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GRANITOPHILE ELEMENTS IN GRANITOID ROCKS IN PRECAMBRIAN BASEMENT WINDOWS IN NORDLAND, NORTHERN NORWAY, WITH SPECIAL REFERENCE TO THE RARE-ELEMENT ENRICHED GNEISS AT BORDVEDÅGA, HØGTUVA WINDOW

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Tittel:
Granitophile elements in granitoid rocks in Precambrian basement windows in Nordland, Northern Norway, with special reference to the rare-element enriched gneiss at Bordvedåga, Høgtuva window.

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Fylke: Nordland	Kommune: Rana, Lurøy, Meløy
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1974, 1985 og 1986	01.02.87	1900	I. Lindahl

Sammendrag:

This report gives the results of field work done on the Precambrian Høgtuva window, with minor comments on the other similar windows in Nordland.
The aim for this work was to attain a better understanding of the genesis of the rare-element mineralization at Bordvedåga, and thereby provide guide lines for further prospecting for granitophile element deposits.
In the Høgtuva windows, a peraluminous, high-silica, biotite-granitic gneiss is host for a rare-metal deposit at Bordvedåga. This and similar smaller metalliferous occurrences other places in the window are characterized by enhanced values of a suite of trace elements.
A large portion of the rocks in the Høgtuva window is thought to be of supra-crustal origin, both volcanic and sedimentary origin.
It is suggested that the mineralization in the extensive, mineralized gneiss was caused by synvolcanic enrichment of the trace element suite by magmatic/late-magmatic differentiation.

Emneord		
	geokjemi	beryllium
malmgeologi	tinn	fagrapport
uran	prekambrium	

Contents

1. INTRODUCTION

The area, to which the present report refers, consists of Precambrian basement rocks exposed in several tectonic windows in Nordland, from the arctic circle north to Ofoten (fig. 1). This choice of target was motivated by (1) the discovery of a granitophile element enriched gneiss at Bordvedåga in the Høgtuva window, (2) the fact that the basement in the area is enriched in Sn, U and Be (NGU's database on the geochemistry of Norwegian granites) and (3) the results of regional stream sediment geochemistry showing enrichment in granitophile elements in the same area.

A uranium province in northern Sweden is situated along a presumed plate collision zone (Adamek & Wilson 1979). The granitoids in this region are anomalous in U and Th. The enrichment in radioelements in the Norwegian basement windows from Rombaken in the north to Høgtuva in the south, indicates a continuation of the Swedish uranium-province beneath the Caledonides (Lindahl 1984).

In this ore province in the Swedish Precambrium occur epigenetic uranium deposits in uranium-rich granites, and a stratabound uranium mineralization in rhyolitic ignimbrites (Adamek & Wilson 1979). Several types of Mo-W mineralization are also present: in granitic aplites, in pegmatites, in metamorphosed, altered, mostly volcanic supracrustals, and in skarn. Mo-W-Sn mineralization also occurs in greisen.

Greisen is not known to occur in the Norwegian basement windows studied. Uranium occurrences are, however, present in pegmatites at Orrefjell (Lindahl et al. 1985) and in epigenetic veins at Harelifjell (Lindahl 1984). Several Mo and W mineralizations are found within the basement and above the contact to the tectonically overlying units.

The economically interesting elements Be, Sn, U, REE, Zr, Nb and Y are not today exploited by Norwegian mining industry.
The aim for the present study was to get a better understanding of the genesis of the mineralization at Bordvedåga, and thereby formulate guide lines for further prospecting for granitophile elements. Such guide lines

could be useful in the search for areas with differentiated rocks where also post-magmatic or later events have caused further enrichment. Locali-zation of areas with more differentiated granitoids with economic potential is another object for this work.

Rock samples were collected regionally in 1974 (not reported). The present report is mainly restricted to the work on the Høgtuva window.

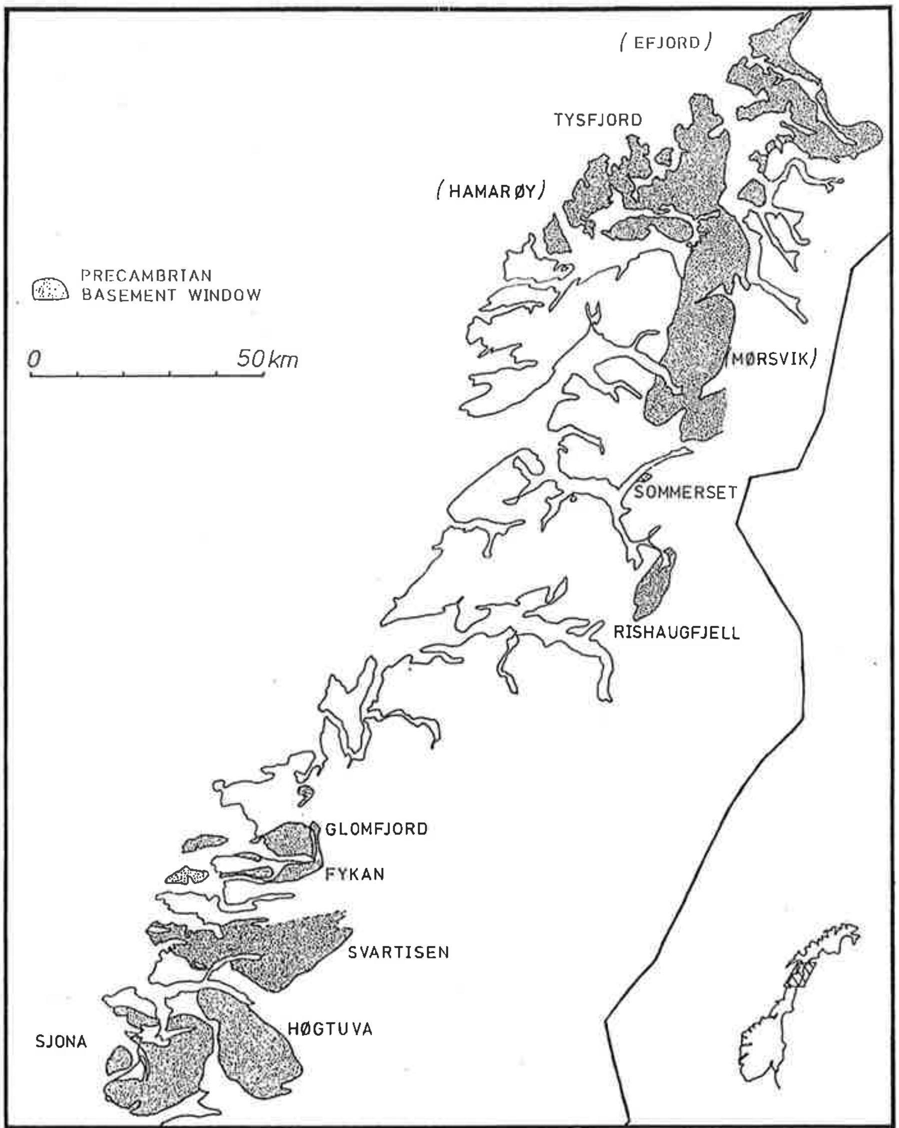


Fig. 1

Precambrian basement windows NORDLAND

2. PREVIOUS WORK

The Precambrian windows in Nordland are discussed by several workers, especially the middle and northern parts (Glomfjord-Tysfjord), e.g. Holmsen (1916, 1932), Rekstad (1929), Foslie (1941, 1942), Cooper & Bradshaw (1980), Wilson & Nicholson (1973), Rutland & Nicholson (1965), Rutland et al. (1960), Andresen & Tull (1986), Stephens et al. (1984), Gustavson & Gjelle (1981), Gabrielsen et al. (1981), Holmes (1965).

Relatively new geological maps cover most of the area (Gjelle et al. 1985, Søvegjarto et al. 1987, Gustavson 1985, Johnsen 1983).

The eastern and northern part of the Høgtuva window and the Caledonian nappes have been mapped in scale 1:50 000 (Gjelle et al. 1985, Søvegjarto et al. 1987); the Precambrian basement rocks are, however, mostly undifferentiated.

Since the study of this area was initiated in 1980-81, prospecting work include helicopter - borne radiometric, magnetic, and VLF surveys (Håbrekke 1983), car-borne radiometric surveys (Berge & Hatling 1983), ground radiometric survey (Furuhaug 1984), and panning of stream sediments (Hatling 1983, Wilberg 1987). Lindahl & Furuhaug (1987) discuss major and trace element geochemistry, detailed geology and core drilling.
Other reports from the area are Røste (1984 and 1986).

3. REGIONAL GEOLOGY

Precambrian basement is exposed in domal tectonic windows in the Caledonian high metamorphic nappes. The numerous windows in Nordland may be divided into two groups. An eastern set lies partly along the Norwegian-Swedish border and includes Rombak, Nasafjell and Børgefjell, while a western set lies along the Norwegian coast and includes Tysfjord, Glomfjord, Svartisen, Høgtuva and Sjona (fig. 1).

Large-scale gravity lows are associated with the eastern basement domes. This, combined with studies of magnetic data and lineament patterns, indicates that the eastern basement windows represent large-scale culminations, whereas the western domes represent smaller bodies, perhaps granitic cores in recumbent folds within the nappes. They are not associated with detectable lows and demonstrate only negligible mass deficiencies (Gabrielsen et al. 1981).

The domal character of these bodies, has by several authors been thought to be connected with the Caledonian deformation (Rutland & Nicholson 1965, Nicholson & Rutland 1969). However, Cooper & Bradshaw (1980) tend to ascribe basement doming to gravity instabilities due to inhomogenities in the basement. Ramberg (1966, 1973) interpreted them to be due to diapiric processes by density inversion in the crust.

The western coastal windows consists of granite gneisses which show penetrative, early deformation structures and metamorphism in common with their allochthonous (Gustavson 1978) metasedimentary cover (Nicholson & Rutland 1969).

The Glomfjord window has been dated (Rb-Sr whole-rock) at 1696 ± 73 Ma, a date interpreted as the age of intrusion, while the deformation and metamorphism has been interpreted to be of Caledonian age (Wilson & Nicholson 1973). This dating is comparable with ages obtained from basement rocks throughout northern Norway, which range from 1800 to 1650 Ma (Cribb 1981), and comparable to values obtained from the adjacent Swedish Baltic Shield.

Such ages also apply to rocks included in the high grade nappe sequence (Wilson & Nicholson 1973, Reymer et al. 1977). In addition to this late Svecofennian event, a younger Sveconorwegian event has also been reported (Gabrielsen et al. 1981).

The foliation of the granitic gneisses is parallel to the boundary and the schistosity of the deformed schists in the Caledonian nappes. It seems obvious that the shearing and foliation in the gneisses are related to the Caledonian tectonism. Gustavson & Gjelle (1981) note that the uppermost part of the basement seems to be affected by Caledonian deformation. Towards the center of the windows the granite becomes massive. This, however, is not in agreement with the present study of the Høgtuva window, or with the Tysfjord area (Andresen et al. 1986) where Caledonian foliation is recognized several kilometres below the basement-cover contact.

Autochthonous/parautochthonous Cambro - Silurian rocks overlay the contact of the eastern basement windows; in the coastal windows no autochthonous cover has been found (Gustavson 1978). Rutland et al. (1960) were of the opinion that the contacts between basement and cover do not show a sedimentary relationship.

The concordant, foliated contact between the granite gneisses and the overlying metasediments is complexly folded, and in several places being overturned in fold nappe style (Rutland & Nicholson 1965, Nicholson 1973). This structural style for the western windows is shown in fig. 2 (from Stephens et al. 1985).

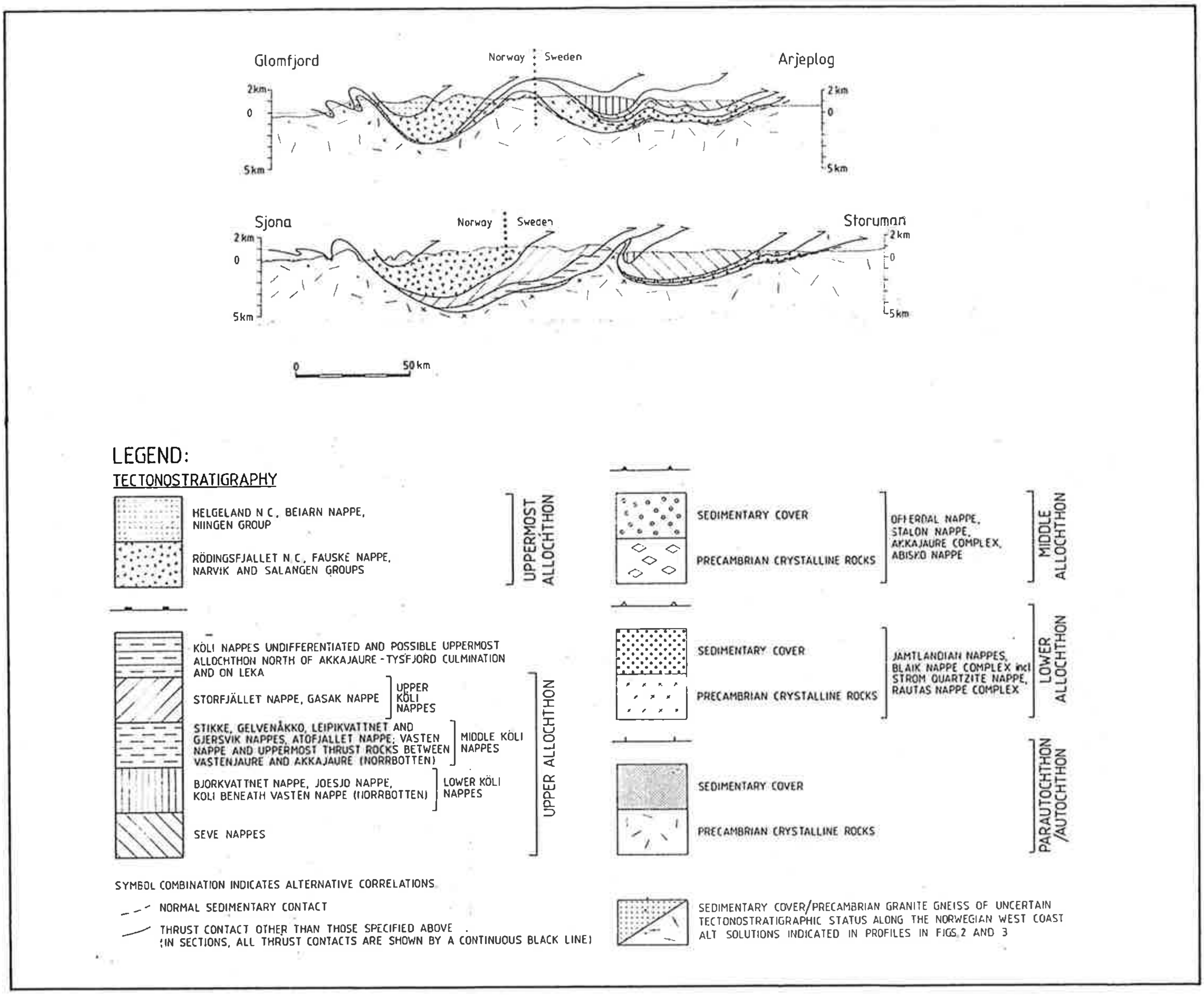


Fig. 2: Schematic profiles through the central - north Scandinavian Caledonides. Model 1 structural solution. Subdivision of the metasedimentary cover into Middle Allochton and Uppermost Allochton equivalents is based on Gustavson (1978). From Stephens et al. (1985).

In an alternative structural interpretation (fig. 3) the granitic gneisses of the western coastal windows are integral parts of the Uppermost Allochthon and are completely detached from the underlying Precambrian crystalline rocks (Stephens et al. 1985).

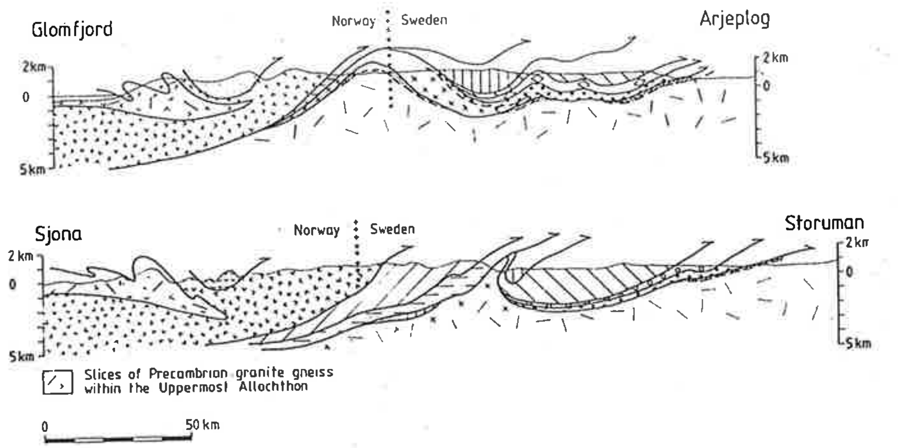


Fig. 3: Alternative schematic profiles incorporating model 2 structural solution for the western granite gneiss massifs. From Stephens et al. (1985).

It has been suggested that the gneisses in the core of the windows are autochthonous (e.g. Hossack 1976, Nystuen 1981). However, the gneisses in the core of the Nasafjell window (east of Høgtuva) have resently been shown to be allochthonous (Thelander et al. 1980). Other windows may also have allochthonous gneiss cores underlain by blind sole thrusts (e.g. Chapman et al. 1985), and Hossack & Cooper (1986) suggest the windows to be allochthonous.

Gustavson & Gjelle (1981) point to similarities between the Precambrian rocks in the coastal area and those of the national border area, and to radiometric determinations which suggest that they were of the same age, about 1700 - 1800 Ma (Wilson & Nicholson 1973). The area between forms a synclinorium resting on a probably continuous granitic basement. The more intense deformation and higher degree of incorporation of basement in the assumed Caledonian fold structures in the coastal district can be regarded as a consequence of closer proximity to the central Caledonian area of deformation (Gustavson & Gjelle 1981).

According to Rutland and Nicholson (1965), the coastal gneiss windows may be regarded as a belt of "internal massifs" comparable to the Pennine cores in the Penninic zone in the Alps, but they must be regarded as basement like the windows along the Swedish-Norwegian border.

The basement consists predominantly of homogeneous, well foliated, leucocratic granitic-gneiss. The cover is miogeosynclinal metasediments (with some metavolcanics), previously thought to be of Caledonian age. However, recent work based on Rb-Sr whole-rock dates indicates that much of the cover sequence is in fact Precambrian (Råheim & Ramberg 1978, Styles 1979, Cooper & Bradshaw 1980, Cribb 1981). Within this metasedimentary sequence conjunctive nappes have been recognized (Rutland & Nicholson 1965). These are in turn overlain by the disjunctive Beiarn Nappe (Cooper & Bradshaw 1980).

4. GEOCHEMISTRY OF ROCKS IN THE PRECAMBRIAN BASEMENT (PCB) WINDOWS

82 samples from the eight PCB windows Høgtuva, Sjona, Svartisen, Glomfjord, Fykan, Rishaugfjell, Sommerset and Tysfjord (the latter consisting of the Efjord, Tysfjord, Hamarøy and Mørsvik areas) were analysed for major and trace elements, for a comparative study. The mean values are listed in tab. 1.

Tab. 1: Average trace element abundances in the PCB windows. n = number of samples.

	Høgtuva	Sjona	Svartisen	Glomfjord	Fyken	Rishaug-fjell	Sommerset	Tysfjord, Hamarøy, Mørsvik	Efjord
n	19	22	11	5	4	5	3	24	4
Nb	22	19	18	17	21	29	22	17	28
Zr	319	345	294	280	248	490	210	354	197
Y	40	34	40	27	37	77	46	39	51
Sr	182	238	275	209	127	91	63	165	74
Rb	221	167	183	168	211	255	247	176	197
Zn	42	58	71	47	36	79	51	80	42
Cu	6	7	5	4	3	4	3	6	4
V	12	20	19	12	8	3	3	7	4
Ba	508	812	1099	841	398	501	240	851	305
Sn	7	6	6	6	6	6	6	6	6
Mo	3	3	3	3	3	4	3	4	3
U	10	7	7	6	9	10	11	6	8
Th	24	17	17	14	28	32	25	16	21
Pb	32	27	27	25	20	46	29	27	22
Co	4	5	6	4	3	4	3	5	3
Ce	137	126	132	97	102	214	172	152	159
La	53	46	54	35	47	98	83	73	76
Ni	3	4	5	3	4	3	3	3	3
Cr	21	21	23	12	11	7	7	21	12
Sc	4	6	7	4	4	4	3	6	3
Rb/Sr	1,2	0,7	0,7	0,8	1,7	2,8	3,9	1,1	2,7
Ba/Rb	2,3	4,9	6,0	5,0	1,9	2,0	1,0	4,8	1,5

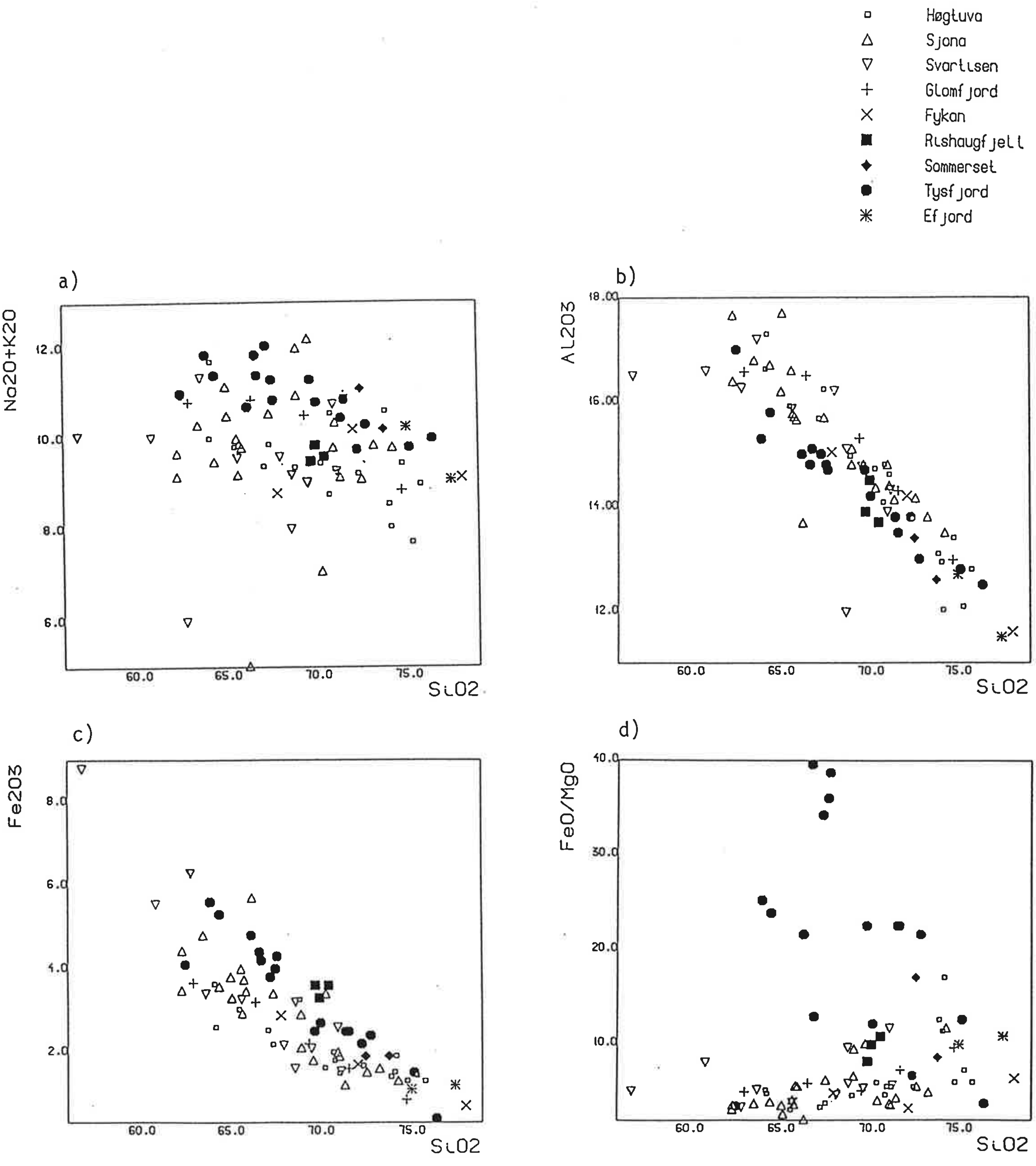
All samples are of acid rocks (SiO_2 from 63 - 75 %). In variation diagrams, they show the same trends: decrease in all the main elements together with Zr, Sr, Ba and Zn with increasing SiO_2, while Rb, Th, Y and Nb increase with increasing SiO_2. Pb, Ce, La and Cr show no correlation with SiO_2.

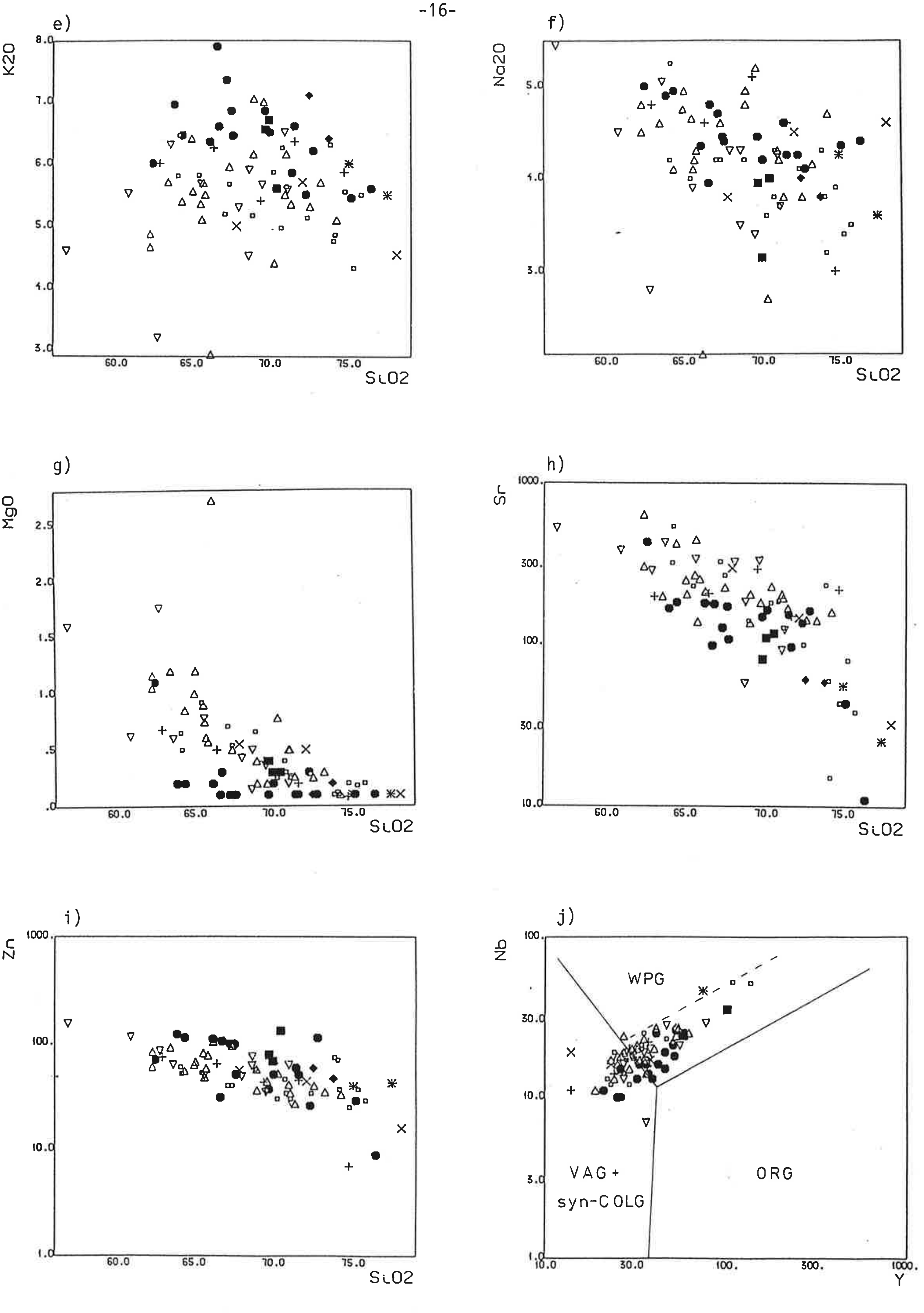
There is a distinct trend of compositional variation from north to south in the area. Alkalinity increases to the north, shown in fig. 4a. Tab. 2 gives the average agpaitic indices, and shows that the mean values from the six southernmost windows group in the metaluminous field while the remaining three northernmost are slightly peralkaline (see also fig. 5). Generally Sommerset, Efjord, Tysfjord and partly Rishaugfjell have distinctly lower contents of Al_2O_3, MgO and Sr, and higher Fe_2O_3, FeO/MgO, K_2O, Na_2O, Y/Nb and Zn (fig. 4a-j).

Tab. 2: Average molecular - alkali ratios in the PCB windows

	$\frac{Al_2O_3}{K_2O+Na_2O+CaO}$	$\frac{Al_2O_3}{K_2O+Na_2O}$
Høgtuva	0.937	> 1
Sjona	0,921	> 1
Svartisen	0,887	> 1
Glomfjord	0,921	> 1
Fykan	0,933	> 1
Rishaugfjell	0,931	> 1
Sommerset	0,875	< 1
Tysfjord, Hamarøy, Mørsvik	0,867	< 1
Efjord	0,889	< 1

Fig. 4: Variation diagrams showing major oxides and trace elements for samples from the 9 PCB windows. The fields in fig. 4j , within plate granite (WPG), volcanic arc granite (VAG) and ocean ridge granite (ORG), are from Pearce et al. (1984).




e)
K2O
8.0
7.0
6.0
5.0
4.0
3.0
60.0
65.0
70.0
75.0
SiO2
f)
Na2O
5.0
4.0
3.0
60.0
65.0
70.0
75.0
SiO2
g)
MgO
2.5
2.0
1.5
1.0
.5
.0
60.0
65.0
70.0
75.0
SiO2
h)
Sr
1000.
300.
100.
30.0
10.0
60.0
65.0
70.0
75.0
SiO2
i)
Zn
1000.
100.
10.0
1.0
60.0
65.0
70.0
75.0
SiO2
j)
Nb
100.
30.0
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syn-COLG
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100.
300.
1000.
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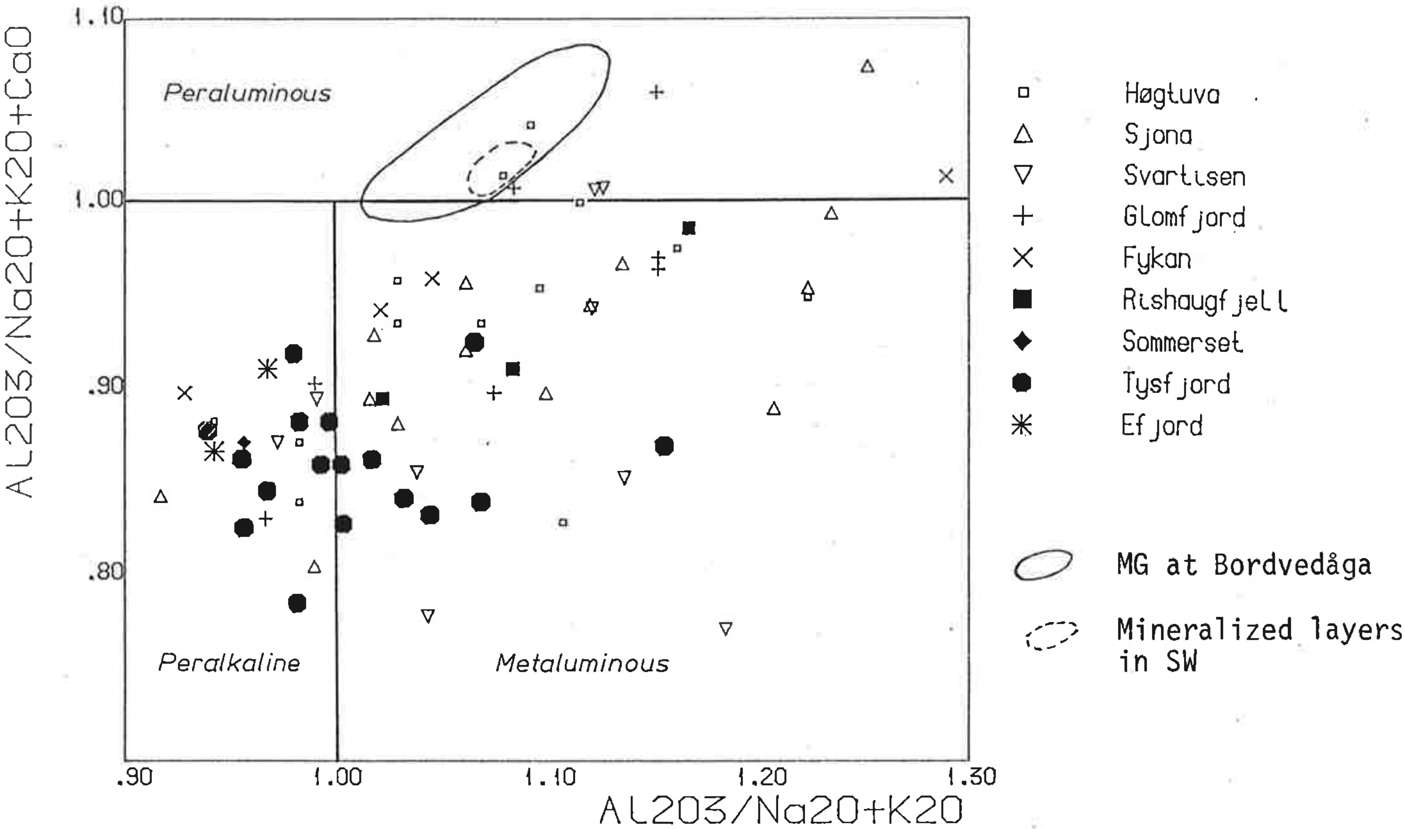


Fig. 5: Aluminium vs. alkali ratios from the PCB windows, including the mineralizations in the Høgtuva window.

Rb, Sr, Ba and their ratios points to more differentiated granitoids in Høgtuva, Rishaugfjell, Fykan, Sommerset and Efjord (tab. 1). These also show a slight increase in incompatible trace elements.

Though this trend may seem speculative because of the low number of samples and also the high detection limit for some of the trace elements, the data is partly confirmed by samples from earlier work.

The granitoid rocks in all the windows fall in the "within plate" field, with a "volcanic arc" contribution, on various discrimination diagrams (Pearce et al. 1984) as exemplified by fig. 4j and 6. However, they do not show the marked Ba depletion of the modern, more mature alkali provinces.

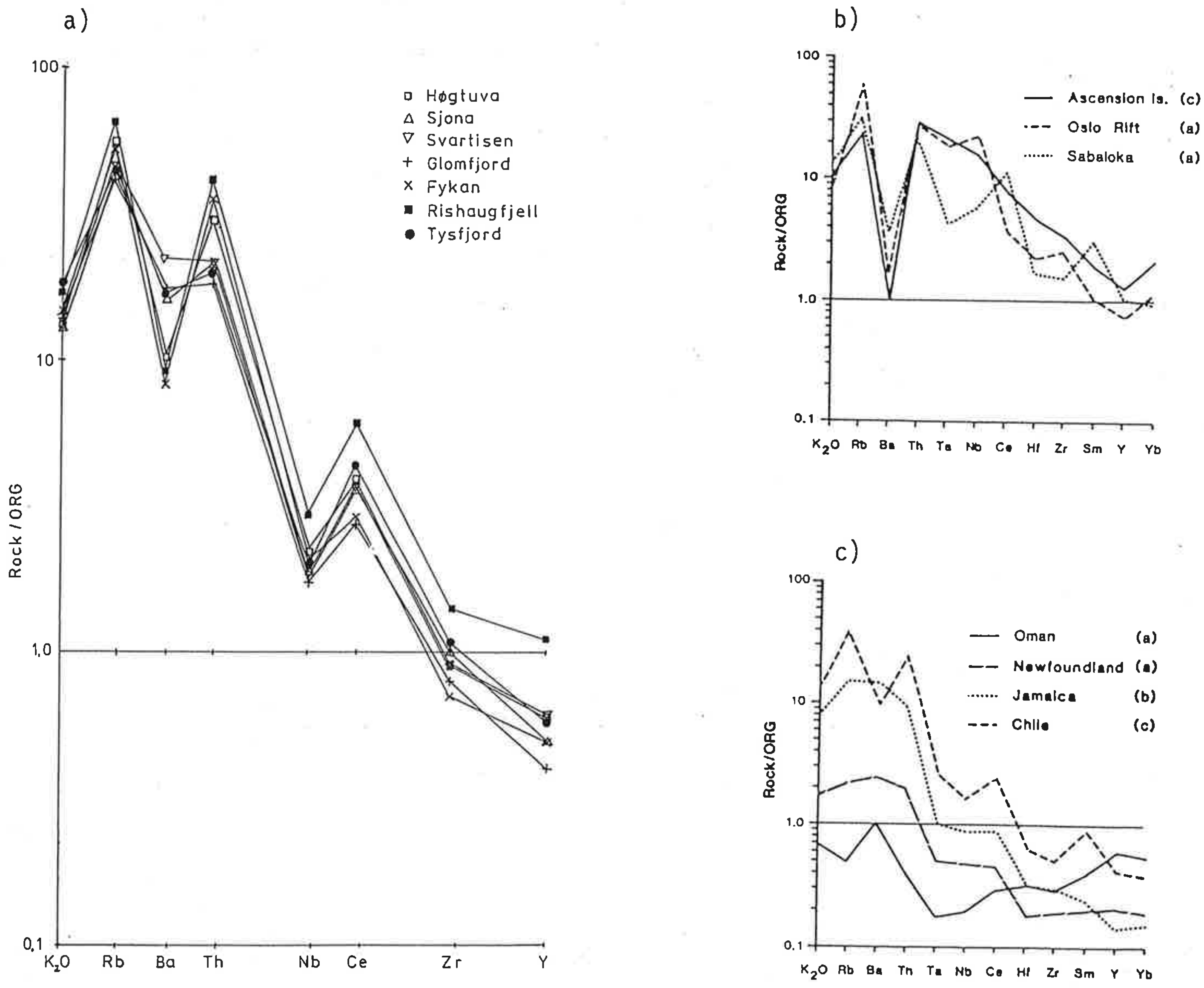


Fig. 6: Ocean ridge granite (ORG) normalized geochemical patterns for representative analyses from the PCB windows (6a). Examples from within plate granites (6b) and volcanic arc granite (6c) from Pearce et al. (1984), for comparison.

5. THE HØGTUVA WINDOW

5.1 Introduction

Høgtuva is the southernmost of the tectonic windows in Nordland, with an outcrop-area is about 290 km^2 and altitudes varying from sea level to 1294 metres. Bedrock is well-exposed (80-90 % outcrop) except for the approximately 20 km^{2}Høgtuva glacier. The area is accessible by sea in the north, and by road in northeast and southwest.

5.2 Geological characteristics

The available data does not establish whether the Høgtuva window is autochton basement or a completely detached slice in the nappes. It has form of a mega-lens, with the axes NW-SE, dipping to the east, with the western and southern basement/cover contact inverted. The foliation in the window is parallel to the cover contact, flatlying in the central part, with dips outwards to the north and east, and overturned dips in west and south.

The Høgtuva window differs from the other windows further north in the coastal region in that the rocks are more fine grained and more foliated. The rocks in the Høgtuva window are dominated by quartzo - feldspathic, well-foliated, acid biotite gneisses classified according to Streckeissen (1976) as granite/ rhyolite, with minor units of intermediate (hornblende - quartz - monzonite) and basic (dolerite dykes) rocks.

Three major textural units are generally recognized: (1) fine grained granitic gneisses, (2) medium grained granitic gneisses, sometimes porphyritic, and (3) coarse grained gneiss granite.

Nine different rock types may be recognized, based mainly on textural and mineralogical criteria:

1. Coarse grained gneiss granite
2. Coarse grained hornblende - quartz - monzonite

3. Fine grained, biotite-rich granitic gneiss
4. Fine grained, light grey to pink granitic gneiss
5. Medium grained granitic gneiss
6. Porphyritic granitic gneiss
7. Granitic, deaply weathered gneiss
8. Dolerite dykes (amphibolite and biotite schist)
9. Carbonate - fluorite layers

The rocks are dominantly fine grained gneisses in the southern part of the window and medium grained gneisses in the north, with intercalated, minor coarse grained varieties sometimes only weakly foliated. Amphibolites and associated biotite schists, which originally was dolerite dykes/sills, are widespread.

Alternation with sharp boundaries between fine and medium grained gneisses are characteristic. Though the gneisses have been highly tectonized and foliated, the extensive persistence of this supposed lithological banding, together with local intercalations of carbonate-horizons with associated of minor quartzite, makes it probable that the gneisses were derived from original volcanics and/or sediments.

In the major thrust zone, Søvegjarto et al. (1987), noted the occurrence of cover sediments consisting of quartzite and sulphide bearing mica schist and graphite schist. A new discovery of these sediments at the eastern margin of the window confirms the interpretation of Søvegjarto et al. (1987) of a minor thrust zone in the uppermost part of the basement gneisses (plate 1 and 2). In general only a few metres of cover metasediments are preserved in the thrust zones, where they often occur as intensely deformed lense shaped bodies. Similar features are found in other Precambrian windows, as in the Nasafjell, Børgefjell and Tømmerås windows to the east and south. The metasediments are considered to be parautochthonous or allochthonous.

Upper Cambrian radioactive black shales are well known on the stable Baltoscandian Platform and along the margins of the basement windows in Northern Norway, with characteristic high contents of elements like U, V and

Mo (Gee 1980). The graphite schist bordering the Høgtuva window is also enriched in these elements, with values of 12, 265 and 35 ppm respectively.

5.3 Rock types and geochemical characteristics

Average chemical analyses of the Høgtuva rocks are given in tab. 3.

Coarse grained, massive to weakly foliated rocks, supposedly of intrusive origin, occur as two main types: (1) high-silica, metaluminous gneiss granite and (2) hornblende - quartz - monzonite.

Tab. 3: Mean chemical composition of rocks in the Høgtuva window. (-means below detection limit.)

	Fine grained granitic gneiss	Fine grained, rusty granitic gneiss	Fine grained, muscovite bearing granitic gneiss	Fine grained, biotite rich gneiss, above minor thrust zone	Medium grained granitic gneiss (central)	" " , NW	" " , W	" " , NE	Coarse grained granite, Bordvedåga	" " " , Melfjordbotn	" " " , Hyttan	Hornblende - quartz monzonite	Porphyritic gneiss	Deeply weathered granitic gneiss	Amphibolite	Biotite schist
n	13	2	2	2	2	3	3	2	1	1	1	3	2	1	3	3
SiO_2	74,22	73,32	76,41	61,42	68,95	63,64	67,98	74,53	74,09	76,83	70,02	56,13	62,19	74,27	46,19	44,22
Al_2O_3	13,09	13,64	12,41	17,40	15,08	16,99	15,34	13,12	12,18	11,61	14,95	16,11	18,30	11,35	14,68	13,84
Fe_2O_3	1,50	1,33	1,05	5,26	1,97	4,00	3,29	1,43	2,55	0,90	1,71	7,53	2,86	3,93	13,76	15,40
TiO_2	0,20	0,24	0,15	0,73	0,38	0,57	0,49	0,19	0,21	0,10	0,29	0,62	0,60	0,32	2,00	1,68
MgO	0,21	0,22	0,15	1,62	0,46	0,87	0,58	0,17	0,01	0,05	0,29	4,05	0,66	0,06	6,38	7,91
CaO	0,53	0,71	0,50	2,63	1,00	2,30	1,59	0,55	0,48	0,34	0,87	6,60	1,46	0,21	9,67	6,95
Na_2O	4,1	4,2	3,8	4,7	4,0	4,8	4,0	3,3	3,9	2,6	4,0	3,6	4,8	3,5	2,5	2,4
K_2O	5,35	4,37	3,58	4,39	5,62	5,40	5,60	5,56	5,21	5,56	5,62	3,00	6,82	4,96	1,85	4,28
MnO	0,04	0,02	0,03	0,12	0,05	0,08	0,06	0,05	0,05	0,02	0,05	0,14	0,05	0,09	0,26	0,24
P_2O_5	0,01	0,01	-	0,19	0,03	0,13	0,10	-	0,01	-	0,02	0,18	0,10	-	0,53	0,32
Nb	20	24	37	42	20	16	20	18	33	13	17	13	13	30	18	16
Zr	230	225	250	321	293	412	298	189	719	269	286	131	506	954	158	190
Y	30	39	53	50	30	30	42	24	107	39	37	22	22	55	38	75
Sr	75	75	57	503	223	426	236	56	28	49	161	491	350	7	253	259
Rb	210	180	150	138	186	152	198	181	421	251	160	98	112	258	90	583
Zn	30	30	23	137	33	70	58	32	135	15	35	73	36	165	163	223
Cu	-	70	-	14	-	13	6	-	-	-	-	38	-	-	14	36
V	-	6	-	62	13	35	15	-	-	-	-	150	14	-	194	225
Ba	200	270	40	1564	570	1100	671	178	70	157	517	681	1350	35	285	540
Sn	-	-	-	-	-	-	-	-	15	-	-	-	-	-	-	-
Mo	-	15	-	-	-	-	-	-	-	-	-	-	-	9	-	-
U	-	45	21	-	-	-	-	-	-	-	-	-	-	-	-	-
Th	30	35	60	25	23	21	15	25	34	74	16	-	-	19	-	-
Pb	33	125	35	71	27	26	35	25	52	32	33	18	20	22	44	49
Co	-	-	-	9	-	9	-	5	-	-	-	27	13	-	55	45
Ce	90	150	130	100	118	165	147	65	201	44	104	54	46	345	60	45
La	23	70	42	47	44	82	60	-	88	18	44	23	19	96	30	25
Ni	-	-	-	15	-	5	-	-		-	-	28	-	6	93	143
Cr	22	15	25	41	19	18	46	18		9	10	85	22	18	88	330
Sc	-	-	-	10	-	7	-	-		-	-	20	6	-	25	26

1) Coarse grained, massive to weakly foliated granite is found in three localities (Hyttan, Melfjordbotn and Bordvedvatnet) as lenses in the finer grained, foliated gneiss.

Quartz, microcline and plagioclase occur in slightly varying amounts together with c. 5 % biotite and minor amphibole. The granite associated with the mineralized gneiss at Bordvedåga is described in chapter 6.2.2.

2) Hornblende - quartz - monzonite occurs in one locality (Gjervaldalen) where it forms a lense-like body about 1 km long.
The texture varies from massive in the middle to foliated along the margins. Particularly in the central part the coarse grained rock has porphyritic texture, with up to 1 cm hornblende + biotite porphyries in an inequigranular quartz - feldspar groundmass. A typical modal composition is 15 % quartz, 35 % plagioclase, 15-20 % microcline, 25 % hornblende, 5 % biotite and 3 % epidote with accessory apatite, zircon, muscovite, chlorite, sphene, fluorite, allanite and opaques. Patch - antiperthite and myrmekitic intergrowths are common. Some of the feldspar is strongly altered, and the plagioclase is occasionally zoned.

The upper contact between monzonite and granitic gneiss is sharp, but complex with interfingering. The lower contact is transitional from monzonite to a hornblende-bearing reddish granitic gneiss.

3) Fine grained, biotite rich, granitic gneiss. This gneiss-type occurs in the peripheral parts of the window, and is separated from the underlying gneisses by a minor thrust zone. It is continuous along the eastern margin of the window, and is distinguishable from other gneiss types by its higher biotite content and intermediate character (61 % SiO_2).

4) <u>Fine grained granitic gneiss</u> occupies most of the southern part of the window, and is also intercalated in the dominant, medium grained gneiss in other parts of the window. The fine grained gneiss can be differentiated in three groups in addition to the mineralized gneiss (MG) which is described in 6.2.2.: (1) light grey to pink, fine grained gneiss, (2) rusty, fine grained gneiss and (3) muscovite bearing, fine grained gneiss.

The fine grained gneiss is typically light grey and well foliated, the foliation being defined by parallel orientation of biotite and elongation of feldspar and quartz. Pink to red varieties also occur, and are typical in the south. Because of a small biotite content, as low as 1-2 % in some areas, the gneiss appears rather massive, specially in the south. The typical fine grained gneiss consists of the following minerals: 35 modal % quartz, 35-40 % microcline, 20 % plagioclase (An_{20-25}), 4-6 % biotite, with accessory sericite, chlorite, zircon, allanite, epidote, zoisite, fluorite and opaques (up to 4 %). The texture varies from equigranular to inequigranular, with average grain size 0.4 - 0.5 mm.

Flaky muscovite is a typical constituent (5-10 %) in the fine grained gneiss in the Snøfjellet area. In the gneisses of other areas, muscovite occurs only as microscopical widespread accessory sericite. The muscovite gneiss has a slightly different chemistry as compared to the normal fine grained gneiss. Increase in modal quartz, and decrease in biotite, is reflected in the increased SiO_2, while the other main constituents are less.
The muscovite gneiss is higher in uranium (10-33 ppm) and incompatible element content.

In the same area at Snøfjellet, and also at Melfjellet as an extensive (2-3 km long), thin (1 m) layer, a typical rusty fine grained gneiss occur. Major and trace element contents are similar to the normal, fine grained gneiss, except for a higher Cu, Mo, U, Pb and REE content (tab. 3).
With only 2-3 % dark minerals (biotite), the rock is very light, and it has an equigranular (0.2 - 0.3 mm) texture. Typical characteristics are

the increased radiation and the rusty weathered surface, due to oxidation of pyrite. Pyrite is disseminated as discrete euhedral to subhedral grains and occurs also at grain boundaries and in joints.

5) Medium grained gneiss, which dominates in the northern part of the massif, is heterogeneous both in texture, mineralogy and chemistry. Due to deformation it shows variation in texture from equigranular to porphyritic, and in the degree to which biotite (and hornblende) is aggregated and parallel oriented. Quartz and feldspar are elongated and biotite is concentrated in thin layers. The colour is mostly grey to pink.

Although the composition is variable quartz and K-feldspar, generally occur in about equal amounts and dominate over plagioclase. Biotite is the main mafic mineral (6-10 %), while hornblende and epidote are found in some rocks in amounts of up to 10 % and 3 % respectively. Accessories are sphene, apatite, zircon, allanite, chlorite, muscovite, garnet, zoisite and opaques (mainly magnetite). Myrmekitic texture and flame-perthite are common.

The number of rock analyses is too small to make conclusions about regional variance in composition. When grouped in NW, NE, W and central part, however, a pattern emerges. The gneisses in NW are lowest in silica, and are typically higher in Ca, P, Ba and Sr, and lower in Rb than the other gneisses. The high-silica (74 %) gneisses in NE have the opposite trend. The medium grained gneisses from W and central part group compositionally in between.

In contrast to its homogeneous character elsewhere, the gneiss has a banded appearance in a restricted area in the north, along Melfjorden. Dark and light bands alternate in dm-scale (fig. 17).

6) Porphyritic granitic gneiss is easily distinguishable in a 4 km long fold structure between Melfjellet and Høgtuva. The thickness is up to 100 m, including a minor darker gneiss underneath the porphyritic gneiss. The modal composition of the porphyritic gneiss is 30 % quartz,

45 % microcline, 15 % plagioclase, 8 % biotite and abundant sphene, with trace amounts of apatite, muscovite, epidote, chlorite, calcite (in joints and along grain boundaries in plagioclase) and small amounts of opaques. Biotite flakes bend around quartz - feldspar grain aggregates. The porphyritic gneiss (62 % SiO_2) has a high K_2O content (6.8 %), while the underlying biotite rich gneiss (56,8 % SiO_2) is characteristically higher in Fe_2O_3, MgO, CaO, MnO and P_2O_5 as compared to the porphyritic gneiss.

7) Granitic, deeply weathered gneiss. A characteristic zone (30-50 m wide, often consisting of two separate, parallel zones) runs along the NE- and NW margin at a distance of about 1 km from the cover contact. At the surface the rock is almost completely weathered to gravel, occasionally with remaining resistant quartzo-felspathic bands with abundant magnetite. Representative sampling is thus difficult.

The modal composition is about equal amounts of quartz, microcline and plagioclase, 4 % biotite, 6-8 % hornblende and 2 % riebeckite. Zircon is abundant, both as euhedral grains and as anhedral aggregates, together with epidote, apatite, allanite and opaques. Average grain size is 1.5 mm.

Na-amphibole is rare. In addition to the weathered gneiss, it is found only in or near the mineralized gneiss at Bordvedåga and in the mineralized dark pegmatitic band at Trolldalsaksla.

In fact, the weathered gneiss bears other similarities to the mineralized gneiss. The major element content is similar, as is the trace element pattern: typically high Zr, Zn, REE and Rb, and low Ba and Sr.

Explanation for the strong weathering is not found microscopically. Similar weathered gneiss is found at the same level in the neighbouring Svartisen window to the north.

8) Dolerite dykes. Amphibolite and biotite schist are not differentiated on the geological map. They are abundant throughout the entire window,

mostly as thin (less than 1 m thick and several km long), extensive, concordante layers of biotite schists in depressions in the benching of the gneiss, and as thicker (up to 20-30 m) amphibolite layers cutting the foliation.

They are supposed to have the same origin: dolerite dykes or sills; the amphibolite being tectonically transposed to biotite schist. Evidence for this is seen in several places, where amphibolite passes graditionally into biotite schist with breccia fragments along strike, or where amphibolite boudins occur in biotite schist. Recumbent folds thrusted along biotite schist are observed (fig. 7).

As discussed in chapter 5.4., amphibolite dykes dipping in the opposite direction to the regional foliation are seen to pass into thin biotite schists concordantly in between gneiss benches. The thrusts probably utilized existing zones of weakness into which dolerite was intruded and then tectonized to complete parallelism.
Xenoliths of granitic gneiss in the dykes are occasionally observed.

The primary mineralogy, and texture, is changed to a metamorphic mineral assemblage consisting of hornblende, actinolite, plagioclase, biotite, epidote and sometimes garnet.

Amphibolite dykes and biotite schists are present in the mineralized gneiss, but are themselves unmineralized. Exceptions are a thin biotite schist layer with locally high contents of purple fluorite, and another containing high Rb (1100 ppm) and Sn (18 ppm). Rb is in fact the only element that show large variation (42-1100 ppm) in the amphibolites and biotite schists. The geochemistry of the dolerites is influenced by the deformation and alteration to biotite schist (see tab. 3).

The typical modal composition of the amphibolite is 50-60 % hornblende, 15-25 % plagioclase, 5-10 % biotite, subordinate quartz and K-feldspar, up to 10 % each of epidote and sphene, and accessory apatite, garnet, allanite, chlorite and opaques. Grain size is 5-6 mm. Hornblende and garnet are commonly poikilitic with inclusions of quartz, feldspar,

biotite and opaques. Chlorite is the alteration product of biotite and hornblende. Plagioclase is frequently strongly sericitized.
The biotite schists commonly contain more than 50 % biotite.

9) Carbonate - fluorite layers. Carbonate occurs in thin cm-thick layers in the gneiss, locally up to 1 m thick. Locally fold structures are observed. Extensive layers can be traced over lengths of up to 2 km. These carbonates are found only in the southeastern part of the window, and show a typical weathered surface (fig. 9).

The carbonate bands consists of calcite with varying amounts of quartz, K-feldspar, plagioclase and biotite, often with mylonitic texture and alternating layers of massive calcite and recrystallized, microcrystalline quartz, feldpar and biotite. Accessories are muscovite, sericite, epidote, zoisite, allanite, zircon and opaques.

Commonly fluorite occurs, from trace amounts up to massive fluorite layers (fig. 10), usually as white fluorite, but locally purple. Fluorite is interstitial between quartz, feldspar and biotite, often in long, continuous grain-aggregates (along the foliation plane). Feldspar is sericitized, specially along grain boundaries and typically in contact with fluorite.

This fluorite-carbonate rock is interpreted as hydrothermal replacement. Irregular enrichments of fluorite (and beryl) in zones of weakness in the gneisses are occasionally found near fluorite horizons, and perhaps represent pathways for the fluor-rich fluids. The carbonate layers may have acted as reducing or precipitating agent for fluor and other elements. The fluorite horizons contain up to 175 ppm Be, 17 ppm U, 16 ppm W and 0.19 % Ce. In places visible amounts of chalcopyrite (and malachite), pyrite and molybdenite occur.

In carbonate host-rocks, replacement deposits of fluorite-rich skarn enriched in Be, W, Sn etc. may be expected. The dark pegmatitic band at Trolldalsaksla, just W of the MG (mineralized gneiss) and below the granite, can possibly be interpreted as a skarn-mineralization in carbonate-fluorite rock. It is extremely enriched in many trace elements, among them Be (1.19 %), Sn (0.16 %), U (1.01 %), Th (3.19 %), Zr (4.39 %), Nb (1.25 %), Y (0.65 %), Ce (0.24 %), Zn (3.71 %), Pb (0.30 %) and Cu (0.07 %). Identified minerals are hornblende, riebeckite, tremolite, diopside, quartz, feldspar, calcite, fluorite, x-mineral (Ca, Fe, Be silicate), epidote, allanite, biotite, chlorite, apatite, sphene, danalite, helvite, gadolinite, zircon, aenigmatite, pyrochlor, thorite, galena, pyrite, chalcopyrite, magnetite and an intergrowth of a Nb-oxide and Y-silicate.

5.4 Structural geology and metamorphism

The structural interpretation of the Høgtuva window is based on scarce mesoscopic fold patterns and on foliation and lineation as defined by biotite and elongation of quartz-feldspar grains.
The predominant structural elements are believed to be the result of interference between two superimposed major fold phases.
An alternative interpretation is however presented.

The first fold phase observed (F_1)[1)] produced isoclinal folds with axes striking E-W associated with an axial plane schistosity, and was accompanied by sliding and thrusting predominantly along amphibolites and biotite schists (fig. 7). The thrusting accompanying F_1 possibly caused the transformation of the amphibolite to biotite schist (see chapter 5.3.). In the most carefully studied area, around the MG the amphibolites are sometimes discordante with the foliation in the gneiss, and sometimes dip in the opposite direction, while the biotite schists at the bottom of the gneiss benches are always concordant. This is shown in fig. 8. E-W trending lineation is coaxial with the F_1 folds, and in the east part of the window plunges 20^{0} to the east.

[1)] The three fold phases observed are denoted F_1, F_2 and F_3 in the following description, although earlier, nonrecognized deformation episodes may have taken place.

Fig. 7: Recumbent F_1-fold above biotite schist. To the right, gneiss fragments incorporated in the biotite schist.

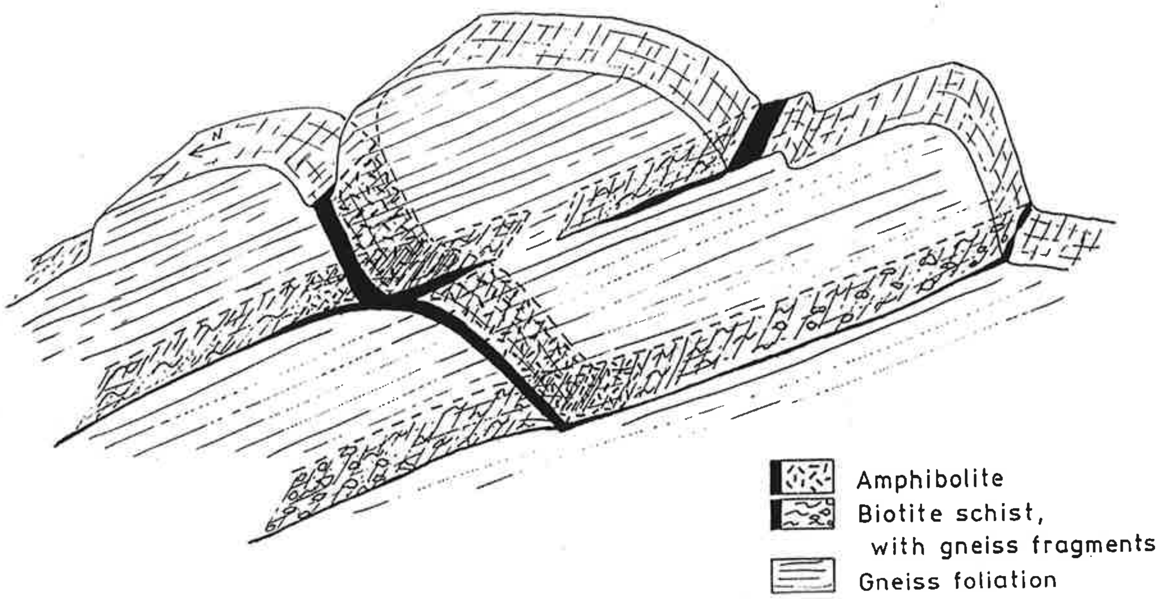


Fig. 8: Block diagram showing the relation between amphibolite (dolerite) and tectonized biotite schist.

The carbonate layers in the east show the first fold phase, with fold axes at 90° (E-W) and plunge 20° E (fig. 9).

Fig. 9: Folded carbonate layers.

Gneiss intercalations within carbonate-fluorite layers, are occasionally asymmetrically folded during F_1 (fig. 10).

Fig. 10: Asymmetrical folding of gneiss layer in massive, banded fluorite - the banding, with accompanying mylonitization, in the fluorite produced by layer - parallel shear.

In some localities biotite bands (primary structures ?) cut the present foliation in the MG, and are folded on F_1 axes. F_1 axial plane schistosity is prominent in these layers (fig. 11).

Fig. 11: Folded biotite bands cutting the present foliation (perpendicular to the pencil).

The second fold phase (F_2) caused the regionally pervasive foliation with approximately concordante or small angle discordante relationship to the bedding and compositional banding in the supracrustal rocks. F_2 folds are open to tight, NW-SE to E-W trending with overturned axial planes parallel to the cover contact.

Large scale F_2 folds are expressed by the amphibolites/biotite schists around the margins of the window, with axes parallel to the cover-contact (fig. 12).
An example of small-scale F_2 folding is shown in fig. 13, which is from the MG.

Fig. 12: F_2-folded biotite schist.

Fig. 13: Small F_2-fold in the mineralized gneiss at Bordvedåga. Fold axes is marked by the pencil.

A third fold phace (F_3) produced open large scale folds with steep axial planes trending NW-SE.

Characteristic interference patterns occur on various scales, usually forming heart and anchor (crescent and mushroom) structures. Examples are the large crescent - shaped structure at Melfjellet (plate 1 and fig. 14), the fold structure at the roadcut at Melfjellet (fig. 15), and in small scale in the MG at Bordvedåga (fig. 16). The interference is locally expressed as dome and basin-like structures in the banded gneiss at Sandvika (fig. 17).

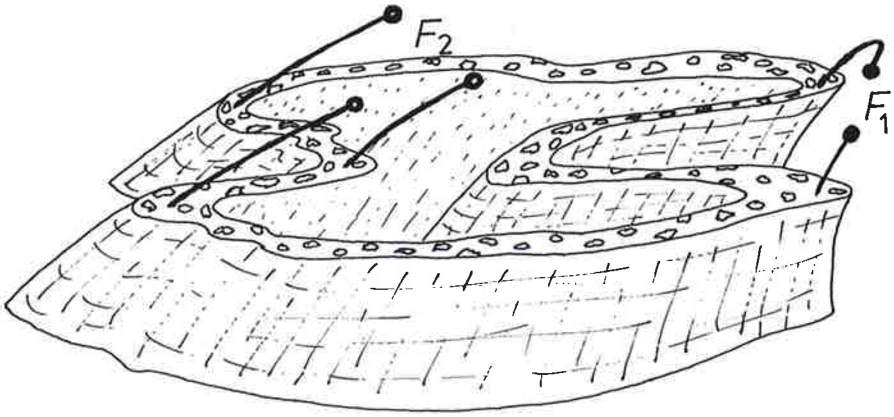


Fig. 14: Fold structure in porphyritic gneiss at Melfjellet.

Fig. 15: Folded dark gneiss at Melfjellet.

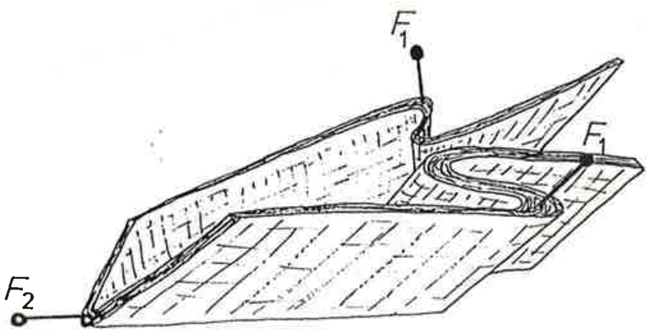


Fig. 16: Folded biotite band in the mineralized gneiss at Bordvedåga.

Fig. 17: Dome and basin-like structures in banded gneiss.

Nicholson & Walton (1963) described folds with circular and elliptical profile sections from the nearby Glomfjord area, which they, similarly to the present interpretation from the Høgtuva window, considered to be the result of refolding. According to Cobbold & Quinquis (1980) such folds are now termed sheath folds, and are regarded as diagnostic structures of shear zones. Their results are therefore applicable to such geological situations as the bases of nappes and diapir margins, where deformation is perhaps complex, but where the major component is simple shear. The model tests of Cobbold & Quinquis (1980) producing passive folds in shear regimes result in noncylindrical, sheathlike folds similar to the discussed examples from Høgtuva (fig. 14, 15, 16 and 17) - all of which occur along the margin of the window.

The foliation defined by the biotite is mostly parallel to the cover contact, but inside the window different biotite orientations occur: 1) One parallel to the axial plane of the second fold phase and 2) one forming girdles around local mesoscopic fold axes (as in the porphyritic gneiss).

All the biotite is believed to be of the same age. The different orientations are regarded as passive re-orientation dependent on the local strain conditions during deformation.

The latest sign of deformation is faults, generally with two directions, NESW and minor N-S, with little or no displacement. They do not influence the tectonostratigraphy. It is breccia zones and fractures, sometimes containing thin selvedges of biotite and chlorite, with no enrichment of U or other rare elements. In places, however, quartz crystals, hematite, magnetite, pyrite, muscovite, fluorite and carbonate are present in the fractures. The most pronounced fracture zone in the window is several kilometres long and cuts the MG at Bordvedåga. It shows both brittle and ductile deformation.

Metamorphism

Most of the rocks contain the assemblage quartz - plagioclase - K-feldspar - biotite ± hornblende, which is of high variance in any component system and has a wide field of stability. The association hornblende - quartz - plagioclase - K-feldspar - biotite ± garnet is present in amphibolite. This is diagnostic of the amphibolite facies. Subordinate secondary (retrograde) chlorite, biotite and muscovite are present in some rocks. The cover rocks are metamorphosed to amphibolite facies.

6. ECONOMIC GEOLOGY

6.1 Types of beryllium deposits

Mineral deposits associated with granitoids may be conveniently grouped in two wide categories on the basis of geological and geochemical features: porphyry-type deposits and granitophile deposits.

Certain lithophile elements such as B, Be, W, Sn, Mo, Nb, Ta, Y, U, Th, REE, Li and Rb tend to be concentrated in residual magmas, and are often concentrated in pegmatites together with complexing agents, such as F^-, CO_3^{2-} or Cl^-. The salic portions of many rhyolites, trachytes and phonolites represent residual magma that can be especially rich in these lithophile elements.

Although rare elements may be concentrated in magmatic, late-magmatic and postmagmatic processes, most models for the transport and deposition of rare elements require postmagmatic (autometasomatic and/or hydrothermal) processes in the final stage to form economic concentration of these elements.

Granitophile deposits tend to concentrate toward the contact zones of the granitoids, and occur as disseminations or pegmatites in the pluton (endocontact) and as veins and stockworks developed upwards or outwards from it (exocontact).

Deposits of granitophile elements have been classified using a variety of criteria such as tectonic setting or associated mineral assemblages (Taylor 1979), or by relating deposit characteristics to depth of emplacement (Varlamoff 1978).

Beryllium deposits can be divided into two broad categories: (1) pegmatitic, and (2) non-pegmatitic or hydrothermal. The hydrothermal deposits that are of most interest as a potential resource of beryllium are mainly hypothermal types formed at high temperatures and at least moderate depths, and epithermal types formed at shallow depths (Griffitts 1973). These deposits have until recently provided a very small proportion of the total supply of beryllium, but will become dominant in the future.

Mulligan (1968) subdivided the non-pegmatitic deposits in pneumatolytic - hydrothermal (high-temperature quartz vein, greisen - quartz vein, skarn (contact metasomatic)), hydrothermal veins ("Alpine veins" and manganese veins), and the loosely grouped "disseminated (replacement) deposits" (probably of metasomatic origin).

Theories regarding the genesis of beryllium deposits are based on the assumption that beryllium is concentrated in the residual fluids of crystallizing igneous rocks, together with the alkalis, silica, and alumina, various rare-earth- and other lithophile metals, and the typical anion-forming (volatile) elements and complexes such as fluorine, hydroxyl, carboxyl (CO_3^{2-}), boron, and phosphorous. The relative atomic concentration of these substances, the temperature - pressure conditions prevailing, the structural developments, and the type of wall-rock encountered at various late and post-magmatic stages are considered to be the dominant influences in determining the degree of concentration that may be attained, the depositional type, and mineral form in which beryllium may occur (Mulligan 1968).

With regard to distribution and age of Be deposits Mulligan (1968) concludes that they conforms in a general way with the concepts of metallogenic provinces and epochs, although these can be only loosely defined in most areas. Most of the known beryllium occurrences of the world are in granite pegmatites, which are widely distributed in the Precambrian Shield areas and some belts of younger crystalline rocks. Most beryllium produced has come from granite pegmatites in the Shield areas of southern Africa, eastern South America, and India.
The non-pegmatitic occurrences are mainly in the younger orogenic belts, related to large granitic intrusions.

The close spatial relationship of beryllium deposits, both pegmatitic and non-pegmatitic, to granitoids is widely recognized. A primary factor in the genesis of beryllium deposits is the original concentration of Be in the parent igneous body.

Beryllium occurs as an essential constituent in 40 minerals. Beryl was earlier the only beryllium mineral known to form economic deposits. Bertrandite, phenacite, helvite, chrysoberyl and barylite, previously thought to be rare, are now known to occur in large bodies of present or potential economic value.

The largest known beryllium deposits are the epithermal mantos at Spor Mountain, Utah, in which a rhyolite ash bed (water-laid tuff) containing as much as 65 % of limestone and dolomite in pebbles and cobbles was replaced by montmorillonite, silica, fluorite, adularia and bertrandite (Griffitts 1973). This is the only non-pegmatitic deposit in present operation, with 5 mill. tonnes of proved bertrandite ore with 0.22 % Be. Similar topazrock related deposits occur in Utah (Honey Comb Hills, Juab County).

On the Western Seeward Peninsula, Alaska, large replacement deposits of beryllium and fluorine contain fluorite, chrysoberyl, diaspore, muscovite and tourmaline, with trace to small amounts of euclase, bertrandite, helvite, phenacite, todorokite and hematite. Beryl occurs sparsely in late veins of quartz and fluorite. Chrysoberyl is the earliest and most common beryllium-mineral, followed by euclase and bertrandite, phenacite and beryl. Helvite is restricted to banded skarns which consist of magnetite and fluorite (Sainsbury 1969).

The Aguachile deposit in Mexico is similar. This bertranditebearing fluorite deposit consists of several disconnected fluorite bodies formed by replacement of brecciated, fractured and altered Cretaceous limestone along the contact of a ring-dyke of rhyolite porphyry (Van Alstine 1962).

At Seal Lake, Labrador, barylite and eudidymite occur in a zone of alkaline soda-rich paragneiss associated with syenitic intrusions. Beryllium mineralization occurs also in heterogeneous migmatites and metasomatized shear zones. Nb is also enriched, together with minor zinc and rare-earth minerals. The beryllium and other rare-element minerals are fine grained (about 0.2 mm), and occur in aggregates or lenses parallel the gneiss foliation (Mulligan 1968). A similar alkali metasomatic deposit is Vishnevye Gora, USSR (Zhabin et al. 1960).

The Thor Lake rare-metal deposit (Be, Y, REE, Nb, Ta, Zr, Ga) in Yellowknife, Canada is associated with the core of an Archean alkaline-syenite intrusive complex. The ore deposit is interpreted by Trueman et al. (1985) to represent the late stage, highly fractionated magmatic and metasomatic products of the complex. The unusual alteration assemblage appears to be the product of metasomatic replacement of earlier-formed syenitic and granitic types.

Another potential source of beryllium is a 40 million ton deposit in Central France, a granite containing beryllium as well as lithium, niobium, thantalum and lead (Farr 1980).

Examples of beryllium-skarn deposits are Iron Mountain and Victoria Mountains, New Mexico, Quitman Mountains, Texas and Långban, Sweden. At Iron Mountain the contact metamorphic beryllium deposits occur in irregular bodies of tactite formed by replacement of Paleozoic limestone, generally at or near contacts with small intrusive masses of aplite and fine grained granite.

In Norway Be-occurrences are known from both plumasitic and agpaitic pegmatites, the former mainly in Precambrian rocks, the latter as nephelinesyenite pegmatites in the Oslo region, where also contact metasomatic (skarn) deposits containing beryllium minerals occur. Other minor types are disseminations and cavity fillings in nordmarkite, Drammens granite, elpidite granite, granitic gneiss, quartz, porphyry, albitefels and mica schist, and as hydrothermal vein occurrences.

The beryllium deposit at Bordvedåga fall in the group "disseminated deposits" of Mulligan (1968), and has several analogues in other countries with different ore mineralogy and host rock (e.g. (sodic) paragneiss, rhyolitic tuff and porphyry, schist and limestone).

6.2. Occurrences of granitophile elements in the Høgtuva window

6.2.1. Introduction

The Høgtuva window consists mainly of metaluminous and minor peraluminous biotite-granitic gneisses; the mineralization is associated with the peraluminous biotite-granitic gneisses. The metalliferous occurrences are characterized by enhanced values of a suite of elements such as Fe, Na, Be, Nb, Zr, Y, Rb, Zn, Pb, U, Th, Sn, W, Ce, La, Co and depletion of K, Si, Ca, Ba and Sr.

6.2.2. The mineralization at Bordvedåga

Geology

The geological map of the eastern part of the window including the mineralization is presented in plate 2. The rare-metal enriched gneiss (MG) is stratiform, 7-8 km long and up to 400-500 m wide (as defined by content of Zr > 0.2 %) slice dipping 25-30^{o} to the NE in a suite of fine (to medium) grained gneisses in which thin (dm-thick) carbonate layers are interbedded. The gneisses are intruded by dolerite dykes or sills. Beneath the mineralization occurs a coarse grained massive to weakly foliated granite. Both lithological contacts and mineralization are concordant to foliation, while some of the amphibolites are discordant.

The MG is divided into a weakly mineralized zone (WMZ), defined by values > 15 000 counts/min. as measured with a Knirps scintillometer or > 0.2 % Zr - a highly mineralized zone (HMZ) defined by > 40000 counts/min. or > 0.8 % Zr, and a beryllium-mineralized zone (BMZ).

The MG is grey and well foliated like the surrounding fine grained gneisses. In addition to higher radioactivity, the mineralized zone is distinguished by the restricted occurrence of amazonite in concordant to low-angle discordant pegmatite lenses inside the zone. Pegmatite lenses are found throughout the whole window, but amazonite is found only in a few localities in addition to MG-area.

Segregations of zircon, quartz, biotite, thorite and monazite occur sporadically in the mineralized gneiss.

The gneiss shows little sign of alteration, except for weak sericitization of feldspar, occurrence of riebeckite around hornblende and magnetite. Weak chloritization of biotite and hornblende is seen, and also some secondary biotite. Flame-perthite and myrmekite are common.

Deformation has caused extensive granulation, producing lobate and sutured grain boundaries. Mortar texture is commonly observed - the quartz showing undulatory extinction and polygonization. Other signs of deformation are ductile bending of feldspar and biotite, and brittle deformation of feldspar and quartz.

The mineralized gneiss is unique in its rare metal enrichment and complex mineralogy. The modal composition is quartz (40-50 %), microcline (20-30 %), plagioclase (15-20 %) and biotite (8-10 %). Accessory minerals include fluorite, phenacite, zircon, opaques, allanite, epidote, calcite, apatite, hornblende, riebeckite, sphene, chlorite, and at least six unidentified phases (some of which are REE carbonates). The opaque assemblage includes large (up to 2 mm) magnetite crystals that host pyrophanite, cassiterite, a Zn-rich oxide phase and uraninite (Grauch & Lindahl 1984). Other opaques are pyrite, sphalerite, ilmenite, molybdenite, cassiterite and thorite. In the beryllium zone, phenacite is present as rock-forming mineral together with a new beryllium-mineral (here termed x-mineral).

The relationships between the various oxide phases are interpreted as the results of modification of the primary Fe-Ti-Mn-Zn-Sn-oxide phase during cooling of an igneous unit and subsequent regional metamorphism (Lindahl & Grauch 1986).

The HMZ is restricted to 100 m in width, and has the same element covariance as the WMZ, while the BMZ is 130-140 m long and 10-15 m thick and consists of several beryllium-enriched bands. This zone has a somewhat different element distribution.

In addition to this Be-zone other minor beryllium occurrences are found in the area. Some tens of meters to the NW a narrow, weaker beryllium-mineralization can be followed 200 m along the strike with the berylometer. 300 m north of the BMZ, and outside the radioactive mineralized gneiss, approximately 30 m^2 of high-grade beryllium mineralization was found to contain the x-mineral. Analyses from this zone are not yet available.

As previously stated (5.3.), beryllium is enriched in the fluorite - carbonate layers below the granite.
Poikilitic beryl crystals occur locally in fluorite-rich gneiss close to a fluorite layer. At Trolldalsaksla, in about the same stratigraphic position, a small mineralization occur in a dark carbonate-rich, pegmatitic band. It is extremely rich in rare metals (5.3.). The beryllium occur in danalite, helvite, gadolinite and the new beryllium mineral.

Geochemistry

The mineralized gneiss is slightly peraluminous, in contrast to the nearby granite and gneisses which is metaluminous. This is due to lower K_2O content rather than higher Al_2O_3.

Only one sample of the granite has as yet been analysed, showing it to be a metaluminous, differentiated, high-Si granite (74 % SiO_2) with very low MgO (< 0.01 %) and CaO (0.48 %), high FeO/MgO (229), and trace element characteristics analogue to the MG (tab. 4). The granite, in similarity to the mineralization, is low in Cu and V compared to average granite (Taylor 1964). Unlike the mineralization it is low in Mo and Co. Weak molybdenite impregnation is, however, seen along both upper and lower contact, both in the granite and the gneiss.

Tab. 4: Mean major and trace element compositon in the beryllium mineralized zone, highly mineralized zone, weakly mineralized zone, host rock, and the mineralized layers from the SW margin of the window.

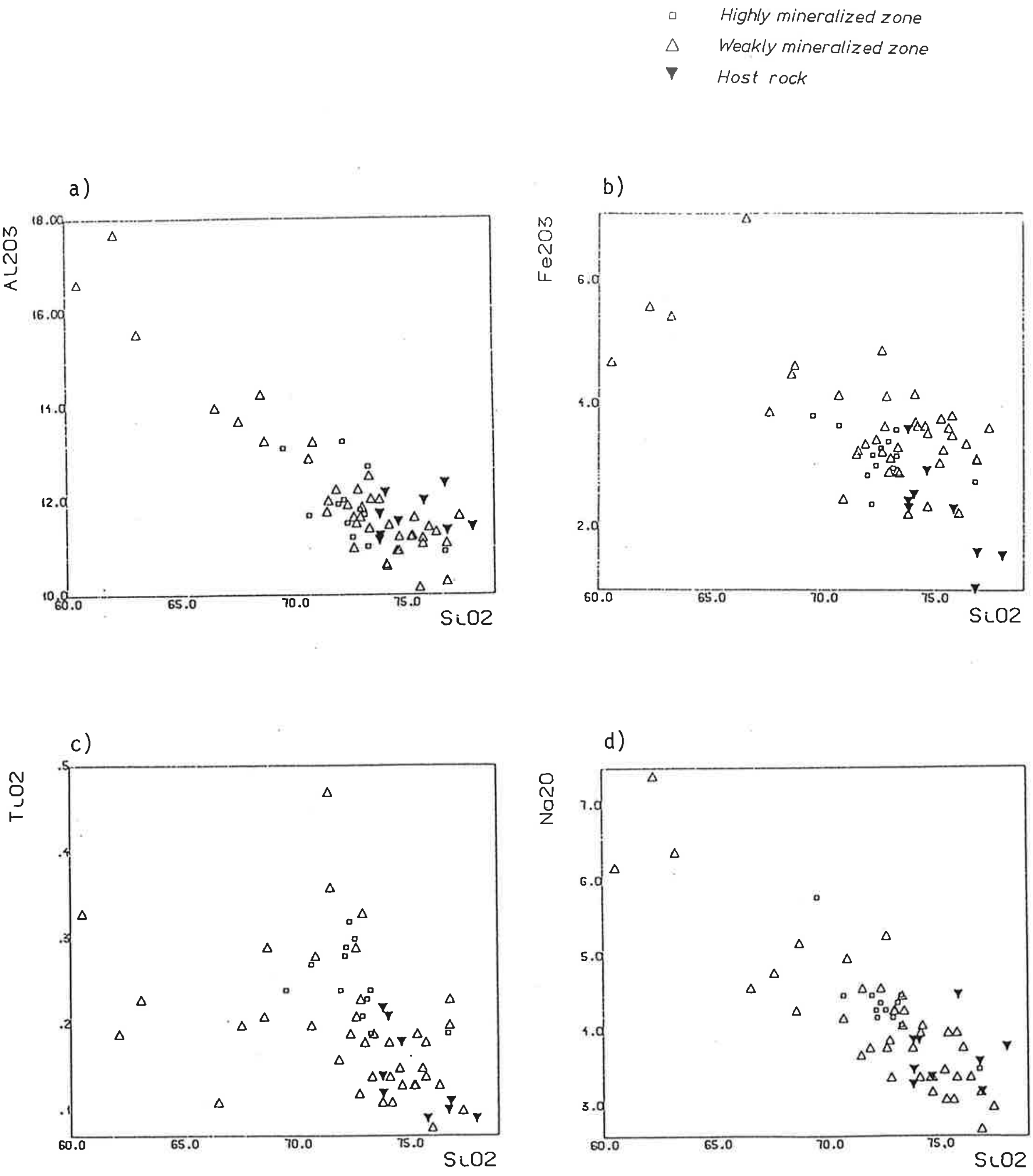
	BMZ	HMZ	WMZ	Host rock: Hanging wall gneiss	Host rock: Foot wall granite	Mineralized layers from the SW margin of the window
n	3(11 trace el.)	58	12	4	1	3
SiO_2	69,93	72,60	72,85	75,12	74,09	75,28
Al_2O_3	12,62	11,92	12,16	11,50	12,18	10,49
Fe_2O_3	4,21	3,18	4,10	2,32	2,55	4,66
TiO_2	0,24	0,25	0,16	0,15	0,21	0,25
MgO	0,13	0,25	0,38	0,26	0,01	0,12
CaO	0,28	0,35	0,45	0,42	0,48	0,48
Na_2O	5,30	4,39	3,85	3,65	3,90	3,20
K_2O	2,93	3,60	4,41	4,53	5,21	4,13
MnO	0,02	0,04	0,03	0,03	0,05	0,04
P_2O_5	< 0,01	< 0,01	< 0,01	< 0,01	0,01	< 0,01
K_2O/Na_2O	0,55	0,82	1,15	1,24	1,34	1,29
Nb	606	457	101	42	33	60
Zr	12800	10000	2900	850	719	2500
Y	1160	1100	383	167	107	300
Sr	17	26	20	19	28	40
Rb	1287	1200	670	433	421	280
Zn	255	373	154	71	135	85
Cu	11	11	8	< 5	< 5	7
V	< 5	< 5	< 5	6	< 5	< 5
Ba	25	28	28	70	70	150
Sn	192	125	45	< 10	15	22
Mo	23	9	13	< 5	< 5	18
U	321	232	56	16	< 10	40
Th	702	408	114	47	34	50
Pb	215	482	72	54	52	35
Co	81	70	19	8	< 5	18
Ce	448	457	420	309	201	540
La	152	144	157	140	88	220
Be	3700	30	16	< 10		17

The MG has a SiO_2 range of 65-78 %, with the HMZ restricted to 70-73 %, and the host rock clearly higher at 74-78 % SiO_2.

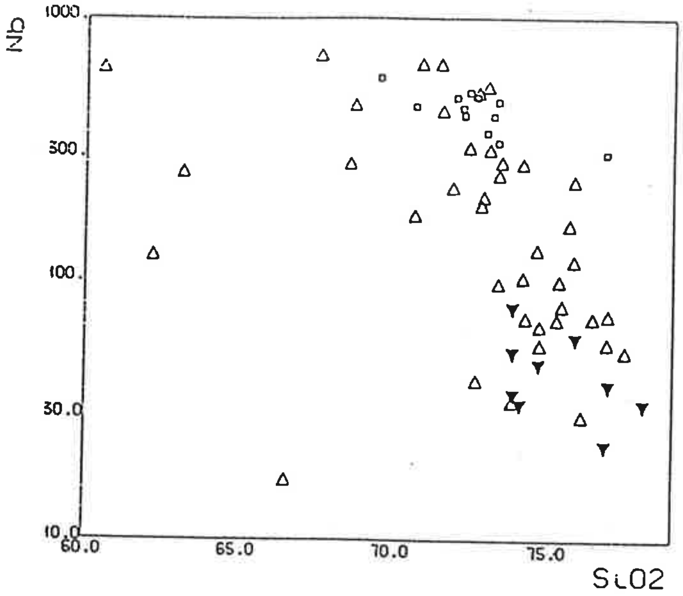
Most characteristic for the MG is lower abundances of SiO_2, K_2O and Al_2O_3, and higher Fe_2O_3, Na_2O and the suite of trace elements (tab. 4).

Harker diagrams (fig. 18), including both MG and host rock, show straight negative correlation against silica for Al, Fe and Na, while no element shows positive correlation with silica. However, most of the trace elements including Ti have a maximum content corresponding to about 72 % SiO_2, and decrease with increasing silica. K, Fe/Mg, Sr and Ba show the opposite trend. This maximum/ minimum at about 72 % SiO_2 is due to the fact that silica reaches a minimum value in the richest part of MG (SiO_2 decreases with increasing mineralization).

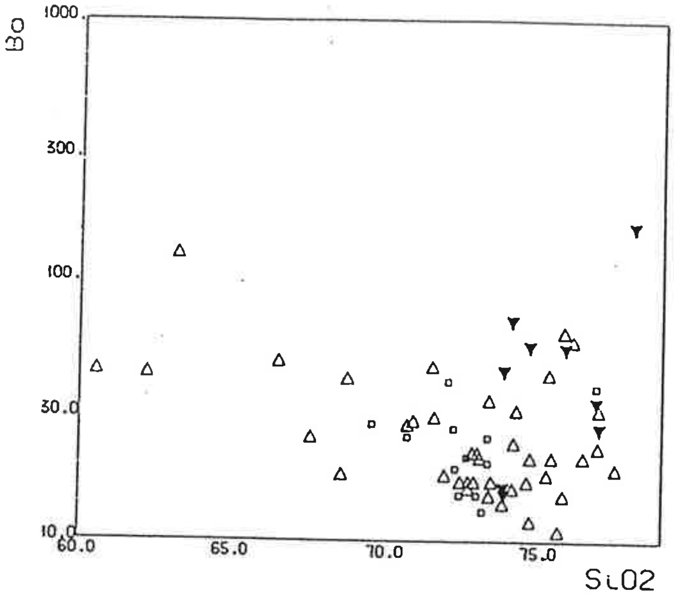
Fig. 18: Selected major and trace element plots for samples from the mineralized gneiss (HMZ and WMZ) and immediate host gneiss at Bordvedåga.



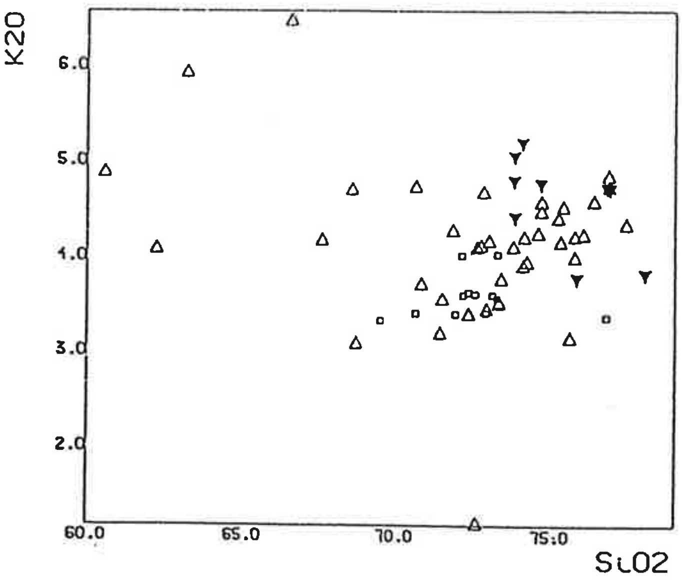
e)


Nb
1000.
300.
100.
30.0
10.0
60.0
65.0
70.0
75.0
SiO2

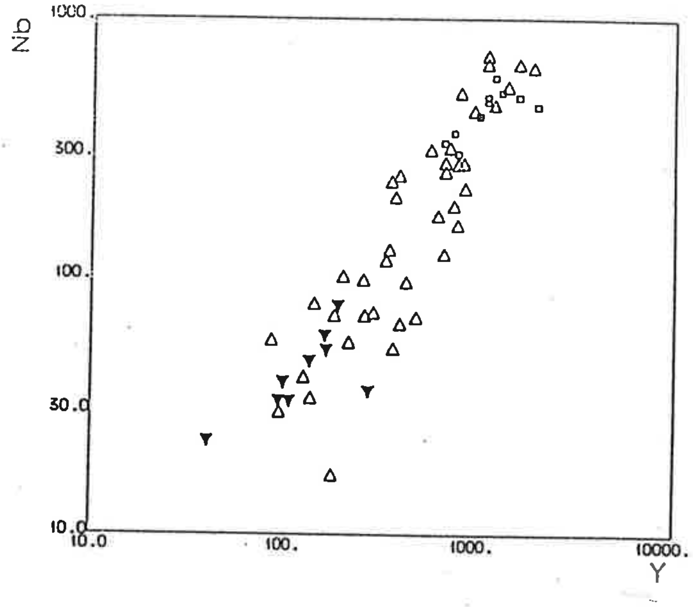
f)


Ba
1000.
300.
100.
30.0
10.0
60.0
65.0
70.0
75.0
SiO2

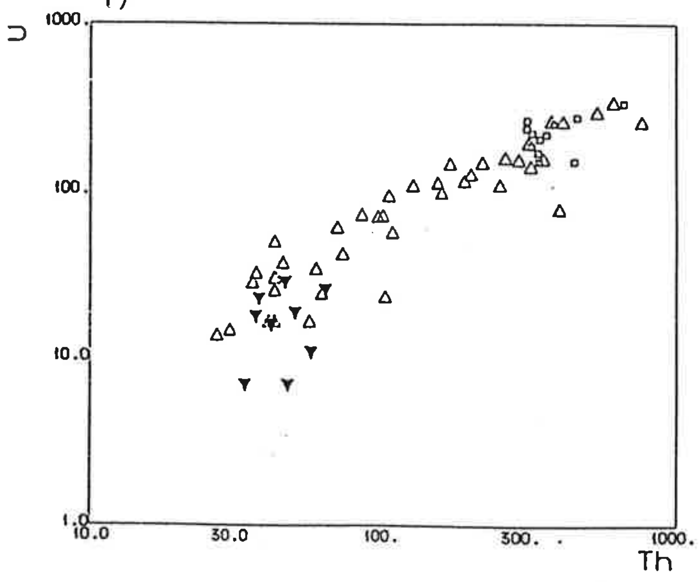
g)


K2O
6.0
5.0
4.0
3.0
2.0
60.0
65.0
70.0
75.0
SiO2

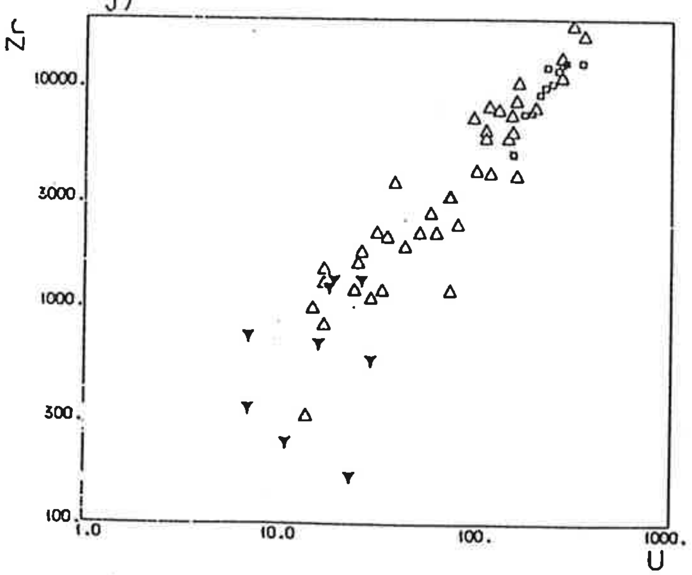
h)


Nb
1000.
300.
100.
30.0
10.0
10.0
100.
1000.
10000.
Y

i)


U
1000.
100.
10.0
1.0
10.0
30.0
100.
300.
1000.
Th

j)


Zr
10000.
3000.
1000.
300.
100.
1.0
10.0
100.
1000.
U

Continuous evolution from the footwall granite (and partly the hangingwall gneiss), through weakly and highly mineralized zones to the beryllium zone, can be seen in the main constituents: SiO_2, K_2O and CaO are depleted while Na_2O follows an enrichment trend (tab. 4). The depletion in K is counterbalanced by increase in Na, so that the total alkalies are nearly constant. The other main elements do not follow these continuous trends. Al_2O_3 and Fe_2O_3 have highest content in the BMZ, while the distribution of TiO_2 and MnO are irregular. MgO has a continuous negative evolution trend (decreases with increased mineralization) in the MG, but is extremely low in the footwall granite and in some samples from the hangingwall gneiss.

Most trace elements show increasing enrichment from the footwall granite through WMZ to the HMZ, e.g. Nb, Zr, Y, Rb, Zn, Cu, Sn, U, Th, Pb, Co and Be, with a corresponding depletion in Ba and Sr. Mo, Ce and La do not follow any of these trends, but they are highest in the WMZ.

The BMZ is further enriched in the elements Nb, Zr, Y, Rb, Sn, U, Th, Co and Be, while Ba, Sr, Zn, Pb, Ce and La are depleted compared to the surrounding HMZ. Mo does not follow a regular trend, and are richest in the BMZ.

Through comparison of the drill-core analyses (fig. 19), a more detailed correlation can be made between the BMZ and the immediate host, the HMZ. It can be seen that the BMZ is enriched in Be, Sn, Mo, U, Th and Nb, and depleted in Ce, La, Y, Zn, Rb and Sr relative to the HMZ. Zr, Ba, Cu, Pb and Co show no preferred enrichment or depletion in any of the zones.

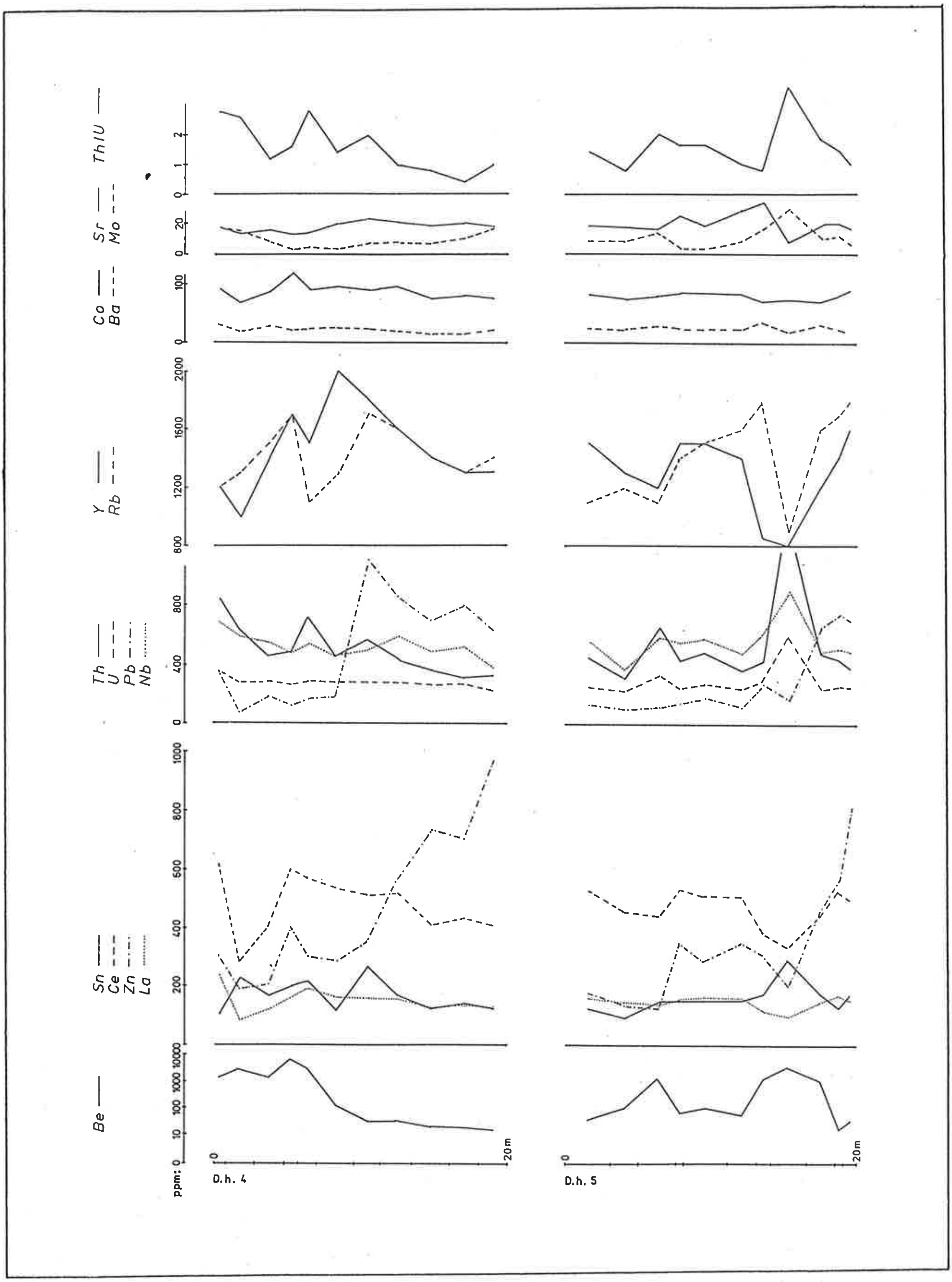


Fig. 19: Vertical profiles of drill holes 4 and 5, with trace element composition.

In the BMZ beryllium show a positive correlation only with Sn, negative correlation with Ce, La, Y, Sr, Ba, Zn and Pb, and is independent of Cu, Co and Rb. Some elements, U, Th, Nb, Zr and Mo, show positive correlation with Be in one drill-hole and negative in the other.

This observation is not in accordance with the element associations from the mineralized gneiss outside the BMZ, where all the trace elements show positive correlation, except Ba and Sr which have a negative correlation with the others. The similarity in trace element distribution in the MG suggests that all were concentrated by the same mineralizing fluids under similar conditions. The different situation in the BMZ, lack of covariation between Be, Sn and the other elements, suggests changing conditions during deposition, possibly because of transition from magmatic to increased post-magmatic activity from which the latter gave rise to beryllium mineralization together with enhanced Sn, Mo, U, Th and Nb.

The beryllium mineralization associated with the carbonate - fluorite layers seems to have resulted from hydrothermal fluids following zones of weakness and interacting with carbonate.

The transporting agent(s) for the lithophile elements is another problem. The complexing ability of F is mentioned, but insufficient analyses have been made of the volatiles and anionic complexes F, Li, B and Cl. These elements do not, however, show a clear relationship to the incompatible elements, though the amount of fluorite in the MG is high.

The variation in chemistry is shown schematically in a profile across the mineralized gneiss at Bordvedåga in fig. 20.

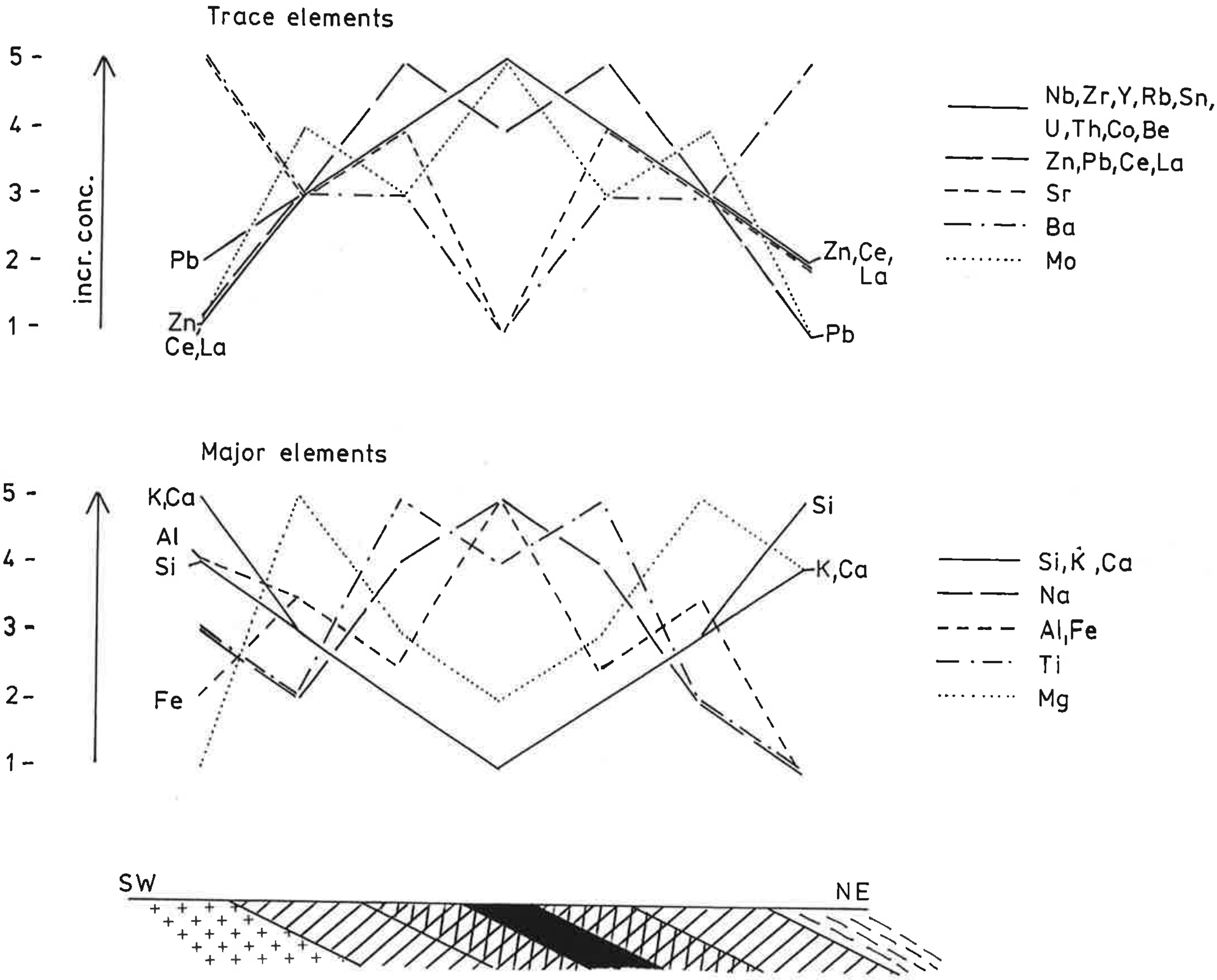


Fig. 20: Geochemical profile across the mineralization at Bordvedåga. Distribution of major and trace elements. Same symbols as plate 2.

Chondrite normalized REE patterns are presented in fig. 21 (from Grauch & Lindahl 1984), and show that the MG is enriched in heavy REE - up to 1000 times the chondrite norm - but has significantly by low Eu.

Recent research results seem to indicate that REE distribution could be a useful tool in recognizing favourable volcanic cycles (Thurston 1981, Campbell et al. 1984). Ore bearing volcanic rocks show a flat REE pattern and a pronounced Eu anomaly against the steep REE development of barren felsic volcanic rocks.

Enrichment in heavy REE points to derivation by magmatic differentiation from granite. Residual hydrothermal fluids can also result in enrichment of the heavy REE by autometasomatic processes, but tend to obliterate the Eu-anomalies (Bowden & Whitney 1974).

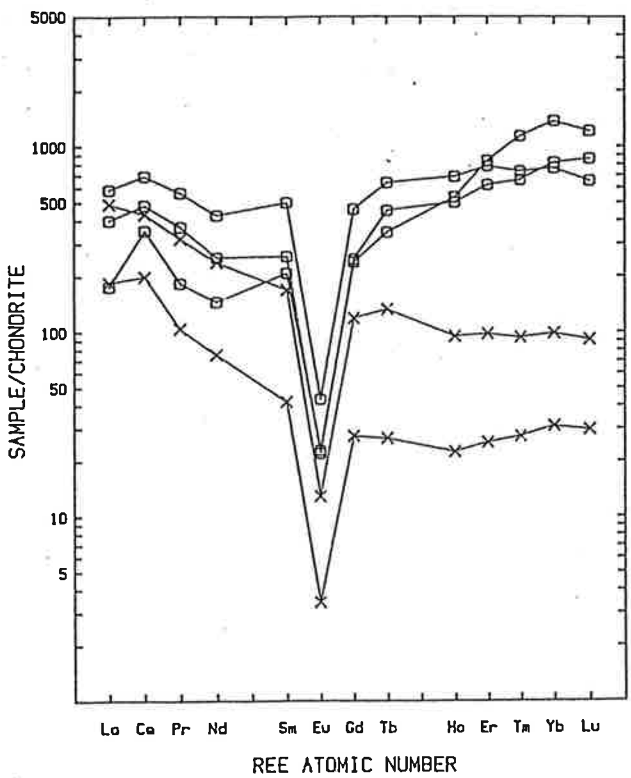


Fig. 21: Chondrite-normalized rare earth element patterns for the mineralized gneiss at Bordvedåga (□) and for host rock gneiss (x).

6.2.3. Other mineralizations

Mineralizations similar to the MG occur (up to 15 km distant) on the southwestern margin of the Høgtuva window (plate 1). They occur as three separate 0.5-1 m thick concordante layers of fine grained (0.4-0.5 mm), leucocratic gneiss, with characteristic increase in radioactivity and Zr content compared to the wallrock. The mineralized rocks have massive appearance due to low biotite content (c. 4 %). Quartz, microcline and plagioclase occur in amounts of respectively 35 %, 25 % and 20 %. Opaques, mainly magnetite, are abundant (3-4 %), as is also a typical intergrowth of epidote minerals, known from the MG, with epidote, zoisite and allanite. Common accessories are chlorite, muscovite, zircon and apatite. Also present is a Nb-oxide, identified with microprobe from the MG, which is often zoned with a clear core and yellow rim.

These layers closely resembles the MG, and specially the WMZ, in chemical composition (tab. 4) and peraluminous character (fig. 5). They are slightly higher in SiO_2, Fe_2O_3, Ba, Sr, Mo, Ce and La, and slightly lower in Al_2O_3, alkalies and the other trace elements. The U/Th ratio is lower than in the WMZ (1.3 compared to 2.0). Rb, Ba and Sr abundances point to less differentiation for these layers compared with the WMZ. Rb/Sr ratio is 7 compared to 46 for WMZ.

Another meter-thick, radioactive layer at Melfjellet is more enriched in Ce and La (645 and 259 ppm) than the MG, and has a U/Th ratio of 2.

A heavy mineral survey done by NGU in 1983 (Wilberg 1987) has indicated an assemblage fairly similar to the MG in the NW-corner of the window.

A completely different occurrence is the molybdenite impregnation in the contact zone of the granite underlying the MG. This is low-grade and restricted; there are no indications that molybdenite deposits of any size occur in the window.

6.2.4. Some aspects of selected elements

Beryllium

Most of the samples were treated with HNO_3. Solubility of phenacite in HNO_3 is, however low. When decomposed with HF the solubility of Be increases in the order of 8-10 times in the phenacite disseminated gneiss. Analyses was done by atomic absorption.

Several Be-minerals are present in different rock types and localities in the Høgtuva window.

These are:
- Phenacite ($Be_2 SiO_4$) disseminated in the granitic gneiss at Bordvedåga.
- Beryl ($Be_3Al_2Si_6O_{18}$) in fluorite-rich gneiss bordering a carbonate-fluorite layer in the Bordvedåga area.
- Danalite ($Fe_8(BeSiO_4)_6 S_2$), helvite ($Mn(BeSiO_4)_6 S_2$) and gadolinite ($Y_2FeBe_2O_2(SiO_4)_2$) in a carbonate-fluorite rich, dark pegmatitic band at Trolldalsaksla.
- X-mineral in the same band, and associated with the phenacite in the gneiss at Bordvedåga.
- Gadolinite in pegmatites, both in the Høgtuva window and in the cover.

Detection limit is 5 ppm for Be. Most samples outside the MG have less than 5 ppm.

The average beryllium content of biotite granite, according to Beus (1962), is 4.1 ppm.

One drill profile through the BMZ intersected two zones of Be mineralization averaging respectively 0.31 % Be over 5.4 m and 0.19 % over 2.6 m, and separated by 5.5 m of barren rock. The average for the whole 13.5 m section is 0.16 % Be. The highest value is 0.82 % Be in 0,6 m of drill core.

In the small, carbonate-hosted occurrence at Trolldalsaksla most of the beryllium is contained in the complex Be-silicates danalite and helvite. These are the most abundant Be-minerals in skarns. Their restricted formation can be explained in terms of sulfur pressure and the necessity for numerous other cations, notably Fe, Mn and Zn, to be available during their formation. The sulphides in the Trolldalsaksla occurrence include minerals of Cu, Pb and Zn.

The other carbonate-fluorite layers in the Bordvedåga area contain up to 175 ppm Be. The rare-metal enriched layers of fine grained gneiss at the western margin contain 15-20 ppm Be. Other rock types show only few scattered values above detection limit. These higher values are always accompanied by anomalous Sn.

Tin

In the MG tin is known to occur as cassiterite in ilmenite lamellae in magnetite. A very few discrete cassiterite grains have been identified. Most of the Sn is thought to be bound to biotite.
Analyses of whole-rock and biotite concentrate from one sample from the HMZ gave respectively 66 and 466 ppm Sn. With 10 % biotite in the gneiss, this means that 70 % (46 ppm) of total Sn is bound to biotite, either in the lattice or as discrete cassiterite grains.

Heavy mineral concentrates (with magnetite removed) from samples taken downstream from the deposit gave up to 174 ppm Sn, while one would expect higher values if considerable amounts of the Sn occurred as discrete cassiterite grains.

While most of the rocks have Sn contents lower than the detection limit (10 ppm), the layers of fine grained gneiss contain 15-45 ppm, and the fluorite zones 12-20 ppm. Average values for the BMZ, HMZ and WMZ are 192, 125 and 45 ppm.

The Caledonian cover sequence in the Rana region contains several massive sulphide deposits where cassiterite and stannite occur as accessory minerals (Vokes 1960 and 1963, Saager 1967). Tin is known to be a minor component in many submarine exhalative ore deposits. In deposits of Archean age in Canada, Mulligan (1975) cites tin contents of 0.14 % for Kidd Creek and 0.25 % for South Bay. The acid volcanic Proterozoic deposits of Boliden in Sweden contains 0.006 % Sn as stannite (Grip & Wirstam 1970), whereas the sediment-hosted Proterozoic deposit of Sullivan (Canada) is well known for its tin content which, although averaging 0.05 %, contains locally up to 2 % (Mulligan 1975). Both cassiterite and stannite occur in the Paleozoic sediment-hosted Bleikvassli deposit (0.03 % Sn), about 90 km south of the Høgtuva window (Vokes 1960 and 1963).

Badham (1982) argues that there is an association between Fe-Zn exhalative ores and high tin values, while Plimer (1983) suggests that the association of minor amounts of Sn in exhalative ores is more closely correlated with the presence of F-minerals (e.g. Kidd Creek, Canada and Iberian Pyrite Belt) or tourmaline (e.g. Sullivan and possibly Bleikvassli). Ishihara & Terashima (1983) point out that the Kuroko type cassiterite bearing, massive sulphide deposits in the Canadian Shield are associated with Sn-rich volcanic rocks, while tin is very poor in the volcanic rocks around the Sn-free Kuroko deposits in Japan.

In norwegian rocks tin enrichments are rare. Cassiterite and stannite as accessory minerals in massive sulphide deposits are reported only from four deposits, except in the Rana region where cassiterite and stannite are found in several deposits (Vokes 1960 and 1963, Saager 1967). Cassiterite is found in a Li-Be bearing pegmatite at Ågskaret in the same region. These occurrences in the Caledonian nappes and the tin-mineralized gneiss in the Høgtuva window indicates a tin-province in the Rana region. Sn from the basement could have been mobilized and redeposited in the overlying Caledonian nappes, which is also suggested for beryllium by Lindahl et al. (in prep.) who have defined a Be-province in the basement and basal Caledonian nappes in the Rana region.

Fluorine

Very few analyses of fluorine in the Høgtuva rocks exist, but fluorite is a common accessory mineral, specially abundant in the MG at Bordvedåga.

F enrichments:

- Fluorine is enriched in the MG, with an average of 1320 ppm from 20 samples, compared to the averages for rhyolite and granite of 480 and 870 ppm respectively (Fleischer et al. 1963).
- Fluorite is abundant in the amazonite pegmatite lenses in the MG and in the underlying granite.
- Massive, dm-thick fluorite layers and dispersed fluorite in carbonate layers.

- Discordant fluorite enrichments in the gneiss close to these layers.
- Tectonic biotite schist rich in fluorite (20-30 % fluorite in one locality).
- Late movement of fluorine, precipitated as fluorite in late fractures.

Uranium

Uranium is enriched in the more fine grained gneisses, notably the thin fine grained gneiss layers (40 ppm), muscovite gneiss (21 ppm), the rusty gneiss (45 ppm) and the MG (tables 3 and 4), and also in pegmatites and biotite-zircon-rich segregations. The highest value detected is 1.01 % U from the mineralization at Trolldalsaksla.

Silver

Silver is low in the gneisses, typically about 0.5 ppm. It is slightly enriched in the MG to about 1.0 ppm. The massive fluorite layers have 3-4 ppm silver.

Lithium

Abundances are very variable, and seem more related to the amount of biotite present than to the concentration of incompatible elements.

6.3. Discussion of ore geology

A large portion of the rocks in the Høgtuva window is thought to consist of supracrustals of both volcanic and sedimentary origin. Especially in the area of mineralization, the occurrence of carbonate layers with thin intercalations of quartzite suggests this.

The main foliation appears to be almost concordant with the original stratification, and the Bordvedåga mineralization is parallell to this foliation, i.e. the mineralization is stratiform. However, locally it can be seen that a mineral banding which could be primary cross-cuts the foliation (see 5.4.).

The intense deformation and metamorphism of these rocks has destroyed most of the primary textures. However, the mineral composition, with abundant feldspar, and the chemical composition, point to a magmatic, extrusive protolith of granitic composition for the mineralized gneiss.

The very fine distribution of trace elements (e.g. cassiterite in ilmenite-lamellae in magnetite) points to a fine grained protolith, possibly a rhyolitic flow or tuff.

It is suggested that the mineralization in the extensive WMZ was deposited as synvolcanic enrichment of the trace element suite (all of which have good positive correlation) by magmatic/late magmatic differentiation. The mineralization might be ascribed to synvolcanic, autometasomatic processes. The possibility of magmatic differentiation and primary enrichment of the rare metals is not critical. The Thor Lake rare-metal deposit is an example of this mecanism.

Syngenetic mineralization (or possibly autometasomatic alteration) is consistent with widespread regional occurrence. The MG is 7-8 km long and similar mineralizations are found up to 15 km away on the other side of the window. The extent of the MG also indicates an extrusive origin.

Alternative hypotheses for the origin of the protolith are suggested by Lindahl & Grauch (1986): metal-rich (dyke-like) intrusive unit, or mineralization along granite contacts or alang fault zones.

The mineralization in the BMZ might be ascribed to a second postmagmatic stage of mineralization, where hydrothermal fluids give rise to enrichment in Be, Sn, U, Th, Nb and F, and depletion of Ce, La, Y, Zn, Rb and Sr compared to the WMZ and HMZ. Interactions of these solutions with carbonate could have led to the deposition of fluorine together with minor amounts of beryllium.

Fluorine tends to be associated geochemically and form stable complexes with the lithophile elements, including U, Be, Li, Mn, Nb, Sn, W and Y. Fluorine is concentrated in alkalic and silicic hypabyssal and extrusive rocks and related hydrothermal deposits. It is well known that the distribution of Be in volcanic rocks is closely related to the distribution of fluorine (Coats et al. 1962).

There is no visible hydrothermal alteration in the MG. All observed mineral convertions are due to later metamorphic processes.

A problem which remains unsolved is the MG's close resemblance to peralkaline rocks (extreme enrichment in trace elements such as Nb, Ta, Zr, Mo, Zn). Peralkaline rocks are, however, not known to occur in the Høgtuva window (except a few samples). The major-element composition of the mineralized gneiss resembles that of subalkaline granites associated with Proterozoic peralkaline granites and syenites in Labrador and the Northwest Territories, Canada, which host Be and REE deposits.

The MG is highly differentiated, with a high abundance and covariation of a range of trace elements, and very low Ba and Sr content. However, the positive correlation of Zr with other lithophile elements is not typical of a magmatic differentiation process. The high Zr content is some granites can be attributed to the solubility and relative mobility of zircon in the presence of excess alkalies (Watson 1979). Fluorine is also believed to have similar effects on zircon crystallization in the presence of excess alkalies, forming alkalie-zirconofluoride complexes.

The slightly peraluminous MG has many chemical and mineralogical characteristics in common with varieties of pantellerites (Washington 1913), but contains less alkalies, Fe_2O_3 + TiO_2 and more Al_2O_3 and SiO_2 (which make it more similar to commendites). The trace element content, however, strongly resembles pantellerite.

Peraluminous granitic rocks contain a variety of aluminium rich phases which include muscovite, biotite, garnet, aluminosilicates (andalusite, sillimanite, kyanite, topaz, staurolite), cordierite and tourmaline. This characteristic mineral assemblage is not present in the peraluminous MG, with the sole exception of biotite. The peraluminous character is conferred by low K_2O contents rather than by high Al_2O_3.

The occurrence of Na-amphibole locally in the MG and underlying granite is propably due to late, local movement of Na.

Though there has been found no petrographic evidence for this, the question may nevertheless be raised whether an early Na-metasomatism (with which the mineralization was associated) has occurred in the MG. The association of alkali metasomatism with U-mineralization in the previously mentioned U-province in Sweden has been noted by Adamek and Wilson (1977), and this is also the case in several rare-metal deposits. Major-element variation during sodic metasomatism (albitization or riebeckitization) shows increase in Na and Fe and decrease in K, which is also the case in the rocks at Bordvedåga.

In case of magmatic differentiation causing the increase in trace element content in the MG, the behaviour of K is in disagreement with one of the principal laws regulating the geochemistry of alkalies, which is that K concentrates during magmatic differentiation as does Rb (Heier & Adams 1964).

Relationships between magmatic differentiation, volcanological evolution and occurrence of some trace elements in alkali-potassic volcanics have been pointed out by Locardi et al. (1967). In the cases studied by them, the more differentiated rocks have lower K contents and are oversaturated in alumina. This oversaturation is not only caused by secondary processes, but can be explained by a loss of alkalies in superficial differentiations related to the volcanic conditions.
Locardi et al. (1967) inferred two principal differentiation processes, one of which causes silica increase (normal differentiation) and the other silica decrease. The desilication is due to pneumatolytic differentiation.

Correlation between desilication and increase in pneumatophile trace elements (U, Th, Be, Zr, Rb) is verified in the MG at Bordvedåga. Up to a certain point the concentration of pneumatophile trace elements increases as the pneumatolytic differentiation increases. As differentiation processes proceed the major (Si, K) and trace elements follow two different paths.

The K/Rb ratio is not constant, but in the higher differentiates this ratio decreases because a relative increase of Rb versus K. However, according to Tischendorf (1977) this behaviour of K is a general trend in differentiation, and the ratio K/Rb can be used for characterizing the stage of differentiation. Because of the different behaviour of the elements in the evolution of a normal and a specialized granitoid rock, a different correlation behaviour (positiv, negative) can be observed. Furthermore, the absolute values can be used to disinguish both granite types. Tischendorf (1977) proposes the distinguishing ratio-value of less than 100 for specialized granites. The values for the MG and the underlying granite at Høgtuva are respectively 50 and 130.

As with K/Rb there is a pronounced decrease in the ratios K/Th and K/U from the hostrock-gneiss through WMZ and HMZ to BMZ.

7. POTENTIAL FOR GRANITOPHILE ELEMENT DEPOSITS

A combination of geological mapping, geochemical studies of the rocks and panning of stream sediments have delineated anomalous concentrations of economically interesting elements of different associations:

- Bordvedåga-type rare-metal mineralization (WMZ to HMZ) at various localities in the window.
- REE-enriched Bordvedåga-type depleated in Sn (see 6.2.2.).
- Molybdenite mineralization at the granite-gneiss contact.
- Beryllium mineralization of various types.

Only the stratiform Be mineralization appears to have economic potential in the Høgtuva window so far, and the trace element enrichment associated with the Bordvedåga-type makes this type even more interesting.

Potential types of beryllium mineralization are (1) dissemination in the gneiss, both stratiform in the fine grained gneiss and discordant mineralization bound to fluorite-enriched gneiss near fluorite layers, (2) in the carbonate-fluorite layers, and (3) in pegmatites.

The WMZ at Bordvedåga is a large rare-metal deposit of too low grade to be utilized. Of possible economic interest is the Be-enriched part, with perhaps byproducts from Nb, Y, REE and U.

The BMZ has been outlined by use of berylometer (a neutron activation field instrument) and near-surface drilling in 9 profiles. The zone has been traced 130 m along the strike, and is about 14 m thick. Average grade for the whole thickness is about 0.2 % Be. The ore consists of several parallel, narrow bands rich in phenacite and x-mineral.
Proved ore is 40 000 tonnes with c. 0.2 % Be down to a depths of 10 m. Assuming a depths of 100 m an estimate of probable ore is in the order of 0.4 mill. tonnes.

The geology and variety and distribution of the known beryllium (± other rare elements) mineralizations in the PCB windows and cover rocks, suggests the possibility of finding additional deposits. By applying modern prospecting techniques both on the ground and in the air, new deposits of the "invisible" Bordvedåga-type may be found.

8. PROSPECTING METHODS

Airborne gamma-ray spectrometric surveys are of use in delineating potential target areas for further exploration for granophile element specialization.
The MG at Bordvedåga shows up on the total radiation map as 8-10 times back-ground radiation, with 8-10 x increase in the uranium channel and 6-7 x in the thorium channel (Håbrekke 1983).
The MG also produces a significant magnetic anomaly, due to the increased magnetite content. Therefore, a combined radiometric and magnetic anomaly is considered indicative of mineralization.

The MG is characterized by anomalous radioactivity, generally increasing together with the trace element abundance towards the central part with its BMZ. However, smaller high-grade beryllium occurrences without accompanying radiometric anomalies are also known. The potential for beryllium mineralization is therefore not dependent on the presence of significant radiometric anomalies.

The berylometer is most helpful in mapping beryllium mineralization at targets delineated by regional exploration.

Stream sediment sampling was not used in the Høgtuva window due to the high degree of outcrop and limited amounts of detritus.

Heavy mineral panning gave good results. The principal features of the results, which are given in detail in a seperate report (Wilberg 1987) are summarized below:

- Zr, Sn, Mo and Pb show significant positive anomalies related to the MG.
- Sc, V and Co show negative anomalies downstream from the mineralization.
- The other analysed elements show little tendency to vary in a manner relatable to the mineralization.

Other field methods that have revealed small beryllium mineralizations are scintillometer-survey and mineralogical studies in the field, the latter because the Be-bearing x-mineral is the only visible sign of beryllium mineralization in the gneiss.

It is realized that there is a genetic and geochemical relation between the types of granitophile element deposits and the associated granites. This consept of geochemical specialization is commonly used in exploration for such deposits, with concentration of certain elements taken as indicative of potential mineralization (Tischendorf et al. 1978).

Sn-W mineralization is commonly associated with peraluminous granites (Stemprok & Sulcek 1969), although there are exceptions (Hesp 1974). The worldwide association of uranium with peralkaline rocks is well known. The success in the discovery of disseminated type beryllium deposits in Utah has led to exploration in other areas where high-siliceous, alkaline and peralkalic igneous rocks occur.

The extreme enrichment in such trace elements (as Nb, Ta, Zr, Mo, Zn) characterize peralkaline volcanics. No peralkaline rocks are, however, known in the Høgtuva window, which makes it somewhat unique among volcanic uranium and granitophile element deposits.

9. CONCLUSION

The 1700-1800 Ma Precambrian basement rocks exposed in numerous windows in the coastal district of Nordland, are dominated by granitic composition. They have a geochemical trend towards increased alkalinity from south to north in the region.
The regional rock-sample programme indicated five of the windows to contain more differentiated rocks favourable for rare-metal mineralization. In one of them, the Høgtuva window, a peraluminous, high-silica (72-73 % SiO_2) biotite-granite gneiss is host for a rare-metal deposit at Bordvedåga. This and similar, smaller metalliferous occurrences other places in the massif are characterized by enhanced values of a suite of elements such as Fe, Na, Be, Nb, Zr, Y, Rb, Zn, Pb, U, Th, Sn, W, Ce, La, Co and depletion of K, Si, Ca, Sr and Ba.

Most characteristic for the mineralized gneiss is lower abundances of SiO_2, K_2O and Al_2O_3, and higher Fe_2O_3, Na_2O and trace elements.

A continuous evolution can be seen in most main and trace elements from the footwall granite and hangingwall gneiss through weakly and highly mineralized zones to the beryllium zone.

It is suggested that the mineralization in the long mineralized gneiss zone was deposited as synvolcanic enrichment of the trace element suite (all of which have good positive correlation) by magmatic/late magmatic differentiation. Syngenetic mineralization is consistent with the widespread regional occurrence of these mineralizations.
The difference in correlation between the elements in and outside the BMZ suggests changing conditions during deposition.

A large portion of the rocks in the Høgtuva window is thought to be supra-crustals of both volcanic and sedimentary origin. The occurrence of fine grained quartz-feldspar gneisses, carbonate layers and very minor quartzites suggests volcano-sedimentary origin for the gneisses.
The protolith for the mineralized gneiss was a homogeneous and probably fine grained rock of extrusive origin.

Beryllium is the element considered as having the best potential for forming economic deposits in these Precambrian windows. Several beryllium-minerals are present in various rock types in the Høgtuva window, and also in the cover rocks as well as in other windows (in pegmatites).
Elements such as Nb, Y, REE, U, etc. seem, at least in the Høgtuva window, to be dependent on the presence of beryllium to be of economic potential.

Trondheim 1.2.1987

Rune Wilberg

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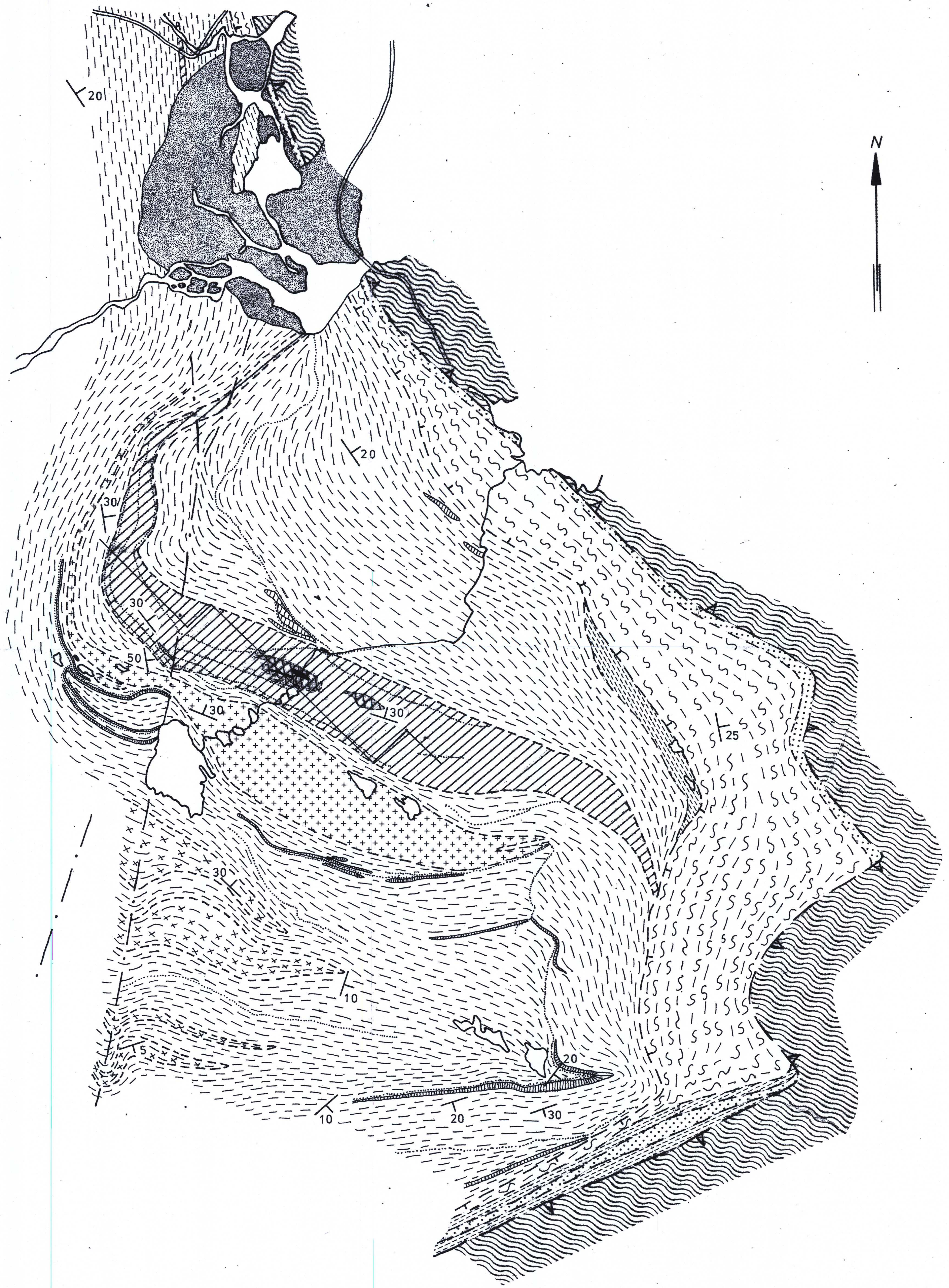
MELFJORDEN
Hyttan
Sandvika
Melfjordbotn
Gjervaldalen
Melfjellet
Høgtuva
Trolldalsaksla
Bordvedåga
Isvatnet
Snøfjellet
SJONA
N
LEGEND:
Caledonian cover
SNØFJELLET NAPPE
FEMFJELLET GROUP
q Quartzite
m.s. Mica schist, graphite bearing and rusty
HØGTUVA COMPLEX
Fine grained, biotite rich granitic gneiss
Coarse grained (>2 mm) granite
Hornblende - quartz monzonite
Medium grained (1-2 mm) granitic gneiss
Fine grained (<1mm) granitic gneiss
Mineralized gneiss, > 0,2 % Zr
at Bordvedåga
in thin layers other places
Porphyry gneiss
Granitic gneiss, deaply weathered
Amphibolite and biotite schist
Thrust boundary
Minor, local thrust boundary
Fault, joint, sometimes with brecciation
Glacier
Scale 1: 50000
0 1 2 3 km
PRELIMINARY GEOLOGICAL MAP
HØGTUVA MASSIF
MÅLESTOKK 1:50 000
OBS. R.W.
TEGN.
TRAC. R.W.
KFR.
NORGES GEOLOGISKE UNDERSØKELSE TRONDHEIM
TEGNING NR. 87.043-01
KARTBLAD NR. 1927 I og IV, 1928 III



LEGEND

- Superficial deposits, Quaternary
- Caledonian cover

SNØFJELLET NAPPE

FEMFJELLET GROUP

- Quartzite
- Mica schist, graphite bearing and rusty

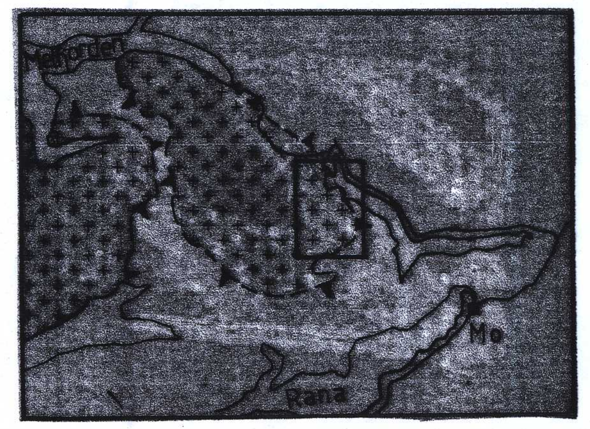
HØGTUVA COMPLEX

- Fine grained, biotite rich granitic gneiss
- Coarse grained granite
- Medium grained granitic gneiss
- Fine grained granitic gneiss
- Weakly mineralized zone, 0,2 % $<$ Zr $<$ 0,8 %
- Highly mineralized zone, $<$ 0,8 % Zr
- Beryllium mineralization
- Carbonate - fluorite layers
- Biotite schist and amphibolite

- Thrust boundary
- Minor, local thrust boundary
- Joint with brecciation
- Baseline for grid

Scale

0 1 2 km



GEOLOGICAL MAP
HØGTUVA

NORGES GEOLOGISKE UNDERSØKELSE
TRONDHEIM

MÅLESTOKK	MÅLT	R.W.
1:20 000	TEGN	
	TRAC	R.W.
	KFR.	
TEGNING NR.	KARTBLAD (AMS)	
87.043-02	1927 I	