11	2.11

Table 5. Whole-rock and mineral Sm and Nd data of Hunnedalen East dyke rocks.

age of 835 ± 47 Ma and a MSWD value close to 1 is obtained.

Sm-Nd age calculations of whole-rock Hunnedalen dyke samples, either of unaltered rocks, or unaltered combined with altered rocks, reveal quite inconsistent results, with enormous errors in age and $Nd_{(i)}$ values, due to the 30 times smaller spread in the $^{147}Sm/^{144}Nd$ ratio of whole-rocks compared to minerals.

The $^{143}Nd/^{144}Nd$ ratios of individual samples from Hunnedalen East, recalculated for an assumed 'age' of 835 Ma, plotted against their position within the dyke show a striking regular sinuous configuration (Fig. 19) with variations larger than (the present) analytical uncertainty. A similar pattern is produced when present-day $^{143}Nd/^{144}Nd$ ratio values are plotted against the positions of the samples across the dyke.

The $Nd_{(i)}$ pattern of Fig. 19a shows no concordance with the variations in chemistry and mineralogy across the dyke (Fig. 5a,b), the total Sm (Fig. 19b), the total Nd (Fig. 19c) and the Sm/Nd bulk ratios (Fig. 19d). If we assume a uniform $Nd_{(i)}$ of 0.51166 (Fig. 18), the variation in $^{143}Nd/^{144}Nd$ demands an age difference of at least 19 Ma for the crystallisation of the more primitive, earlier formed crystals as in Rog 479 compared to the more evolved, later crystallised material as in Rog 477 and 481. If the assumption of a uniform $Nd_{(i)}$ ratio is true, it would imply the intrusion of a crystal mush, with only a limited amount of liquid, at least for the central sample Rog 479. In the case of a higher liquid/crystal ratio during intrusion, the difference in age between first formed, more primitive crystals and younger, more evolved material should be considerably larger than 19 Ma.

On the other hand, crystallisation of all dyke material within a relatively short period of time implies a non-uniform but symmetrical distribution of $Nd_{(i)}$ values across the dyke, possibly due to flow differentiation or intrusion of a non-homogeneous magma. Inhomogeneity of $Nd_{(i)}$ values can also be explained by some (selective?) contamination or by (minor) metasomatic changes of the Nd-isotope systematics during alteration. In both options, however, the regularity and symmetry of the $Nd_{(i)}$ pattern across the dyke seem difficult to explain.

The positive initial ε_{Nd} values of the Hunnedalen dyke rocks (Table 5) indicate that these rocks were derived from a
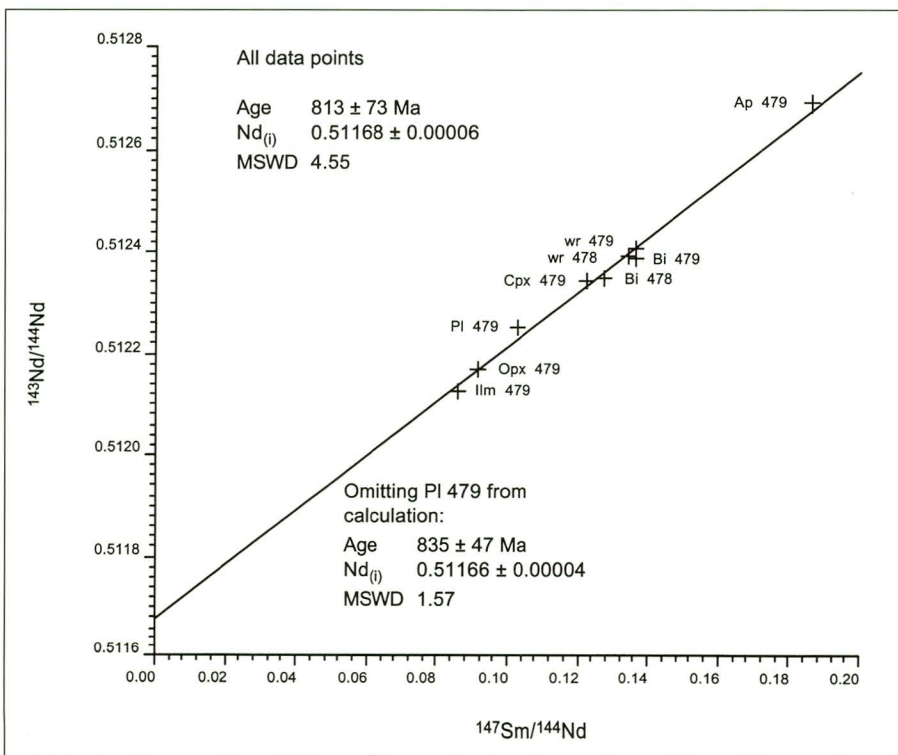
Fig. 18. Nd mineral - whole-rock isochron plot of the Hunnedalen East dyke rocks Rog 478 and Rog 479.
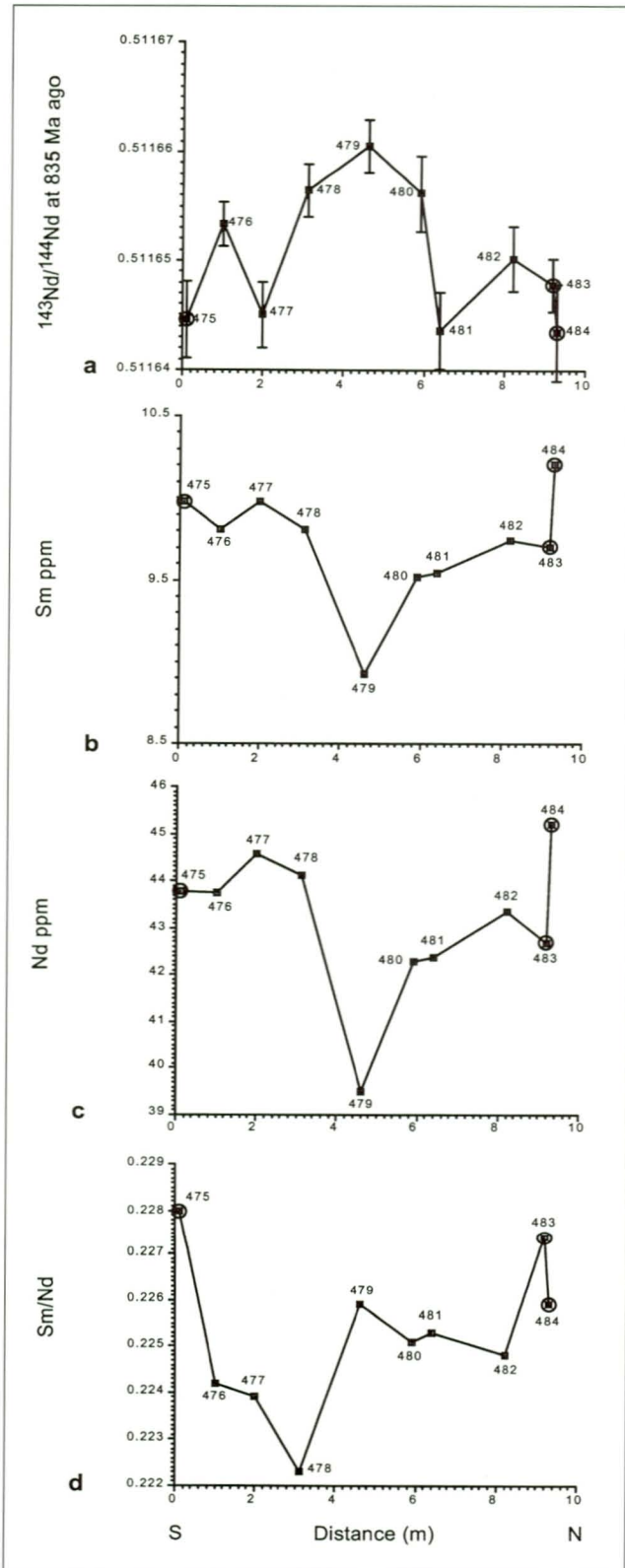
Fig. 19. (a) Plot of $^{143}Nd/^{144}Nd$ values (at 838 Ma ago) of the Hunnedalen East monzonorites versus their positions within the dyke. Altered rocks are circled. (b) Plot of the Sm contents (ID values) of the Hunnedalen East monzonorites versus their positions within the dyke. Altered rocks are circled. (c) Plot of the Nd contents (ID values) of the Hunnedalen East monzonorites versus their positions within the dyke. Altered rocks are circled. (d) Plot of the Sm/Nd bulk ratios (ID values) of the Hunnedalen East monzonorites versus their positions within the dyke. Altered rocks are circled.

depleted source. The variation in $\mathcal{E}_{Nd}$ values of the rather homogeneous Hunnedalen dykes is much smaller compared to those of many other Rogaland rock series (Fig. 20). For instance, the $\mathcal{E}_{Nd}$ of pyroxene-poor and pyroxene-rich anorthosites of the Håland anorthosite vary from + 3.5 and 4.0 to 1.9 and 0.0, respectively (t = 1 Ga, Demaiffe et al. 1986); the $\mathcal{E}_{Nd}$ of 3 anorthosites from one locality in the Håland anorthosite varies from 2.1 to 4.4, while the $\mathcal{E}_{Nd}$ of 4 leuconorites, sampled about 50 m from the anorthosites varies from - 0.5 to + 2.3 (t = 1.05 Ga, Menuge 1988); the $\mathcal{E}_{Nd}$ of augen gneisses varies from - 0.4 to + 1.3 (t = 1.05 Ga, Bingen et al. 1993); the $\mathcal{E}_{Nd}$ values of one megacyclic unit of the Bjerkreim-Sokndal Layered Intrusion varies from + 1.0 to + 2.1 (t = 0.955 Ga, Nielsen et al. 1996).

Menuge (1988) proposed a model in which the anorthosites and leuconorites of the Håland anorthosite were derived by major crustal contamination of, and fractional crystallisation from, a picritic magma derived from isotopically depleted mantle. Both Menuge's anorthosites and leuconorites appear chemically and mineralogically rather uniform rocks, but show large $\mathcal{E}_{Nd}$ variations. The anorthosites contain 90-95 % plagioclase, the remainder being chlorite, sericite and ore minerals, the leuconorites comprise 80-90 % plagioclase, with orthopyroxene partly altered to chlorite and minor sericite and opagues. The high Sr concentrations and low Nd concentrations of anorthosites and leuconorites preclude that the difference in initial Sr and Nd isotope ratios between these rocks could be explained by bulk assimilations of any known country rocks (Menuge 1988).

After variable assimilation of pre-existing crust by a picritic, depleted mantle-derived, parental magma near the base of the crust and fractional crystallisation of mafic minerals (Ol, Opx, Cpx), Sr and Nd concentrate in a residual Al-rich magma. Plagioclase starts to crystallise from this relatively light Al-rich residual magma rising diapirically through the lower crust, while contamination of the residual liquid continued, causing the Sr isotope ratio to rise and the Nd isotope ratio to fall. The residual magma continued to exchange Sr and Nd with plagioclase crystals, probably continuing over a geologically significant time span. Due to the high Sr-concentrations and low Nd-concentrations of plagioclase relative to the magma, the effect of this interaction was negligible on Sr isotope ratios but substantial on Nd isotope ratios. The large variation in initial $\mathcal{E}_{Nd}$ between different samples of anorthosite and of leuconorite therefore resulted from variable proportions of intercumulus liquid and variation in the amount of Nd exchanged between crystals and residual magma prior to separation of crystals from magma (Menuge 1988). Thus, Menuge's model is based on the interaction of plagioclase crystals with increasingly contaminated residual liquids and therefore especially true for plagioclase-dominated rocks like anorthosite and leuconorite.

Another, more general model is that proposed by De Haas (1997a,b) to explain the variable Sr and Nd isotope composition and variable LILE and LREE enrichment of
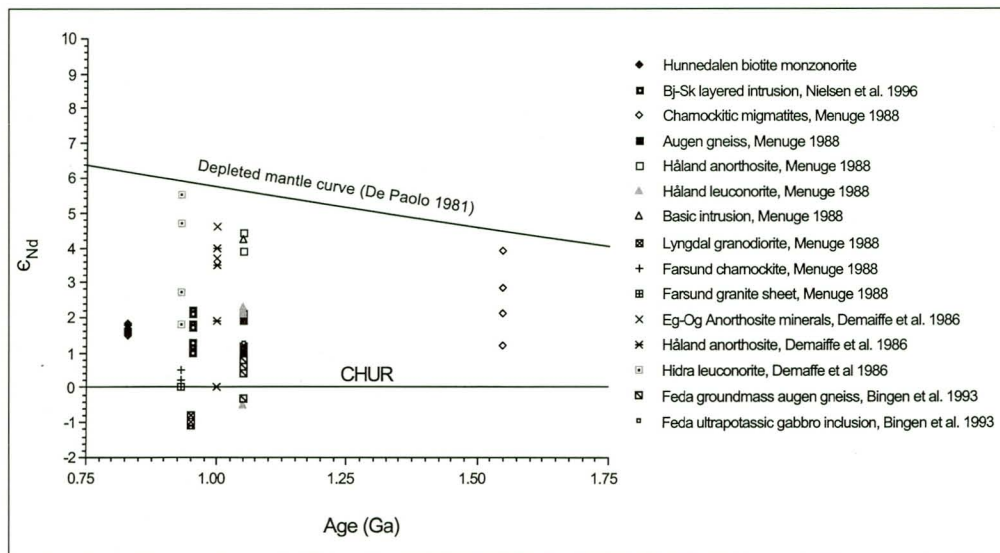
Fig. 20. Sm-Nd evolution diagram, showing (Nd versus age for the Hunnedalen dyke rocks and other magmatic rocks of SW Norway.

Proterozoic coronitic gabbros of the Bamble Sector, southern Norway. Also, De Haas rejected the idea of bulk assimilation of continental crust. His model involves the mixing of depleted asthenosphere-derived magmas with variable but, in general, small amounts (1-3%) of ultrapotassic/lamproitic melts derived from a metasomatically altered lithospheric mantle (below the Proterozoic Bamble continental crust).

K-Ar ages: results and discussion

Dyke rocks

Table 6 gives the analytical K-Ar data of the sieved whole-rock fractions and the two biotites of the Hunnedalen mafic dyke. Analytical details are given in the Appendix.

The calculated K-Ar ages of the sieved whole-rock fractions show a considerable spread, varying from 603 to 960 Ma. This agrees with calculated K-Ar ages (recalculated with the I.U.G.S recommended constants) - 570, 655, 848 and 940 Ma - of 4 sieved whole-rock fractions of the Hunnedalen dyke published by Versteeve (1975).

The large spread of the sieved whole-rock K-Ar ages is not at all surprising in view of the severely modified K-contents of the altered marginal samples. Unaltered samples from the centre of the dyke - Rog 476-482 - exhibit far less spread in their calculated ages: i.e., 660-763 Ma, average 715 ± 30 Ma (n = 15). Nevertheless, this age is much lower than the best age estimates - about 835 Ma - obtained by the Rb-Sr and Sm-Nd methods.

Compared with the calculated ^{40}Ar contents at 835 Ma, the ^{40}Ar loss of petrographically unaltered Hunnedalen mafic dyke rocks varies from 10 to 31 %, with an average of 19.3 %. Loss of ^{40}Ar due to the grinding process is improbable, because loss of ^{40}Ar starts with grinding to grain sizes < 100 μm (e.g. Gentner & Kley 1957) and the Hunnedalen rocks were measured on 125-250 (m sieve fractions.

K-Ar ages of 2 biotite fractions vary from 877-910 Ma

Sample Nr.	Rock/mineral	K [% Wt]	^{40}Ar* [ppb]	calculated K-Ar age Ma
Rog 475	Monzonorite	1.180	70.930	708 ± 8
			69.613	697 ± 8
Rog 476	Monzonorite	1.427	88.076	724 ± 8
			90.819	742 ± 8
			93.965	763 ± 8
Rog 477	Monzonorite	1.361	81.050	701 ± 8
			78.163	682 ± 8
Rog 478	Monzonorite	1.385	77.760	669 ± 8
			76.452	660 ± 8
Rog 479	Monzonorite	1.356	82.391	714 ± 9
			84.460	729 ± 9
Rog 480	Monzonorite	1.422	90.948	745 ± 9
			91.612	750 ± 9
Rog 481	Monzonorite	1.406	85.201	713 ± 8
		1.398	82.516	698 ± 8
Rog 482	Monzonorite	1.468	88.562	710 ± 8
			89.891	719 ± 8
Rog 483	Monzonorite	0.574	46.730	905 ± 10
			50.424	960 ± 10
			48.793	936 ± 10
Rog 484	Monzonorite	0.986	44.503	603 ± 7
			44.609	605 ± 7
Rog 485	Biotite garnet gneiss	2.928	137.96	577 ± 7
			129.45	546 ± 7
			133.07	630 ± 8
Rog 486	Biotite garnet gneiss	3.333	191.06	681 ± 8
			191.82	683 ± 8
Rog 490	Pyroxene amphibolite	1.611	86.608	645 ± 8
			86.717	646 ± 8
Rog 491	Pyroxene amphibolite	1.126	61.549	654 ± 8
			61.410	653 ± 8
Rog 478	Biotite	7.599	6157.3	901 ± 10
			6138.3	899 ± 10
Rog 479	Biotite	7.677	6009.5	877 ± 10
			6175.1	896 ± 10

Table 6. Whole-rock and biotite K-Ar data (sieved fractions 125-250 mm) of Hunnedalen East dyke and country rocks.

with an average of 893 Ma. These ages are distinctly higher than K-Ar whole-rock ages and the best Rb-Sr and Sm-Nd

age estimates. The K-minerals orthoclase and biotite are the most important source of radiogenic Ar in the Hunnedalen dyke. Probably orthoclase lost (part of) its ^{40}Ar, because K-feldspar looses Ar by diffusion already at ambient temperatures (e.g. Dickin 1995).

The discrepancy between the whole-rock and biotite K-Ar ages might be explained as follows: During intrusion of the Hunnedalen dyke magma ^{40}Ar is set free in the heated country rocks. The released ^{40}Ar moves partly into the dyke, where it becomes incorporated in all minerals when they cool down below their Ar-blocking temperatures T_{bl}, as in the Avanavero dolerite in Surinam (Hebeda et al. 1973). Dalrymple et al. (1975) also showed convincingly that Jurassic dolerite dykes incorporated excess ^{40}Ar from Precambrian basement gneisses in Liberia and not when they transected K-poor Palaeozoic, sedimentary rocks. The experiments of Karpinskaya (1961) and Karpinskaya et al. (1961) had earlier showed that muscovite subjected to argon pressures at high temperatures incorporated significant amounts of argon.

Alteration of the margins of the Hunnedalen dykes and their country rocks, occurred most probably during the M4 phase of metamorphism, related to the Caledonian orogeny, c. 400 Ma ago. The alteration reached prehnite-pumpellyite to lowest greenschist-facies conditions, indicative of regional temperatures of 300 to 350°C. This temperature remains below the Ar-blocking temperature of biotite, T_{bl}^{Bt} being 350-400°C (Berger & York 1981), or 345°C for a biotite of composition Ann 56 with a cooling rate of 100°C/Ma (Harrison et al. 1985). The cooling rate of the Hunnedalen dykes certainly was higher than 100°C/Ma. It implies that the excess Ar may well have remained incorporated in the biotite. This probably holds too for pyroxenes. On the other hand, regional temperatures of 300-350°C seem higher than the Ar-blocking temperatures of plagioclase and K-feldspar - T_{bl}^{Pl} and T_{bl}^{Kfs} being strongly dependent on composition and microstructure (Hanes 1991), scattering from about 300°C to less than 100°C - implying loss of excess Ar and pre-M4 produced Ar from plagioclase and K-feldspar. In this scenario, the Hunnedalen dykes will indeed show lower whole-rock K-Ar ages, in between the intrusion age of about 835 Ma and the Caledonian M4 age of c. 400 Ma, but higher biotite (and pyroxene?) K-Ar ages, in accordance with calculated ages.

Of the altered rocks, one sample [Rog 475] has an apparent K-Ar age in the range of the unaltered samples, while the other altered rocks have apparent K-Ar ages higher [brecciated Rog 483, with lowest K$_2$O content !] or lower [strongly sheared Rog 484] than the average age of the unaltered samples. Obviously, the altered rocks suffered loss or gain of Ar independently of the loss or gain of K.

The severely altered marginal dyke sample Rog 483 shows a considerable reduction in both Ar and K and it yields 'high' calculated K-Ar ages between 905 and 960 Ma (average 933 Ma, n = 3). Apparently, K-depletion, in this sample, is even stronger than Ar depletion.

Country rocks

The calculated whole-rock K-Ar ages of sieved country rock fractions [biotite-garnet gneisses Rog 485 (severely altered), 486 (nearly fresh) and pyroxene amphibolites Rog 490, 491 (both altered)] vary from 546 to 683 Ma (Table 8), again indicating 'low' ages that are due to a considerable loss of Ar.

Discussion

The chemistry of the Hunnedalen monzonorites compares well with the chemical characteristics of other monzonoritic dykes and intrusions in Rogaland/Vest Agder, as presented by Duchesne et al. (1989), Duchesne (1990) and Vander Auwera et al. (1998). Especially the chilled margins of the Hidra monzonorite and those of the Bjerkreim-Sokndal layered intrusion at Tjørn (HITJ or HT in Duchesne's tables and figures, = primitive jotunites of Vander Auwera et al. (1998)) are remarkably similar and overlap in major and trace elements, element ratios, REE patterns (Fig. 10), Eu/Eu*, Sr$_{(i)}$ value and on average MORB-normalised (spidergram Fig. 11) or chondrite-normalised, trace element abundance diagrams. The troughs at Th and Sr in the chondrite-normalised spidergram of Hunnedalen are also seen in Thompson spidergrams of most other monzonorites from Rogaland (Duchesne et al. 1989). The troughs at Sr presumably result from fractional crystallisation of plagioclase (Thompson et al. 1983). Plagioclase occurs as phenocrysts in the chilled margins of the Hunnedalen dykes and is argued to have crystallised first. The Hunnedalen monzonorites plot in, or close to, the HT field in AFM and Harker diagrams given by Duchesne et al. (1989), and Duchesne (1990) and close to the primitive jotunites, or in between primitive and evolved jotunites in variation diagrams presented by Vander Auwera et al. (1998).

The Hunnedalen monzonorites are only slightly higher in SiO$_2$, Na$_2$O, K$_2$O, Zr and (La/Yb)$_N$ and slightly lower in TiO$_2$, FeO$_{tot}$, MgO, P$_2$O$_5$, Ba, K/Rb and Mg# than the average of 7 HITJ rocks, = primitive jotunites of Vander Auwera et al. (1998). This is indicative of slightly increased fractionation of the Hunnedalen monzonorites compared to the average HITJ rock. This chemical similarity supports a common origin of the Hunnedalen dykes and other monzonorites of the Rogaland/Vest Agder region.

The Hunnedalen Dyke Swarm crops out 10 to 75 km north to northeast of the monzonorites of the large Egersund Igneous Complex. The southwestern termination of some Hunnedalen dykes occurring only c. 10 km from the Bjerkreim-Sokndal Layered Complex raises the possibility of a lateral SW to NE intrusion trend. However, the variation in composition of individual dykes given by Drent (1982), and that of the two sampled profiles of the one dyke studied here is insufficient to permit conclusive arguments in favour of such a scenario of lateral emplacement. Moreover, the position of the Hunnedalen dykes in Thompson's (1982) normative (Ne-)Di-Hy-Ol(-Qtz) diagram

(Fig. 6) is indicative of a considerable vertical intrusion component.

Besides a difference in location, there is also a difference in age between the monzonorites of Hunnedalen (c. 835 Ma) and those of the Egersund Igneous Complex (920-930 Ma, similar to the ages of massif-type anorthosites; Schärer et al. 1996, Vander Auwera et al. 1998). This time gap raises the question of the meaning of the age of 835 Ma for the Hunnedalen dykes: the time of intrusion or the time of metamorphic resetting. The age of 835 Ma is younger than the regional biotite Rb-Sr and K-Ar ages, indicating temperatures below biotite Rb-Sr and K-Ar blocking temperatures, i.e. below c. 350°C at 835 Ma ago.

Under such low temperature conditions it is extremely unlikely that both the Rb-Sr and the Sm-Nd system - of rocks with fresh appearance and clear igneous texture - were reset, producing equal ages defined by good linear alignments of most of the constituent magmatic minerals. Moreover, the relatively low $Sr_{(i)}$ value of about 0.7043, even slightly lower than for other Rogaland monzonoritic bodies, e.g. the Bjerkreim-Sokndal Layered Intrusion: 0.7055 (Nielsen et al. 1996) and the Hidra Leuconorite: 0.7055 (Demaiffe & Hertogen 1981), preclude a substantial crustal residence time before 835 Ma ago. Therefore, it is much more plausible to regard the Rb-Sr and Sm-Nd age of 835 Ma as the age of intrusion.

The differences in place and time between the monzonorites of the Hunnedalen dykes and those of the Egersund Igneous Complex appear to form serious objections to their supposed genetic relationship. However, the massif-type anorthosites have earlier been considered the results of very long-lasting, multiphase emplacement histories (Duchesne 1984, Duchesne et al. 1985, Duchesne & Maquil 1987). Later the 930 ± 2 Ma zircon and baddeleyite age of several massif-type anorthosites was interpreted by Schärer et al. (1996) as indicating a short duration for the anorthosite emplacement process. However, Maijer & Verschure (1996) argued that this age of zircon and baddeleyite - in leuconoritic veinlets and aggregates cutting deformed orthopyroxene megacrysts- represents the solidification of final residual liquids during the long-lasting emplacement history of the anorthosites. In the case of a prolonged emplacement history the time difference is no objection any more. In this view, the Hunnedalen monzonoritic magma may be explained - as for the origin of other Rogaland monzonoritic magmas (Duchesne et al. 1989, Duchesne 1990) - by decompressional melting of basic and intermediate rocks in the lower crust. The melting possibly occurred in response to updoming of the crust-mantle interface due to underplating by late manifestations of the anorthositic magmatism. Although speculative, such an updoming will create an extensional tectonic setting in the overlying crust resulting in ductile and brittle normal faulting which made possible the emplacement of the Hunnedalen monzonoritic dykes.

Conclusions

1. The Hunnedalen dykes are biotite monzonorites. Crystallisation started with plagioclase which was successively accompanied by orthopyroxene, pigeonite + Ca-rich clinopyroxene, biotite and perthite + quartz. Olivine was found in one dyke (Drent 1982).

2. Different dykes have variable mineralogical and chemical compositions (Drent 1982), but individual dykes are homogeneous. Thicker (> 5m) dykes - as the one studied here - exhibit some mineralogical and chemical variation due to flow differentiation with some concentration of first formed plagioclase and Fe-Mg minerals with highest Mg# in the centre of the dyke. Major and trace elements indicate an anorogenic, within-plate character of the dyke swarm.

3. Alteration affecting small dykes and the margins of thicker dykes took place in two steps (Drent 1982). Locally, amphibolites formed first. This alteration was thought to have taken place directly after intrusion and crystallisation. A second alteration was a low-grade chloritisation that seriously affected both the mineralogical and the chemical composition. This alteration is correlated with a regional phase of low to very low-grade metamorphism related to the Caledonian orogeny at about 400 Ma ago. Only the effects of the latter phase are recorded in our samples.

4. The Hunnedalen dykes were emplaced approximately 835 Ma ago in country rocks of Late Proterozoic (Gothian) age that had been metamorphosed under high-grade conditions (transitional amphibolite to granulite facies). The intrusion age is based on consistent Rb-Sr (834 ± 9 Ma) and Sm-Nd (835 ± 47 Ma) mineral isochrons.

5. The isotopic systems were variably disturbed.
   Rb-Sr: In the altered marginal zones the Rb-Sr system was disturbed by changes in the Rb and Sr contents and in the Sr-isotopic ratio. In the petrographically unaltered samples the Rb-Sr-system was only slightly altered, i.e. of those minerals most sensitive to alteration.
   Sm-Nd: The Sm-Nd systematics were hardly disturbed, or not disturbed at all, in either the unaltered or the altered samples.
   K-Ar: The K-Ar systematics have been changed twice. During intrusion, Ar set free from heated country rocks was incorporated as excess Ar in all mineral components. During low-grade alteration Ar was lost from those minerals - especially feldspars - with closure temperatures below the temperature of low-grade alteration, resulting in whole-rock K-Ar ages lower than the assumed intrusion age. On the other hand, biotite with a blocking temperature above the temperature of low-grade alteration did not loose its excess Ar, resulting in biotite ages higher than the assumed intrusion age.

6. Chemically, the Hunnedalen dykes show a striking similarity (major and trace elements, element ratios, REE pat-

terns, spidergrams, Eu/Eu* ratios, $Sr_{(i)}$ values) with other monzonorites in Rogaland, especially with the chilled margins of the Hidra leuconorite and the Bjerkreim-Sokndal layered intrusion. This similarity supports a genetic relation of the Hunnedalen monzonorite magma to the Rogaland anorthositic-leuconoritic magmatism, as suggested by Drent et al. (1983).

7. The differences in place (the Hunnedalen Dyke Swarm crops out 10 to 75 km north to northeast of the large Egersund Igneous Complex) and especially in time (a time gap of about 100 Ma between the emplacement age of the Hunnedalen Dyke Swarm 835 Ma ago and the, in our opinion, terminal phase of the prolonged emplacement of the Egersund Igneous Complex at 930 Ma ago) are serious but not insurmountable obstacles for the supposed genitic relationship.

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Appendix

Experimental procedures and constants

Before processing, the samples were first thoroughly cleaned by brushing with deionized water.

Pulverised whole-rock fractions were obtained by the subsequent use of: Hydraulic pressure rock cutters. A modified Schriever jaw crusher. A Bico Brown rock pulveriser with revolving plates. A Tema ring mill. A Fritsch planetary agate ball mill. For XRF and INA determinations the samples were further comminuted in a Spec tungsten high-speed impact mill.

Rock fragments of the < 250 μm range resulting from the grinding with the revolving plate grinder were split by means of a rotary sample splitter. From such splits, sieved and deslimed 250-125 μm fractions have been prepared for K-Ar measurements and mineral separation. All sample splits of whole-rock powder were made after homogenisation in a Turbula mechanical mixer.

Mineral separation was achieved using a Faul-type dry shaking table, heavy liquid separation by means of an overflow centrifuge (IJlst 1973a, 1973b), a modified Frantz magnetic separator (Verschure & IJlst 1969) and hand-picking. The quality of the biotite, plagioclase, ilmenite and apatite concentrates was about 99 % pure; the pyroxenes contained inclusions.

Major elements of whole-rock samples were measured on beads by X-ray fluorescence spectrometry, using a Philips PW 1450/AHP automatic spectrometer. Rb and Sr contents of whole rocks were measured on pressed-powder pellets and are considered to be within 0.5 %. Mass-absorption corrections for both sample and external standard are based upon the Compton scattering of the $Mo-K_a$ primary beam (Verdurmen 1977). Whole-rocks of the samples of both the Hunnedalen West and East sites were analysed also for trace elements by XRF and for the Hunnedalen East dyke rocks also by Induced Neutron Activation at the Interfacultair Reactor Instituut (IRI), Technische Universiteit Delft. The or-

der of magnitude of the values obtained by both XRF and INA is identical, the correlation between both values varies from perfect correlation to no correlation. Bad or no correlation is mainly due to restricted spread of data points.

The chemical separation and mass spectrometry procedures are those routinely used at the former NWO Laboratory for Isotope Geology (e.g. Valbracht 1991).

The Rb-Sr and Sm-Nd isotope analyses on the Hunnedalen East and West whole-rocks and minerals were performed on a Finnigan-Mat 261 mass spectrometer. The analytical accuracy for the measurements are considered to be within 0.5 % for isotope dilution of Rb, Sr, Sm and Nd and average 0.003 % and 0.0015 % for $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$, respectively.

Ar was extracted in a bakeable glass vacuum apparatus and analysed by isotope dilution techniques in a Varian GD-150 mass spectrometer; the measurements were made by the static method. The K contents of the sieved sample fractions for the K-Ar analyses were determined by flame photometry with a Li internal standard and a $CsCl-Al(NO_3)_3$ buffer. The analytical accuracy is believed to be within 1 % for K and 2 % for radiogenic Ar.

All the estimated overall limits of relative error for the Finnigan-Mat 261 and the Varian GD-150 mass-spectrometers are the sum of the known sources of possible systematic error and the precision of the total analytical procedures.

Best-fit lines through the Rb-Sr and Sm-Nd data-points were calculated by means of a least-squares regression analysis computer program according to York (1966, 1967). The errors quoted for the isochron ages and intercepts are 95% confidence limits inclusive a correction for the MSWD.

The age calculations are based upon the I.U.G.S. recommended constants (Steiger & Jäger 1977).

The notations of ε_{Nd} follow DePaolo & Wasserburg (1976).