NGU

Norges geologiske undersøkelse

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Geological Survey of Norway 1858–1983

125-year Anniversary Volume

Universitetsforlaget 1983

Trondheim – Oslo – Bergen – Tromsø

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Geological Survey of Norway

Norges geologiske undersøkelse (Geological Survey of Norway), Leiv Eirikssons vei 39, Trondheim. Telephone: national (07) 91 58 60, international + 47 7 91 58 60. Postal address: Box 3006, N–7001 Trondheim, Norway.

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EDITOR
Førstestatsgeolog Dr. *David Roberts,* Norges geologiske undersøkelse, P.O.Box 3006. N–7001 Trondheim, Norway.

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Geological Survey of Norway 1858–1983 125-year Anniversary Volume

A collection of papers on Norwegian geology to commemorate the 125-year anniversary of the founding of the Geological Survey

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Preface

The Geological Survey of Norway 1858–1983

The official date of foundation of the Geological Survey of Norway (NGU) is February 6th 1858. On that date a Royal Resolution was passed authorizing the geological mapping of the country according to a plan submitted by Theodor Kjerulf in 1856. NGU is therefore one of the oldest institutions of this kind in the world.

Theodor Kjerulf was the first director of NGU, a position he held for 30 years, and to date only 6 persons have held this post. At the start a budget of 10,000 Spd. was given for a three year period and the staff consisted of the director and one assistant, both employed on a half time basis. At the time of the centenary in 1958, 34 persons were employed, half of them geologists. The annual budget was about 1,000,000 kroner. A new impetus was given to the Survey in 1961 when it was relocated to Trondheim and merged with two other institutions concerned with geophysics and geochemistry. Today NGU consists of four branches, bedrock geology, softrock (Quaternary) geology, geophysics and geochemistry, with a common administration. 220 people are employed, about half of which have a university degree. The budget for 1983 is 56,000,000 kroner.

NGU is located at Østmarkneset in Trondheim with a small office in Oslo.

Although NGU had a rather modest beginning with only two persons in part-time positions, it nevertheless achieved a lot in the first years. This was because Kjerulf was able to employ a number of temporary assistants, mainly university students, during the summer field season. This policy of cooperation with the universities continues to this day and will hopefully continue in the future. It has been of great benefit to NGU and I hope also to the universities concerned. This anniversary volume contains several papers by university scientists and I wish to thank them for honouring us in this way.

The role of a geological survey in society is constantly changing and its profile should therefore never be static. The concept of geology is also changing and today we involve ourselves with all aspects of earth sciences; bedrock and softrock (Quaternary) geology, ore geology, hydrogeology, geophysics and geochemistry. Basically, the main function of a survey is to produce geological maps. The maps which we produce provide industry with information about mineral resources, and local authorities with details about groundwater resources and other features important in regional planning and development: and they provide the earth sciences with information relating to the composition and structure of the crust, as well as on the processes which mould the crustal surface. A completely revised edition of the 1:1,000,000 geological map of Norway is being issued this year. When this map is compared with the previous edition from 1960, which was a revised edition of the 1953 map, the remarkable improvement in our knowledge of the bedrock geology is evident. This is equally true for the areas of Precambrian basement as for the metamorphic allochthon of the Caledonian mountain belt. This improvement in our knowledge over the past 25 years, since our centenary, is in no small part due to the work and efforts of the staff of NGU.

NGU does not have the responsibility for the geological mapping of our continental shelf. The wisdom of delegating the responsibility to a different and newly formed state institution may be disputed but it has allowed us to concentrate our activities on land. Nevertheless our field activity today also includes the surface geology of the coastal and fiord areas, and some of our most recent geological maps also portray aspects of the geology of the sea floor.

In conclusion I wish to thank all employees of the Geological Survey for their diligence and skill and I look forward to seeing the results of the next 25 years.

Trondheim, 6 February 1983

Knut S. Heier
Director

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Polyphase Caledonian metamorphism in the Precambrian basement of Rogaland/ Vest-Agder, Southwest Norway

P. C. C. SAUTER, G. A. E. M. HERMANS, J. B. H. JANSEN, C. MAIJER, P. SPITS & A. WEGELIN

Sauter, P. C. C., Hermans, G. A. E. M., Jansen, J. B. H., Maijer, C., Spits, P. & Wegelin, A. 1983: Polyphase Caledonian metamorphism in the Precambrian basement of Rogaland/Vest-Agder, SW Norway. *Norges geol. Unders. 380*, 7–22.

In the Sveconorwegian basement of Rogaland/Vest-Agder, SW Norway, two incipient Caledonian metamorphic overprints are assumed on the basis of petrological textures and regional extension of the metamorphic mineral distributions. The inferred M4a is a burial metamorphism mainly in prehnite-pumpellyite facies and the M4b a metamorphism mainly in greenschist facies, restricted to the vicinity of the Caledonides. In order to propose a model sequence of retrograde metamorphic events in the basement of the whole of South Norway the available mineral records of newly formed green biotite, stilpnomelane, prehnite and pumpellyite are integrated with relevant age determinations and other geological information.

P. C. C. Sauter, G. A. E. M. Hermans, J. B. H. Jansen, C. Maijer, Instituut voor Aardwetenschappen. Rijksuniversiteit Utrecht, Budapestlaan 4, 3508 TA Utrecht, P.O. Box 80.021, The Netherlands

P. Spits, Shell Internationale Petroleum Maatschappij, E. P. 22.3, Postbus 162, Den Haag, The Netherlands

A. Wegelin, Pan Canadian Petroleum Ltd., P.O. Box 2850, Calgary, Alberta T2P 2S5, Canada

Introduction

The Precambrian basement of Rogaland/Vest-Agder, SW Norway, which is part of the Sveconorwegian of the Baltic Shield, consists mainly of anorthositic masses, the lopolith of Bjerkreim–Sokndal and a polymetamorphic high-grade envelope of charnockitic and granitic migmatites, augen-gneisses and metasupracrustals (Fig. 1; Hermans et al. 1975, Birkeland 1981). The metasupracrustals comprise the garnet-bearing migmatites (Kars et al. 1980, Huijsmans et al. 1981) and the Faurefjell formation, which consists mainly of quartzites, marbles and calc-silicate rocks (Sauter 1981, Sauter 1983).

A synopsis of the apparent ages of the rocks in the basement demonstrates Pre-Sveconorwegian ages of about 1500 Ma for the M0 stage of metamorphism, Sveconorwegian ages of about 1200 Ma for the M1 stage and about 1050 Ma for the granulite facies M2 stage of metamorphism (Versteeve 1975, Pasteels & Michot 1975, Wielens et al. 1981). Late-Sveconorwegian ages

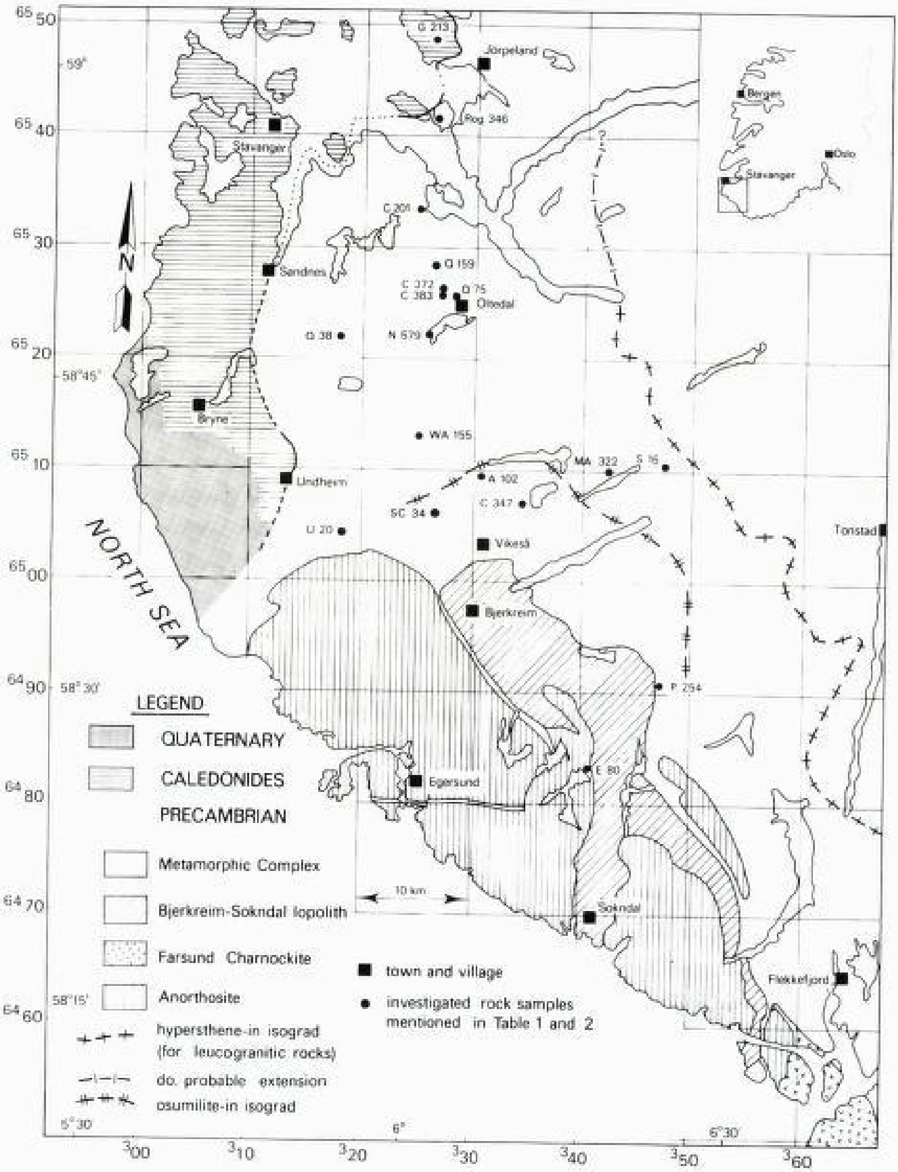

Fig. 1. Outline of the regional geology of Rogaland, SW Norway; with locations of the investigated samples.

of about 950 Ma are related to the M3 stage of metamorphism in the amphibolite facies, which affected most of the intrusive complexes and the envelope, and biotite ages in the range of 900–850 Ma represent the time of a regional cooling in a late M3 stage (Dekker 1978, Wielens et al. 1981). After uplift and erosion, the sub-Cambrian peneplain was formed between 700 and 600 Ma ago and subsequently Palaeozoic and locally even older sediments were deposited (Sigmond 1978, Andresen & Faerseth 1982). The main Caledonian basal thrust-plane must have been situated approximately along or just above the pre-existing peneplain (Sigmond 1978). During the overthrusting the sediments were squeezed out and subsequently metamorphosed together with the upper levels of the bedrock in a greenschist facies metamorphism. The index minerals for this metamorphism in the parautochthonous units and in the basement are green biotite, stilpnomelane, chlorite, actinolite and epidote (Sigmond 1978, Maijer 1980). This metamorphism in Rogaland has been reported as Caledonian in age by Verschure et al. (1980). They dated green biotite from the basement near Stavanger with Rb–Sr methods at about 400 Ma old. The last metamorphic episode in allochthonous rocks in the Sauda area is mentioned by Sigmond (1978) as occurring 410 Ma ago (Andresen et al. 1974). Nevertheless some lower intercept ages of zircons (285–300 Ma) from the basement close to the Caledonian front are given by Pasteels & Michot (1975). Recently Wielens et al. (1981) have reported lower intercept ages in a range from 310 to 410 Ma at a distance of about 50 km east of the Caledonian front. Van den Haute (1977) obtained apatite fission track ages in the basement of Rogaland in the range of 220 to 260 Ma and he interpreted this result as a reflection of an uplift period after a burial of the basement under Cambro–Silurian sediments.

In Rogaland and Vest-Agder many occurrences of retrograde minerals are found and one of the purposes of this study is to elucidate Caledonian incipient metamorphic overprints on the basement. It is also the aim to give an outline of the retrogradation all over southern Norway and to intergrate it with some geological information into a model sequence of low-grade metamorphic events.

Retrograde mineral records in Rogaland

Retrograde alteration of high-grade minerals and rocks can be found all over the map area of Fig. 2, but it varies notably in type and intensity. The retrogradations include the formation of typical metamorphic minerals such as stilpnomelane, green biotite, pumpellyite, prehnite and actinolite-tremolite together with more common alteration minerals. A selection of primary and secondary mineral assemblages of some Precambrian basement rocks of Rogaland is listed in Table 1.

In charnockitic and granitic rocks and in augen-gneisses plagioclase may be altered into sericite, albite or saussurite; orthoclase and sanidine into albite or microcline; biotite into chlorite, titanite and ilmenite; and hypersthene into serpentine, uralite amphibole, chlorite and talc. The garnetiferous migmatites show, in addition to the alterations mentioned, pinite from cordierite and osumilite; and diaspore, magnetite and hematite from hercynite. In mesocratic parts of the migmatites the mafic constituents retrograde to actinolite, chlorite, pumpellyite and prehnite. In marbles of the Faurefjell formation forsterite may be altered into serpentine and partly into clinohumite (Fig. 3A). Diopside shows rims of tremolite. Phlogopite and spinel may alter into Mg-chlorite, and some spinel into hydrotalcite. Locally tremolite, talc, clinohumite and calcite are growing at the expense of serpentine. Hedenbergite-bearing rocks and quartz-diopside gneisses from the same formation show epidote, zoisite, sericite, actinolite and hydrogrossular as retrograde minerals; pumpellyite and prehnite are present. K-feldspar-rich rocks contain albite, microcline and riebeckite as replacement minerals. In the anorthosite complexes An-rich plagioclase is locally altered into kaolinite, prehnite, laumonite and margarite.

Table 1. List of mineral contents of analysed rock samples
(x: mineral content > 10%, o: mineral content < 10% and > 5%, •: mineral content < 5%, _: totally or for the greater part decomposed.
mi: microcline, or: orthoclase, 10: An% plagioclase; 2: hydrogrossular).
The accessory minerals apatite, zircon and opaque are present in most of the samples.

sample		quartz	alkali feldspar	plagioclase	forsterite	orthopyroxene	clinopyroxene	amphibole	phlogopite	brown biotite	spinel	garnet	carbonate	tremolite	actinolite	green biotite	chlorite	serpentine	sericite, talc, sauss.	stilpnomelane	pumpellyite	prehnite	clinohumite	epidote, zoisite	titanite
		primary											retrograde												
A102	leucogranite	x	or	10						o̲							o			o					
P254	do	x	mi	29											•		o					•		•	
E00	do	x	or	24				o̲		o̲									•	o		•			
ROG346	do	x	or	30						o̲						o				o				•	•
Q38	garnet granite	x	or	x̲						x̲		x̲					o	o	o		•			•	•
M579	do	x	or	28						o		o	•				o		o		o			•	
S16	garnet granofels	x		10								x	•				•		o			o		•	
MA322	hbl. tonalite	x		x̲				x̲					o		•		o					o		•	
U20	bi. enderbite	x	mi	18		o				o̲			•				o		o		•	o		•	
SC34	norite			x̲		x̲						•[2]			o		x		x		o			•	o
MA155	Qu.hedenb.rock	x		x̲			x						o		•				o		o	o		o	•
C201	blastomylonite	x	mi										o			x	o		x	x				o	
Q213	do	x	mi	7								o				x			o	o				o	•
C347	di.-fo.rock	o			x		x						o	o											
C372	fo-sp-di marble				x		•		•		o		x				x	•					•		
C387	fo-sp marble				x				•		x		x	o			o	•							
Q75	fo marble				x		o		x				x				o						o		
Q159	di marble				x̲		x		x̲				x	o			o	•							

Stilpnomelane commonly occurs as small fibrous aggregates in cracks and cleavages of altered plagioclase (Fig. 3B), or as small crystals around dark minerals. Mostly it is associated with chlorite and albite. This stilpnomelane is found all over Rogaland and the adjacent parts of Vest-Agder. The proportion and crystal size of the stilpnomelane increase from SE towards NW. Especially in the mylonitic rocks, which are related to the Caledonian front, the stilpnomelane occurs as large radiating, euhedral postkinematic crystals (Fig. 3C). Mostly it is associated with calcite, white mica and chlorite but it is evidently younger than the synkinematic green biotite, white mica and chlorite. Locally in some mylonitic rocks of the basement, in some parautochthonous phyllites and rarely in allochthonous Precambrian rocks within the Caledonides the stilpnomelane is also observed to be oriented together with green biotite, white mica and chlorite along the plane of the schistosity.

A new generation of green biotite is formed in the Sveconorwegian basement in an approximately 15 km wide zone along the Caledonian front (Fig. 2). The proportion of this biotite increases toward that front (Verschure et al. 1980). Northwest of the green biotite–in isograd it becomes a common constituent, just as brown, titanium-rich biotite is in fresh basement rocks outside the zone. Towards the Caledonian front the brown biotite is progres-

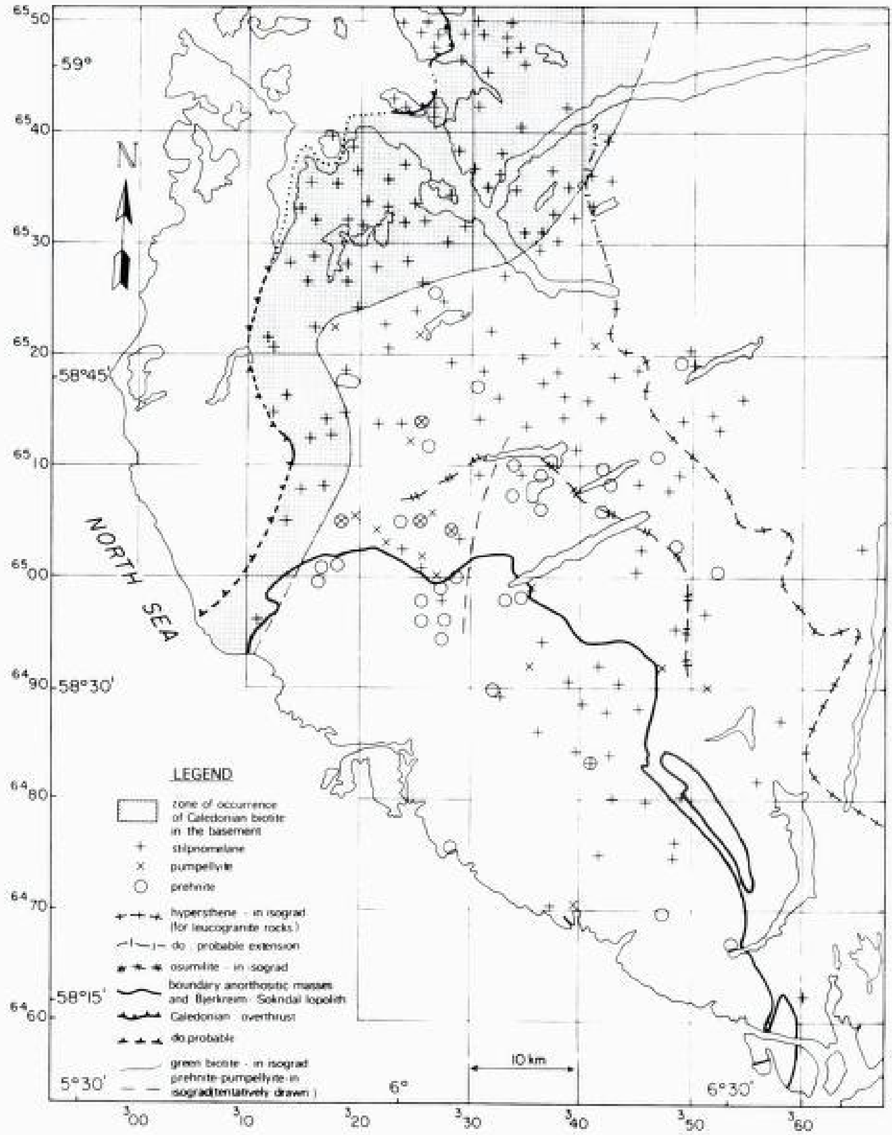

Fig. 2. Regional distribution of the minerals stilpnomelane, green biotite, pumpellyite and prehnite in Rogaland/Vest-Agder, SW Norway.

sively more decomposed and within the zone mentioned above, small granules of titanite appear with simultaneous discoloration and chloritization (Verschure et al. 1980). Close to the front most brown biotites have been replaced by pseudomorphs consisting of a mixture of green biotite, chlorite and titanite, locally with albite and white mica. Green biotite forms fine-grained aggregates of Ti-poor flakes also at the expense of amphibole and opaques. In mylonitic rocks in the basement and in phyllites of the Caledonides, for example near Jörpeland, the green biotite occurs mainly as minute flakes parallel to the foliation.

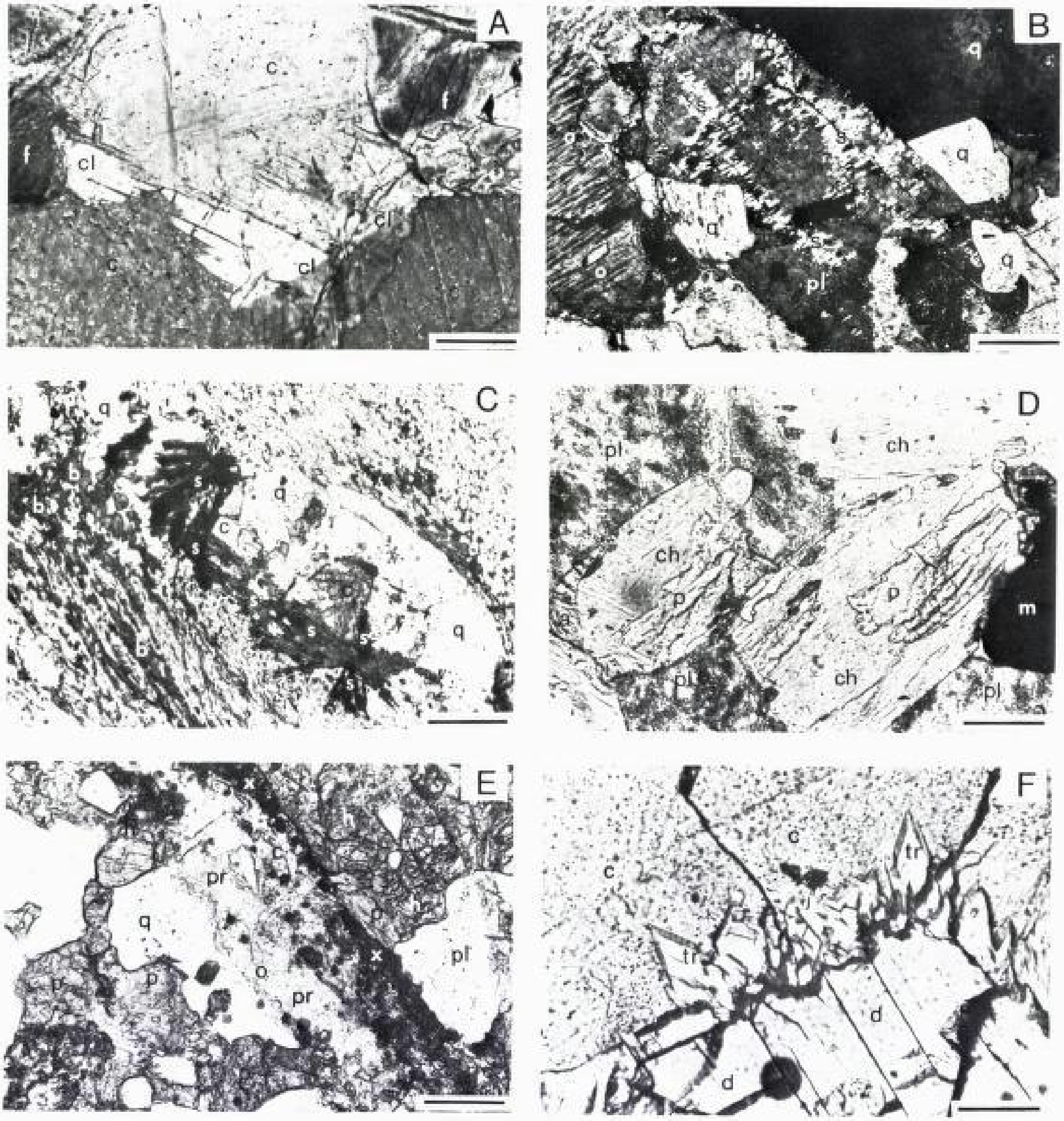

Fig. 3 (A) Clinohumite (cl) with lamellar twins in forsterite (f) marble; sample Q75, scale bar: 0.1 mm.

(B) Flakes of stilpnomelane (s) in cracks and cleavages through plagioclase (pl); sample E80, scale bar: 0.2 mm.

(C) Postkinematic stilpnomelane (s) in mylonitic rock; sample C201, scale bar: 0.2 mm.

(D) Pyroxene totally altered into chlorite (ch) and pumpellyite (p) with rims of actinolite (a); plagioclase (pl) is saussuritic; sample SC34, scale bar: 0.1 mm.

(E) Vein of prehnite (pr) with an aggregate of pumpellyite, epidote and titanite (x); hedenbergite (h) is partly also altered into pumpellyite (p); sample WA155, scale bar: 0.2 mm.

(F) Tremolite (tr) rims in diopside (d) marble; sample Q159, scale bar: 0.04 mm. (o = orthoclase, q = quartz; c = calcite; b = green biotite; m = magnetite; t = titanite).

Pumpellyite is rarely found. It occurs only in ca. 20 out of several thousands of samples. It mainly replaces plagioclase and pyroxene, and forms small anhedral aggregates in the matrix of saussuritic plagioclase. Brown biotite may also be replaced by pumpellyite, chlorite, epidote and titanite. In more mafic rocks hypersthene can be completely altered into chlorite aggregates in which pumpellyite and titanite are observed, while along the margins of the chlorite crystals new actinolite is developed (Fig. 3D).

Prehnite is observed in about 40 thin-sections. While pumpellyite is mostly an alteration product, the prehnite fills cracks and veinlets. Where these veinlets intersect altered clinopyroxene or plagioclase the prehnite may be attended by epidote, chlorite, pumpellyite and titanite (Fig. 3E). In some samples located SE of the green biotite–in isograd the coexistence of prehnite and pumpellyite is observed, and tentatively a prehnite–pumpellyite–in isograd is drawn in Fig. 2. Neither pumpellyite nor prehnite is found in the zone of green biotite, which is due to the fact that they require lower grade metamorphic conditions. Although the occurrence of stilpnomelane overlaps the area where prehnite and pumpellyite are found, the pumpellyite is not, and the prehnite only once, observed in association with stilpnomelane. This phenomenon presumably is due to differences in appropriate rock chemistry.

Actinolite and tremolite are more or less frequently found as secondary minerals, particularly in marbles and diopside gneisses. A few samples in the pumpellyite-prehnite zone contain actinolite with pumpellyite, chlorite and carbonate (Fig. 3D). The actinolite is not found in association with stilpnomelane or green biotite. Actinolite rims around brown or deep green amphiboles are mostly related to the M3-cooling stage (Dekker 1978), while newly formed actinolite needles replacing chlorite and serpentine are assumed to be very late. Tremolite is in particular observed together with talc, newly formed at the expense of serpentine, or as rims around diopside (Fig. 3F).

The increase in size and proportion of the metamorphic minerals stilpnomelane and green biotite in a NW direction towards the Caledonian front are considered to bear some relation to the Caledonian orogenic belt. The regional temperature of retrograde metamorphism in Rogaland may have ranged from about 400°C for the green biotite-in isograd (Verschure et al. 1980) to about 300°C for the prehnite-pumpellyite-in isograd (Liou 1971), while the pumpellyite-actinolite assemblages point to a cover thickness of several kilometres (Nitsch 1971, Coombs et al. 1976). The Caledonian origin of metamorphism is also supported by the parallelism of the metamorphic zonation in the Precambrian basement along the present front of the Caledonian nappe system. Due to the fact that the metamorphic minerals involved are found in the basement as well as in the Caledonian units, especially in the parautochthonous Lower Palaeozoic sediments, at least a part of the formation of these minerals post-dates the Caledonian thrusting. However, it is questionable if all the Caledonian retrogradations in the basement of SW Norway are developed subsequent to the thrusting.

Table 2. Electron microprobe analyses [1)]

	Pumpellyite			Prehnite			Stilpnomelane			Clino-zoisite
	N 579	SC 34	U 20	P 254	S 16	U 20	E 80	C 201	G 213	N 579
SiO_2	38.3	37.5	38.1	44.7	43.35	45.6	45.25	47.6	45.6	39.6
Al_2O_3	24.5	21.7	22.7	20.3	24.9	24.55	7.9	6.7	7.25	30.2
TiO_2	b.d.	0.1	0.4	-	-	-	0.1	-	b.d.	b.d.
Fe_2O_3	4.9	6.7	5.3	3.8	2.3	0.1	32.3	33.5	35.2	1.8
MnO	b.d.	0.1	0.1	-	0.05	b.d.	0.2	1.0	1.7	-
MgO	1.5	3.7	3.2	-	2.5	b.d.	6.1	3.65	1.6	-
CaO	23.0	22.4	22.8	25.85	21.8	26.1	0.5	b.d.	0.1	24.3
Na_2O	0.25	0.1	0.2	0.2	0.5	b.d.	0.1	0.35	0.6	0.2
K_2O	b.d.	b.d.	b.d.	0.9	b.d.	b.d.	1.5	0.9	1.5	b.d.
Total	92.45	92.3	9[illegible].8	95.75	95.4	96.35	93.95	93.7	93.55	96.1

	Tremolite		Green-biotite	Chlorite			Serpentine		Clinohumite	
	C 347	Q 159	Rog 346	P 254	C 372	C 383	C 372	C 372	Q 75	C 372
SiO_2	57.94	56.47	35.4	30.3	28.72	29.07	35.89	39.30	38.32	37.48
Al_2O_3	0.14	0.28	15.4	18.6	20.86	20.17	7.70	1.87	b.d.	b.d.
TiO_2	b.d.	b.d.	1.2	b.d.	0.06	0.12	b.d.	0.12	b.d.	3.24
FeO	2.91	1.17	28.3	22.3	1.74	2.23	1.71	1.79	4.73	2.24
MnO	0.26	0.16	0.21	0.1	-	-	0.08	0.17	0.42	1.41
MgO	22.46	22.82	5.8	15.0	32.64	33.41	38.33	39.88	53.77	51.86
CaO	13.09	13.84	b.d.	0.9	-	-	-	-	-	-
Na_2O	0.11	0.10	b.d.	0.2	b.d.	b.d.	-	-	-	-
K_2O	0.13	0.08	9.3	0.8	b.d.	b.d.	-	-	-	-
Total	97.04	94.95	95.6	88.2	84.02	85.00	83.71	83.13	97.24	96.23

1) The analyses were performed at the Instituut voor Aardwetenschappen of the Rijksuniversiteit Utrecht, and that of the Vrije Universiteit Amsterdam. Both institutes were provided with financial and personnel support by ZWO-WACOM (Research group for analytical chemistry of minerals and rocks, subsidized by the Netherlands Organisation for the Advancement of Pure Research).

b.d.: below detection limit.

CHEMISTRY OF SOME RETROGRADE MINERALS

To illustrate the composition of the retrograde minerals, some microprobe analyses are listed in Table 2. Pumpellyite contains significant amounts of ferric iron. The pleochroic colours of pumpellyite, ranging from almost colourless to deep-green, seem to be related to this Fe^{3+} content. Prehnite too contains Fe^{3+}, up to about 11 mol% of the Fe^{3+} end-member. Relatively high Mg contents and low Ca contents in sample S16 suggest a substitution of Ca by Mg. The stilpnomelanes mainly show a compositional variation in Fe, Mg and Mn, suggesting substitution of Fe+Mn by Mg. The brown colour of the stilpnomelanes indicates an appreciable amount of Fe^{3+} present. No compositional differences have been found between stilpnomelane associated with Caledonian mylonitization and those in the Precambrian basement. The green biotite is rather Fe-rich and Ti-poor. Chlorite, as a secondary product in plagioclase (P254), is iron-rich. Chlorites and other secondary minerals in the marbles have a composition close to their Mg end-members.

Retrograde mineral distribution in South Norway

The regional extension of stilpnomelane, green biotite, pumpellyite and prehnite in the basement of South Norway, observed in our samples and encountered in the literature, is given in Fig. 4. The widespread occurrences

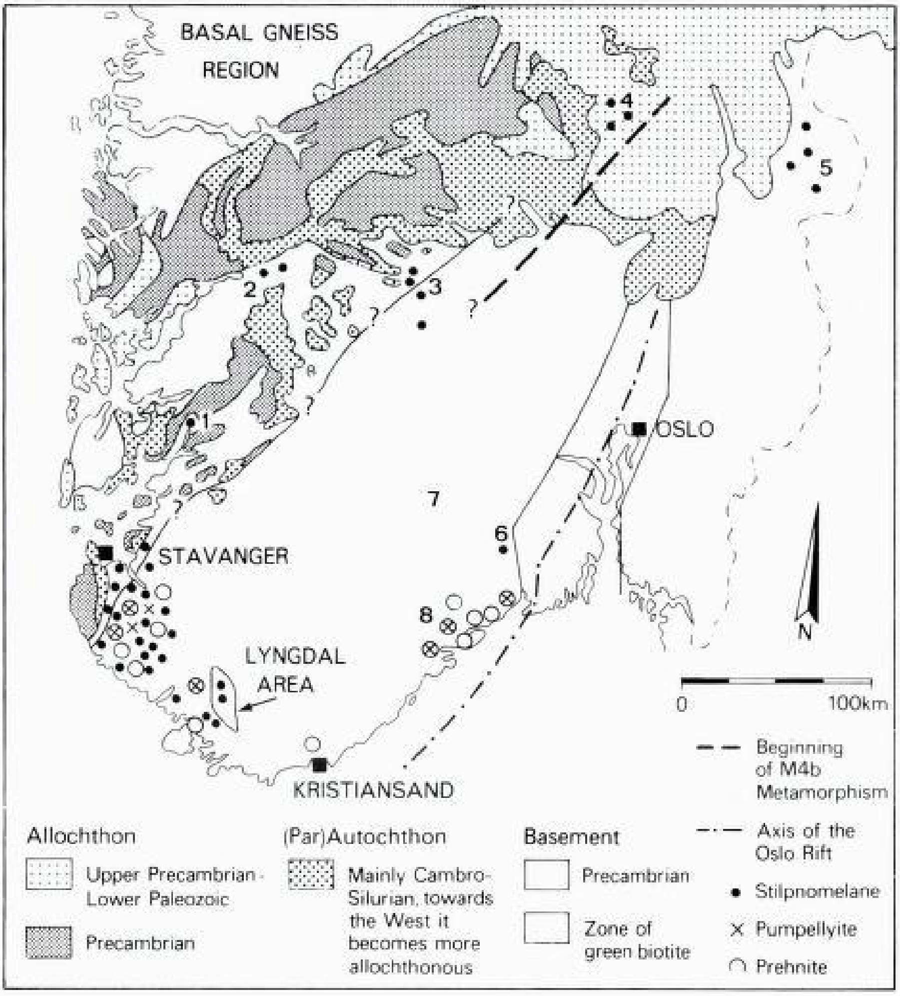

Fig. 4. Schematized distribution of low-grade metamorphic minerals in South Norway; geology after Reymer (1979). Axis of the Oslo Rift after Huseby & Ramberg (1978). Beginning of M4b metamorphism interpreted after Bryhni & Brastad (1980). 1: Sauda area; 2: Eidfjord area; 3: Geilo area; 4: Vinstra area; 5: E Hedmark; 6: Fen area; 7: E Telemark; 8: Bamble area.

in the Precambrian of South Norway are records of metamorphic overprinting, but they cannot all be attributed to the influence of the Caledonian nappe system.

The stilpnomelane distribution in the basement is largely restricted to the vicinity of the Caledonian orogenic belt. Most occurrences in the basement near Stavanger (this paper), in the Sauda area (Sigmond 1978), in the Eidfjord granite (Priem et al. 1976), in the neighbourhood of Geilo (Priem et al. 1972) and, of course, those in the lower units of the Caledonides NNW of Oslo, near Vinstra (Englund 1973) and in Rogaland (this paper) may be attributed to the Caledonian metamorphism, post-dating the main thrusting. This metamorphism, so far called M4 in Rogaland, will be referred to as M4b. The formation of stilpnomelane in the sparagmites near Vinstra seems to indicate the beginning of this Caledonian metamorphism towards the NW, while the SE part of these allochthonous units is practically unmetamorphosed (Englund 1973, Bryhni & Brastad 1980). Priem et al. (1970) reported stilpnomelane in sub-Jotnian rocks of the basement of eastern Hedmark, NNE of Oslo, beneath unmetamorphosed allochthonous sparagmites. These occurrences are restricted to mylonitic rocks and they are presumably strain-induced and related to the Caledonian thrusting events. In the Dala region, being the eastward continuation of this basement in Sweden, a burial metamorphism has been recorded with an increasing grade from prehnite-pumpellyite facies via pumpellyite-actinolite facies to greenschist facies towards the east (Nyström & Levi 1980, Nyström 1982). These authors assume that this burial metamorphism took place before 1220 Ma, but biotites of this basement in east Hedmark, Norway, yield Sveconorwegian K-Ar and Rb-Sr ages of around 925 Ma with a youngest age of 842 Ma (Priem et al. 1970). Whole-rock K-Ar ages as low as 605 Ma in the sub-Jotnian porphyries in the Dala region are interpreted as a function of argon leakage due to mild Caledonian events (Priem et al. 1968). Verschure (1981) therefore suggested that this burial, retrograde metamorphism is likewise related to the Caledonian orogeny. The distribution of the metamorphic grade, however, shows no clear relation with the present position of the Caledonian front. This fact, and the considerable distance between the Caledonian nappe system and the greenschist facies zone, suggests that the basement underwent the burial metamorphism before the Caledonian overthrusting. The stilpnomelane in the Fen area belongs to a late stage of the fenitization, while the carbonatite itself is dated as about 600 Ma old (Verschure et al. 1983). The occurrences of stilpnomelane in the Lyngdal area and most of the occurrences in Rogaland are not related to mylonitization. Although the original southeastward extent of the Caledonian nappe system some 400 Ma ago is unknown, it is unlikely that these stilpnomelane occurrences are induced by Caledonian thrusting or by Caledonian M4b metamorphism, in view of the long distance to the present front position.

Green biotites, which are newly formed in the basement rocks, can be distinguished in three different areas: –

1) In the Basal Gneis Region (fig. 4) it occurs as a secondary mineral in the greenschist facies overprint with epidote, actinolite, white mica, chlorite, albite, microcline and often titanite or ilmenite.
   Priem et al. (1973), for example, described this greenschist facies assemblage as an overprint on the Sveconorwegian Hestbrepiggan granite, which contains relict brown biotite with reset K-Ar and Rb-Sr ages of about 390 Ma. The effects of greenschist facies metamorphism were also reported by Bryhni et al. (1971) in the mineral content of their samples used for dating work. No pumpellyite, prehnite or stilpnomelane is known from this region.

2) In the area between and just southeast of the Caledonides the green biotite is not associated with actinolite, but it frequently occurs with stil-

pnomelane. The same holds for the green biotite occurrences in the lower nappe units of the Caledonides in Rogaland. These assemblages are indicative for a lower grade part of the greenschist facies than those of the Basal Gneiss Region.

3) In the Lyngdal area the secondary green biotite is associated with actinolite, stilpnomelane, chlorite, epidote, white mica, albite and microcline. In this area the metamorphic grade seems to be a bit higher than in Rogaland. This is supported by the occurrences of retrograde corundum instead of diaspore as an alteration pruduct of hercynite (Pieters, pers. comm.).
   As already stated for the stilpnomelane findings in this area, these green biotites are not developed during the M4b metamorphism. They are evidently later than the relict, olive-green biotites which are supposed to be Sveconorwegian.

The prehnite and pumpellyite occurrences in South Norway are scarce. In Rogaland the pumpellyite and prehnite-pumpellyite assemblages are observed SE of the green biotite isograd. Only a few prehnite-pumpellyite assemblages are observed in the Lyngdal area and they are probably also formed outside the green biotite zone (Fig. 4). The latter assemblages and the prehnite occurrences in the Lyngdal area and near Kristiansand, Vest-Agder (Falkum 1966), are in fact situated about a hundred kilometres away from the present Caledonian front. Therefore, these occurrences and presumably also the southeasterly occurrences in Rogaland do not belong to the M4b metamorphism. It is practically impossible, for the same reason, that the occurrences of prehnite and pumpellyite reported from SE Norway (Mitchell 1967, Field & Rodwell 1968, O'Nions et al. 1969) have originated during the M4b metamorphism. Due to the continuation of the axis of the Oslo rift valley along the Norwegian coast in a southwestward direction (Husebye & Ramberg 1978), the basement in this area may have been affected by a Permian event which may be responsible for the low-grade metamorphic effects (Fig. 4). The Permian thermal influence can also be recognized in K-Ar ages of 260 Ma on whole-rock samples of pre-Sveconorwegian rhyolites in eastern Telemark (Priem et. al., 1967) and in lower intercept ages of 260 Ma on zircons and sphenes from the Bamble region (O'Nions & Baadsgaard 1971).

Discussion

A large part of the retrograde metamorphism in the Precambrian basement of SW Norway is clearly attributed to the Caledonian orogeny. This is supported by Caledonian ages of green biotite (Verschure et al. 1980) and of lower intercepts of zircons (Pasteels & Michot 1975, Corfu 1978, 1979, Schärer 1980, Wielens et al. 1981).

Petrological textures in the basement of Rogaland are suggestive of a polyphase retrograde metamorphism. Indications are, for instance, the growth of green biotite at the expence of chlorite, which is pseudomorphous after brown biotite, the occurrence of synkinematic and postkinematic generations of stilpnomelane, green biotite, chlorite and white mica and the development of tremolite, clinohumite and talc at the expense of serpentine, which is an alteration product of forsterite or diopside. In addition, the geographical distribution of the retrogradations in the basement of Southern Norway in relation to the present position of the Caledonian front also suggests that not all retrogradations can be fitted into only one phase of Caledonian incipient metamorphism.

The occurrences of stilpnomelane, green biotite and prehnite + pumpellyite in the Precambrian basement close to the Caledonian front are assumed to be the expression of a Caledonian M4b metamorphism. In Rogaland the distribution of these minerals seems to suggest an inversed thermal gradient. Metamorphism showing inversed gradients is well known in the Scandinavian Caledonides (e.g. Andréasson & Gorbatschev 1980, Andréasson & Lagerblad 1980). In allochthonous units this inversion can be caused by a telescoping effect of superposed nappes, but in the autochthonous basement mainly by a downwards directed heat flow from relatively hot nappes. As a result of intense Caledonian mylonitisation the retrograde overprint may locally be strain-induced (Andréasson & Gorbatschev 1980). Generally, such zones are relatively thin. In Rogaland pure strain-induced retrogradation may be excluded. Although stilpnomelane growth may be favoured in mylonitic zones, it also occurs in undeformed rocks throughout the mapped area (Fig. 2). The green biotite is not particularly related to deformation.

Low-grade to very low-grade metamorphism at greater distances from the Caledonian front can clearly not be related to the Caledonian overthrust. Therefore it is assumed that these metamorphic effects are related to a burial type of metamorphism (M4a), possibly prior to Caledonian thrusting, as a result of a burden of sediments in late Precambrian or early Palaeozoic times. Thick late Precambrian sediments are known from the sparagmite region in central south Norway (Bjørlykke 1978), and Lower Palaeozoic sediments in the parautochthonous (Vidda group) and allochthonous (Holmasjö allochthon) units at the base of the Caledonian nappe pile in SW Norway (Andresen & Faerseth 1982). Evidence from central Norway and Sweden shows that sediment thicknesses vary laterally but they may amount to several kilometres (Bjørlykke 1978, Gee & Zachrisson 1979).

The absolute timing of the Caledonian retrograde metamorphic events in Rogaland is uncertain. Generally in the Caledonian orogen a Finmarkian and a Scandian phase are recognized, dated to late Cambrian to early Ordovician time and Middle to late Silurian time, respectively (Roberts & Sturt 1980, Bryhni & Brastad 1980). In the Caledonides of SW Norway both phases are recognized (Roberts & Gale 1978). The only isotopically dated retrograde mineral from the basement in Rogaland close to the front is green biotite, yielding K-Ar ages of 557 Ma and 469 – 444 Ma and Rb-Sr ages from

421 to 397 Ma (Verschure et al. 1980). The ages around 400 Ma evidently refer to the Scandian phase which is the most commonly reported deformation event in the Caledonides of SW Norway (Sturt & Thon 1978, Roberts & Sturt 1980, Roddick & Jorde 1981). Tentatively, the high K-Ar ages can be attributed to the Finnmarkian phase, although Verschure et al. (1980) suggest incorporation of excess argon as the cause of the high ages. Most of the relict biotites in the basement SE of the Caledonian front in southern Norway remain Sveconorwegian (900–850 Ma) in age (O'Nions et al. 1969, Priem et al. 1970, Brueckner 1972, Priem et al. 1972, Versteeve 1975, Verschure et al. 1980, Wielens et al. 1981); only a few brown biotites just beneath the Caledonian thrust front near Stavanger form an exception in showing reset K-Ar and Rb-Sr ages down to 700 Ma (Verschure et al. 1980). In the Basal Gneiss Region, where most biotites show reset Caledonian ages in contrast to biotites from the basement SE of the Caledonian front, the K-Ar mineral datings are also higher than the Rb-Sr datings. Brueckner (1972) ascribed this feature to a more ready reequilibration of Sr than to the idea that Ar is expelled from the mineral. This could mean that some of the K-Ar mineral datings are a real reflection of the Finnmarkian event. Bryhni et al. (1971) reported three $^{39}Ar/^{40}Ar$ ages of 524, 488 and 405 Ma in the Basal Gneiss Region. Further indications for an early Caledonian event in SW Norway possibly may be found in a date of about 540 Ma on the parautochthonous phyllites in the Jæren district, south of Stavanger (Roddick & Jorde 1981) and in a lower intercept age of zircons of about 540 Ma in autochthonous basement near Undheim (Pasteels & Michot 1975). Whole-rock K-Ar ages, as low as 564 Ma in Sveconorwegian dike systems in Rogaland and Vest-Agder, sampled about 50 km east from the Caledonian nappe front, seem to reflect argon leakage due to post-Sveconorwegian thermal events (Versteeve 1975). Thus, in the basement of SW Norway there are isotopic indications for two Caledonian events.

Based on the above reasoning we assume a Finnmarkian metamorphism (M4a), mainly in the prehnite-pumpellyite facies, affecting the basement in SW Norway. This weak, up to now radiometrically not definitely dated event, was a regional burial metamorphism. Because of the interference of the M4a and M4b metamorphic phases in the basement the mineral records of the M4a in Rogaland were formerly ascribed to the M4b phase, which was induced about 400 Ma ago by the Caledonian nappe system (Maijer 1980, Verschure et al. 1980, Verschure 1981). The retrograde metamorphic events in the basement of southern Norway may be summarized in the following model sequence:

M3 – The regional Sveconorwegian cooling stage is reported in K-Ar ages on amphiboles and biotites over the whole of South Norway and it lasted to about 850 Ma. Low-grade retrogradation may have been operative. After denudation, Upper Precambrian and Palaeozoic sediments were deposited on the peneplain.

M4a – A burial Finnmarkian metamorphism took place, mainly in prehnite–

pumpellyite facies. Locally, where the sediment pile was thicker, it may have produced a higher metamorphic grade. It may have been active in several periods. Probably this metamorphism extended over a great part of the Baltic Shield.

M4b -This Scandian metamorphic phase in the basement is restricted to the vicinity of the Caledonides. It mostly developed with an inversed gradient and it was locally strain-induced in specific mylonitic zones. It is about 400 Ma old and waning towards the southeast from greenschist facies near the Caledonian front into prehnite-pumpellyite facies. Late M4b thermal reequilibration lasted to about 300 Ma ago.

M5 – The Permian reheating of about 250 Ma ago was active in the basement only in a zone marginal to the Oslo Rift valley and along Permian dike systems. Its conditions reached prehnite-pumpellyite facies. Erosion of the basement started again and at least 220 Ma ago the Sveconorwegian basement, as it is exposed at the present day, passed through the 100°C isograd.

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REFERENCES

Andréasson, P. G. & Gorbatschev, R. 1980: Metamorphism in extensive nappe systems: a study of the Central Scandinavian Caledonides. *Geol. Fören. Stockholm Förh. 102,* 335–357.

Andréasson, P. G. & Lagerblad, B. 1980: Occurrence and significance of inverted metamorphic gradients in the western Scandinavian Caledonides. *J. Geol. Soc. London 137,* 1–12.

Andresen, A. & Faerseth, R. 1982: An evolutionary model for the southwest norwegian Caledonides. *Am. J. Sci. 282,* 756–782.

Andresen, A., Heier, K. S., Jorde, K. & Naterstad, J. 1974: A preliminary Rb/Sr geochronological study of the Hardangervidda-Ryfylke Nappe system in the Røldal area, SW Norway. *Norsk geol. Tidsskr. 54,* 35–47.

Birkeland, T. 1981: The geology of Jæren and adjacent districts. A contribution to the Caledonian nappe tectonics of Rogaland, south west Norway. *Norsk geol. Tidsskr. 61,* 213–235.

Bjørlykke, K. 1978: The eastern marginal zone of the Caledonide orogen in Norway. *Geol. Surv. Can. 78-13,* 49–55.

Brueckner, H. K. 1972: Interpretation of Rb-Sr ages from the precambrian and paleozoic rocks of southern Norway. *Am. J. Sci. 271,* 334–358.

Bryhni, I. & Brastad, K. 1980: Caledonian regional metamorphism in Norway. *J. Geol. Soc. Lond. 137,* 251–259.

Bryhni, I., Fitch, F. J. & Miller, J. A. 1971: $^{40}Ar/^{39}Ar$ Dates from recycled Precambrian rocks in the gneiss region of Norwegian Caledonides. *Norsk geol. Tidsskr. 51,* 391–406.

Coombs, D. S., Nakamura, Y. & Vuagnat, M. 1976: Pumpellyite-actinolite facies schists of the Taveyanne Formation near Loèche, Valais, Switzerland. *J. Petrol 17,* 440–471.

Corfu, F. 1978: Comparison of U-Pb systems in zircons, monazites and sphenes. A contribution to the geochronology of the Precambrian basement of central south Norway. *U.S. Geol. Surv. Open file rep. 78-701,* 81–83.

Corfu, F. 1979: Response of U–Pb and Rb–Sr systems to Precambrian and Palaeozoic events in central south Norway. Abstracts of ECOG VI.

Dekker, A. G. C. 1978: Amphiboles and their host rocks in the high-grade metamorphic Precambrian of Rogaland/Vest-Agder, SW Norway. (Thesis Utrecht) – *Geol. Ultraiect. 17,* 277 pp.

Englund, J. O. 1973: Geochemistry and mineralogy of pelitic rocks from the Hedmark Group and the Cambro-Ordovician sequence, southern Norway. *Norges geol. Unders. 286,* 1–60.

Falkum, T. 1966: Geological investigations in the Precambrian of southern Norway. I. The complex of metasediments and migmatites at Tveit, Kristiansand. *Norsk geol. Tidsskr. 46,* 85–110.

Field, D. & Rodwell, J. R. 1968: The occurrence of prehnite in a high grade metamorphic sequence from South Norway. *Norsk geol. Tidsskr. 48,* 55–59.

Gee, D. G. & Zachrisson, E. 1979: The Caledonides in Sweden. *Sveriges geol. Unders. C, 769,* 48 pp.

Haute, P. van den. 1977: Apatite fission track dating of Precambrian intrusive rocks from the southern Rogaland (south-western Norway). *Bull. Soc. belge Géologie, 86,* 97–110.

Hermans, G. A. E. M., Tobi, A. C., Poorter, R. P. E., & Maijer, C. 1975: The high-grade metamorphic Precambrian of the Sirdal–Ørsdal area, Rogaland/Vest-Agder, South-west Norway. *Norges geol. Unders. 318,* 51–74.

Husebye, E. S. & Ramberg, I. B. 1978: Geophysical investigations. In: Dons, J. A. & Larsen, B. T. (eds.). The Oslo paleorift. A review and guide to excursions. *Norges geol. Unders. 337,* 41–53.

Huijsmans, J. P. P., Kabel, A. B. E. T., & Steenstra, S. E. 1981: On the structure of a high-grade metamorphic Precambrian terrain in Rogaland, S. Norway. *Norsk geol. Tidsskr. 61,* 183–192.

Kars, H., Jansen, J. B. H., Tobi, A. C. & Poorter, R. P. E. 1980: The metapelitic rocks of the polymetamorphic Precambrian of Rogaland SW Norway. Part II: mineral relations between cordierites hercynite and magnetite within the osumilite-in isograd. *Contrib. Mineral. Petrol. 74,*235–244.

Liou, J. G. 1971: Synthesis and stability relations of prehnite, $Ca_2Al_2Si_3O_{10}(OH)_2$. *Am. Min. 56,* 507–531.

Maijer, C. 1980: Caledonian retromorphism of the Precambrian of Rogaland/Vest-Agder, SW Norway (abstract). Colloquium high-grade metamorphic Precambrian and its intrusive masses, Utrecht, Holland.

Mitchell, R. H. 1967: The precambrian rocks of the Telemark area in South central Norway. V. The Nissedal supracrustal series. *Norsk geol. Tidsskr. 47,* 295–332.

Nitsch, K-H, 1971: Stabilitätsbeziehungen von Prehnit- und Pumpellyit-haltigen Paragenesen. *Contrib. Mineral. Petrol. 30,* 240–260.

Nyström, J. O. 1982: Post-Svecokarelian Andinotype evolution in cntral Sweden. *Geol. Rundsch. 71,* 141–157.

Nyström, J. O. & Levi, B. 1980: Pumpellyite-bearing Precambrian rocks and post-Svecokarelian regional metamorphism in central Sweden. *Geol. Fören. Stockholm Förh. 102,* 37–39.

O'Nions, R. K. & Baadsgaard, H. 1971: A radiometric study of polymetamorphism in the Bamble region Norway. *Contrib. Mineral. Petrol. 34,* 1–21.

O'Nions, R. K., Morton, R. D. & Baadsgaard, H. 1969: Potassium-argon ages from the Bamble sector of the Fennoscandian shield in south Norway. *Norsk geol. Tidsskr. 49,* 171–190.

Pasteels, P. & Michot, J. 1975: Geochronologic investigation of the metamorphic Terrain of Southwestern Norway. *Norsk geol. Tidsskr. 55,* 111–134.

Priem, H. N. A., Boelrijk, N. A. I. M., Hebeda, E. H., Verdurmen, E. A. Th. & Verschure, R. H. 1972: Rb–Sr investigations in the Uvdal granitic-gneissic complex: evidence of an "Elsonian" event in the Precambrian of southern Norway? In: Progress report of the ZWO Laboratorium voor Isotopen-Geologie, p. 111–116.

Priem, H. N. A., Boelrijk, N. A. I. M., Hebeda, E. H., Verdurmen, E. A. Th. & Verschure, R. H. 1973: A note on the geochronology of the Hestbrepiggan granite in West Jotunheimen. *Norges geol. Unders. 289,* 31–35.

Priem, H. N. A., Boelrijk, N. A. I. M., Hebeda, È. H., Verdurmen, E. A. Th. & Vershure, R. H. 1976: Isotope geochronology of the Eidfjord granite, Hardangervidda, West Norway. *Norges geol. Unders. 327,* 35–39.

Priem, H. N. A., Mulder, F. G., Boelrijk, N. A. I. M., Hebeda, E. H., Verschure, R. H. & Verdurmen, E. A. Th. 1968: Geochronological and palaeomagnetic reconnaissance survey in parts of central and southern Sweden. *Phys. Earth Planet. Int. 1,* 373–380.

Priem, H. N. A., Boelrijk, N. A. I. M., Verschure, R. H., Hebeda, E. H., Verdurmen, E. A. Th., Svinndal, S. & Barkey, H. 1967: Age studies in Telemark, southern Norway. In: Progress-report on the isotopic dating project in Norway, p. 15–17.

Priem, H. N. A., Verschure, R. H., Verdurmen, E. A. Th., Hebeda, E. H. & Boelrijk, N. A. I. M. 1970: Isotopic evidence on the age of the Trysil porphyries and granites in Eastern Hedmark, Norway. *Norges geol. Unders. 266*, 261–276.

Reymer, A. P. S. 1979: Investigations into the metamorphic nappes of the Central Scandinavian Caledonides on the basis of Rb–Sr and K–Ar age determinations. Ph.D. thesis, Rijksuniversiteit Leiden. 123 pp.

Roberts, D. & Gale, G. H. 1978: The Caledonian-Appalachian Iapetus Ocean, in Tarling, D. H. (ed.): Evolution of the Earth's Crust. Academic Press, London, p. 255–342.

Roberts, D. & Sturt, B. A. 1980: Caledonian deformation in Norway. *J. Geol. Soc. London 137*, 241–250.

Roddick, J. C. & Jorde, K. 1981: Rb-Sr whole rock isochron dates on rocks of the Jæren district, SW Norway. *Norges geol. Unders. 365*, 55–67.

Sauter, P. C. C. 1981: Mineral relations in siliceous dolomites and related rocks in the high-grade metamorphic Precambrian of Rogaland, SW Norway. *Norsk geol. Tidsskr. 61*, 35–45.

Sauter, P. C. C. 1983. Metamorphism of siliceous dolomites in the high-grade Precambrian of Rogaland, SW Norway. (Ph.D. Thesis, Rijksuniversiteit Utrecht), *Geol. Ultraiect. 32*, 143 pp.

Schärer, U. 1980: U-Pb and Rb-Sr dating of a polymetamorphic nappe terrain: the Caledonian Jotun nappe, Southern Norway. *Earth Plan. Sc. Letters 49*, 205–218.

Sigmond, E. M. O. 1978: Beskrivelse til det berggrunnsgeologiske kartbladet Sauda 1:250.000. *Norges geol. Unders. 341*, 1–94.

Sturt, B. A. & Thon, A. 1978: Caledonides of southern Norway. *Geol. Surv. Canada Paper 78–13*, 39–47.

Verschure, R. H. 1981: The extent of Sveconorwegian and Caledonian metamorphic imprints in Norway and Sweden. *Terra Cognita, 2*, 64–65.

Verschure, R. H., Andriessen, P. A. M., Boelrijk, N. A. I. M., Hebeda, E. H., Maijer, C., Priem, H. N. A. & Verdurmen, E. A. Th. 1980: On the thermal stability of Rb-Sr and K-Ar Biotite Systems: Evidence from coexisting Sveconorwegian (ca. 870 Ma) and Caledonian (ca. 400 Ma) Biotites in SW Norway. *Contrib. Mineral. Petrol. 74*, 245–252.

Verschure, R. H., Maijer, C., Andriessen, P. A. M., Boelrijk, N. A. I. M., Hebeda, E. H., Priem, H. N. A. & Verdurmen, E. A. Th. 1983: Dating explosive volcanism perforating the Precambrian basement in southern Norway. *Norges geol. Unders. 380*, 35–49.

Versteeve, A. J. 1975: Isotope geochronology in the high-grade metamorphic Precambrian of SW Norway. *Norges geol. Unders. 318*, 1–50.

Wielens, J. B. W., Andriessen, P. A. M., Boelrijk, N. A. I. M., Hebeda, E. H., Priem, H. N. A., Verdurmen, E. A. Th. & Verschure, R. H. 1981: Isotope geochronology in the high-grade metamorphic Precambrian of Southwestern Norway: New Data and Reinterpretations. *Norges geol. Unders. 359*, 1–30.

Subdivision of the Jotun Nappe Complex between Aurlandsfjorden and Nærøyfjorden, South Norway

INGE BRYHNI, KJARTAN BRASTAD & VIGDIS W. JACOBSEN

Bryhni, I., Brastad, K. & Jacobsen, V. W. 1983: Subdivision of the Jotun Nappe Complex between Aurlandsfjorden and Nærøyfjorden, South Norway. *Norges geol. Unders. 380*, 23–33.

In the area between Aurlandsfjorden and Nærøyfjorden the Jotun Nappe Complex can be subdivided into a lower thrust unit of mainly gneisses with mangerites, jotunites and minor supracrustals (the Flåm unit), and an upper unit of mainly gabbroic to anorthositic rocks (the Stiganosi unit). The basal thrust of the Nappe Complex and the contact between the two internal units are blastomylonitic and characterized by retrogression of the original granulite-facies assemblages. Exrension of investigations into a larger part of the 'Faltungsgraben' indicates that the tectonic subdivision of the Jotun Nappe Complex into a lower and an upper part has regional significance.

Inge Bryhni, Kjartan Brastad, Vigdis W. Jacobsen[1]
Mineralogisk-Geologisk Museum, Sars gt. 1, Oslo 5, Norway.

[1] *Present address: Oljedirektoratet, N-4000 Stavanger*

Introduction

The area between Aurlandsfjorden and Nærøyfjorden belongs geologically to the Upper Jotun Nappe, which is shown to have a composite structure and therefore from here on will be called the *'Jotun Nappe Complex'*. It is covered in early rough maps and descriptions by Rekstad (1905), Reusch (1908) and Goldschmidt (1916), and has not been restudied since until quite recently when several informal reports have been written (Bryhni 1976, 1977, Bryhni et al., 1977, Qvale 1982).

In general, this part of the Caledonides has thick crystalline Precambrian thrust nappes resting on Cambro-Ordovician and possibly older phyllites or schists, which in turn overlie the Precambrian basement. The general relations of this 'nappe region' have been studied in adjacent areas by Hødal (1945), Landmark (1949), Skjerlie (1958), Kvale (1948, 1960), Lacour (1969, 1971), Fareth (1977), Heim et al. (1977), Roberts (1977), Bryhni et al., (1977), Henry (1977), Henry and Lacour (1978), Sigmond (1978), Corfu (1980) and others. Major problems are related to the *tectonostratigraphy, age* of the internal movement zones and the *derivation* of the thrust nappe complex: do the internal subdivisions, made in scattered areas (Skjerlie 1958, Lacour 1969, Battey & McRitchie 1973, Bryhni 1976 etc.), have any regional significance in the sense that the Jotun rocks can be subdivided into several independent thrust nappes? And if so, was the internal thrust sequence established during or prior to the Caledonian orogeny? Is the Jotun Nappe Complex rooted below the 'Faltungsgraben', as favoured in recent years by Smithson & Ramberg (1970), Battey & McRitchie (1973), Smithson et al. (1974) and Banham et al. (1979), or has it been derived from far off the coast as advocated by Holtedahl (1936), Kvale (1948, 1960), Naterstad et al. (1973) and recently by Hossack (1976), Heim et al. (1977) and Henry (1977)?

The area between Aurlandsfjorden and Nærøyfjorden is important for the solution of some of these problems, since the Jotun Nappe Complex here appears to be composed of two separate mappable, tectonic units (Fig. 1). The

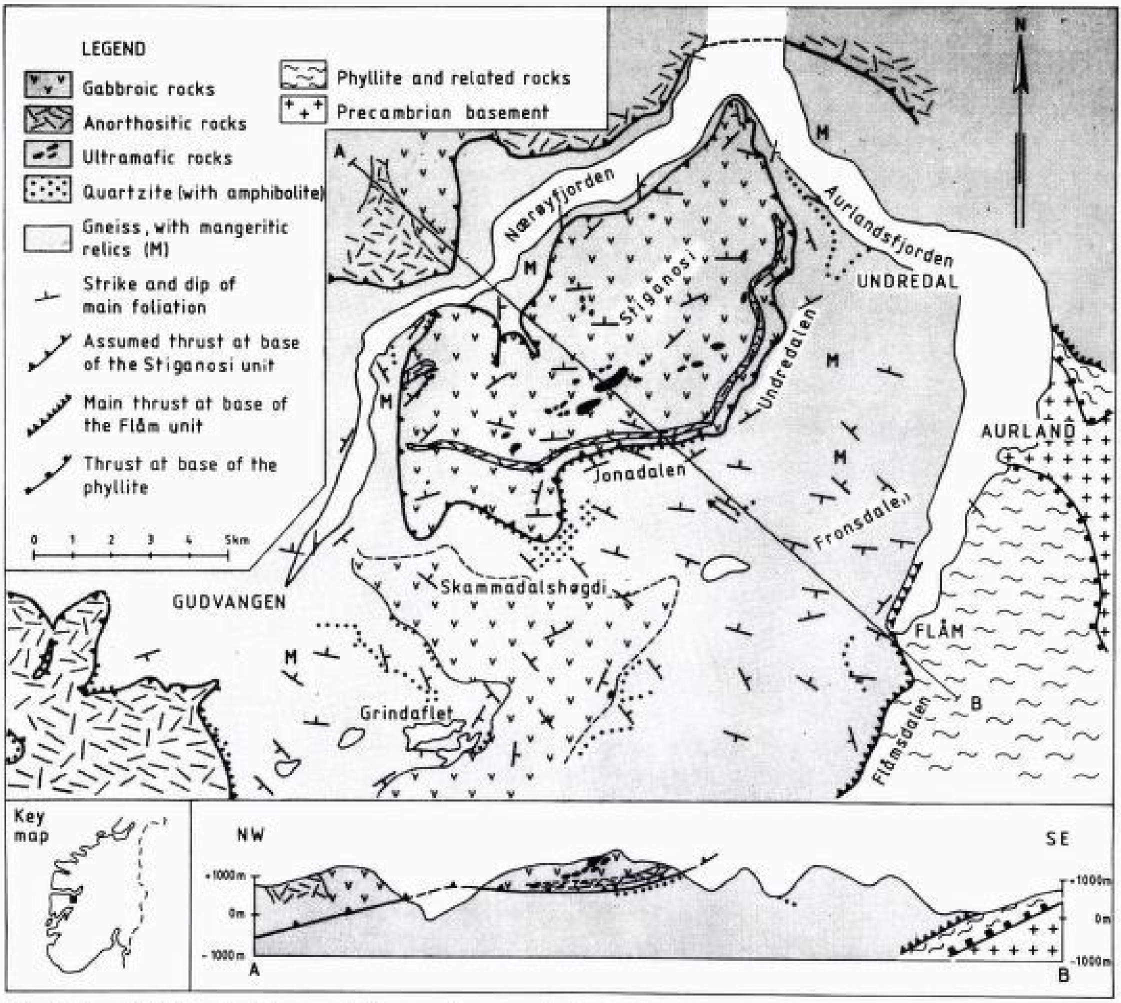

Fig. 1. Simplified geological map of the area between Aurlandsfjorden and Nærøyfjorden.

aim of the present paper is first to give a general description of these units and second to extend the tectonostratigraphy into a larger part of the 'Faltungsgraben'.

General geology

The two tectonic units between Aurlandsfjorden and Nærøyfjorden can be named after localities exposing their characteristic development:

1. *Stiganosi unit,* with gabbroic rocks, anorthosites and ultramafites.
2. *Flåm unit,* with gneisses, hornblendite, relics of mangeritic-jotunitic rocks and narrow bands of quartzite with amphibolite. Its basal thrust zone is characterized by schistose gneisses, blastomylonites and irregular intercalations of metamorphosed psammitic rocks (sparagmites?).

The Stiganosi is separated from the underlying Flåm unit by a crushed or cataclastically deformed zone, assumed to represent a major thrust fault. Minor low-angle faults are parallel to this boundary and quartzitic rocks often occur at intervals below it. Rb/Sr radiometric dating indicates that rocks within the lower unit are more than 1600 m.y. old, but that Sveconorwegian regeneration with amphibolite-facies metamorphism took place 1000–1200 m.y. ago (Bhanumathi & Bryhni, unpulished).

A few dykes of granodioritic composition have been observed, but only in the upper unit. They are related to the igneous activity so prominent in the Årdal–Kaupanger area farther to the north and where a Rb/Sr date of about 450 m.y. was recorded by Barthomier et al. (1972). Recent studies by Koestler (1982) indicate, however, that these late- or post-tectonic intrusives are actually about 900 m.y. and thus late Sveconorwegian in age.

FLÅM UNIT

The lowermost thrust unit is well exposed in the area close to Flåm railway station and in the lower part of Flåmsdalen where it overlies a more than 800 metres thick tectonic unit of phyllite (Fig. 1). The most important rocks are described below.

A. *Cataclastic Rocks*

The cataclastically deformed, often schistose zone at the base of the Flåm unit varies in thickness from a few metres to a few hundred metres. The effect of flattening within this thrust zone is demonstrated locally by the presence of boudins and by the realignment of originally cross-cutting pegmatite dykes into parallelism with the main foliation.

The rocks closest to the phyllite (0–100 m above the contact north of Aurland) are developed as compositionally layered blastomylonitic gneiss with a planar foliation parallel to the contact. There is no distinct lineation within the closely spaced foliation surfaces and the banding is often transected by dykes of black, flinty ultramylonite and cataclasite. A sample of such an ultramylonite dyke from only a few metres above the phyllite contact has a distinct fluxion structure defined by alignment of flat, irregular domains with different coloration formed by diminuation of individual minerals of an original medium-grained rock. Few mineral fragments in this rock exceed 0.2 mm in their longest dimension and most of the grains in the matrix are too small to be resolved under the microscope. Another sample, collected along the road below Skjerdal about 100 m from the contact, is massive and with only a faintly visible fluxion structure. Grain outlines indicative of a pre-existing medium-grained fabric can be seen in plane-polarized light, while inspection with crossed polarisers reveals that the average grain-size is less than 0.1 mm. Chlorite and calcite have formed in the matrix and quartz occurs in minor veins cutting through the rock fabric.

It is difficult to evaluate the exact extent of recrystallization in cataclastically deformed rocks, but the lack of easily visible strain effects in the matrix grains indicates that most of the investigated samples were strongly recrystallized. A rock from the thrust sole at Fronsdalen apparently consists of 70% recrystallized grains less than 1 mm in size and 30% of porphyroclasts about 0.2 mm across. It is a typical blastomylonite derived from similar feldspathic rocks as those occurring higher up in the nappe, possibly with some addition of new sericitic mica during cataclasis. Some have a laminated or finely layered structure falsely reminicent of supracrustal rocks, which may occur locally, however. Thus, a quartzose, grey-black, quartz net-veined blastomylonite in the contact zone along the west side of the fiord north of Flåm railway station represents a severely altered psammitic rock. It is composed mainly of quartz (c. 60%) and sericite (c. 40%), and makes up a layer only a few metres thick above black phyllite. Feldspathic sandstones with primary flood structures like coarse graded bedding, cross-bedding and cross-lamination have been recorded by Bryhni et al. (1977) in a similar tectonic position just outside our map-area (Ljosberget, southeast of Flåm). On the basis of our evidence, it is likely that also the sericite-quartz blastomylonite at Flåm is, in fact, a remnant of Valdres Sparagmite-type psammites which have become overridden by the Jotun Nappe Complex and are only preserved in places within the thrust zone.

B. *Gneisses and deformed plutonic rocks*

The Flåm unit consists mainly of gneisses which sometimes appear to have formed by cataclastic deformation, recrystallization and neomineralization from original plutonic rocks. The rocks have various mixtures of feldspar (microperthite, microcline, plagioclase), quartz, orthopyroxene, clinopyroxene, biotite, amphibole and garnet as essential minerals. Quartz rarely makes up more than 20% of the modes. Thus they classify in the field between syenite and gabbro, and can often be termed mangeritic or jotunitic. It is difficult to distinguish the different types in the field because the extent of secondary alteration is so variable. – Relatively unaltered mangeritic rocks are massive, medium- or coarse-grained with pale-brown, pink or grey feldspar. The feldspar is string microperthite or mesoperthite broken down in gneissic varieties to a more fine-grained mosaic of microline and acid plagioclase. This transformation is accompanied by complete alteration of original pyroxenes to amphibole, biotite and epidote.

C. *Quartzite with amphibolite*

A rock composed essentially of quartz (c. 70–90%) occurs in intimate association with amphibolites as regionally distributed layers, up to a few hundred metres thick. Apart from quartz the rock contain potash feldspar (including microperthite), plagioclase, biotite and minor white mica, sphene and epidote. Grain-size varies from medium-grained to very fine-grained, but the rock is mesoscopically often a 'glassy' white and grey orthoquartzitic rock. Hødal (1945) described corresponding rocks as 'silexites' of inferred magmatic origin, but we would rather interpret them as originally supracrustal rocks, either older than the associated gneissified mangerites-jotunites or tectonically interlayered with them. Fareth (1977) has described similar quartz rocks from his 'Klovafjell Group', and we must conclude that metasedimentary rocks have a much wider occurrence in the Jotun Nappe Complex than hitherto recognized.

CONTACT BETWEEN FLÅM AND STIGANOSI UNITS

The contact between the two units is distinct on the western side of Nærøyfjorden where it is a cataclastic zone, a few metres thick, which also cuts obliquely into the overlying Stiganosi unit. It is less conspicuous to the east of the fiord where it often is merely a cataclastically deformed contact zone between lithologically different rocks. To the southwest of Gudvangen it has also been mapped over extensive areas outside the present map-area. There the contact is a strongly deformed zone that appears to be affected by regional folding along an ESE-axis. Gabbroic rocks near the base of the Stiganosi unit above Jonadalen are strongly flattened and gneissified towards this cataclastic zone, which here follows closely a zone of quartzite with amphibolite. Quartzite also occurs in the strongly tectonized marginal zone of anorthositic rocks southwest of Gudvangen where also dykes of granodiorite are involved in the deformation (Ottesen, pers. comm. 1978). Schistose meta-anorthosites in the contact zone have retrograde mineral assemblages (sodic plagioclase down to An_{11}, epidote, chlorite and white mica).

STIGANOSI UNIT

The uppermost thrust unit is typically developed in the high ground west of Undredalen–Jonadalen where it underlies the mountain Stiganosi and the plateaux southwestwards toward Skammadalshøgdi (Fig. 1). The body of essentially gabbroic rocks at Stiganosi can be correlated with bodies of mainly anorthositic rocks northwest of Nærøyfjorden and southwest of Gudvangen. A large body of gabbro-norite at Grindaflet–Skammedalshøgdi bounded by blastomylonitic rocks with local quartzite can possibly be related to the same unit. The most important rocks are described below.

A. *Anorthostic rocks*

The largest anorthostic body, southwest of Gudvangen, is exposed in 1,200 metres high, almost vertical cliffs, above a sole of amphibolite, schistose gabbro, quartzite and blastomylonites. Another occurrence, a layer 100 m thick in gabbroic rocks between Undredalen and Gudvangen, has a blastomylonitic base and grades upwards into the gabbroic rocks or their amphibolitized derivatives. Small anorthostic bodies or layers less than 1 m thick also occur within the gabbro-norites, especially in the Grindaflet area. Primary igneous layering is locally preserved, but the rocks have more often a foliation parallel to the axial planes of tight folds. It would therefore appear that the more than 1,200 metres thickness of the Gudvangen anorthositic body is due to tectonic repetition of thinner layers.

The relative contents of dark minerals vary, so that there is a gradual transition from true anorthosite to anorthositic gabbro. The main minerals of such anorthositic rocks are plagioclase An_{60-75} (with inverse zoning up to An_{81}), olivine, clinopyroxene, garnet, amphibole, corundum and spinel.

The first foliation with flattened coronas of pyroxenes and garnet around original olivine (Griffin 1971 a) is cut by gabbroic dykes with marginal enrichment of garnet. Such dykes are locally strongly flattened in the foliation surface and altered to conformable amphibolitic layers. Schistose zones in the anorthositic rocks have fine-grained retrograde mineral assemblages, and contain a penetrative ESE–WNW trending lineation which is not found in the associated massive rocks. Thus, there are probably various sets of foliations formed under different metamorphic conditions spanning from the granulite to the greenschist facies.

B. *Gabbroic rocks*

The gabbroic rocks comprise banded as well as massive gabbro-norites or basic granulites with their amphibolitized derivatives. Plagioclase, orthopyroxene, clinopyroxene, amphibole and garnet are the main minerals.

A regular alternation of light and dark layers is characteristic for large parts of the gabbroic massif between Undredal and Nærøyfjorden, where the rocks often have a gneissose aspect. The light bands are relatively enriched in plagioclase, and locally contain dark red garnets in crystals up to several centimetres across. The dark layers are relatively enriched in amphibole. Other varieties are massive or display igneous graded layering, as is typically developed in the Grindaflet–Skammadalshøgdi body.

Blastomylonitic zones, up to 50 metres thick, transect the gabbroic rocks and locally include varieties that can be termed plagioclase augen gneisses.

C. *Ultramafite*

About 25 isolated occurrences of ultramafite have been mapped as inclusions in the gabbroic rocks. To this can be added many minor disrupted layers or ultramafic basal parts of graded beds in gabbro-norite. Post-crystalline tectonic movements within the main rock body led to disruption of the layers into fragments which become enclosed within the gabbroic rocks.

The main minerals of the ultramafite are olivine, orthopyroxene, clinopyroxene, amphibole, spinel and plagioclase. Orthopyroxene is always present in the contact zone between plagioclase and olivine, and the whole ultramafite body is usually bounded by an amphibolized reaction rim of enstatite, clinopyroxene and spinel towards the surrounding rocks, as described by Griffin (1971 b) from Jotunheimen. Garnet may be concentrated in the outer part of the reaction rim.

Igneous layering is sometimes developed in the ultramafite, grading upwards from olivine-rich to pyroxene-spinel-rich layers. The thickness of such graded layers is usually several metres, and locally there may be continuous gradation from the upper pyroxene-rich part of the layer into gabbronorite.

Linear structures

The primary igneous layering in the original Jotun plutonites has been obliterated in most places by strong transposition and isoclinal folding along axes trending between E–W and ESE–WNW. Stereograms of poles to the foliation and compositional layering show diffuse girdles around the same ESE–WNW axis, which is transverse to the axial trend direction of the Caledonian 'Faltungsgraben'. Pegmatite dykes sometimes cut foliated rocks with transverse linear structures, and limbs of major folds with this orientation are cut at the contact to the underlying phyllite. Such evidence – seen in conjunction with the limited radiometric data – suggests that the transverse axial structure in most cases pre-date the Caledonian emplacement of the Nappe Complex. There is, however, also evidence that transverse linear structures have been formed subsequent to emplacement. For example, the basal thrust zone south of Flåm is folded along an ESE–WNW axis, and similar trending lineations can be found in the phyllites and in the uppermost, 'Caledonized' part of the basement. We conclude, therefore, that the transverse linear structures have a composite origin both pre-dating and post-dating the Caledonian thrusting in the area.

A later deformation is recognizable as regional folds along the NE–SW Caledonian trend. It has affected the trust zone and is also visible as open, upright folds in schistose zones in the interior of the Nappe Complex. Structure-contour maps show that the basement/phyllite contact forms a series of regional elongated domes and basins with a similar Caledonian trend (Fareth 1977), and the 'Faltungsgraben' itself could be related to this phase of deformation.

In the area to the southeast of Aurland, Fareth (1977) found either calcareous schist or a zone of cataclasis and retrograde-metamorphosed rocks between groups which we correlate with our Flåm and Stiganosi units. Reconnaissance geological mapping by us in the area between Nærøyfjorden and Sognefjorden has indicated that the two major tectonostratigraphic units are developed over a large area (Fig. 2, and Bryhni et al. 1977, Qvale 1982). At

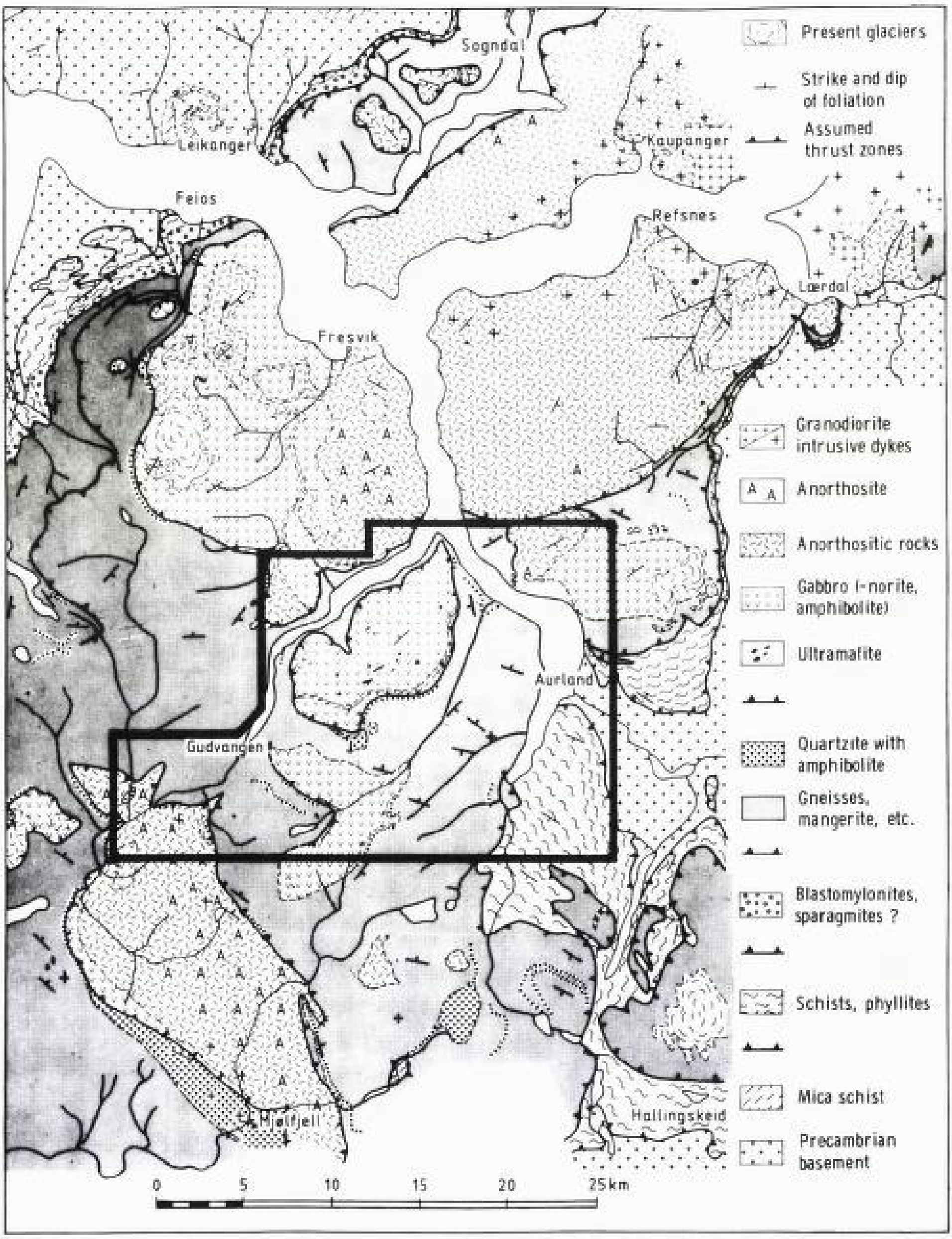

Fig. 2. Geological map of the 'Faltungsgraben' in Sogn. Area of Fig. 1. is outlined. (Mainly after Bryhni et al. 1977).

Sogndal the contact can be tied up with the boundary described by Skjerlie (1958) and Lacour (1969) between lower groups of essentially alkalifeldspar-bearing rocks and upper groups of rocks with abundant plagioclase. The Flåm unit is, however, in these western parts of the 'Faltungsgraben' probably underlain by one or possibly two thrusts which are not found in the Aurlandsfjorden–Nærøyfjorden area. The lower of these units has even recently been considered (Roberts 1977) to represent psammitic rocks ('Valdres Sparagmite') but is more likely to represent cataclastically deformed gneisses with only minor and locally distributed sediments. Its position at the base of the Jotun Nappe Complex probably indicates that it represents an analogue to the Valdres Nappes east of the 'Faltungsgraben', albeit in a very cataclastic state and with more basement than cover sediments represented.

A summary of the tectonostratigraphy is given here as Fig. 3. The apparent continuation of at least some of the tectonic units across the 'Faltungsgraben' indicates that the contacts are thrusts of regional significance. Movements along the contact between the two major units of the Jotun Nappe Complex have probably been extensive in this part of the 'Faltungsgraben' since dykes of granodiorite here appear to be present only in the upper unit (Bryhni et al. 1977, Henry 1977). The dykes are usually undeformed, but tend to become strongly deformed in the contact zone between the two units (Ottesen, pers. comm. 1978). Thus, at least some of the movement along the contact zone post-dates the intrusive date of about 900 m.y. recorded by Koestler (1982) and must then be either late Sveconorwegian or Caledonian in age. The contact south of Gudvangen shows local inversions which indicate a stronger folding than can reasonably be expected in the Jotun Nappe Complex during

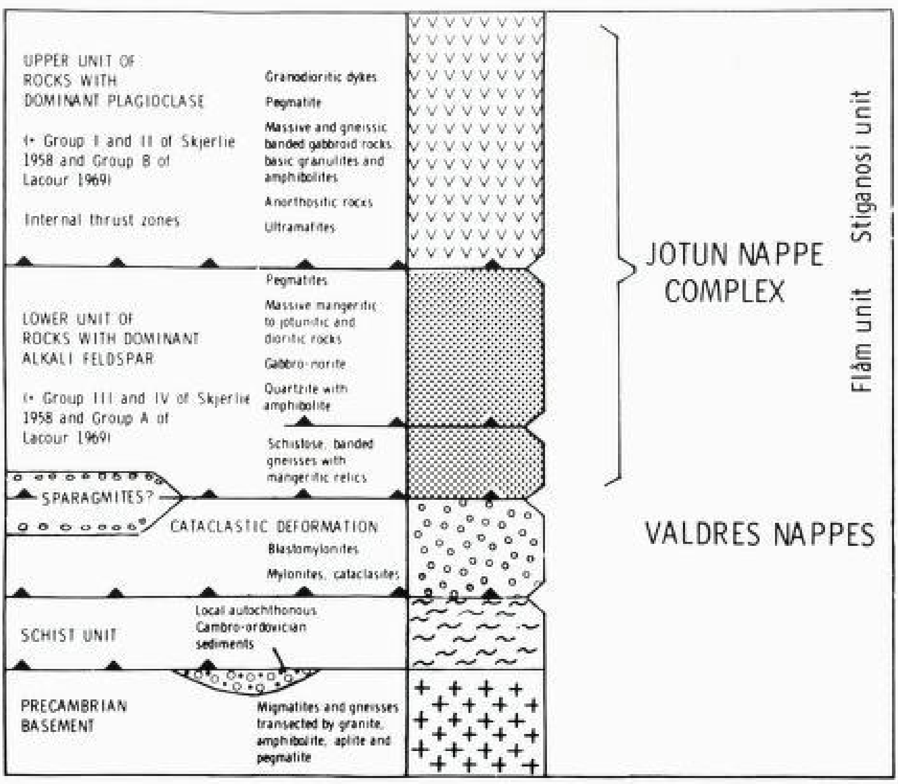

Fig. 3. Tectonostratigraphic subdivision of the Jotun Nappe Complex in Sogn.

any of the Caledonian events, and the emplacement of the Stiganosi unit upon the Flåm unit thus most likely pre-dates the Caledonian orogeny.

At Fillefjell, Corfu (1980) has subdivided the Jotun Nappe Complex into three subunits which were emplaced as nappes in a tectonic sequence and later metamorphosed *prior* to the movement of the whole pile as a Caledonian nappe with negligible internal deformation. The story in the area around Sognefjorden may turn out to be similar to that at Fillefjell although there is as yet no radiometric evidence.

The geometry of the internal thrust in the Jotun Nappe Complex and the presence also of supracrustals leads us to believe that the units are relatively shallow sheets which have been derived by horizontal thrusting rather than upthrusted from a root below the 'Faltungsgraben'.

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REFERENCES

Banham, P. H., Gibbs, A. D. & Hopper, F. W. M. 1979: Geological evidence in favour of a Jotunheimen Caledonian suture. *Nature 277*, 289–291.

Battey, M. H. & McRitchie, W. D. 1973: A geological traverse across the pyroxene-granulites of Jotunheimen in the Norwegian Caledonides. *Norsk geol. Tidsskr. 53*, 237–265.

Berthomier, C., Lacour, A., Leutwein, F., Maillot, J. & Sonet, J. 1972: Sur quelques trondhjemites de Norvège: étude géochronologique et géoshemique. *Sciences de la Terre 18*, 341–351.

Bryhni, I. 1976: Geologisk undersøkelse av tunnelprosjekter på stamveg Bergen–Oslo mellom Flåm og Styvi (Flåm-Gudvangen). *Rapport til Statens Vegvesen, Sogn og Fjordane* (Unpubl. report) 29 pp.

Bryhni, I. 1977: Geologisk undersøkelse av tunnelprosjekter på stamveg Bergen–Oslo mellom Flåm og Gudvangen. *Rapport til Statens Vegvesen, Sogn og Fjordane* (Unpubl. report), 20 pp.

Bryhni, I., Brastad, K. & Jacobsen, V. W. 1977: Jotundekket og dets underlag i Sogn. *Norges geol. Unders.*, Rapport 1560/28, 73 pp.

Corfu, G. N. 1980: Geologie der Jotun-decke sowie U/Pb und Rb/Sr Geochronologie des darunterliegenden Precambrischen Schieldes, Filefjeld, Suednorwegen. *Eidgenossischen Techn. Hochschule Zürich, Doktorabhandlung*, 137 pp.

Fareth, E. 1977: Petrography and structure of the Aurlandsdalen area, western Norway - a portion of the Caledonian margin. *Norges geol. Unders. 334*, 59–81.

Goldschmidt, V. M. 1916: Geologisch-petrographische Studien im Hochgebirge des südlichen Norwegens. IV. Übersicht der Eruptivgesteine im Kaledonischen Gebirge zwischen Stavanger und Trondhjem. *Skr. Norske Vidensk.-Akad. i Oslo, Mat.-Naturv. Kl., 1916, No. 2*, 140 pp.

Griffin, W. L. 1971a: Genesis of coronas in anorthosites of Upper Jotun Nappe, indre Sogn, Norway. *J. petrol. 12*, 219–243.

Griffin, W. L. 1971b: Mineral reactions at a peridotite-gneiss contact, Jotunheimen, Norway. *Min. Mag. 38*, 435–445.

Heim, M., Schärer, U. & Milnes, A. G. 1977: The nappe complex in the Tyin–Bygdin–Vang region, central southern Norway. *Norsk geol. Tidsskr. 57*, 171–178.

Henry, A. 1977: La nappe du Jotun et les nappes de socle connexes. *Revue de geogr. phys. et geol. dynamique XIX*, 433–442.

Henry, A. & Laccour, A, 1978: Les grand traits structuraux de la region de Sogndal, Caledonides de Norvège centro-méridionale. *Bull. Soc. géol. France XX*, 71–79.

Holtedahl, O. 1936: Trekk av det skandinaviske fjellkjedestrøks historie. *Skand. Naturforsk. Möte 19, Helsingfors,* 129–145.

Hossack, J. R. 1976: Geology and structure of the Beito Window, Valdres. *Norges geol. Unders. 327,* 1–33.

Hødal, J. 1945: Rocks of the anorthosite kindred in Vossestrand, Norway. *Norsk geol. Tidsskr. 24,* 129–243.

Koestler, A. G. 1982: A Precambrian age for the Ofredal granodiorite intrusion, Central Jotun Nappe, Sogn, Norway. *Norsk geol. Tidsskr. 62,* 225–228.

Kvale, A. 1948: Petrologic and structural studies in the Bergsdalen quadrangle, Western Norway. Part II, structural geology. *Bergens Museum Årbok 1946–1947. Naturv. rekke no. 1,* 155 pp.

Kvale, A. 1960: The nappe area of the Caledonides in western Norway. *Int. geol. Congress Norden 1960, Guide to excursions no. A7 and no. C4,* 43 pp.

Lacour, A. 1969: Données pétrologiques sur la nappe de Sogn–Jotun et son substratum dans la region de Sogndal (District de Sogn, Norvège centro-meridionale). *Bull. Soc. geol. France 7, XI,* 354–366.

Lacour, A. 1971: Observations sur la mylonization des roches de la nappe du Jotun dans la région de Sogndal, Norvège centro-meridionale. *Extrait du C. R. Sommaire des séances de la Soc. Geol. de Grance 1971. Fascicule 3,* 150–151.

Landmark, K. 1949: geologiske undersøkelser Luster–Bøverdalen. *Univ. i Bergen Årbok 1948, Naturv. rekke no. 1,* 57 pp.

Naterstad, J., Andresen, A. & Jorde, K. 1973: Tectonic succession of the Caledonian nappe front in the Haukelisæter–Røldal area, Southwest Norway. *Norges geol. Unders. 292,* 1–20.

Qvale, H. 1982: Jotundekkets anorthositter: geologi, mineralogi og geokjemi. *Norges geol. Unders.,* Rapport 1560/32, 98 pp.

Rekstad, J. 1905: Fra indre Sogn. *Norges geol. Unders. 43, no. 7,* 53 pp.

Reusch, H. 1908: Tekst til Geologisk kart over fjeldstrøkene mellom Jostedalsbreen og Ringerike. *Norges geol. Unders. 47,* 40 pp.

Roberts, J. L. 1977: Allochthonous origin of the Jotunheim Massif in southern Norway: a reconnaissance study along its northwestern margin. *Geol. Soc. London* 351–362.

Sigmond, E. M. O. 1978: Beskrivelse til det berggrunnsgeologiske kartbladet Sauda 1:250 000. *Norgs geol. Unders. 341,* 94 pp.

Skjerlie, F. S. 1958: Geological investigations between Fjærlandsfjord and Sogndalsdalen, Western Norway. *Univ. i Bergen Årbok 1957. Naturv. rekke no. 5,* 67 pp.

Smithson, S. B. & Ramberg, I. B. 1970: Geophysical profile bearing on the origin of the Jotun Nappe in the Norwegian Caledonides. *Geol. Soc. Amer. Bull. 81,* 1571–1576.

Smithson, S. B., Ramberg, I. B. & Grønlie, G. 1974: Gravity interpretation of the Jotun Nappe of the Norwegian Caledonides. *Tectonophysics 22,* 205–222.

Dating Explosive Volcanism Perforating the Precambrian Basement in Southern Norway

R. H. VERSCHURE, C. MAIJER, P. A. M. ANDRIESSEN, N. A. I. M. BOELRIJK, E. H. HEBEDA, H. N. A. PRIEM & E. A. Th. VERDURMEN

Verschure, R. H., Maijer, C., Andriessen, P. A. M., Boelrijk, N. A. I. M., Hebeda, E. H., Priem, H. N. A. & Verdurmen, E. A. Th. 1983: Dating explosive volcanism perforating the Precambrian basement in southern Norway. *Norges geol. Unders. 380*, 35–49.

K-Ar and Rb-Sr dating have been performed on vaarious hypabyssal diatreme-facies volcanic rocks perforating the Sveconorwegian zone of the Precambrian basement in southern Norway. The dated rocks were taken from the Fen peralkaline–carbonatitic area, the Gardnos and Hjölmodalen explosion breccias, the marginal breccia of the Fjone calcite–syenite plug, damtjernites from Gulbrandstjern, Presteöya and Brånan, and the carbonatized damtjernitic explosion breccia with abundant xenoliths from Tveitan and Hönstjern in the Bamble region. The results reveal two episodes of explosive volcanic activity prior to the Permian magmatism, one about 600 Ma ago (latest Precambrian) and another probably about 350–300 Ma ago (Carboniferous). These episodes of volcanic activity possibly reflect precursor phases of the Caledonian orogeny and the Permian epeirogeny, respectively. A dolerite dike in the Fen area may be of Permian age.

R. H. Verschure, P. A. M. Andriessen, N. A. I. M. Boelrijk, E. H. Hebeda, H. N. A. Priem & E. A. Th. Verdurmen, Z. W. O. Laboratorium voor Isotopen-Geologie, De Boelelaan 1085, 1081 HV Amsterdam, The Netherlands

C. Maijer, Instituut voor Aardwetenschappen, Rijksuniversiteit Utrecht, Budapestlaan 4, 3508 TA Utrecht, The Netherlands

Introduction

Explosive outbursts of volcanic and related subvolcanic activity have penetrated the Precambrian continental crust at numerous points in southern Norway (Figs. 1 and 2). This part of the Baltic Shield is characterized by ages between 1200 and 800 Ma, which link the cratonization to the Sveconorwegian (Grenville) Episode of tectonism, metamorphism and magmatism. Evidence of the existence of an older continental crust is provided, for example, by the relict ages of about 1500 Ma reported from the Levang gneiss dome in the Bamble region (O'Nions & Baadsgaard 1971), the Telemark supracrustals (Priem et al. 1973) and the high-grade charnockitic country rocks of the anorthositic complex in Rogaland (Versteeve 1975, Priem & Verschure 1982).

The best known example of the explosive volcanic phenomena in southern Norway is the peralkaline–carbonatitic complex of the Fen area (e.g. Brögger 1921, Sæther 1957, Bergstöl 1960, 1979, Bergstöl & Svinndal 1960, Barth &

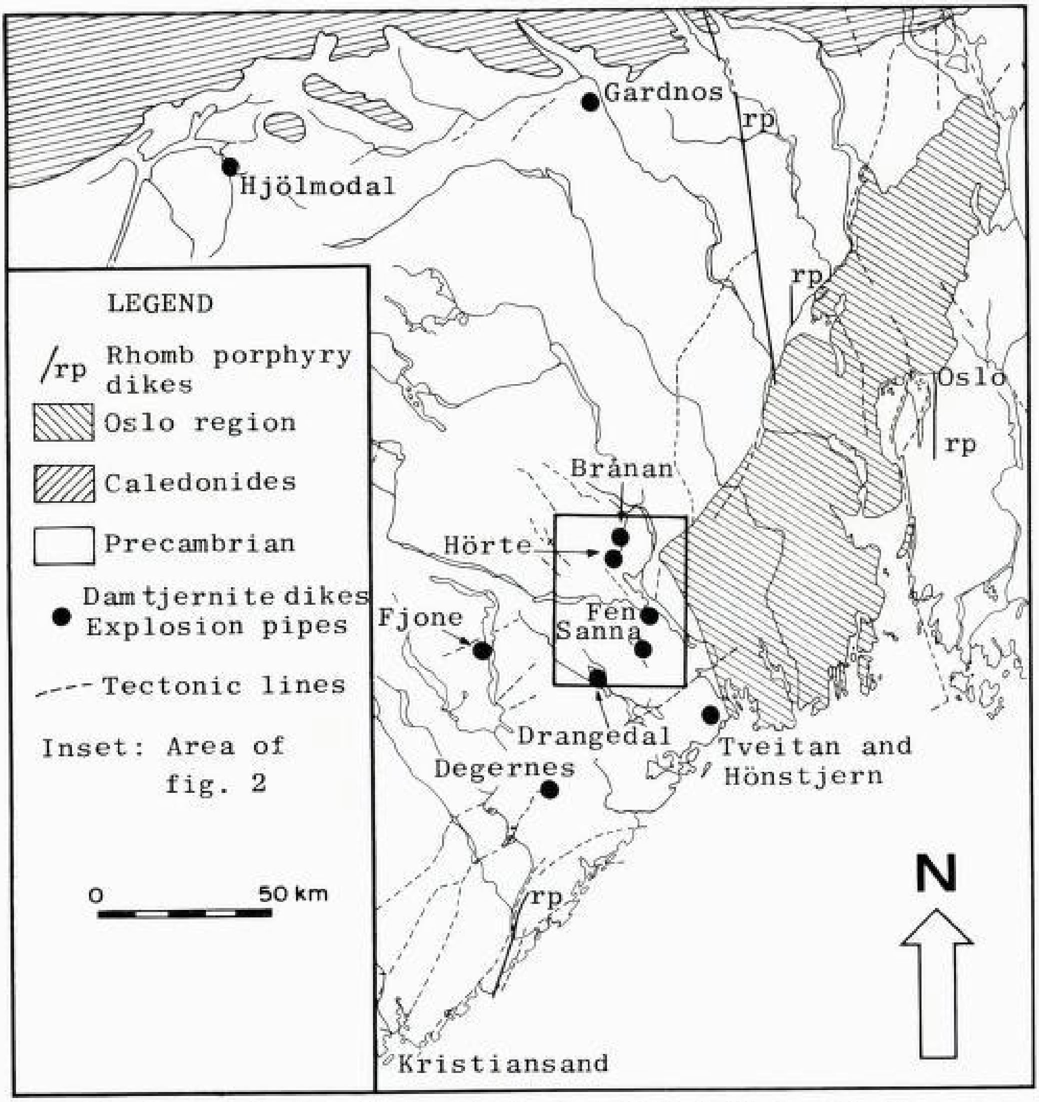

Fig. 1. Map showing the locations of the investigated rocks (modified after Ramberg & Barth 1966).

Ramberg 1966, Ramberg & Barth 1966, Mitchell & Crocket 1970, Griffin 1973). Less well-known are the alkaline vents and dikes, explosion breccias, and ultramafic damtjernitic plugs and dikes at many other points (Figs. 1 and 2). All these (sub)volcanic phenomena are supposedly related to faults and basement fractures. Damtjernite is a porphyritic hypabyssal rock with kimberlitic and alnoitic affiniTIES (Brögger 1921). The rock varies widely in mineral composition, but mostly it consists of a groundmass of carbonate, biotite, pyroxene and magnetite in which are embedded phenocrysts of biotite, clinopyroxene and brown hornblende in varying proportions. In diatreme facies the rock contains varying amounts of xenoliths and xenocrysts of crustal and upper-mantle origin (Griffin & Taylor 1975).

A number of age data have been published for the peralkaline-carbonatitic complex of the Fen area: early 'chemical' U-Th-Pb ages of koppite (Sæther 1957), Pb-α ages of zircon (Neumann 1960) and K-Ar ages of biotites (Faul et al. 1959, Neumann 1960, Kulp & Neumann 1961, Broch 1964). On the basis of these ages, all between 600 and 560 Ma, an Eocambrian age was assigned to the Fen volcanism (Ramberg & Barth (1966).

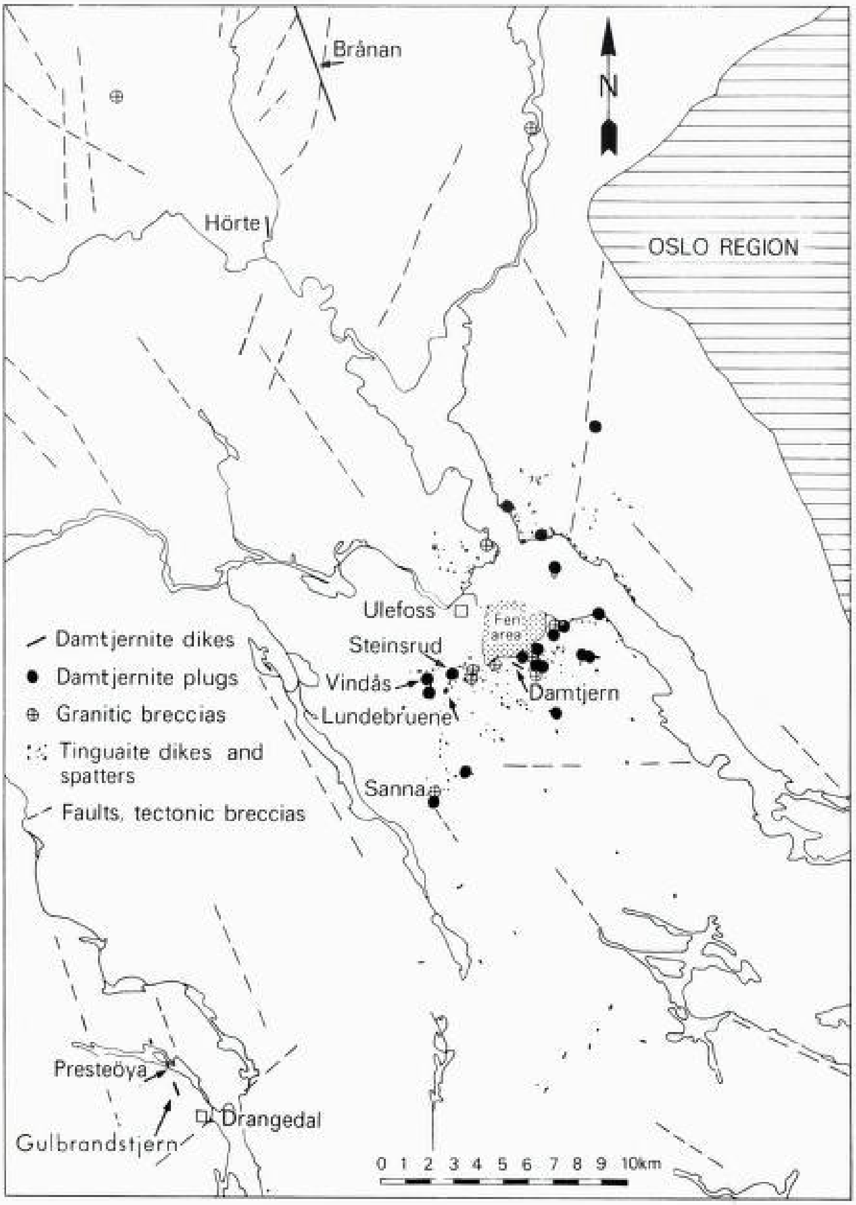

Fig. 2. Enlarged insert of Fig. 1 (modified after Ramberg & Barth 1966), showing the locations of the investigated rocks close to the Fen peralkaline–carbonatitic complex (indicated on the map as 'Fen area'). The locations of the samples from within the Fen complex are not shown.

The present study reports the results of K-Ar and Rb-Sr age measurements on separated minerals and whole-rocks from various types of explosive and related subvolcanic rocks in southern Norway (Figs. 1 and 2). A preliminary report of part of the data has been presented before (Verschure et al. 1977). The analytical data and calculated ages are listed in Tables 1, 2 and 3. More dating work is in progress.

Experimental procedures and constants

The usual techniques were applied for the analysis of potassium, argon, rubidium and strontium (see, for example, Wielens et al. 1980). The analytical accuracy is believed to be within 1% for K, 2% for radiogenic Ar, 1% for XRF Rb/Sr, 1% for isotope dilution Rb and Sr, and 0.05% for $^{87}Sr/^{86}Sr$ (0.2% for the Fen 6 biotite, which represents an older analysis; Second Progress-Report 1967). The errors are the sum of the estimated contributions of the known sources of possible systematic error and the precision (2σ) of the total analytical procedures. The ages are based upon the IUGS recommended set of constants:

$\lambda_e(^{40}K) = 0.581 \times 10^{-10}a^{-1}$, $\lambda_\beta(^{40}K) = 4.962 \times 10^{-10}a^{-1}$, abundance ^{40}K = 0.01167 atom percent total K and $\lambda(^{87}Rb) = 1.42 \times 10^{-11}a^{-1}$.

Results and discussion

FEN AREA

The investigated samples belong to different types of hypabyssal rocks. They come from six locations (Fig. 2):

1. The type-locality damtjernite forms a dike in diatreme facies near Damtjern, just outside the Fen complex. It is a dark porphyritic rock with phenocrysts of Ti-augite, brown hornblende and biotite (up to several cm in diameter), embedded in a fine-grained groundmass rich in carbonate and biotite. The damtjernite contains abundant xenoliths and xenocrysts of crustal rocks and upper-mantle spinel lherzolite. K-Ar analysis of biotite and hornblende yield ages of 564 ± 20 Ma and 597 ± 20 Ma, respectively.

2. Within the Fen complex, 0.5 km east of Söve, there is a dark, carbonatized damtjernite in diatreme facies. The rock contains abundant biotite flakes (up to 4 cm in diameter), which give a K-Ar age of 578 ± 20 Ma and a Rb-Sr model age between about 555 and 580 Ma, depending on the assumed initial $^{87}Sr/^{86}Sr$ ratio (0.705 or 0.702, respectively).

3. A less dark-coloured damtjernite in diatreme facies near Sanna (sannaite, Brögger 1921), 6 km SW of the Fen complex, consists of about 30% of alkali feldspar and a few sericitic aggregates (allegedly pseudomorphs after nepheline) in a groundmass rich in aegirine, carbonate, chlorite and

Fig. 3. Hybrid diatreme-facies damtjernite near Sanna (sannaite, Brögger 1921), with abundant fragments of country rock gneiss and phenocrysts of brown hornblende and biotite.

apatite (fig. 3). Phenocrysts of augite, large brown hornblende and some biotite are also present, as well as abundant fragments of country rock. Biotite and hornblende produce K-Ar ages of 580 ± 20 Ma and 598 ± 20 Ma, respectively.

4. The diatreme-facies damtjernite near Steinsrud, 1 km SW of the Fen complex, is likewise a less dark-coloured rock. Phenocrysts of clinopyroxene, large brown hornblende and biotite, along with phenocrysts or xenocrysts of feldspar (both sodic plagioclase and alkali feldspar) and aggregates of feldspar or feldspar-quartz are embedded in a groundmass of alkali feldspar, minor pyroxene, green amphibole, biotite, opaques, titanite and quartz. Biotite and hornblende give K-Ar ages of 523 ± 20 Ma and 597 ± 30 Ma, respectively.

5. The tinguaite dike near Lundebruene yields a K-Ar whole-rock age of 665 ± 20 Ma. According to Bergstöl (1960, 1979) tinguaite dikes and plugs occur only outside the actual Fen complex, but are especially frequent close to its borders (Fig. 2). He concludes that intrusion of the tinguaites is an early event in the evolution of the carbonatitic and peralkaline complex, a conclusion which is supported by the K-Ar date. The older age of the tinguaites is also confirmed by a xenolith of tinguaite observed in the type-locality damtjernite dike.

Table 1. K-Ar data and calculated ages.

Sample nr.	UTM-Coordinates	Rock type*	Material	K** (% Wt)	Radiogenic ^{40}Ar (ppm Wt)	Atmospheric ^{40}Ar (% total # ^{40}Ar)	Calculated age (Ma)***
Damtjern diatreme-facies damtjernite (type locality)							
Fen 260	$^{5}172$–$^{65}692$	breccia	biotite	6.54	0.300	10	564
Fen 260	$^{5}172$–$^{65}692$	breccia	hornblende	1.51	0.0740	4	597
Söve carbonatized diatreme-facies damtjernite							
Fen 6	$^{5}168$–$^{65}714$	breccia	biotite	8.07	0.381	10	578
Sanna hybridic diatreme-facies damtjernite							
Fen 225	$^{5}134$–$^{65}623$	breccia	biotite	6.84	0.324	10	580
Fen 225	$^{5}134$–$^{65}623$	breccia	hornblende	1.94	0.0935	2	598
Steinsrud hybridic diatreme-facies damtjernite							
Fen 281	$^{5}147$–$^{65}685$	breccia	biotite	6.63	0.279	19	523
Fen 281	$^{5}147$–$^{65}685$	breccia	hornblende	1.99	0.0977	2	597
Lundebruene tinguaite dike							
Fen 92	$^{5}145$–$^{65}678$	tinguaite	whole-rock	2.87	0.160	5	665
Fen carbonatized dolerite dike (Hydro quarry)							
Fen 1	$^{5}161$–$^{65}716$	dolerite	whole-rock	3.68	0.0694**	12	253
Gulbrandstjern hypabyssal-facies damtjernite							
Gul 1	$^{5}027$–$^{65}522$	damtjernite	biotite	6.27	0.310	4	601
Prestöya hypabyssal-facies damtjernite							
Pre 1	$^{5}024$–$^{65}534$	damtjernite	biotite	7.32	0.357	5	594
Brånan hypabyssal-facies damtjernite							
Brå 2	$^{5}077$–$^{65}951$	damtjernite	biotite	7.50	0.366	5	594
Hörte lamprophyre							
Hør 1	$^{5}073$–$^{65}873$	cpx-hbl lamprophyre	biotite	6.25	0.148	6	313
Fjone syenitic breccia							
Fjo 1	$^{4}694$–$^{65}591$	breccia	hornblende	0.846	0.0229	21	354
Gardnos explosion breccia							
Gar 9	$^{5}015$–$^{67}230$	breccia groundmass	whole-rock	3.71	0.103	25	362
Hjølmodalen explosion breccia							
Hjö 1A	$^{4}986$–$^{66}955$	breccia groundmass	whole-rock	2.93	0.152**	3	626
Hönstjern carbonatized damtjernite-like explosion breccia							
Htj 5	$^{5}354$–$^{65}418$	breccia	biotite	6.86	0.324	2	578
Tveitan carbonatized damtjernite-like explosion breccia							
Tve 1	$^{5}356$–$^{65}419$	breccia groundmass	whole-rock	371	0.0888**	5	316
Tve 4	$^{5}356$–$^{65}419$	breccia	chloritized biotite	3.44	0.0722**	26	280
Tve 9A	$^{5}356$–$^{65}419$	xenolith (gn-bio gneiss)	whole-rock	1.57	0.0626**	3	499
Tve 10A	$^{5}356$–$^{65}419$	xenolith (amphibolite)	whole-rock	2.26	0.0662**	5	380
Tve 11A	$^{5}356$–$^{65}419$	xenolith (bio amphibolite)	whole-rock	2.39	0.0823**	5	439
Tveitan country rock							
Tve 31	$^{5}351$–$^{65}422$	amphibolite	whole-rock	1.03	0.0582	17	675

* cpx, clinopyroxene; hbl, hornblende; gn, garnet; bio, biotite.
** Mean of duplicate or more analyses.
*** Error estimated at 3%, based upon estimated errors of 1% for K and 2% for Ar.

All three hornblende ages lie close to 600 Ma. The four biotite ages are lower, ranging from 580 to 523 Ma. This suggests that the hornblende ages approach the age of the volcanism, whereas the biotite K-Ar systems reflect varying degrees of resetting by post-volcanic processes.

There is also some evidence for a much younger magmatic event within the complex:

6. A strongly carbonatized dolerite from the Hydro quarry inside the Fen complex gives a K-Ar whole-rock date of about 255 Ma. If it is accepted as a true age, we are dealing either with a Permian dolerite, or with an older dolerite in which the K-Ar system has been completely reset in Permian time. Permian rejuvenation is conspicuously absent for biotites and hornblendes in the Fen area, however, so the K-Ar age is tentatively interpreted as reflecting the intrusion age of the dolerite. The dike may then be related to the Permian magmatism in the Oslo Graben. Sæther (1957) described several dolerites in the Fen area, all of them strongly altered, but he attributed the alteration to hydrothermal autometamorphism, being different from the alterations in other rocks of the Fen complex. On geological grounds both Sæther (1957) and Brögger (1921) regarded the dolerites in the Fen area as related to the Permian magmatism in the Oslo Graben, which is thus supported by the K-Ar date.

DAMTJERNITES OF THE SAME AGE OUTSIDE THE FEN AREA

Three damtjernites from outside the Fen area contain biotites which yield K-Ar ages concordant with the age of the Fen complex (Fig. 2):

1. A dark damtjernite dike near Gulbrandstjern (Kläy 1965), some 20 km SW of the Fen complex, consists of abundant phenocrysts of zoned pyroxene (Ti-augite, occasionally with aegirine-rich cores), biotite (strongly replaced by chlorite and apatite), opaques, zoned melanite and apatite in a groundmass of carbonate, chlorite, epidote, white mica, opaques, titanite, apatite and minor alkali feldspar. Biotite yields a K-Ar age of 601 ± 20 Ma.

2. Very similar in mineral composition, but containing less altered, zoned biotite, is the damtjernite dike on the nearby island Presetöya in Hoseivatn (Kläy 1965), which probably forms the continuation of the Gulbrandstjern dike. Biotite gives a K-Ar age of 594 ± 20 Ma.

Table 2. Rb-Sr biotite data and calculated model ages.

Sample nr.	UTM-Coordinates	Rock Type	Rb* (ppm Wt)	Sr* (ppm Wt)	$^{87}Sr/^{86}Sr$*	Calculated model age (Ma)**
Söve carbonatized diatreme-facies damtjernite						
Fen 6	$^{5}168-^{65}714$	breccia	413	161	0.7642	582 or 554
Tveitan carbonatized damtjernite-like explosion breccia						
Tve 4	$^{5}356-^{65}419$	breccia	511	387	0.71881	310 or 255

* Isotope dilution. Mean of duplicate analyses.
** Assuming an initial $^{87}Sr/^{86}Sr$ ratio of 0.702 or 0.705, respectively.

3. The damtjernite dike near Brånan (Werenskiold 1910), 20 km NNW of the Fen complex, is also similar in appearance. The rock, which has been described by Brögger (1921) and Griffin & Taylor (1975), consists of phenocrysts of zoned augite, zoned biotite and aggregates of carbonate in a groundmass of pyroxene, biotite, opaques, melanite, sericitic pseudomorphs after nepheline, apatite and carbonate. Biotite produces a K-Ar age of 594 ± 20 Ma.

The biotites from the three damtjernite dikes thus give similar ages close to 600 Ma, confirming the age of about 600 Ma assigned to the damtjernitic volcanism in the Fen area.

OTHER VOLCANIC ROCKS AND EXPLOSION BRECCIAS OUTSIDE THE FEN AREA

A lamprophyric dike, a syenitic plug and four explosion breccias have been investigated (Fig. 1):

1. The fine-grained, inhomogeneous, melanocratic to mesocratic, calcite-bearing lamprophyric dike near Hörte (Werenskiold 1910) contains abundant augite and brown hornblende, and varying contents of biotite and alkali feldspar. Locally, euhedral sericitic aggregates are observed (allegedly pseudomorphs after nepheline). Biotite produces a K-Ar age of 313 ± 10 Ma.

2. The syenitic plug near Fjone, along Nisservatn (Dons 1965) resembles the less dark-coloured parts of the lamprophyric dike near Hörte. Hornblende collected near the marginal breccia of this reddish, medium-grained, calcite- and quartz-bearing syenite plug gives a K-Ar age of 354 ± 10 Ma.

It may also be relevant to note that Touret (1970) has reported a K-Ar age of about 310 Ma for a biotite from the plug of hornblende periodotite at Degernes, along the great breccia zone which runs northeast from Kristiansand towards the Oslo Graben. All this evidence, although admittedly still rather scarce, points to a phase of subvolcanic alkaline magmatism in southern Norway about 250–300 Ma ago (Carboniferous).

Two of the investigated explosion breccias do not contain a magmatic component:

3. The Gardnos breccia (Broch 1945) forms an oval-shaped diatreme measuring about 4 by 5 km. It is filled with breccia of unsheared, angular country rock and mineral splinters, ranging in size from fragments of about 10 cm in diameter to dust-like particles, and an extremely fine-grained chloritic graphitic groundmass. According to Broch, fragments of folded phyllites of supposedly Cambrian age are present, implying that the diatreme has formed subsequent to Caledonian folding movements. The sample investigated in this study is the fine-grained groundmass filling the spaces

Table 3. Rb-Sr whole-rock data.

Sample nr.	UTM-Coordinates	Rock type*	Rb** (ppm Wt)	Sr** (ppm Wt)	Rb/Sr** (Wt/Wt)	$^{87}Sr/^{86}Sr$	$^{87}Rb/^{86}Sr$
Xenoliths of the Tveitan carbonatized damtjernite-like explosion breccia							
Tve 6	$^{5}356-^{65}419$	albitite	90.4	440	0.2055	0.71199	0.56
Tve 7	$^{5}356-^{65}419$	gn gneiss	137	110	1.247	0.77819	3.6
Tve 8	$^{5}356-^{65}419$	bio gneiss	133	235	0.566	0.73179	1.6
Tve 9	$^{5}356-^{65}419$	gn-bio gneiss	136	507	0.2673	0.71236	0.77
Tve 10	$^{5}356-^{65}419$	amphibolite	178	357	0.497	0.71466	1.4
Tve 11	$^{5}356-^{65}419$	bio amphibolite	188	258	0.728	0.72575	2.1
Tve 12	$^{5}356-^{65}419$	bio gneiss	99.2	365	0.2718	0.71903	0.78
Tve 13	$^{5}356-^{65}419$	bio gneiss	115	413	0.2766	0.71898	0.80
Tve 14	$^{5}356-^{65}419$	bio-hbl granite	76.7	86.7	0.884	0.74057	2.6
Tve 15	$^{5}356-^{65}419$	sheared gneiss	103.5	153	0.679	0.74853	2.0
Tve 16	$^{5}356-^{65}419$	bio-hbl gneiss	193	103	1.883	0.81751	5.5
Tve 18	$^{5}356-^{65}419$	hbl-bio gneiss	101	93.2	1.083	0.74824	3.1
Tve 19	$^{5}356-^{65}419$	gn gneiss	82.8	247	0.336	0.72219	0.97
Tveitan country rocks							
Tve 28	$^{5}351-^{65}422$	gn-bio gneiss	90.9	238	0.3831	0.72247	1.1
Tve 29	$^{5}351-^{65}422$	pegmatitic hbl-bio gneiss	93.9	197	0.4749	0.72751	1.7
Tve 30	$^{5}351-^{65}422$	pegmatitic hbl-bio gneiss	95.7	186	0.5159	0.72915	1.5
Tve 31	$^{5}351-^{65}422$	amphibolite	69.9	258	0.2711	0.71222	0.78
Tve 32	$^{5}351-^{65}422$	bio gneiss	80.3	249	0.3231	0.72191	0.94
Tve 33	$^{5}351-^{65}422$	bio gneiss	112	372	0.3003	0.72121	0.87
Tve 34	$^{5}351-^{65}422$	gn-hbl-bio gneiss	78.3	287	0.2737	0.71836	0.79

* gn, garnet; bio, biotite; hbl, hornblende.
** X-ray fluorescence spectrometry. Mean of duplicate analyses.

between larger fragments of presumably Sveconorwegian gneisses. A K-Ar whole-rock analysis of the groundmass yields a date of 362 ± 10 Ma. This age may be interpreted as reflecting a complete resetting of the K-Ar system, or nearly so, during the fragmentation due to the explosive brecciation. The K-Ar date may then be taken as setting a maximum to the age of the explosive activity, which supports Broch's conclusion that the Gardnos breccia is post-Caledonian.

4. The Hjölmodalen breccia (Svinndal & Barkey 1967) closely resembles the Gardnos breccia. The rock shows less alteration, however, and the size of the diatreme is much smaller, some 130 x 150 m. A K-Ar analysis of the groundmass gives a date of 626 ± 20 Ma. If the explosive volcanism leading to the Hjölmodalen breccia was contemporaneous with that of the Gardnos breccia, this date may be interpreted as reflecting a much less advanced resetting of the K-Ar system of the gneiss mateial than in the case of the Gardnos breccia.

The age of the Gardnos and Hjölmodalen breccias is thus not yet firmly established. On the basis of the K-Ar age of the Gardnos breccia, however, this phase of explosive activity may tentatively be correlated with the alleged phase of subvolcanic alkaline magmatism about 350–300 Ma ago.

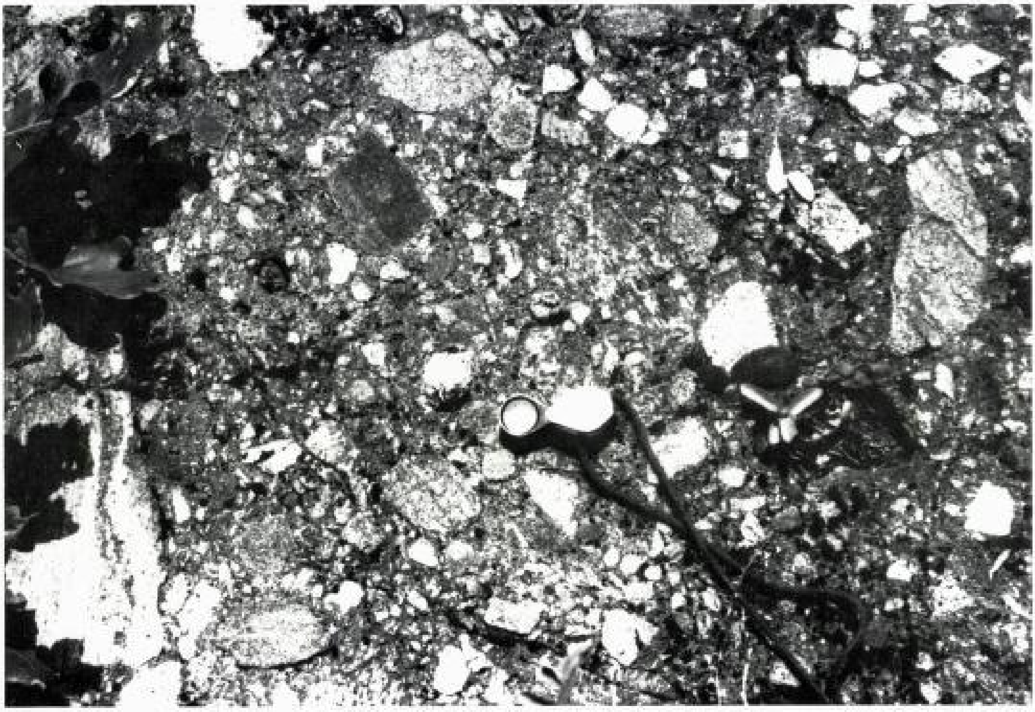

Fig. 4. Carbonatized damtjernite-like explosion breccia near Tveitan, Bamble region, with numerous fragments of country rock.

Two explosion breccias produce ambiguous and conflicting age data:

5. In the Bamble region two carbonatized damtjernite-like explosion breccias have been studied, the Hönstjern breccia and the Tveiten breccia (Fig. 4). They are situated less than 0.5 km apart, in an area dominated by anatectic paragneisses with intercalated amphibolites and metagabbros (Morton et al. 1970). The breccias lie about 10 km W of the nearest exposure of Permian intruisives of the Oslo Graben. The breccias are very similar; they consist of a wide variety of xenoliths and xenocrysts in a very fine-grained groundmass consisting mainly of carbonate, green biotite, opaques and apatite (Fig. 4). Among the xenoliths three groups can be distinguished: (1) small, rounded fragments (up to 0.5 cm in diameter) of ultramafic, occasionally porphyritic rocks with phenocrysts of biotite or brown hornblende; (2) larger, angular fragments (up to 10 cm in diameter) of crustal gneisses, amphibolites, granites and metagabbros; and (3) occasional fragments of a similar damtjernitic breccia. Many of the xenoliths and xenocrysts are strongly altered, but the abundant apatite phenocrysts and the cores of biotite phenocrysts and perthite xenocrysts do not show any alteration. Numerous veinlets of carbonate transect both the xenoliths and the groundmass.

Biotite phenocrysts from the Hönstjern breccia yield a K-Ar age of 578± 20 Ma, concordant with the age of the Fen complex. The partly chloritized biotite booklets from the Tveitan breccia give much younger ages, however: a K-Ar age of 280 ± 10 Ma and a Rb-Sr model age between about 310 Ma and 255 Ma, depending on the assumed initial $^{87}Sr/^{86}Sr$ ratio (0.702 and 0.705, respectively). The early Permian age of the Tveitan biotite is supported by four K-Ar whole-rock dates obtained from the same breccia: an age of 316 ± 10 Ma for the groundmass and ages between 500 and 380 Ma for three crustal xenoliths. The latter three ages could very well be interpreted as reflecting varying degrees of resetting of the K-Ar systems of Sveconorwegian crustal fragments during transport by the exploding magma in Permian time.

A suite of 14 gneiss and amphibolite xenoliths from the Tveitan breccia has been investigated (for petrographic details see Table 4). The Rb-Sr datapoints scatter widely (Fig. 5), but all lie between boundary lines corresponding to ages of 1.4 Ga and 0.2 Ga. Five gneisses and one granite define a crude linear correlation along a 0.9 Ga line; these may be interpreted as signalling derivation from (possibly slightly reset) Sveconorwegian source rocks. The

Table 4. Petrography of 14 crustal xenoliths from the carbonatized damtjernite-like explosion breccia near Tveitan, Bamble region[1]

		texture								primary minerals								secondary minerals				
		hypidiomorphic	fine-grained	medium-grained	leucocratic	mesocratic	melanocratic	foliated	banded	quartz	K-feldspar perthite	plagioclase	anorthite %	garnet	hornblende	hbl colour	biotite (primary)	biotite (secondary)	chlorite	carbonate	sericitization (plagioclase)	albitization (plagioclase)
Tve 6	Albitite	*		*	*							xx	0						x	x		**
Tve 7	Gn gneiss		*		*			*		xx	xx	xx	20	x					x		*	*
Tve 8	Bio gneiss		*			*		*		xx	xx	xx	20					x	x	x	*	*
Tve 9	Gn-bio gneiss		*	*		*		*		xx		xx	30	x					x	x	*	
Tve 10	Amphibolite		*				*	*		x		xx	40		xx	gb		x			*	
Tve 11	Bio amphibolite			*			*	*				xx	35		xx	pg	xx	x	x		*	*
Tve 12	Bio gneiss		*		*	*		*	*	xx		xx	30		x	g	xx	x	x	x	*	*
Tve 13	Bio gneiss		*	*	*			*		xx		xx	30				x		x		*	*
Tve 14	Bio-hbl granite	*	*	*	*					xx	xx	xx	20		x	dg	x					
Tve 15	Sheared gneiss		*		*			*		xx	xx	xx	0				x		x	x	*	**
Tve 16	Bio-hbl gneiss		*		*			*		xx	xx	xx	20		x	dg	x			x	*	
Tve 18	Hbl-bio gneiss			*	*	*		*		xx	xx	xx	20		xx	dg	x		x	x		
Tve 19	Gn gneiss		*		*			*		xx		xx	20	x							*	*
Tve 21	Gn-bio gneiss		*	*		*		*		xx	xx	xx	20	x			x		x		*	*

[1] *, Applicable; xx, major component; x, minor component; g, green; b, brown; p, pale; d, dark; gn, garnet; bio, biotite; hbl, hornblende.

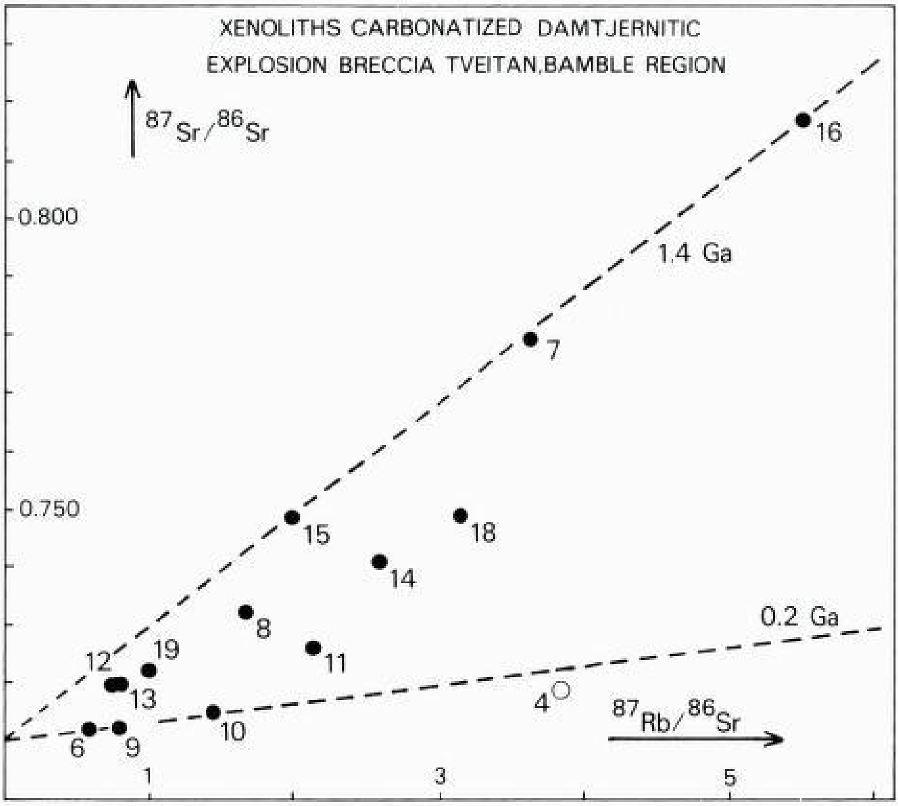

Fig. 5. Plot of the Rb–Sr whole-rock data of xenoliths (closed circles) and of the biotite (open circle) from the carbonatized damtjernite-like explosion breccia near Tveitan, Bamble region.

three points defining the upper boundary line of 1.4 Ga. (Tve 7, 15 and 16) may testify to the presence of older, pre-Sveconorwegian elements among the source rocks. On the other hand, the three points defining, or close to the lower boundary line of 0.2 Ga could very well reflect complete resetting, or nearly so, of the Rb-Sr systems of fragments of Sveconorwegian or older rocks during transport by the exploding magma in Permian time.

There thus appears to be a difference between the age of the Hönstjern breccia and that of the Tveitan breccia; about 580 Ma for the former and about 280 Ma for the latter. The simplest explanation is that the age difference is real, the Hönstjern breccia having been formed in the latest Precambrian, in relation to the damtjernite volcanism elsewhere, and the Tveitan breccia having formed in the Permian and associated with the magmatism in the nearby Oslo Graben. The similarity between both breccias is then difficult to understand, however. Another explanation could be that both breccias were formed about 280 Ma ago, but that the Hönstjern breccia contains biotite derived from an older rock, carried upwards by the exploding magma.

For comparison, Rb-Sr whole-rock analyses have also been carried out on a suite of six gneisses and one amphibolite from country rock, collected at an

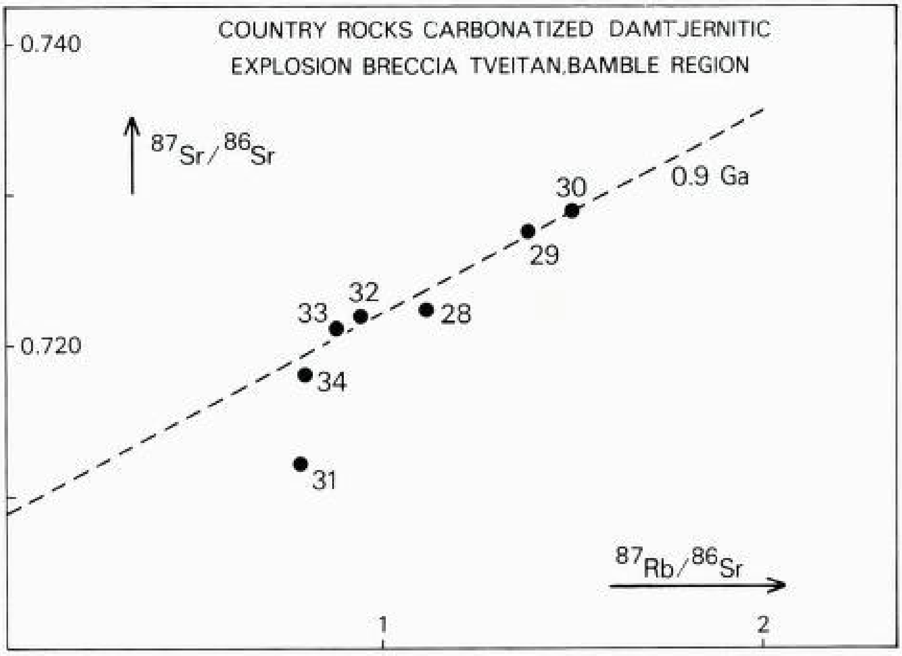

Fig. 6. Plot of the Rb-Sr whole-rock data of country rock gneisses and amphibolite (31) 0.5 km NW of the carbonatized damtjernite-like explosion breccia near Tveitan, Bamble region.

outcrop 0.5 km NW of Tveitan (Table 4). The Rb-Sr data-points (Fig. 6) display a crude linear correlation along a line corresponding to an age of 0.9 Ga, roughly concordant with the six intermediate crustal xenoliths from the Tveitan breccia. The scatter of data-points may be related to the retrogradation and minor carbonate veining in these rocks (Table 5). The amphibolite Tve 31 falls below this line, which may be due to a lower initial $^{87}Sr/^{86}Sr$ ratio; its low K-Ar age may be attributed to a partial resetting.

Conclusions

Two episodes of (sub)volcanic activity, both of explosive nature and associated with alkaline magmatism, are apparent in southern Norway before the Permian magmatism in the Oslo Graben. The first episode is dated at about 600 Ma, but the age of the second episode is still uncertain; it probably lies at about 350–300 Ma. The older episode may be related to a precursor phase of the Caledonian orogeny, as has already been proposed by Brögger (1921) for the Fen complex. Similarly, the younger episode may be related to a precursor phase of the epeirogenic movements which ultimately led to the formation of the Oslo Graben system. The intrusion of the doleritic dikes in the Fen area could be related to the Permian magmatism in the Oslo Graben. Doig (1970) has reported the existence of an extensive 580–570 Ma old alkaline magmatic province on both sides of the Nort Atlantic region. It is of interest to note that Doig also reports some Late Paleozoic ages for alkaline rocks in the western part of this province.

Table 5. Petrography of country rock gneisses and an amphibolite 0.5 km NW of the carbonatized damtjernite-like explosion breccia near Tveitan, Bamble region[1]

	texture									primary minerals							secondary minerals				
	equigranular	inequigranular	fine-grained	medium-grained	leucocratic	mesocratic	melanocratic	foliated	banded	quartz	K-feldspar perthite	plagioclase	anorthite	garnet	biotite	hornblende	chlorite	chlorite + carbonate (pseudomorph hbl?)	carbonate	veinlets (carb + chl + adularia)	sericite
Tve 28: Gn-bio gneiss		*		*		*		*		xx	xx	xx	25	x	xx		x	x	x		x
Tve 29: (Pegm.) hbl-bio gneiss		*	*	*		*		*		xx	xx	xx	25	x	x	x	x	x	x		x
Tve 30: (Pegm.) hbl-bio gneiss		*	*	*		*		*		xx	xx	xx	25		x	x	x				
Tve 31: Amphibolite	*		*				*					xx	70			xx	x	x			xx
Tve 32: Bio gneiss		*	*	*		*		*	*	xx		xx	30		x		xx			x	x
Tve 33: Bio gneiss		*	*	*	*			*		xx		xx	30		x		xx		x		x
Tve 34: Gn-hbl-bio gneiss		*	*	*		*		*		xx	xx	xx	25	x	x	x	xx	x	x	x	x

[1] *, Applicable; xx, major component; x, minor component; carb, carbonate; chl, chlorite; bio, biotite, gn, garnet; hbl, hornblende; pegm, pegmatitic.

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REFERENCES

Barth, T. F. W. & Ramberg, I. B. 1960: The Fen circular complex. In: Carbonatites. Tuttle, O. F. & Gittins, J. eds. Interscience Publishers, New York, 225–257.

Bergstöl, S. 1960: En petrografisk og mineralogisk undersökelse av bergartene rundt Fensfeltet, med hovedvekt gangbergartene. Unpublished internal report, University of Oslo, 55 pp.

Bergstöl, S. 1979: Tinguaite dikes adjacent to the Fen alkaline complex in Telemark, Norway. *Norsk geol. Tidsskr. 59,* 115–124.

Bergstöl, S. & Svinndal, S. 1960: The carbonatite and per-alkaline rocks of the Fen Area. *Norges geol. Unders. 208,* 99–110.

Broch, O. A. 1945: Gardnosbreksjen i Hallingdal. *Norsk geol. Tidskr. 25,* 16–25.

Broch, O. A. 1965: Age determination on Norwegian minerals up to March 1964. *Norges geol. Unders. 228,* 84–113.

Brögger, W, C. 1921: Die Eruptivgesteine des Kristianiagebietes IV. Das Fengebiet in Telemark, Norwegen. *Vid. Akad. Skr. 9,* 408 pp.

Doig, R. 1970: An alkaline rock province linking Europa and North America. *Can. J. Earth Sci. 7,* 22–28.

Dons, J. A. 1965: The Precambrian rocks of the Telemark area in South Central Norway IV. Calcite syenite with marginal breccia at Fjone. *Norges geol. Unders. 45,* 57–68.

Faul, H., Elmore, P. L. D. & Brannock, W. W. 1959: Age of the Fen carbonatite (Norway) and its relation to the intrusives of the Oslo region. *Geochim. Cosmochim. Acta 17,* 153–156.

Griffin, W. L. 1973: Lherzolite nodules from the Fen alkaline complex, Norway. *Contrib. Mineral. Petrol. 38*, 135–146.

Griffin, W. L. & Taylor, P. N. 1975: The Fen damkjernite: petrology of a "central complex kimberlite". Physics and Chemistry of the Earth 9. Ahrens, L. H., Dawson, J. B., Duncan, A. R. & Erlank, A. J. eds., Oxford, 163–177.

Kläy, L. 1965: Drangedal i Kilebygda. Unpublished report University of Oslo.

Kulp, J. L. & Neumann, H. 1961: Some K-Ar ages from the Norwegian basement. *Ann. New York Acad. Sci. 91*, 469–475.

Morton, R. D., Batey, R. & O'Nions, R. K. 1970: Geological investigations in the Bamble sector of the Fennoscandian shield in South Norway. 1. The geology of Eastern Bamble. *Norges geol. Unders. 263*, 72 pp.

Mitchell, R. H. & Crocket, J. H. 1972: Isotopic composition of strontium in rocks of the Fen alkaline complex, South Norway. *J. Petrology 13*, 83–97.

Neumann, H. 1960: Apparent ages of Norwegian rocks and minerals. *Norsk geol. Tidsskr. 40*, 151–173.

O'Nions, R. K. & Baadsgaard, H. 1971: A radiometric study of polymetamorphism in the Bamble Region, Norway. Contrib. Mineral. *Petrol. 34*, 1–21.

Priem, H. N. A. & Verschure, R. H. 1982: Review of the isotope geochronology of the high-grade metamorphic Precambrian of SW Norway. *Geol. Rundschau 71*, 81–84.

Priem, H. N. A., Boelrijk, N. A. I. M., Hebeda, E. H., Verdurmen, E. A. Th. & Verschure, R. H. 1973: Rb-Sr investigations of Precambrian granites, granitic gneisses and acidic metavolcanics in central Telemark: metamorphic resetting of Rb-Sr whole-rock systems. *Norges geol. Unders. 289*, 37–53.

Ramberg, I. B. & Barth, T. F. W. 1966: Eocambrian volcanism in southern Norway. *Norsk geol. Tidsskr. 46*, 219–236.

Sæther, E. 1957: The alkaline rock province of the Fen Area in southern Norway. *Det Kgl. Norske Vidensk. Selsk. Skr. 1*, Trondheim, 150 pp.

Second Progress-Report on the Isotopic Dating Project in Norway 1967: Z.W.O. Laboratorium voor Isotopen-Geologie, Amsterdam, 36–38.

Svinndal, S. & Barkey, H. 1967: Contributions to the geology of Hardangervidda (West-Norway) I. An explosion-breccia occurrence in Hjölmodalen. *Norges geol. Unders. 247*, 44–50.

Touret, J. 1970: Deep-seated volcanism along the major Precambrian breccia in South Norway I. Degernes at Lake Vegår. *Norsk geol. Tidsskr. 50*, 249–252.

Verschure, R. H., Boelrijk, N. A. I. M., Hebeda, E. H., Maijer, C., Priem, H. N. A. & Verdurmen, E. A. Th. 1977: Dating explosive volcanism in the Precambrian basement of southern Norway–preliminary results. Fifth European Colloquium of Geochronology, Cosmochronology and Isotope Geology, Pisa 1977, Volume of Abstracts.

Versteeve, A. 1975: Isotope geochronology in the high-grade metamorphic Precambrian of southwestern Norway. *Norges geol. Unders. 318*, 1–50.

Werenskiold, W. 1910: Om Øst-Telemarken. *Norges geol. Unders. 53*, 46–48.

Wielens, J. B. W., Andriessen, P. A. M., Boelrijk, N. A. I. M., Hebeda, E. H., Priem, H. N. A., Verdurmen, E. A. Th. & Verschure, R. H. 1980: Isotope geochronology in the high-grade metamorphic Precambrian of soutwestern Norway: new data and reinterpretations. *Norges geol. Unders. 359*, 1–30.

Note added in proof: Throughout this paper the editors prefer the modern spelling damtjernite instead of the original spelling damkjernite used by the majority of the authors on the Fen area (e.g. Barth & Ramberg 1960, Bergstöl 1979, Brögger 1921, Griffin & Taylor 1975).

Rb-Sr Study of Rapakivi Granite and Augen Gneiss of the Risberget Nappe, Oppdal, Norway

ALLAN G. KRILL

Krill, A. G. 1983: Rb-Sr study of rapakivi granite and augen gneiss of the Risberget Nappe, Oppdal, Norway. *Norges geol. Unders. 380*, 51–65.

Rb-Sr dating of rapakivi granite and associated augen gneiss of the Risberget Nappe suggests that the rocks are post-Svecokarelian intrusions that were strongly foliated during the Caledonian orogeny. Microstructures of the augen indicate igneous crystallization followed by cataclastic and ductile deformation, and the augen are interpreted as porphyroclasts rather than metasomatic porphyroblasts. Rb-Sr data from 56 presumably co-magmatic rapakivi rocks yield an errorchron date of 1618 ± 44 Ma with low initial Sr ratio (.7021 ± 5). The least deformed rapakivi samples are isotopically disturbed and partially reset, while an isochron from an ultramylonite within the Risberget Nappe yields a date within the range of error of the suggested 1618 Ma intrusive date. The dating of a younger pegmatite intrusion (1163 ± 80 Ma) suggests that the rocks were subjected to a Sveconorwegian event. Minerals of unfoliated rapakivi granite are only partially reset to Sveconorwegian and Caledonian dates, while minerals of foliated rocks yield Caledonian isochrons.

Allan G. Krill, Norges geologiske undersøkelse, Postboks 3006, N-7001 Trondheim, Norway

Introduction

The distinctive coarse augen gneisses of the Western Gneiss Region are especially well known from the Oppdal district. Törnebohm (1896) correlated them with the Precambrian granite and augen gneiss of the Caledonian allochthon of Jämtland. Bjørlykke (1905) and Goldschmidt (1916) considered them to be derived from Caledonian granitic intrusions. Carstens (1924, 1925) showed that some of the Oppdal augen gneisses have wiborgitic rapakivi textures, i.e., with thin plagioclase rims mantling the large K-feldspar megacrysts, and that they compare reasonably well in composition, texture and mineralogy with the type rapakivi rocks in Finland. He further demonstrated the isochemical transition of massive rapakivi rock to foliated augen gneiss. Barth (1938) and Holtedahl (1938) developed theories of granitization by suggesting a metasomatic origin for the augen. Rosenqvist (1943) documented the presumed metasomatic changes with descriptions and chemical analyses of the various rock types.

The origin of the augen gneiss by large-scale metasomatism has generally been accepted. This interpretation has strongly influenced previous mapping and stratigraphic/structural interpretation, as augen gneisses were only mapped where 'well developed'. More recently, Eggen (1977) mapped and interpreted them as deformed igenous rocks, and Solheim (1980) interpreted Rb-Sr dating to show that any metasomatism must have been Precambrian. Krill (1980a, 1980b) supported these recent studies and the earlier interpretations of Törnebohm and Carstens, suggesting that the Oppdal augen gneisses are not metasomatic but Precambrian porphyritic granites that were cataclastically deformed during the Caledonian orogeny. They form a major part of the Caledonian Risberget Nappe.

Textures of the granite and augen gneiss

The earlier studies included detailed petrographic and geochemical descriptions, and only the most significant textural features are mentioned here. Large pods of massive or weakly foliated rapakivi granite are preserved within the augen gneiss. In the rapakivi granite (Fig. 1), K-feldspar phenocrysts are mainly single-crystal and Carlsbad-twinned ovoids, ranging up to 15 cm in diameter. Most of the phenocrysts are mantled by a continuous rim of plagioclase, generally 2–5 mm thick and independent of the size of the K-feldspar core. The K-feldspar (Fig. 2) is mainly pink to purple perthitic microcline with inclusions of plagioclase, quartz, biotite and hornblende. The inclusions are randomly oriented, rarely in concentric zonal patterns, and have optical properties and grain sizes similar to those of minerals in the matrix.

Fig. 1. Massive rapakivi granite with shear zone. (UTM coord. 322206). Ruler is 16 cm long.

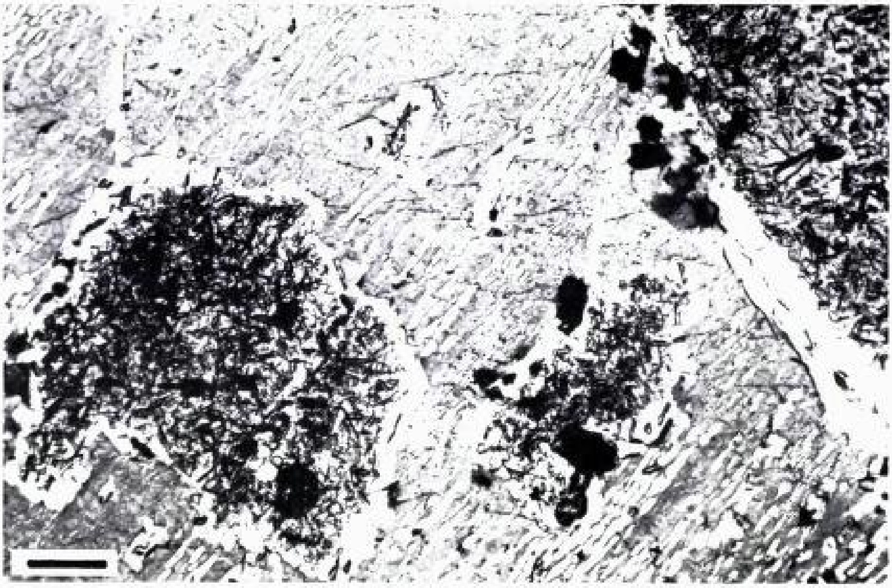

Fig. 2. Rapakivi K feldspar megacryst and matrix. The megacryst is perthitic, and the plagioclase rim (upper right corner) and included grains are saussuritized (note zoisite). Inclusion-free plagioclase separates saussuritized plagioclase from K-feldspar. Bar scale = 0.1 mm. Crossed polars.

The plagioclase rim consists of large albite-twinned grains, nearly optically continuous with each other and with the K-feldspar core. The plagioclase is strongly saussuritized with abundant zoisite inclusions. All the plagioclase in each rock – in the mantle, as augen inclusions, and in the matrix – has the same appearance and composition (An 10–30).

The very dark matrix of the underformed rock is produced by Fe-rich biotite and ferro-paragasitic hornblende, both typical of rapakivi granites (Simonsen & Vorma 1969). Other matrix mineral are idioblastic garnet, blue quartz, saussuritized plagioclase, ilmenite with rims of sphene, zircon and very minor uralitized clinopyroxene. No fine-grained K-feldspar is found in the matrix, suggesting that K-feldspar did not nucleate easily, allowing the few crystals that did nucleate to grow quite large.

Strain textures are apparent even in the least deformed rapakivi granite. K-feldspar megacrysts are commonly fractured, the slightly offset segments separated by margins of granular quartz. Quartz and feldspar are strained with strongly undulose extinction. Matrix biotite may define a weak foliation, but a simple statistical measure of the shape and position of phenocrysts using a method by Ramsay (1967, p. 195) demonstrates the random, massive texture of the granite.

Rapakivi mantled feldspars can develop from various igneous crystallization paths (Tuttle & Brown 1958, Steward & Roseboom 1962), and their exclusively igneous origin is now generally recognized. Many exolved minerals also indicate that the rock first formed at high igneous(?) temperature and partially re-equilibrated at a lower metamorphic temperature. Perthitic K-feldspar demonstrates albite solid solution, and deep colors of K-feldspar indicate a high Fe-content. Zoisite inclusions in plagioclase (Fig. 2) indicate original anorthite solid solution, and blue quartz presumably is due to rutile. Altered relict clinopyroxene also suggests an original temperature higher than that of the regional metamorphism.

Most parts of the rapakivi granite were later deformed into rapakivi augen gneiss with a variety of textures. Some phenocrysts are reduced to thin augen lenses, where the pink K-feldspar cores and greenish plagioclase mantles produce dramatic ribbons (Fig. 3). Cataclastic processes dominate the feldspar deformation, and in thin-section the ribbons are mosaics of equigranular K-feldspar and plagioclase (Fig. 4). Single feldspar crystals with deformed shapes were never found in thin-sections of elongate, flat or folded augen lenses.

Even where no mantled feldspar are found (Fig. 5), textural features of the augen gneiss indicate that augen are relict porphyro*clasts* and not newly formed porphyro*blasts*. The large K-feldspar augen commonly contain inclusions of plagioclase, quartz, biotite and hornblende. The inclusions are always randomly oriented, not foliated as they would be if included during porphyroblastic growth of metasomatic augen in a foliated gneiss. Inclusions are coarse grained as in the undeformed granite, whereas the minerals in the foliated matrix are cataclastically reduced and recrystallised to finer grained foliated textures: The plagioclase inclusions retain albite twins (Fig. 6) and zoisite inclusions, while matrix plagioclase is recrystallized and free of twins or zoisite.

Fig. 3. Strongly foliated rapakivi augen gneiss (304240). Plagioclase mantles are visible despite granulation and deformation of the augen. Scale in cm.

Fig. 4. Augen gneiss of Fig. 3. Note the zoisite needles in lens of fine-grained plagioclase (P). No zoisite in lens of fine grained K-feldspar (K). Bar scale = 0.1 mm. Plane-polarised light.

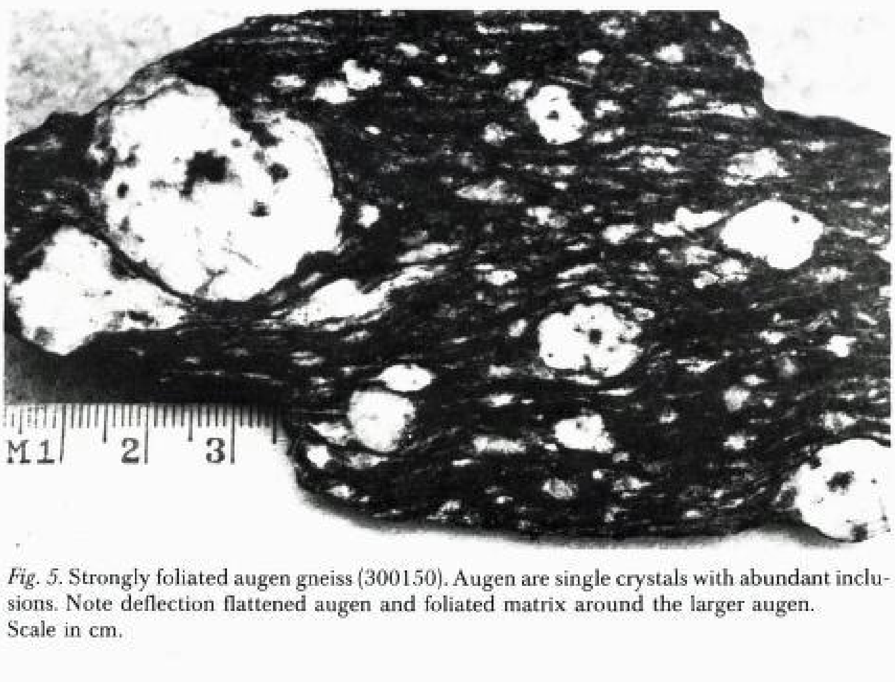

Fig. 5. Strongly foliated augen gneiss (300150). Augen are single crystals with abundant inclusions. Note deflection flattened augen and foliated matrix around the larger augen. Scale in cm.

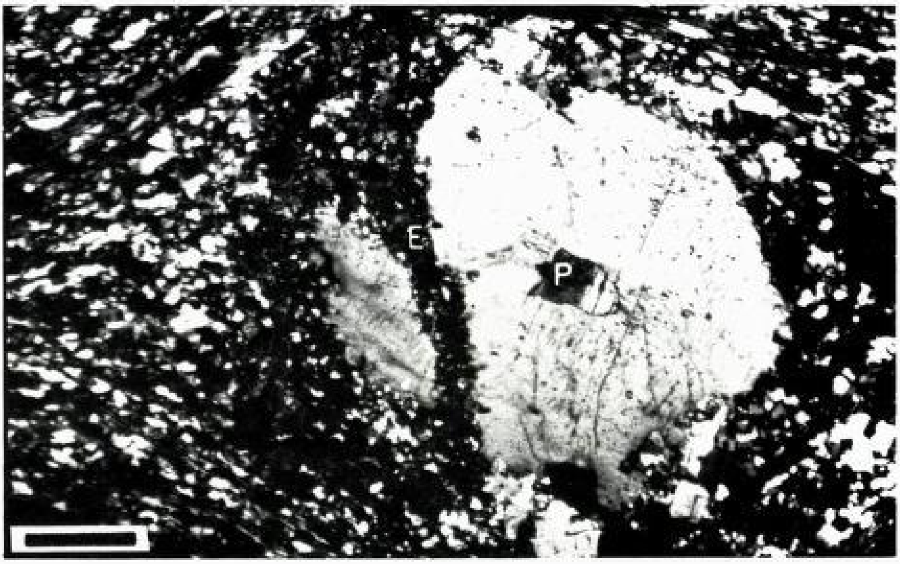

Fig. 6. Small K-feldspar porphyroclast in augen gneiss. Note deflection of matrix around augen. Plagioclase inclusion (P) contains secondary zoisite and shows albite twins. Granular epidote (E) separates K feldspar segments. Bar scale = 1 mm. Crossed polars.

Streaks of flattened augen and matrix minerals bend around the larger, more competent augen. If these augen were porphyroblasts, they would have had to have pushed the matrix minerals aside. Even such dense minerals as garnet apparently do not have such a 'force of crystallization' (Spry 1969) and it seems unlikely that a growing K-feldspar augen could push aside streaks of K-feldspar and denser matrix minerals. Rather the deflection of the matrix is a typical cataclastic texture, as illustrated by Higgins (1971). The rocks could be classified as mylonite or mylonite gneiss, in which the augen are porphyroclasts of normal appearance but unusually large size.

Textures diagnostic of feldspar porphyroblastesis, such as foliation that has been truncated by augen, or overgrown and preserved as helecitic traces in augen, have not been recognized in the Oppdal augen gneisses. The augen cannot be sedimentary clasts, as they are very evenly and widely distributed. No quartz augen (cobbles) are found, and the average quartz content of the augen gneiss (c. 20%) is too low for it to be a normal clastic metasedimentary rock. The feldspar:quartz ratio of the augen is too high for the augen to be an anatectic melt phase, but the normative K-feldspar:plagioclase:quartz ratio of the entire rock is that of eutectic granite (Eggen 1977).

No wiborgitic plagioclase mantles are recognized in much of the Oppdal augen gneiss. The original mantles may have been destroyed by deformation, but it is perhaps more likely that the original igneous rock itself lacked mantles. Such porphyritic rock without wiborgitic mantles is common in rapakivi massifs in Finland, where it is termed pyterlite.

Small pods of light-colored granitic gneiss locally float in the darker augen gneiss. Granulation at the edges of the pods has produced medium-grained grnaitic augen, which are quite different from the large-crystal K-feldspar augen of the typical augen gneiss. Such granite-clast augen gneiss composes only a small part of the Risberget augen gneisses, although inclusion-filled single-crystal augen (Fig. 5) can look like granite clasts.

Granular fine-grained gneiss containing K-feldspar, plagioclase, quartz, epidote, biotite, muscovite and garnet is commonly associated with the Oppdal augen gneiss. It is typically well foliated or flaggy, but a consistently low quartz content (<30%) suggests that it is orthogneiss and not feldspathic psammite. It commonly contains many small, millimeter-sized, feldspar augen and a few large single-crystal augen per outcrop. As in the typical augen gneiss, textural and mineralogical features of the augen indicate that they are porphyroclasts. However, the field appearance of a few large augen in a fine-grained flaggy gneiss might deceptively suggest a metasomatic origin.

Anorthosite and anorthosite-gabbro are closely associated with the Oppdal rapakivi granite, and these rocks fit an interesting global pattern. Post-tectonic of anorogenic rapakivi granite and anorthosite of Mid-Proterozoic age are scattered in an arc across the northern hemisphere from the Ural Mountains to western North America. (Bridgrewater & Windley 1973).

Rb–Sr dating

The first Rb-Sr study of Oppdal augen gneiss and other rocks of the Risberget Nappe was undertaken by Solheim (1980). He presented whole-rock results from seven suites of Risberget rock, including granite, gneiss, augen gneiss and metagabbro. The data points are somewhat scattered on the Rb-Sr plots, and produced errorchrons rather than true isochrons. Some points were arbitrarily deleted to reduce the scatter, and the age results are difficult to interpret in these cases. As presented, the errorchron dates range from 1129 ± 302 Ma to 1737 ± 106 Ma. The initial Sr ratios have large uncertainties and do not faciliate interpretation of the errorchrons. Mineral separates from a single sample of strongly foliated gneiss yielded an isochron date of 392 ± 21 Ma. The whole-rock errorchrons were interpreted as the result of a resetting event following the main metamorphism. The mineral isochron date was interpreted as the age of low-grade heating, or uplift and cooling.

My Rb-Sr study focused on the rapakivi granite and gneisses derived from it. Rocks of three suites were analyzed: massive rapakivi granite, strongly foliated mylonite gneiss, and intensely deformed ultramylonite. Detailed descriptions of the rocks, locations, and analytical methods are presented elsewhere (Krill 1980a) and briefly summarized in the appendix. Analytical data are listed in Tabel 1. The 56 whole-rock analyses define an errorchron corresponding to a date of 1618 ± 44 Ma ($^{87}Sr/^{86}Sr_0$ = .70211 ± 48, MSWD = 43, Fig. 7a. The large MSWD value quantifies the large amount of scatter. If the rocks are all co-magmatic, the scatter must have resulted from geological disturbance or contamination. The initial Sr ratio is low, within the predicted range of mantle-derived intrusive rocks. An isochron with such low initial Sr ratio is usually interpreted as the original igneous age of the rocks (Jäger 1979). However, the present example is an errorchron (Brooks et al. 1972), and must be interpreted with caution. I tentatively suggest that the date of 1618 Ma reflects the age of the rapakivi magmatism. This interpretation is supported by dating of similar rapakivi rock at Flatraket, 150 km to the west, where a zircon date of 1520 ± 10 Ma was interpreted as the intrusive age (Lappin et al. 1979). It also compares well with the dating of porphyritic rocks from the Tännäs Nappe in Jämtland, where a Rb-Sr date of 1610 ± 85 Ma and a zircon date of 1685 ± 20 Ma were interpreted to reflect the intrusion age (Claesson 1980).

RAPAKIVI GRANITE

The samples of unfoliated rapakivi granite which form part of the 1618 Ma errorchron were collected south of Oppdal from a relatively wide part of the Risberget Nappe. Glacially smooted hillside exposures prevent routine sampling, so samples were obtained from large boulders in a rockslide. The distinctive texture and the small sample area suggests that all samples were locally derived and co-magmatic.

Twelve samples averaging about 4 kg each from nine separate boulders yield a highly scattered errorchron: 1237 ± 228 Ma, $^{87}Sr/^{86}Sr_O$ = .7068, MSWD = 16. Eleven smaller samples (c. 0.5 kg) from a single rapakivi boulder give similar isotopic scatter and regression results: 1376 ± 584 Ma, $^{87}Sr/^{86}Sr_O$ = .7060 ± 52, MSWD = 19. There seems to be no significant isotopic difference between these small, closely spaced samples and the larger, widely spaced ones. In Fig. 7b both are plotted together, yielding another errorchron: 1182 ± 115 Ma, $^{87}Sr/^{86}Sr_O$ = .70756 ± 128 MSWD = 23, N = 23.

The 1182 Ma date is considerably younger than the presumed igneous age, and the initial Sr ratio is much higher. Scatter of the points off the best-fit line is large, and these dates cannot be considered meaningful. The isotopic systems have apparently been disturbed or contaminated on the whole-rock scale.

RAPAKIVI MINERALS

Minerals from one of the samples were also analysed. They are also disturbed and do not form an isochron, but mineral-whole rock pairs define the following dates and initial Sr ratios (Fig. 7c):

W.R.-Ksp:	1734 ± 144 Ma,	.69877 ± 206
Ksp-plag:	1134 ± 33 Ma,	.70920 ± 22
W.R.-plag:	740 ± 36 Ma,	.71074 ± 22
W.R.-biot:	368 ± 10 Ma,	.71518 ± 28

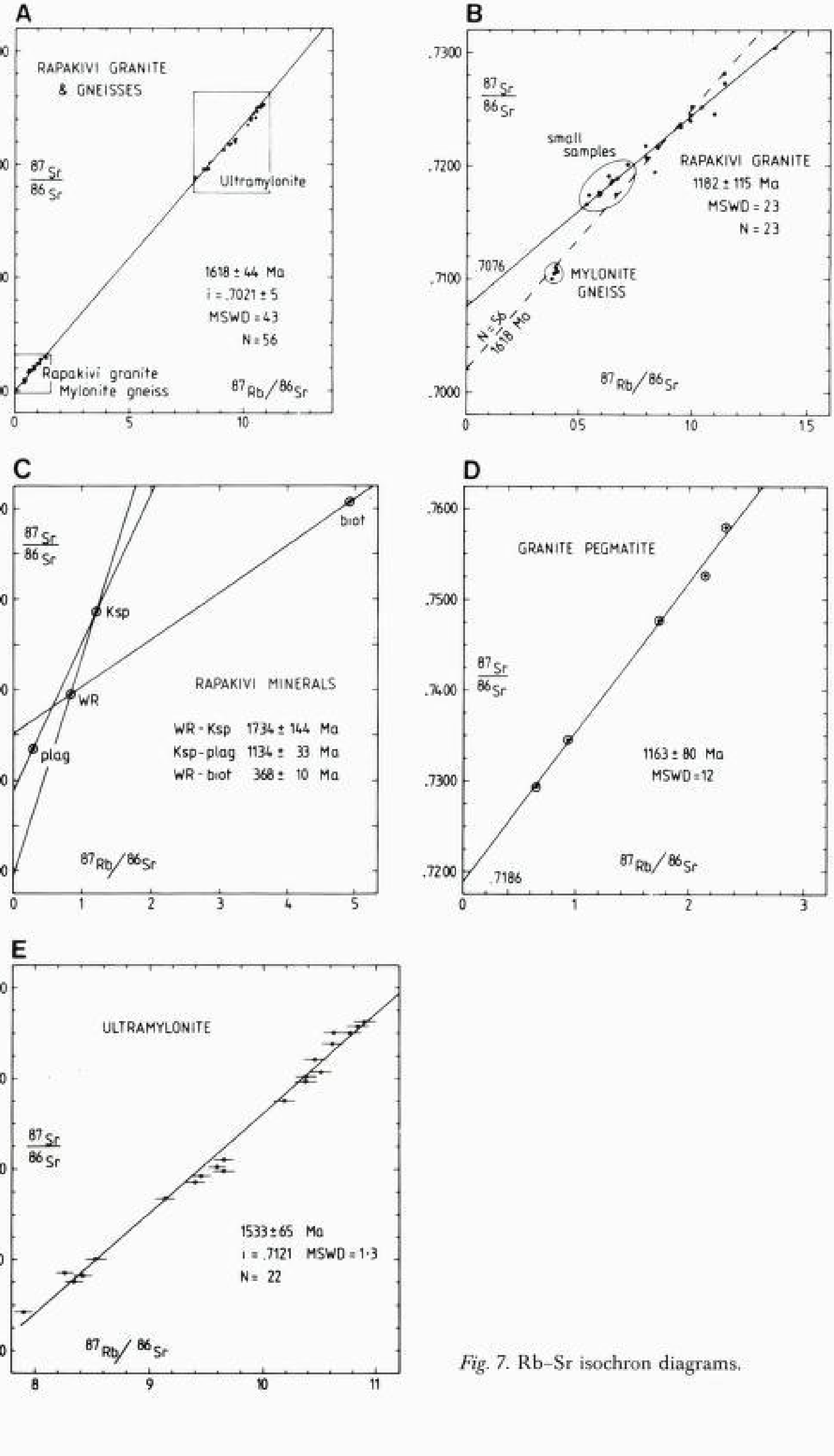

Fig. 7. Rb–Sr isochron diagrams.

Without performing similar analyses on other samples to confirm these dates, only tentative interpretations are possible. The 1734 ± 144 date apparently reflects the date of the rapakivi-granite intrusion. The initial Sr ratio is clearly too low, and therefore the date is probably too old, but both are presumably within the range of calculated errors. The 1134 Ma date may well be meaningless – a partial resetting to a younger age. However, it may record a thermal event, which re-equilibrated the Sr isotopes of the plagioclase mantles and the K-feldspar cores. The 740 Ma date is probably meaningless, since the plagioclase isotopes would more readily reequilibrate with the adjacent K-feldspar than with the whole-rock isotopic value. The 368 ± 10 Ma date probably represents the time of final cooling of the rocks. Although the rock produces no isochron, such disturbed mineral dates must be expected, as thin-sections clearly show secondary metamorphic mineral effects without complete recrystallization.

GRANITIC PEGMATITE

A granite pegmatite within coarse-grained augen gneiss yielded a 5-point whole-rock errorchron of 1163 ± 80 Ma (^{87}/Sr/$^{86}Sr_o$ = .71864 ± 134, MSWD = 12, Fig. 7d). The pegmatite is weakly foliated together with the augen gneiss and is sheared along its contacts. The initial Sr ratio is relatively high, as is typical for pegmatites derived from crustal rocks. The 1163 Ma date may be the intrusive age of the pegmatite, and would support an interpretation that similar dates from the rapakivi granite and K-feldspar–plagioclase pair are meaningful. It is not clear from the field relationships whether the pegmatite actually cuts any earlier foliation in the augen gneiss. The pegmatite is apparently not comagmatic with the repakivi rocks, as it shows no rapakivi texture or characteristic rapakivi minerals. The five samples are not included in the 56-point regression of the 1618 Ma errorchron.

MYLONITE GNEISS

Eleven of the analyses for the 1618 Ma date are from strongly foliated rapakivi mylonite gneiss within about 10 meters of the contact to the Sætra Nappe. No date was possible because of the lack of Rb/Sr spread, but $^{87}Sr/^{86}Sr$ analyses were carried out on all eleven samples to determine the amount of scatter. Scatter of the data is indeed very small; all eleven rocks have very similar ratios. Any secondary processes that may have influenced the isotopic ratios must have been very uniform. The tight cluster of points suggests that any very late non-penetrative disturbance was insignificant.

Because of the medium-grained texture and regular foliation of the mylonite gneiss, cross-cutting fractures, retrogression zones and weathered surfaces are easily identified and avoided in sampling. This rock may be inherently safer to use in geochronologic studies than the very coarse-grained rapakivi granite, in which secondary effects are extremely difficult to identify.

ULTRAMYLONITE

Ultramylonite from a several meter thick zone within the Risberget Nappe provides the remaining analyses for the 1618 Ma errorchron. The rock has a fine-grained granular texture caused by tectonic granulation and recrystallization of the minerals (cf. Higgins 1972, Sibson 1977). The parent rock was apparently a K-rich variety of rapakivi granite, almost completely lacking mafic minerals. Some tiny K-feldspar augen remain, and within the same outcrop some plagioclase-mantled K-feldspar porphyroclasts are preserved in coarser augen gneiss of similar potassium-rich composition. The ultramylonite has a strong foliation and lineation parallel to Caledonian fabrics in the Sætra Nappe, exposed as a tectonic slice about 30 m to the south, and I tentatively interpret it to be a Caledonian ultramylonite, not a relic of Precambrian deformation.

Samples were taken from a single ultramylonite block about 0.5 m^3 in volume. No significant isotopic differences were noted between samples of varying size and relative position within the block. Analyses fall on the same line and are regressed together, yielding an isochron: 1533 ± 65 Ma, $(^{87}Sr/^{87}Sr)_o$ = .7121 ± 1.3, N = 21 (Fig. 7e). Sample 122 was deleted from the regression. It falls below the other points and would completely change the regression line, creating an impossibly low initial Sr ratio. This sample has a distinctive biotite layer which apparently is not in isotopic equilibrium with the other rocks of this suite.

It is interesting that the ultramylonite is the most intensely deformed and recrystallized of the Risberget rocks studied, and yet it yields a true isochron that is within the error of the suggested 1618 Ma age of the igneous source rock. The least deformed rapakivi-granite boulders, on the other hand, yield scattered errorchrons with younger trends.

Discussion of the Rb–Sr dating

This study and the Rb–Sr study of Solheim (1980) provide useful information for interpretation of the Oppdal gneisses. The rocks clearly have a Precambrian origin. None of the whole-rock errorchron dates are as young as the Caledonian event and only a few are within the range of a possible Sveconorwegian event. The very low initial Sr ratio (.7021) of the rapakivi errorchron suggests that the rocks are of igneous origin, and not derived from older acidic crust. Derivation of the augen gneiss by regional metasomatism or granitization is very unlikely. Any K-rich metasomatic fluids would be expected to have relatively long crustal histories with high $^{87}Sr/^{86}Sr$ ratios. If such fluids could produce an errorchron from widely-spaced rocks of variable composition, the resulting initial Sr ratio would predictably be much higher.

The cause of the scatter of the Rb-Sr data is unknown. Rapakivi granites are known for being especially vulnerable to weathering (Eskola 1930), which could easily disturb the Rb and Sr. Even undeformed post-tectonic rapakivi rocks commonly produce anomalously young, scattered mineral dates (Bridgewater & Windley 1973). Data of the ultramylonite and of the massive rapakivi granites each produce lines with shallower slopes and higher initial Sr ratios than the presumed original isochron. This pattern may indicate partial equilibration of Sr isotopes with respect to some average isotopic composition (e.g. Field & Råheim 1979). Complementary U–Pb study of zircons could perhaps demonstrate whether these younger dates are meaningful or only an incomplete resetting to some younger date.

The Precambrian errorchrons do not contradict the interpretation that the rocks were strongly deformed and metamorphosed at amphibolite-facies conditions during the Caledonian orogeny (Krill 1983). Orthogneisses commonly retain older whole-rock dates despite younger tectonic events (Krill & Griffin 1981), and although the Risberget rocks were not reset, Oppdal schists and gneisses that are considered Caledonian (Krill 1980) have yielded Caledonian errorchron dates and Caledonian model ages (Råheim 1977, Solheim 1977, Krill unpubl. data). The fact that foliated rocks give fully reset Caledonian mineral isochrons (Solheim 1980) whereas minerals from massive rocks are only partially reset (Fig. 7c), supports the geological conclusion that the foliation is Caledonian.

Summary

The new Rb-Sr date presented here supplement the initial Rb-Sr study of Solheim (1980) and allow several tentative interpretations to be made. The 56-point errorchron date of 1618 ± 44 Ma is considered to represent the age of rapakivi-granite intrusion. Similar augen gneisses without obvious rapakivi textures may have also intruded at about this time, or they may be younger. Solheim (1980) obtained several errorchron dates of about 1450 Ma with very large uncertainties, and these dates may also represent intrusive ages. The c. 1150 Ma dates from the pegmatite, rapakivi granite and K-feldspar–plagioclase minerals may reflect a Sveconorwegian thermal event with pegmatite intrusion. The c. 395 Ma mineral isochrons from foliated basement and Risberget augen gneiss (Solheim 1980) presumably date the strong foliation, and the 358 Ma date of unoriented biotite in the rapakivi granite (this study) apparently reflects the cooling age.

It is notable that not a single Caledonian whole-rock date has been obtained from the Risberget Nappe, despite the very large number of analyses and the clear geologic and geochronologic evidence that the rocks are Caledonized gneisses. It appears that for the definitive interpretation of these and other problematic Rb-Sr dates from the Western Gneiss Region, additional geochronological data using other decay systems (U-Pb or Sm-Nd) will be necessary.

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REFERENCES

Barth, T. F. W. 1939: Progressive metamorphism of sparagmite rocks of Southern Norway. *Norges geol. Tidsskr. 18*, 54–65.

Bjørlykke, K. O. 1905: Det centrale Norges fjeldbygning. *Norges geol. Unders. 39*, 1–595.

Bridgewater, D. & Windley, B. F. 1973: Anorthesite, post-orogenic granites, acid volcanic rocks and crustal development in the North Atlantic shield during the Mid-Proterozoic. *Spec. Pub. Geol. Soc. S. Africa 3*, 307–317.

Brooks, C., Hart, S. R. & Wendt, I. 1972: Realistic use of two-error regression treatments as applied to rubidium-strontium data. *Rev. Geophys. Space Phys. 10*, 551–557

Carstens, C. W. 1924: Rapakiwigesteine an der westlichen Grense des Trondhjemsgebietes. *Norges geol. Tidsskr. 8*, 81–92.

Carstens, C. W. 1925: Ein aus Rapakiwiegesteinen umgewandelten Augengneis. *Norges geol. Tidsskr. 8*, 235–249.

Claesson, S. 1980: A Rb-Sr isotope study of granitoids and related mylonites in the Tännäs Augen Gneiss Nappe, southern Swedish Caledonides. *Geol. Fören. Stockholm. För. 102*, 403–420.

Eggen, S. S. 1977: *Geologiske studier i Oppdal-feltet, med hovedvekt på petrologi i basal komplex og øyegneiser.* Unpubl. dissertation, Univ. Oslo, 227 p.

Eskola, P. 1930: On the disintegration of rapakivi. *Bull. Comm. Geol. Finl. 92*, 96–105.

Field, D. & Råheim, A. 1979: Rb-Sr total rock isotope studies on Precambrian charnockitic gneisses from south Norway: evidence for isochron resetting during a low-grade metamorphic-deformation event. *Earth Planet. Sci. Lett. 45*, 32–44.

Goldschmidt, V. M. 1916: Geologische-petrographische Studien in Hochgebirge des südlichen Norwegens, IV: Übersicht der Eruptivgesteine und Kaledonischengebirge zwischen Stavanger und Trondheim. *Vidensk. Selsk. Skr. 1, Math. Natur, - Kl. 12*, 140 p.

Higgins, M. W. 1971: Cataclastic rocks. *U.S. Geol. Surv. Prof. Pap. 687*, 97 p.

Holtedahl, O. 1938: Geological and petrological investigations in the Oppdal–Sunndal–Trollheimen district. *Norges geol. Tidsskr. 18*, 29–53.

Jäger, E. 1979: Introduction to geochronology, 1–12. *In* Jäger, E. & Hunziker, J. C. (eds.). *Lecture in isotope geology.* Springer Verlag, Berlin.

Krill, A. G. 1980a: Tectonics of the Oppdal area, central Norway. *Geol. Fören. Stockh. Förh. 102*, 523–530.

Krill, A. G. 1980b: *Tectonics of N.E. Dovrefjell, central Norway.* Unpubl. Dissertation Yale Univ. New Haven, 178 p.

Krill, A. G. 1983: Relationships between the Western Gneiss Region and the Trondheim Region: Stockwerk-tectonics reconsidered. *Proc. 1981 IGCP Symp.*, Uppsala, Wiley (in press).

Krill, A. G. & Griffin, W. L. 1981: Interpretation of Rb–Sr dates from the Western Gneiss Region: a cautionary note. *Norges geol. Tidsskr. 61*, 83–86.

Lappin, M., Pidgeon, R. T. & van Breeman, O. 1979: Geochronology of the basal gneisses and mangerite syenites of Stadlandet, west Norway. *Norges geol. Tidsskr. 59*, 161–181.

Pankhurst, R. J. & O'Nions, R. K. 1973: Determination of Rb/Sr and $^{87}Sr/^{86}Sr$ ratios of some standard rocks and evaluation of X-ray fluorescence spectrometry in Rb-Sr geochemistry. *Chem. Geol. 12*, 127.

Ramsay, J. G. 1967: *Folding and fracturing of rocks.* McGraw Hill Inc. New York, 568 p.

Råheim, A. 1977: A Rb, Sr study of the rocks in the Surnadal syncline. *Norges geol. Tidsskr. 57*, 193–204.

Rosenqvist, I. Th. 1943: Metamorphism and metasomatism in the Oppdal area. *Norges geol. Unders. 22*, 106–202.

Sibson, R. H. 1977: Fault rocks and fault mechanisms. *Journ. Geol. Soc. London 133*, 191–213.

Simonen, A. & Vorma, A. 1969: Amphibole and biotite from rapakivi. *Bull. Comm. Geol. Finl. 238*, 1–28.

Solheim, S. 1977: *Geologiske studier i Oppdal-feltet, med hovedvekt på geokronologi.* Unpubl. dissertation Univ. Oslo, 112 p.

Solheim, S. 1980: Geochronological investigation in the Oppdal area, Central Norway. *Norges geol. Tidsskr. 60*, 175–188.

Spry, A. 1969: *Metamorphic textures.* Permagon Press, Oxford, 350 p.

Stewart, D. B. & Roseboom, E. H. 1962: Lower temperature termination in the three-phase region, plagioclase-alkali feldspar-liquid. *Jour. Petrology 3*, 280–315.

Törnebohm, A. E. 1896: Grundragen af det centrala Skandinaviens bergbygnad. Kgl. Svenska Vetensk. Akad. Handl. 25 (5), 212 p.

Tuttle, O. F. & Brown, N. L. 1958: Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2--H_2O. *Geol. Soc. Amer. Mem. 74,* 151 p.

York, D. 1969: Least squares of a straight line with correlated errors. Earth Planet. Sci. Lett. 5, 320–324.

Appendix

ANALYTICAL TECHNIQUES

The samples were crushed in a steel-jaw crusher and finely ground in a steel-ring mill. Rb/Sr ratios were determined directly by x-ray fluorescence (Pankhurst & O'Nions 1973). Mass spectrometry was performed on a Micromass MS30 at the Mineralogisk-Geologisk Museum, Oslo. Variable mass discrimination in $^{87}Sr/^{86}Sr$ was corrected by normalizing $^{88}Sr/^{86}Sr$ to 8.3752. The ^{87}Rb decay constant used was 1.42×10^{-11} yr.$^{-1}$, and the data were regressed by the method of York (1969). In assigning errors, the coefficient of variation was taken as 1 % for $^{87}Rb/^{86}Sr$. The errors for the $^{87}Sr/^{86}Sr$ measurements are listed in Table 1. Age and intercept errors are quoted at the 2-sigma confidence level.

Table 1. RB–SR DATA

Sample No.	Wt. (kg)	Rb (ppm)	Sr (ppm)	87Rb/86Sr	87Sr/86Sr	1-sigma
Rapakivi granite (boulders)						
25	3.9	123	316	1.14	.72818	13
26	6.6	124	314	1.14	.72732	8
27	2.0	116	304	1.10	.72463	5
28	5.8	109	320	.991	.72404	10
29	2.8	108	314	.995	.72455	5
30	4.1	110	309	1.04	.72535	8
31	1.7	98	210	1.36	.73045	16
32	5.0	104	358	.847	.72170	8
33	5.9	98	359	.792	.72181	11
34	1.8	102	314	.941	.72352	9
35	1.9	98	357	.800	.72075	7
266	3.6	88	307	.834	.71955	10
Small sample (single boulder)						
105	0.5	80	396	.589	.71770	8
106	0.7	80	396	.589	.71745	7
107	0.5	77	377	.591	.71746	6
108	1.0	88	410	.627	.71905	8
109	0.5	83	374	.645	.71878	9
110	0.3	88	383	.670	.71897	5
111	0.2	75	409	.534	.71665	9
112	0.2	84	384	.636	.71843	7
113	0.4	100	407	.713	.72021	8
114	0.3	77	414	.543	.71743	10
115	0.5	85	371	.667	.71752	8
Mylonite gneiss						
14	4.3	56	400	.405	.71073	5
15	1.2	56	401	.406	.71074	5
16	1.3	56	411	.400	.71095	5
17	2.8	54	401	.391	.71148	9
18	1.2	55	399	.403	.71079	4
19	1.5	52	393	.387	.71062	7
20	0.7	55	394	.403	.71060	6
21	1.5	54	409	.387	.71087	4
22	2.0	54	408	.384	.71011	6
23	2.2	55	398	.400	.71104	10
24	1.6	54	403	.393	.71064	6
Ultramylonite						
1	0.5	273	85	9.40	.91731	10
2	0.5	270	84	9.45	.91810	4
3	0.5	278	85	9.65	.91968	8
4	0.5	279	84	9.65	.92224	14
5	0.4	282	87	9.59	.92061	8
6	1.6	264	93	8.34	.89523	16
7	0.8	252	90	8.26	.89736	7
8	3.1	267	93	8.42	.89670	8
9	1.0	265	85	9.14	.91386	9
10	1.7	283	80	10.38	.94063	14
11	1.4	265	91	8.53	.90048	7
12	3.1	284	82	10.19	.93512	4
116	0.4	291	79	10.84	.95192	8
117	0.3	292	79	10.89	.95272	5
118	0.2	281	77	10.77	.95007	10
119	0.3	284	79	10.61	.94789	12
120	0.6	288	80	10.63	.95036	9
121	0.5	265	99	7.90	.88858	8
122	0.6	239	146	4.79	.80402	5
123	0.6	285	80	10.51	.94150	9
124	0.5	280	79	10.46	.94437	5
125	0.5	285	81	10.38	.93950	10
Granitic pregmatite						
184	0.9	115	145	2.32	.75801	3
185	1.0	32	142	.658	.72939	6
186	0.8	127	173	2.14	.75268	5
187	1.0	96	161	1.73	.74775	10
188	1.2	53	166	.931	.73468	8
Minerals sample no. 266						
Ksp		184	444	1.20	.72869	8
plg		50	535	.271	.71361	5
bio		164	97	4.92	.74099	9

SAMPLE LOCATION, DESCRIPTIONS

Rapakivi granite

The sample locality is a boulder rockslide about 1 km south of Drivstua (UTM coord. 32502080). All samples are weakly foliated to unfoliated, but are rusty weathered on outside surfaces and internal fractures. See text and Carstens (1924) for petrographic descriptions.

Mylonite gneiss

About 0.5 m³ in volume, from a newly blasted quarry road-cut south of the southermost Engan flagstone quarry (UTM coord. 29952425). All samples are from a single loosened slab. The rock is very dark, with strong gneissosity defined by lenses of granulated, rotated and recrystallized feldspar from original augen. The lenses are a few mm thick and several cm in length and width. Biotite forms a weak mineral lineation. The rock is c. 20% Ksp, 25% plag, 20% qtz, 15% biot, 10% musc and 10% epid, with accessory sphene and opaque. No chlorite is seen.

Ultramylonite

Samples were collected about 2 km north of Drivstua from a newly blasted road-cut on highway E-6 (UTM coord. 20582400). The samples are from a single block 0.5 m³ in volume in a fine-grained layer. The rock is tan-colored and fractures conchoically. It appears homogenous and is not layered, but has very strong foliation (NI5E, 35SE) and lineation (S75E, 35). Rare biotite layers a few cm wide are parallel to the foliation. Scattered K-feldspar augen are up to a few mm in diameter. The rock is c. 45% Ksp, 20% Ksp, 20% plag, 25% qtz, with accessory epid, gar, biot and musc.

Granitic pegmatite

The dike is exposed along highway 16, 9 km southwest of Lønset (UTM coord. 090385). It contains K-feldspar, plagioclase, quartz, biotite and muscovite. It averages about 1 meter in width, and is foliated together with the augen gneiss.

Nappe and Thrust Structures in the Sparagmite Region, Southern Norway

JOHAN PETTER NYSTUEN

Nystuen, J. P. 1983: Nappe and thrust structures in the Sparagmite Region, southern Norway. *Norges geol. Unders. 380*, 67–83.

The Osen–Røa Nappe Complex is thrust upon a crystalline basement and a thin autochthonous sediment cover. The nappe complex is characterized by a sole thrust cutting up-section toward the SSE, by flats, ramps and contraction faults separating subordinate thrust sheets. Duplex geometry is present in the western part of the leading edge and within some minor thrust sheets. An arcuate fold system developed during the emplacement of the nappe. The Kvitvola Nappe, resting on the Osen–Røa Nappe Complex with a regional disconformity, may have been emplaced on an erosional surface and carried further by piggy-back transport. Another possibility is that the disconformity is the result of out-of-sequence thrusting when the Kvitvola thrust became reactivated. Post-thrust deformation gave rise to the Snødøla–Steinfjellet basement antiform and other undulations in the basement and broad synforms and antiforms in the nappe cover. The driving force of the thrusting is briefly discussed.

J. P. Nystuen, Institutt for geologi, Norges landbrukshøgskole, N-1432 ÅS-NLH, Norway

Introduction

The Caledonian front in Scandinavia includes a series of nappes with varying mutual contacts, lithologies, thrust and fold structures and histories of structural evolution. The *Osen–Røa Nappe Complex* and the *Kvitvola Nappe* are two such major tectonostratigraphical units in the Sparagmite Region in southern Norway (Fig. 1). They belong to the Lower and Middle Allochthon, respectively, of the nappe pile in the Scandinavian Caledonides (Roberts & Gee 1981), and both consist of uppermost Proterozoic to Lower Palaeozoic sedimentary strata and sheets of mid-Proterozoic crystalline basement rocks (Fig 2). The Osen–Røa Nappe Complex has been derived from a fault-bounded cratonic basin and the Kvitvola Nappe from a shelf basin on the western part of the Baltoscandian craton. A displacement towards the SE of 200–400 km has been estimated for the Osen–Røa Nappe Complex and more than 400 km for the Kvitvola Nappe (Nystuen 1981, 1982). The sedimentary evolution of the sequences has been described and discussed by K. Bjørlykke et al. (1976) and Nystuen (1980, 1981, 1982). An outline of the Caledonian tectonostratigraphy and the general geology of the area is given by Bockelie & Nystuen (in press). The present paper deals with the thrust and fold geometry of the nappes, their contact relationships and their structural evolution. Terminology on nappes and thrusts are according to Dahlstrom (1970), Elliot & Johnson (1980), McClay (1981), Boyer & Elliott (1982) and Butler (1982).

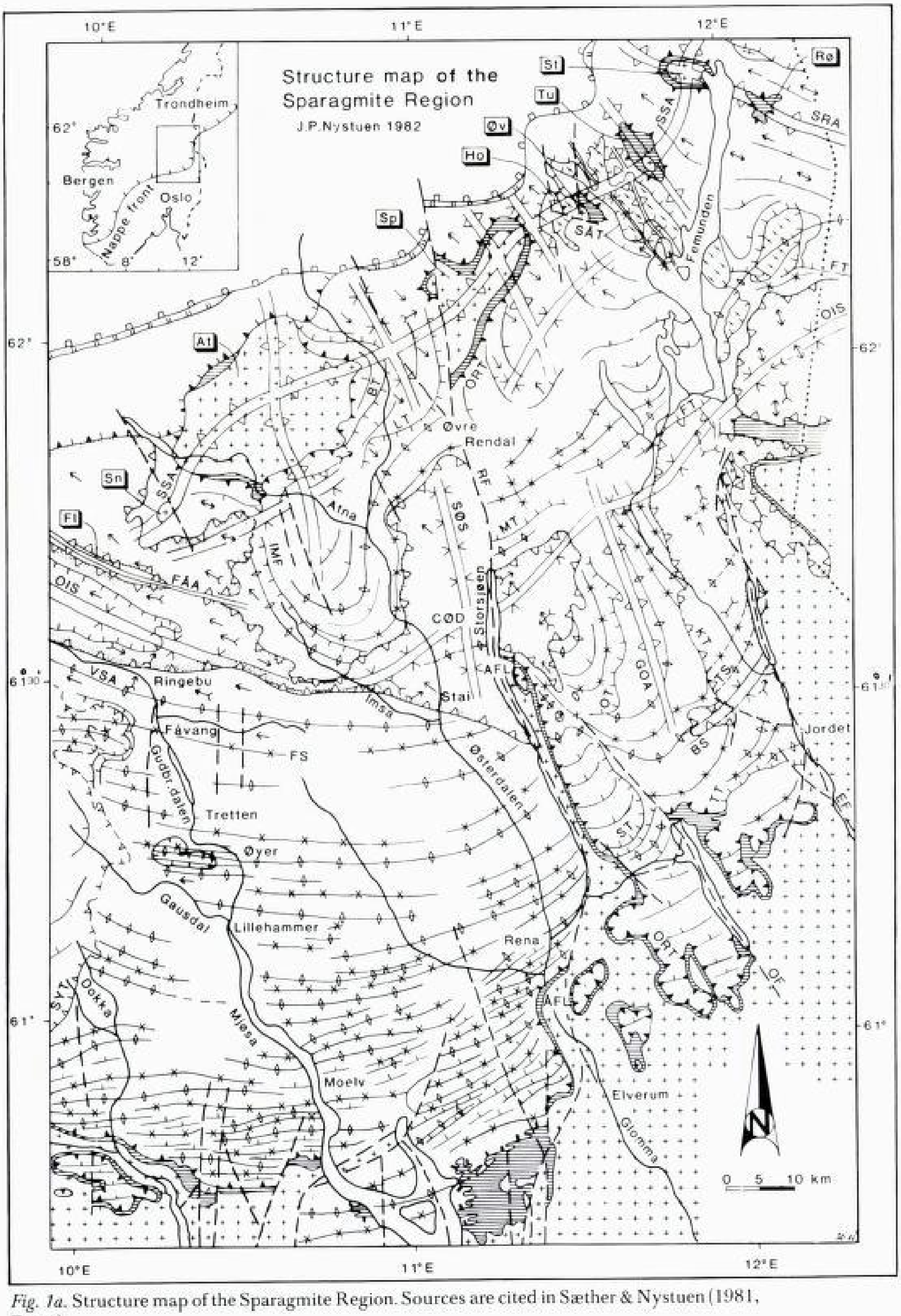

Fig. 1a. Structure map of the Sparagmite Region. Sources are cited in Sæther & Nystuen (1981, Fig. 1).

Osen–Røa Nappe structures

OSEN–RØA THRUST

Adopting the practice of Elliot & Johnson (1980, p. 70), a thrust fault or the surface on which a nappe or thrust sheet rests, is named after the thrust unit. The *sole thrust* of the Osen–Røa Nappe Complex, the *Osen–Røa thrust* (ORT), is exposed along the fringed erosional nappe front in the south and along the rims of several windows in the north (Fig. 1). It is a blind thrust (McClay 1981, p. 9) in most parts of the area; its presence is indicated by folds and subsidiary thrusts in the décollement-moved strata above. The Osen–Røa thrust is conformable with the smooth basement surface, and with the footwall (Butler 1982, p. 239) dominated by Cambrian shales and phyllites. The thrust nappe also rests directly on the basement or on the Vendian tillite or quartzites. The oldest rocks in the hangingwall (Butler 1982, p. 239) are basement sheets in the north at Femunden. The thick Brøttum and Rendalen Formations (Fig. 2) can be traced from overthrust positions at the windows in the north towards the southern part of the Sparagmite Region; thus, the Rendalen Formation is exposed in the hangingwall east of lake Storsjøen (Fig. 1) (see map, Fig. 2 in Nystuen 1982). In the southern 20–50 km of the thrust complex the Osen–Røa thrust cuts up stratigraphically through the Hedmark Group (fig. 2), and

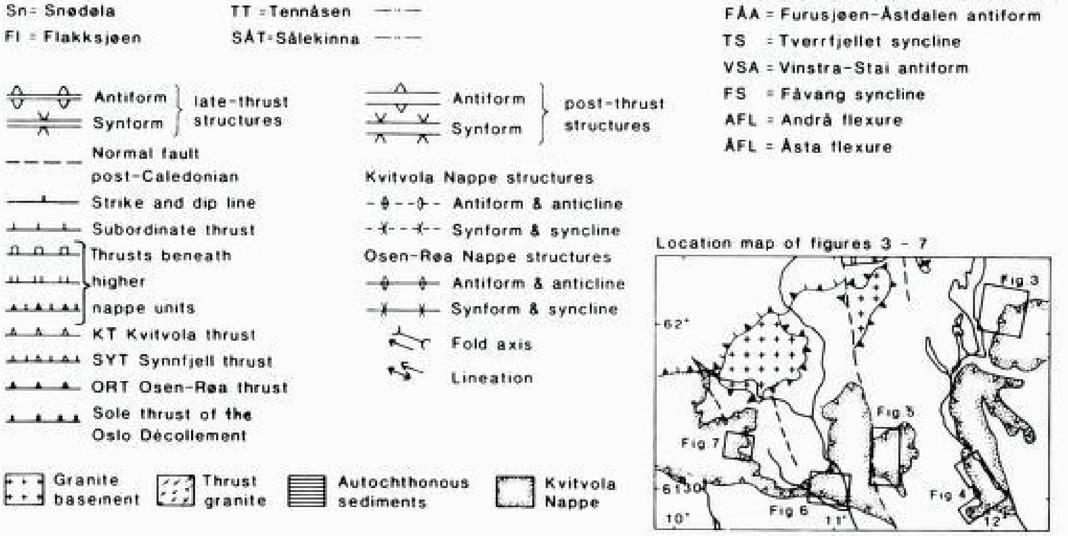

Fig. 1b. Legend to structure map of the Sparagmite Region (Fig. 1a) and location map of Figures 3–7.

in the Mjøsa area the ORT continues into the sole thrust of the décollement-folded Cambro-Silurian sequence of the Oslo Region (A. Bjørlykke et al. 1980). The stratigraphic separation decreases further southwards in the Oslo Region through a series of splay thrusts (Bockelie & Nystuen, in press).

Along mean strike of the Osen–Røa thrust (c. 70°) it cuts up and down stratigraphic section (Dahlstrom 1970) through the Hedmark Group. This is demonstrated in a WSW–ENE traverse from Lillehammer to Jordet (Fig. 1) (see map, Fig. 2 in Nystuen 1982). The hangingwall consists of the Brøttum Formation (or older unknown units) west of the Rendalen fault (RF) and of the Biri Formation and younger formations east of the RF. The post-Caledonian (Permian) Rendalen normal fault is here thought to intersect by acute angle a steep lateral ramp (Butler 1982, p. 240) which trends c. 130°, and which is exposed as the *Imsdalen fault* (IMF) (fig. 1). The IMF dips steeply to the SW and separates turbidites of the Brøttum Formation in the SW from the fluvial Rendalen Formation and younger units in the NE (Sæther & Nystuen 1981, Nystuen 1982).

The hangingwall rocks are deformed by cataclasis along shear planes in the south and by penetrative slaty cleavage, phyllonitization and locally thin mylonitic banding in the north. Locally increased friction by rough gliding has given rise to slicing of the footwall and adherence of small imbricate sheets of basement and sedimentary rocks in the frontal area (Elvsborg & Nystuen, 1978, Sæther 1979).

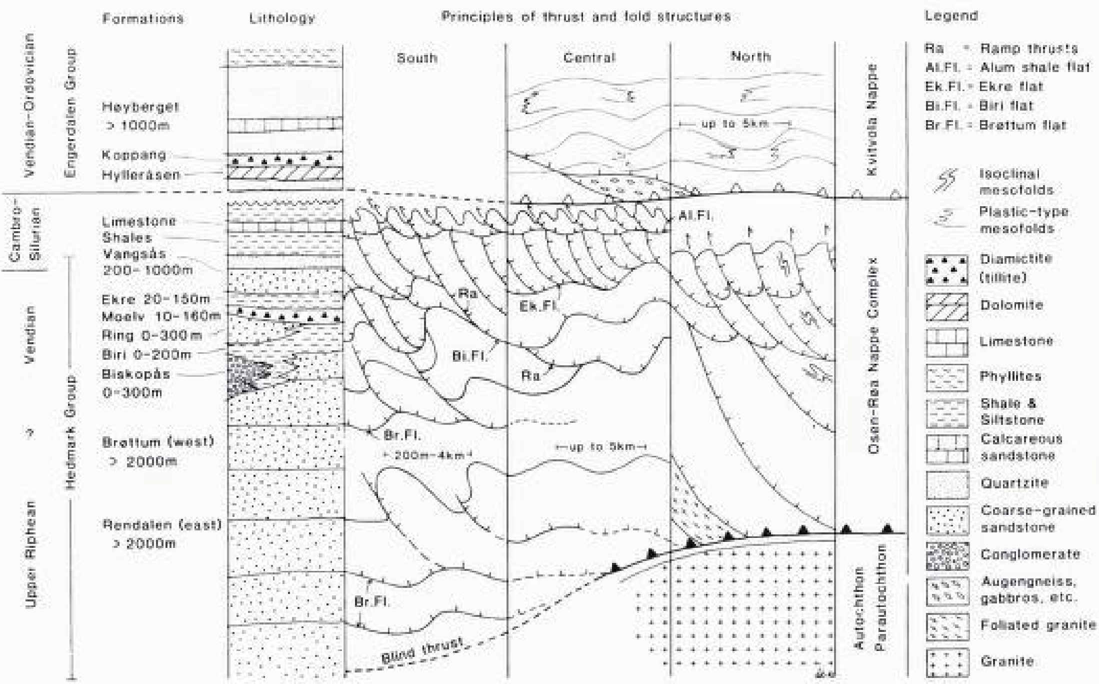

Fig. 2. Simplified stratigraphical sections through the Osen–Røa Nappe Complex (K. Bjørlykke et al. 1976, Nystuen 1982) and the Kvitvola Nappe (Nystuen 1980) and with main types of thrusts and folds in the nappe units.

The basement surface and the Osen–Røa thrust dip 1–2° towards the north at the nappe front but steepen to c. 10° at the Andrå flexure (AFL); a similar flexure probably exists at Åsta (ÅFL) (Fig. 1). These surfaces again rise in the dome-shaped window structure further north (see later discussion on basement geometry). Slip along the ORT in the direction of movement (SSE), measured from the eroded front at Elverum, is 70 km to the Andrå flexure and 125 km to the northern margin of the Spekedalen window.

INTERNAL THRUSTS AND IMBRICATE STRUCTURES

The overall structure of the Osen–Røa Nappe Complex consists roughly of an imbricate frontal zone, a central part dominated by subhorizontal flats (Butler 1982, p. 239) and open, large-scale folds, and a trailing edge, in the north, which displays low- and high-angle contraction faults (McClay 1981, p. 8). Incompetent shale beds have acted as flats at several stratigraphical levels (Fig. 2). In the deeper sections, relative displacements have taken place by gliding along shales in the Brøttum (Englund 1972) and Rendalen (Oftedahl 1943) Formations. There is probably a system of flats and NW-dipping ramps, joining the sole thrust (ORT), thus forming a branching staircase trajectory which is typical of many thrust belts (Dahlstrom 1970, Elliot & Johnson 1980, Butler 1982).

A flat within Biri shales forms the floor thrust (Dahlstrom 1970, p. 357, Butler 1982) of an imbricate stack (Butler 1982, p. 241) in stratigraphically overlying units in the Tretten–Øyer area. The imbrication has been produced by reverse listric faults cutting folds overturned to the south (Englund 1972).

The *Langtjørna thrust* (LT) in the Øvre Rendal area cuts up through the Rendalen Formation and enters the Biri dolomites as a flat, separating a middle and upper thrust sheet. The middle thrust sheet is separated from a lower thrust sheet by the *Brennåsen thrust* (BT) which occurs as a flat in the Moelv Tillite and the Ekre Shale (Nystuen & Ilebekk 1981). The collective slip along these thrusts is at least 25 km. Flats occur in the Ekre Shale beneath imbricate Vangsås Formation in the Tretten–Øyer area (Englund 1972) and in the Fåvang area (Englund 1973).

The *Slemdalen thrust* (ST) and the *Osdalen thrust* (OT) (Fig. 1) are low-angle thrusts following the Ekre Shale as flats for kilometres. The Slemdalen thrust cuts up stratigraphically from the Osen–Røa thrust and appears as a contraction fault with a minimum slip of 14 km. The origin of the Osdalen thrust is uncertain as the hangingwall rocks here are younger than those in the footwall. However, both thrusts form the floor thrusts of imbricate stacks in the Vangsås Formation (Nystuen 1975a, 1975b).

The *Tennåsen thrust* (TT) is a major listric contraction fault along which the Biri Formation is emplaced above the Vangsås Formation (Nystuen 1975c). The TT joins the Osen–Røa thrust in dip-direction, and the minimum slip is about 3 km.

Other mapped major flats follow Cambro-Ordovician phyllites west of Fåvang and separate here three thin sheets with flat-lying Vangsås Formation; the stacking includes a shortening of at least 45 km (Englund 1973, pers. comm. 1982). At a still higher level a flat within Ordovician phyllites probably forms the floor thrust of the Synnfjell Nappe (Duplex) west of Gudbrandsdalen (Hossack et al. in press).

The *Synnfjell thrust* (SYT) acts as the roof thrust (Butler 1982, p. 241) of the frontal imbricate structure and the Osen–Røa thrust as the floor thrust. The imbricate stack, named the *Aurdal duplex* in the Dokka–Valdres area by Hossack et al. (in press), comprises in the Mjøsa area the Vangsås Formation and overlying Palaeozoic strata in the south, whereas progressively older units are included in the imbricate structure towards the north (A. Bjørlykke 1979). North of Moelv the imbrications are replaced by less compressed folds which have their southern limbs overturned to the south (Englund 1978). The Mjøsa imbricate structure, in which imbricate slices are bounded by reverse listric faults dipping north, continues northeastwards to the Rena area (Høy & A. Bjørlykke 1980, K. Bjørlykke 1976).

Imbrications, floored by a blind thrust,occur within the Elstad Formation at Fåvang. This formation,lying structurally beneath the Brøttum Formation, was correlated with the Vangsås Formation by Englund (1973). Adopting this lithostratigraphical correlation, Hossack et al. (in press) interpreted the Elstad structure as a window beneath a thrust with the Brøttum Formation as hangingwall. However, according to J.-O. Englund (pers. comm. 1983), the stratigraphical and structural position of the Elstad Formation relative to the Brøttum Formation is still uncertain, and further studies are needed in order to explain the structural geometry here.

An imbricate stack, with individual slices bounded by NW-dipping reverse listric faults, also occurs within the Rendalen Formation in the Atna area (Sæther & Nystuen 1981). The tectonic shortening provides evidence for the presence of a blind thrust at depth.

High- to low-angle contraction faults underlie basement sheets in the Femunden area (Nystuen 1978, 1979). Some of these sheets have a prismatic shape, being laterally bounded by steep ramps trending 130°. The amount of slip along the *Sålekinna thrust* (SÅT) is a minimum of 30 km. The thrust sheets at the eastern side of Femunden are here referred to the *Femunden thrust zone;* the frontal thrust, *Femunden thrust* (FT), is indicated in Fig. 1. This zone includes several NW-dipping thrust sheets, some of which reveal a duplex geometry (Dahlstrom 1970, p. 352, Boyer & Elliot 1982, p. 1199), arranged in a piggy-back fashion (Nystuen 1979). The whole structure can be interpreted as a giant complex horse (Elliot & Johnson 1980, p. 73), bounded by the ORT as floor thrust and the *Kvitvola thrust* as roof thrust. However, the Femunden thrust zone deforms the Kvitvola thrust (Fig. 3). Hence, in this area the ORT is, at least partly, younger than the overlying Kvitvola thrust, and a piggy-back sequence of thrusting (Elliot 1976, p. 958) can be proved in this section (Fig. 3) (Nystuen 1979).

The *Mistra thrust* (MT) is a contraction fault carrying the overthrust *Mistra basement sheet* (Holmsen & Oftedahl 1956, Sæther 1979). This thrust structure also deforms the overlying Kvitvola thrust (Fig. 5).

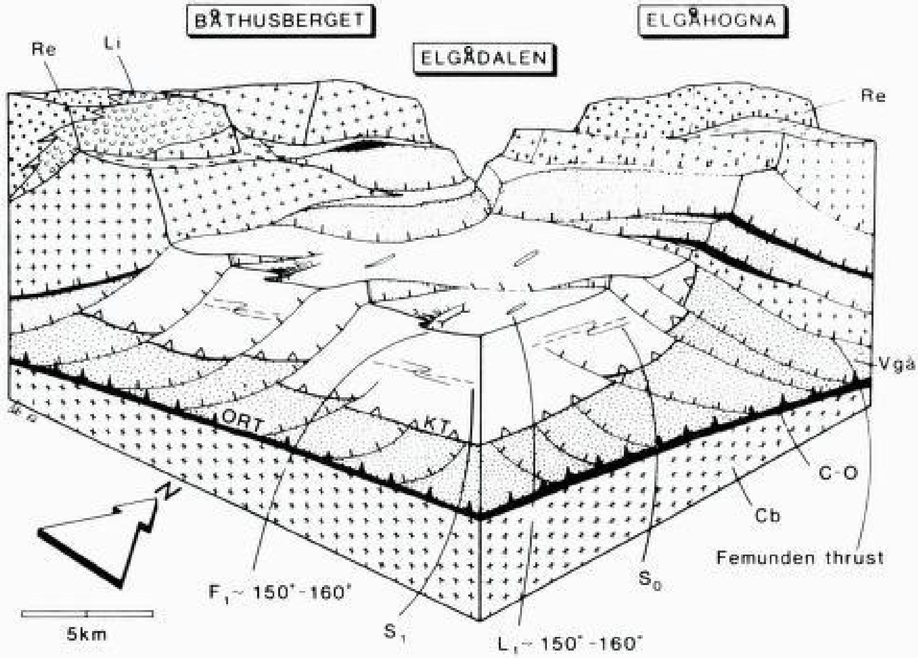

Fig. 3. Femunden thrust zone. Conceptual diagram from the front of the thrust zone. Source: Nystuen (1979). ORT = Osen–Røa thrust, KT = Kvitvola thrust, Cb = Crystalline basement, C–O = Cambro-Ordovician, Vgå = Vangsås Formation, Li = Litlesjøberget Conglomerate, Re = Rendalen Formation. Location shown in Fig. 1b.

FOLD GEOMETRY

Large-scale folds are strongly compressed, overturned to the south and cut by listric faults in the front areas. Open and broad synforms and antiforms dominate in the central areas between Gudbrandsdalen and Østerdalen and between Femunden and Øvre Rendal (Fig. 1). Recumbent folds facing SSW occur in the western part of the *Vinstra–Stai antiform* (VSA) (Englund 1973). Fold hinges are curved in plan view, resulting in arcuate axial surface traces with dimensions ranging from 70–80 km (Dokka–Rena) to about 500 m across. Listric faults are conformably arced. Fold axes plunge slightly away from the central, symmetry lines (c. 165°) of the arcuate structures, and overturned folds verge towards the convex sides. The bending of the hinge lines is continuous without any evidence of superposed folding.

A similar arcuate pattern in the Valdres Nappe was discussed by Nickelsen (1974). He favoured an origin by the rotation of originally NE–SW trending early F_1-folds towards the longest axis of the deformation ellipsoid (NW–SE, nearly parallel to the elongation lineation L_2 of that area) during a D_2-deformation. In the Valdres area, this is favoured by the fact that F_1-folds are transected by later penetrative S_2-cleavages which occur as axial surfaces of F_2-folds (Nickelsen 1974). Structural elements similar to these in the Valdres area occur in the northwestern part of the Osen–Røa Nappe Complex (Englund 1973), but have not been recognized farther east. The deformation phase which has produced the early F_1-folds has probably affected only those sequences which originally were located rather far west on the Baltoscandian craton. The arcuate folds in the Osen–Røa Nappe Complex are here interpreted to be the result of one single phase of deformation. Rotation of folds, which were initiated with a NE–SW trend, may have been brought about by regional variations in rate of tectonic mass transport towards the SSE. The protruding noses of the arcuate fold structures have probably moved with higher velocities than the flanking segments. Such variations in rate of movement may be due to local differences in nappe thickness, surface slope and friction.

MINOR FOLDS, CLEAVAGE AND LINEATIONS

Small-scale folds comprise mainly 'drag folds' with an orientation congruent with that of the host folds. The axial trend of irregular disharmonic folds in incompetent units may deviate considerably from the orientation of regional folds. Folding has taken place by flexural slip along bedding surfaces, but particularly in the northern and northwestern part of the nappe complex this type of folding has been modified by differential flow or shear along cleavage.

Spaced cleavage is developed as an axial surface cleavage. It becomes more pronounced and penetrative towards the northwest and north and passes into a slaty cleavage. Quartz grains and conglomerate pebbles are flattened in the cleavage surface and elongated (110°–160°) (Fig. 1). Local boudins are nearly normal to this direction (Fig. 7). Quartz segregations and quartz veins are common in the northern part of the nappe complex. The cleavage and elongation lineation are interpreted as cogenetic with the regional folds and main phase of thrusting, but cleavage of later origin also occurs, as described below.

Kvitvola Nappe structures

KVITVOLA THRUST

The Engerdalen Group in the Kvitvola Nappe is derived from an open shelf basin, and there is no obvious lateral stratigraphic continuity between the Osen–Røa Nappe Complex and the Kvitvola Nappe, though a general lithostratigraphical correlation exists (Nystuen 1980).

The sole *Kvitvola thrust* (KT) cuts those folds and thrusts which are described in the Osen–Røa Nappe Complex, and the Kvitvola Nappe rests on the lower as well as the higher formations of the Hedmark Group. The nappe also overlies autochthonous basement and Cambrian shales east of the Engerdalen fault (EF). A total absence of structural conformity between the nappes can be directly observed in many outcrops and can be demonstrated regionally (Figs. 1, 4–7). However, in the Ringebu–Vinstra area there is a coincidence between trends of fold axes and lineations in the two nappes (Englund 1973), and local concordance also exists in other places (Fig. 7). On

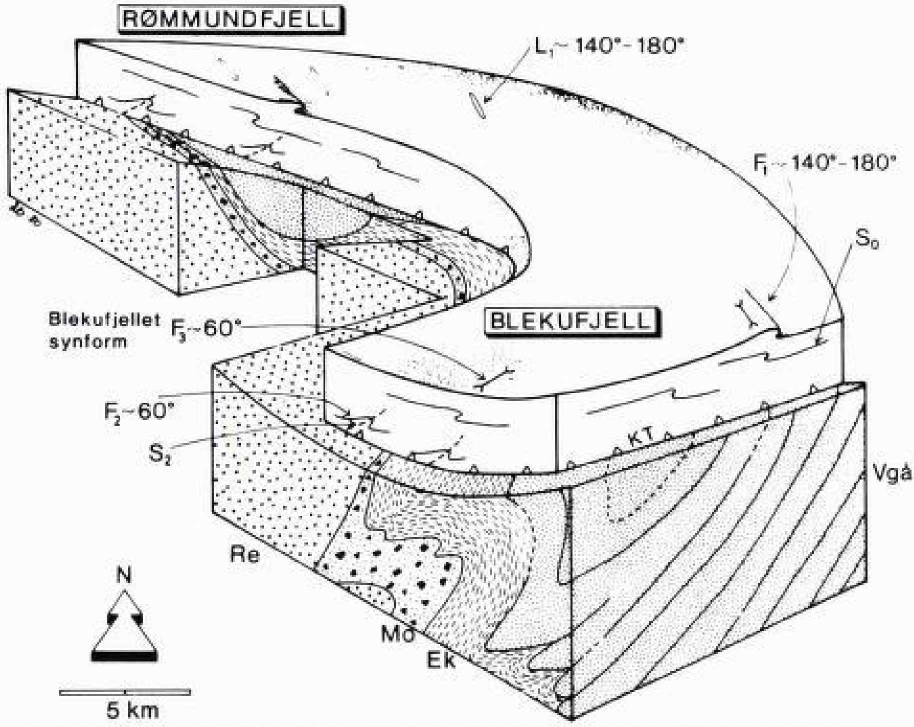

Fig. 4. Blekufjellet synform and structural relations between the Kvitvola Nappe and the Osen-Røa Nappe Complex. Conceptual diagram. Sources: Nystuen (1975a, 1975b, 1975c). KT = Kvitvola thrust, Re = Rendalen Formation, Mo = Moelv Tillite, Ek = Ekre Shale, Vgå = Vangsås Formation. Location shown in Fig. 1b.

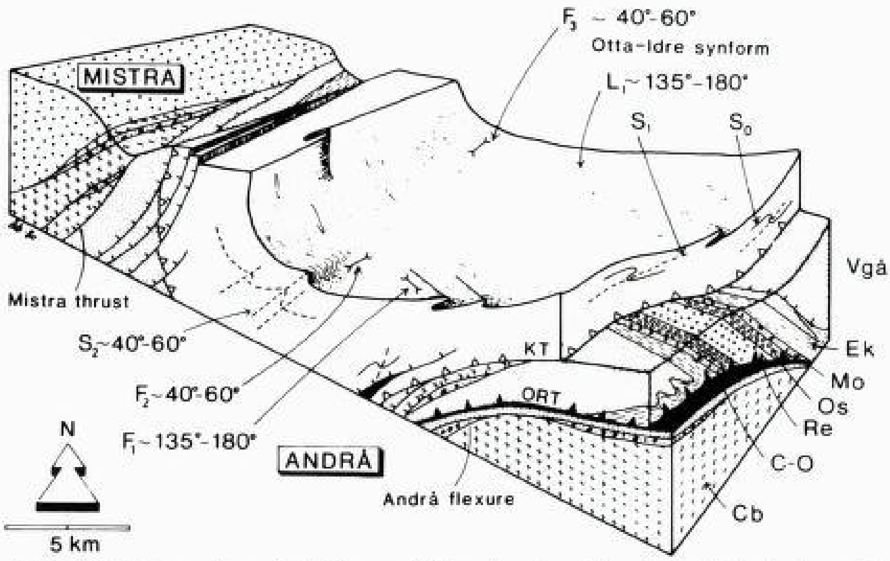

Fig. 5. Otta-Idre synform, Andrå flexure, Mistra thrust and structural relation between the Osen-Røa Nappe Complex and the Kvitvola Nappe. Conceptual diagram. Source: Sæther (1979). ORT = osen-Røa thrust, KT = Kvitvola thrust, Cb = Crystalline basement, C-O = Cambro-Ordovician, Re = Rendalen Formation, Os = Osdalen Conglomerate, Mo = Moelv Tillite, Ek = Ekre Shale, Vgå = Vangsås Formation. Location shown in Fig. 1b.

a regional scale it appears that the thrusting of the Kvitvola Nappe is *younger* than the thrusting of the Osen–Røa Nappe Complex. Nevertheless, the structural features of the Femunden thrust zone (Fig. 3) and the Mistra thrust (Fig. 5) show that in these places the thrusting of the Kvitvola Nappe is *older* than movements along the Osen–Røa thrust.

The Kvitvola thrust cuts the stratigraphic section (Fig. 2) with flats following the base of the incompetent Hylleråsen Formation over large areas. The Høyberget Formation dominates in the hangingwall in the northern part, whereas the lower units are recognized in the southern part; this may indicate that the basal thrust on a regional scale cuts up-section at the trailing edge. However, stratigraphical and structural data are still too limited to allow any definite conclusion to be drawn on the nature of the sole thrust.

Thrusts are developed as footwall imbricates (Butler 1982, p. 242) in several areas. Imbricate sheets derived from the footwall are squeezed into the basal zone of the nappe along contraction faults east of the Engerdalen fault (Nystuen 1974), east of Storsjøen (Sæther 1979) and in Åstdalen (Fig. 7). Fracture cleavage (S_2) and shear folds (F_2) are limited to the basal zone and also penetrate into the footwall within the *Stai fracture zone* (Fig. 6).

EARLY CLEAVAGE, FOLDS, LINEATIONS AND THRUSTS

The rocks of the Kvitvola Nappe reveal a slaty cleavage (S_1) which is penetrative except in the southeasternmost areas (Figs. 1, 3–7). It behaves as an axial plane foliation to recumbent and isoclinal plastic-type folds (F_1). The fold

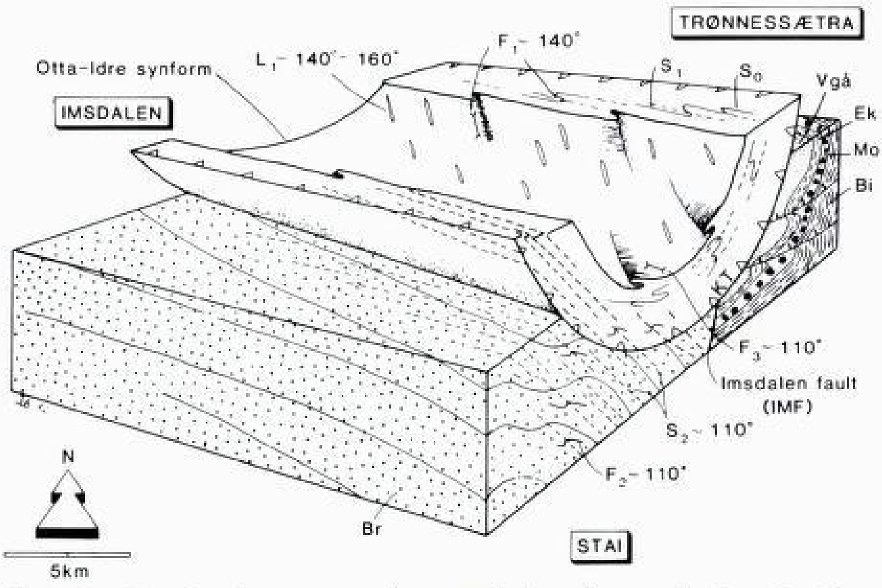

Fig. 6. Otta–Idre, Stai fracture zone and structural relation between the Osen–Røa Nappe Complex and the Kvitvola Nappe. Conceptual diagram. Sources: Elvsborg & Nystuen (1978) and Sæther & K. Bjørlykke (1981). KT = Kvitvola thrust, Br = Brøttum Formation, Bi = Biri Shale, Mo = Moelv Tillite, Ek = Ekre Shale, Vgå = Vangsås Formation. Location shown in Fig. 1b.

trends are parallel (110°–180°) to a mineral elongation (L_1); and the stretched and flattened quartz and feldspar grains lie within the S_1-cleavage. Early formed quartz-microcline-hematite segregations are frequently deformed into S_1 and produce a lineation parallel with L_1.

The incorporation of basement rocks, such as augen-gneiss, granite mylonite, meta-anorthosite and metagabbro, must belong to an early stage in the structural evolution of the nappe. These rocks are now present as sheets at the base of the nappe or within the sedimentary sequence, bounded by folded thrust surfaces. An upper and folded thrust sheet in the Ringebu area, revealing repeated lithostratigraphy, is also of early tectonic origin (J.-O. Englund, pers. comm. 1982).

LARGE-SCALE FOLDS AND ASSOCIATED THRUSTS

The sedimentary strata of the Engerdalen Group are mostly flat-lying. Large folds are open and broad with fold axis orientations varying from about 110° in the west to 040° in the east. In the Engerdalen area, one of these folds, the *Tverrfjellet syncline* (TS), is asymmetric with the northern limb slightly overturned to the southeast.

Towards the Mistra and Femunden thrusts (Fig. 1) listric contraction faults increase in frequency within the Kvitvola Nappe and are well marked topographically at the SE edge of the Femunden thrust zone (Figs. 3 & 5).

Furusjøen–Åstdalen antiform

The *Furusjøen–Åstdalen antiform* (FÅA) deforms the Kvitvola thrust as well as the underlying Brøttum Formation (Figs. 1 & 7). The Brøttum Formation crops out within the *Flakksjøen window* along the antiform axis, and another window is probably present at Furusjøen further to the NW, outside the map area of Fig. 1 (J.-O. Englund, pers. comm. 1982).

At Åstdalen, early structures in the Kvitvola Nappe and slaty cleavage and lineation in the Osen–Røa Nappe Complex are overprinted by a brittle-type deformation along the FÅA (Fig. 7). The deformation appears to be concentrated to a zone of about ± 50 m below and above the Kvitvola thrust, and comprises small-scale tight folds (F_{3a}) overturned to the NNE, crenulation cleavage (S_{3a}) which dips to the SSW and small contraction faults with dip in the same direction. The structural features here can be explained as a local zone of late folding verging NE. It could also be the result of collapse of a lateral ramp (Butler 1982), and hence the axes could lie in the movement direction. Further structural studies are needed along the Furusjøen–Åstdalen antiform.

Otta–Idre synform and Snødøla–Steinfjellet antiform

The erosional remnants of the Kvitvola Nappe are preserved within depressions along the *Otta–Idre synform* (OIS) and the parallel *Blekufjellet synform* (BS) which are folds (F_3 & F_{3b}) defined by contours on the Kvitvola thrust

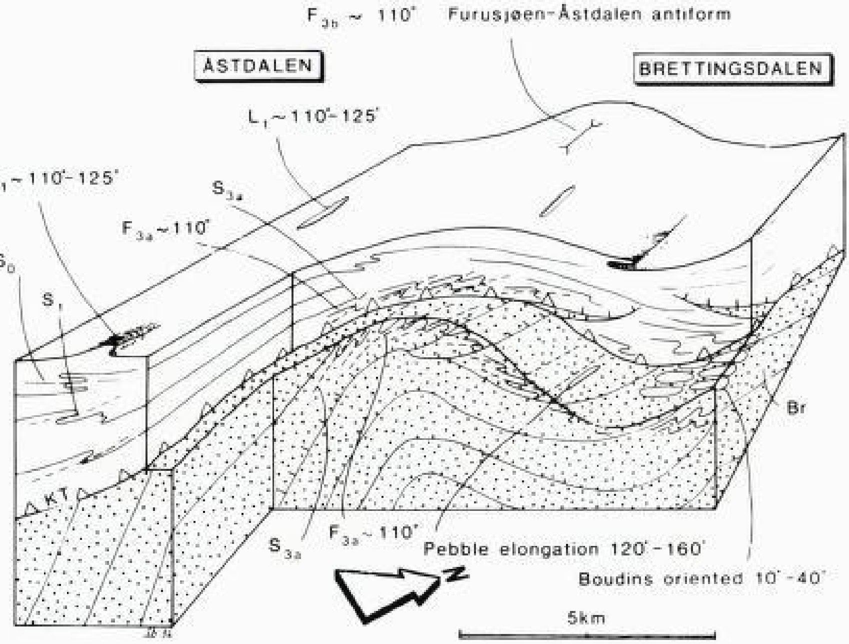

Fig. 7. Part of the Furusjøen–Åstdalen antiform and structural relationship between the Osen–Røa Nappe Complex and the Kvitvola Nappe. Conceptual diagram. KT = Kvitvola thrust, Br = Brøttum Formation. Location shown in Fig. 1b.

(Figs. 1, 4–6). Strand (1960) related the Gudbrandsdalen part of the OIS to a Caledonian cross-folding, but the regional pattern of the axial plane trace displays an arcuate structure, though much more open-arced than those in the Osen–Røa Nappe Complex. The OIS interferes with transverse synforms and antiforms; the *Central Østerdalen Depression* (CØD) is formed at the intersection between OIS and the *Storsjøen–Øvre Rendalen* synform (SØS) (Sæther & Nystuen 1981).

The windows exposing the basement in the north are eroded domes along the *Snødøla–Steinfjellet antiform* (SSA). The domes ans saddles between them can be interpreted as originating by an interference between the NE-trending SSA and several other SSE-trending synforms and antiforms (Fig. 1). The outcrop pattern of the Kvitvola Nappe is controlled by the erosion along such a transverse antiform south of the Atnsjøen window and along the *Grøndalen–Osdalen antiform* (GOA) between Storsjøen and Trysilelva. Strike and dip of bedding and coincident cleavage in sandstones indicate that the SSA is bounded by a broad, parallel synform on its southeastern side. A fracture cleavage which dips steeply to the NW and trends parallel to the SSA, cuts basement and slaty cleavage in the nappe cover and is interpreted to have been formed together with the basement antiform (SSA) (Nystuen & Ilebekk 1981). The basement antiform continues NNE-wards as the 'Riksgrense antiform' which also deforms the Caledonian nappe cover (e.g. Gee 1980). Additional basement antiforms exist still further north in the Scandinavian Caledonides.

Origin of the Snødøla–Steinfjellet antiform

Geophysical studies in the fold belts of the Canadian Rocky Mountains and the Appalachians have demonstrated basement surfaces which dip slightly and evenly for 150–200 km beneath a nappe cover towards the central zone of the orogen (e.g. Price 1981, Brewer et al. 1981, Hatcher 1981). A major implication of this discovery is that basement domes exposed within the nappe regions have been interpreted as allochthonous, underlain by a master thrust fault. Thus, in this way, Hossack et al. (in press) interpreted the basement of the Beito window in the Valdres area to be allochthonous. A major question arises: are all basement antiforms within the nappe region in the Scandinavian Caledonides allochthonous, underlain by a regional sole thrust? The answer is fundamental to all attempts to construct balanced cross sections (Dahlstrom 1969) through the mountain belt.

Limited data are available on the structural behaviour of the autochthonous basement beneath the nappe cover in the Scandinavian Caledonides. From the Sparagmite Region a map showing depths to the magnetic basement has been published by K. Åm (Fig. 6, in Nystuen 1981). In the Caledonian nappe front in Jämtland, Elming (1980) found high magnetic susceptibilities only in the granitic basement, and this is probably valid also for present area. In northern Jämtland, calculated depth to magnetic basement coincided well with depth to crystalline basement found by drilling (Hesselblom 1979). Thus, the magnetic basement of Åm (in Nystuen 1981) is here interpreted to approximate to the crystalline basement surface beneath the nappe cover. The morphology of this surface is characterized by 'lows' and 'highs' which coinside well with the late synforms, antiforms, domes and saddles expressed by the surface geology (Fig. 1). This indicates that the Snødøla–Steinfjellet basement antiform (SSA) was formed in late Caledonian time (Lower to Middle Devonian?) together with similar basement undulations farther south as well as with the Otta–Idre synform and related synforms and antiforms. Basement flexure at Andrå (AFL) and Åsta (ÅFL) may mark a southern limit of this deformation.

Structural evolution: discussion and conclusion

In the Scandinavian Caledonides nappes advanced from the west to the east. In the central and southern parts of the orogen emplacement of thrust sheets on to the Baltoscandian craton commenced by the partial closure of the Iapetus Ocean during the early Ordovician. The main phase of nappe translations occurred during the Middle to Upper Silurian (Roberts & Gee 1981).

The detachment of Kvitvola Nappe from a shelf basin was brought about when higher tectonostratigraphical units approached the shelf from still more western, continental margin areas and eugeosynclinal zones. The initial fracture, starting in the crystalline basement, propagated into the ductile Hylleråsen Formation (carbonate-shale) and mainly followed this stratigraphical level as the sole thrust (KT). Penetrative slaty cleavage (S_1), mineral elongation (L_1), plastic-type folds (F_1) and phyllonite and mylonite zones give evidence of flattening and extension, perhaps related to gravitational collapse (Ramberg 1981) under the overburden of a thick nappe pile. Metamorphism under greenschist facies conditions took place. The laterally extended horizontal stratification indicates that the major thrust movements occurred along a low-frictional sole thrust. However, locally the nappe body collapsed along low- and high-angle contraction faults.

Initial detachment and movement of the Osen–Røa Nappe Complex started when the advancing nappe pile reached the cratonic 'sparagmite basin' some 200–400 km NW of the present Sparagmite Region. In this phase, slaty cleavage, lineations and phyllonite zones were formed due to increase in temperature, confining pressure and simple shear (mesoscale folds (F_1) in the northwest may be still older). The initial fracture originated also in this case in the crystalline basement; it cut up-section and entered the ductile Cambrian shale beds which rested on the eastern foreland basement. A major amount (minimum 125 km) of slip has taken place along this sole thrust (ORT). Imbrication, folding and evolution of the arcuate fold structures took place concurrently with the southeastwards propagation of the active thrust system.

The regional structural disconformity between the Osen–Røa Nappe Complex and the Kvitvola Nappe can be explained in two fundamental different ways (Fig. 8). The imbrications and folds in the Osen–Røa Nappe Complex may have suffered erosion contemporaneously with their formation. The Kvitvola Nappe was emplaced on the erosional surface and became later imbricated and folded during piggy-back transport by the still active Osen–Røa thrust (ORT) (Fig. 8A). The Bruflat Sandstone of upper Llandovery to lower Wenlock age in the Mjøsa area is a possible fore-deep sediment originated from erosion of the advancing Osen–Røa Nappe Complex.

J. R. Hossack (written communication 1982) proposed that the contact relations between the two nappe units were the result of *out-of-sequence thrusting* (Fig. 8B). The ORT and secondary thrusts were formed when the Kvitvola Nappe was emplaced on the 3–4,000 m thick sequence in the 'sparagmite basin'. Further movements along ORT gave rise to piggy-back transport of the Kvitvola Nappe and imbrications in both nappe units (Femunden thrust zone). Folding occurred ahead the nappes, and during reactivation of the Kvitvola thrust (KT) the out-of-sequence thrusting occurred in this area while the ORT still propagated towards the SE. Similar out-of-sequence thrusting has been described from the Moine Thrust Zone in Scotland by McClay & Coward (1981). The mechanism implies that the reactivated KT has truncated the high-angle faults and folds in the Osen–Røa Nappe Complex and thus acts as the roof thrust of a giant duplex in the footwall (see Boyer & Elliott 1982, p. 1209). Both hypotheses (Fig. 8A & B) need further structural analyses.

Late-thrust structures are the open NE-trending folds (e.g. Tverrfjellet syncline), the Stai fracture zone (Fig. 6), the Furusjøen–Åstdalen antiform (Fig. 7), the Mistra thrust (Fig. 5) and the Femunden thrust zone (Fig. 3). Post-

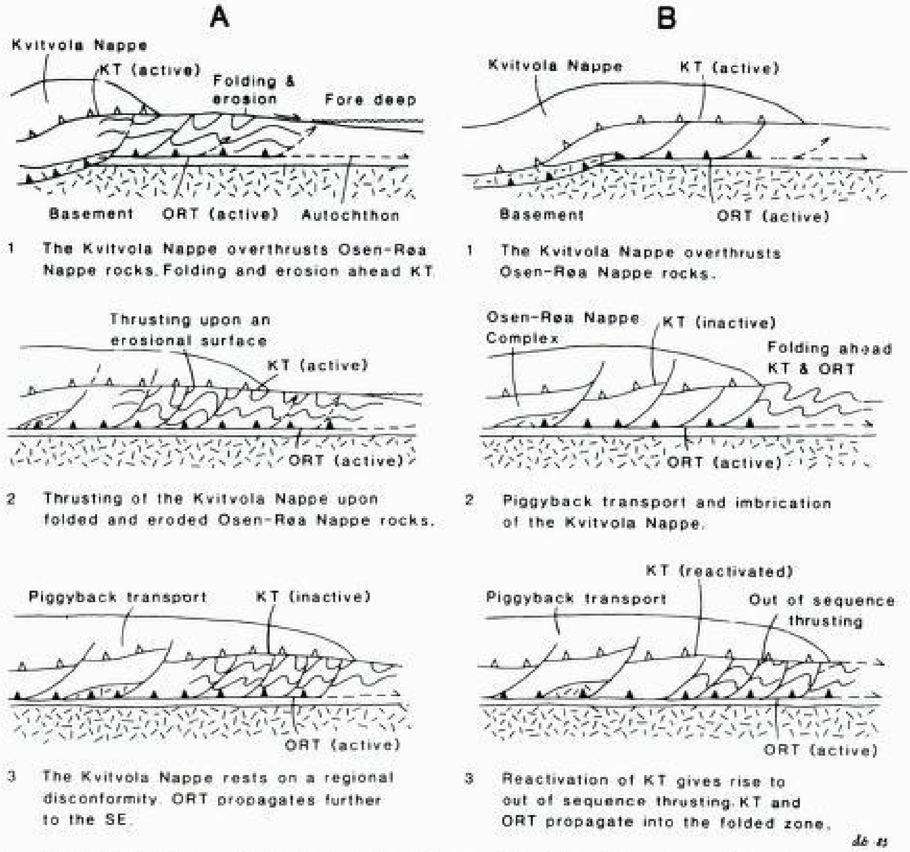

Fig. 8. Interpretations of the structural evolution of the Osen–Røa Nappe Complex and the Kvitvola Nappe. In *A* the structural conformity between the nappe units is explained as an erosional surface and in *B* (after J. R. Hossack, written comm. 1982) as the result of tectonic truncation by KT during out of sequence thrusting.

thrust deformation affected the basement and gave rise to the Snødøla–Steinfjellet antiform, the Otta–Idre synform and related arc and saddle structures (Fig. 1).

The nappe structures of the present area display similarities with thrust structures believed to have formed by gravitational spreading (Elliott 1976, Cooper 1981). However, serious objections have been raised against this hypothesis of thrusting (Chapple 1978, Murrell 1981). Burchfiel & Davis (1972), Price (1981) and Hatcher (1981) maintained that compressional thrusting including formation of low-angle contraction faults into the crystalline basement were brought about by a laterally directed pressure acting from the hot mobile central zone. No definite conclusion can be drawn regarding the driving force for the compressional nappe structures in the Sparagmite Region. However, the great volume of the thrust bodies and the incorporation of basement sheets indicate that the operating stress probably has been higher than that which could have been formed gravitationally only from a surface slope of the nappes (cf. Chapple 1978).

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REFERENCES

Bjørlykke, A. 1979: Gjøvik og Dokka. Beskrivelse til de berggrunnsgeologiske kart 1816 I og 1816 IV, M. 1:50000. (med fargetrykte kart). *Norges geol. Unders. 344*, 48 pp.

Bjørlykke, A., Høy, T. & Skålvoll, H. 1980: Løten, berggrunnsgeologisk kart 1916 I, M. 1:50000. *Norges geol. Unders.*

Bjørlykke,K. 1976: Rena, berggrunnsgeologisk kart 1917 II, M 1:50000. *Norges geol. Unders.*

Bjørlykke, K., Elvsborg, A. & Høy, T. 1976: Late Precambrian sedimentation in the central sparagmite basin of South Norway. *Norsk geol. Tidsskr. 56, 233–290.*

Bockelie, J. F. & Nystuen, J. P. in press: The southeastern part of the Scandinavian Caledonides. *In* Gee, D. G. & Sturt, B. A. (eds.) *The Caledonian Orogen – Scandinavia and related areas.* John Wiley, New York.

Boyer, S. E. & Elliott, D. 1982: Thrust systems. *Am. Assoc. Petrol. Geol. Bull. 66*, 1196–1230.

Brewer, J. A., Cook, F. A., Brown, L. D., Oliver, J. E., Kaufman, S. & Albaugh, D. S. 1981: COCORP seismic reflection profiling across thrust faults. *In* McClay, K. R. & Price, N. J. (eds.) *Thrust and nappe tectonics.* Geol. Soc. London Spec. Publ. 9, 501–511.

Burchfiel, B. C. & Davis, G. A. 1972: Structural framework and evolution of the southern part of the Cordilleran orogen, western United States. *Am. J. Sci. 272*, 97–118.

Butler, R. W. H. 1982: The terminology of structures in thrust belts. *J. Struct. Geol. 4*, 239–245.

Chapple, W. M. 1978: Mechanics of thin-skinned fold-and-thrust belts. *Bull. Geol. Soc. Am. 89*, 1189–1198.

Cooper, M. A. 1981: The internal geometry of nappes: criteria for models of emplacement. *In* McClay, K. R. & Price, N. J. (eds.) *Thrust and nappe tectonics.* Geol. Soc. London Spec. Publ. 9, 225–233.

Dahlstrom, C. D. A. 1969: Balanced cross sections. *Can. J. Earth Sci. 6*, 743–757.

Dahlstrom, C. D. A. 1970: Structural geology in the eastern margin of the Canadian Rocky Mountains. *Bull. Can. Petrol. Geol. 18*, 332–406.

Elliott, D. 1976: The motion of thrust sheets. *J. Geophys. Res. 81*, 949–963.

Elliott, D. & Johnson, M. R. W. 1980: Structural evolution in the northern part of the Moine Thrust Belt. *Trans. Roy. Soc. Edinburgh: Earth Sciences 71*, 69–96.

Elming, S.-Å. 1980: Density and magnetic properties of rocks in the Caledonides of Jämtland, Sweden. *Geol. Fören. Stockh. Förh. 102*, 439–453.

Elvsborg, A. & Nystuen, J. P. 1978: Evenstad, berggrunnsgeologiske kart 1917 I, M. 1:50000. *Norges geol. Unders.*

Englund, J.-O. 1972: Sedimentological and structural investigations of the Hedmark Group in the Tretten–Øyer–Fåberg district, Gudbrandsdalen. *Norges geol. Unders. 276*, 59 pp.

Englund, J.-O. 1973: Stratigraphy and structure of the Ringebu–Vinstra district, Gudbrandsdalen; with a short analysis of the western part of the sparagmite region in southern Norway. *Norges geol. Unders. 293*, 58 pp.

Englund, J.-O. 1978: Lillehammer, berggrunnsgeologisk kart 1817 II, M. 1:50000. *Norges geol. Unders.*

Gee, D. G. 1980: Basement-cover relationships in the central Scandinavian Caledonides. *Geol. Fören. Stockh. Förh. 102*, 455–474.

Hatcher, R. D. 1981: Thrusts and nappes in the North American Appalachian Orogen. *In* McClay, K. R. & Price, N. J. (eds.) *Thrust and nappe tectonics.* Geol. Soc. London Spec. Publ. 9, 491–499.

Hesselblom, Å. 1979: The Tåsjön–Ormsjön magnetic anomaly, Caledonian Front, Sweden. *Geol. Fören. Stockh. Förh. 101*, 349–352.

Holmsen, P. & Oftedahl, C. 1956: Ytre Rendal og Stor-Elvdal. Beskrivelse til de geologiske rektangelkart. *Norges geol. Unders. 194*, 173 pp.

Hossack, J. R., Garton, M. & Nickelsen, R. P. in press.: The geological section from the foreland up to the Jotun thrust sheet in the Valdres area, south Norway. *In* Gee, D. G. & Sturt, B. A. (eds.) *The Caledonide Orogen-Scandinavia and related areas.* J. Wiley & Sons.

Høy, T. & Bjørlykke, A. 1980: Hamar, berggrunnsgeologisk kart 1916 IV, 1:50000. *Norges geol. Unders.*

McClay, K. R. 1981: What is a thrust? What is a nappe? *In* McClay, K. R. & Price, N. J. (eds.) *Thrust and nappe tectonics.* Geol. Soc. London Spec. Publ. 9, 7–9.

McClay, K. R. & Coward, M. P. 1981: The Moine Thrust Zone: an overview. *In* McClay, K. R. & Price, N. J. (eds.) *Thrust and nappe tectonics.* Geol. Soc. London Spec. Publ. 9, 241–260.

Murrell, S. A. F. 1981: The rock mechanics of thrust and nappe formation. *In* McClay, K. R. & Price, N. J. (eds.) *Thrust and nappe tectonics.* Geol. Soc. London Spec. Publ. 9, 99–109.

Nickelsen, R. P. 1974: Geology of the Røssjøkollan-Dokkvatn area, Oppland. *Norges geol. Unders. 314,* 53–100.

Nystuen, J. P. 1974: Engerdal, berggrunnsgeologisk kart 2018 I, M. 1:50000. *Norges geol. Unders.*

Nystuen, J. P. 1975a: Nordre Osen, berggrunnsgeologisk kart 2017 IV, M. 1:50000. *Norges geol. Unders.*

Nystuen, J. P. 1975b: Elvdal, berggrunnsgeologisk kart 2018 III, M. 1:50000. *Norges geol. Unders.*

Nystuen, J. P. 1975c: Jordet, berggrunnsgeologisk kart 2017 I, M. 1:50000. *Norges geol. Unders.*

Nystuen, J. P. 1978: Holøydal, berggrunnsgeologisk kart 1719 III, M. 1:50000. *Norges geol. Unders.*

Nystuen, J. P. 1979: Elgå, berggrunnsgeologisk kart 1719 II, M. 1:50000. *Norges geol. Unders.*

Nystuen, J. P. 1980: Stratigraphy of the Upper Proterozoic Engerdalen Group, Kvitvola Nappe, southeastern Scandinavian Caledonides. *Geol. Fören. Stockholm. Förh. 102,* 551–560.

Nystuen, J. P. 1981: The Late Precambrian "Sparagmites" of southern Norway: a major Caledonian allochthon – the Osen–Røa Nappe Complex. *Am. J. Sci. 281,* 69–94.

Nystuen, J. P. 1982: Late Proterozoic basin evolution on the Baltoscandian craton: the Hedmark Group, southern Norway. *Norges geol. Unders. 375,* 1–74.

Nystuen, J. P. & Ilebekk, S. 1981: Stratigraphy and Caledonian structures in the area between the Atnsjøen and Spekedalen windows, Sparagmite Region, southern Norway. *Norsk geol. Tidsskr. 61,* 17–24.

Oftedahl, C. 1943: Om sparagmiten og dens skyvning innen kartbladet Øvre Rendal. *Norges geol. Unders. 161,* 65 pp.

Price, R. A. 1981: The Cordilleran foreland thrust and fold belt in the southern Canadian Rocky Mountains. *In* McClay, K. R. & Price, N. J. (eds.) *Thrust and nappe tectonics.* Geol. Soc. London Spec. Publ. 9, 427–448.

Ramberg, H. 1981: The role of gravity in orogenic belts. *In* McClay, K. R. & Price, N. J. (eds.) *Thrust and nappe tectonics.* Geol. Soc. London Spec. Publ. 9, 125–140.

Roberts, D. & Gee, D. G. 1981: Caledonian tectonics in Scandinavia. (Abstract) *Terra Cognita, 1,* 69–70.

Sæther, T. 1979: Storsjøen, berggrunnsgeologisk kart 1918 II, M. 1:50000. *Norges geol. Unders.*

Sæther, T. & Bjørlykke, K. 1981: Stor-Elvdal, berggrunnsgeologisk kart, M. 1:50000. *Norges geol. Unders.*

Sæther, T. & Nystuen, J. P. 1981: Tectonic framework, stratigraphy, sedimentation and volcanism of the Late Precambrian Hedmark Group, Østerdalen, south Norway. *Norsk geol. Tidsskr. 61,* 193–211.

Strand, T. 1960: The pre-Devonian rocks and structures in the region of Caledonian deformation. *Norgese geol. Unders. 208,* 170–284.

Devonian Tectonic Deformation in the Norwegian Caledonides and Its Regional Perspectives

DAVID ROBERTS

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Late-Caledonian megascopic folds deforming the Old Red Sandstone sequences of southern Norway are mostly synclines of E–W to NE–SW trend and E to NE plunge. Thrusts along the eastern and southern margins of several ORS areas, and stretching lineations, denote approximate southeastward translations of the sequences from their basinal sites. This contractile movement occurred principally along rejuvenated, syn-depositional, planar or listric-style extension faults.

Development of the basins took place during the imposition of a fundamental NW–SE to NNW–SSE crustal extension following the major Scandian, Silurian orogenic phase. This was also influenced by an orogen-parallel sinistral mega-shear from Svalbard through northern Britain to the Appalachians which assisted in producing a transtensional regime in southern Norway. Thus, during ORS basin development and sedimentation a subordinate dextral strike-slip component has acted in combination with the basic extensile field. During late Devonian, probably Frasnian time, this transtensional regime was replaced by a SE-directed compression, and partly transpression, so producing the macrofolds, late ductile thrusts and other structures which are traceable far outside the confines of the ORS districts.

D. Roberts, Norges geologiske undersøkelse, Postboks 3006, N-7001 Trondheim, Norway

Introduction

The Old Red Sandstone (ORS) basins of southern Norway have received much attention in recent years with emphasis placed on establishing detailed stratigraphies and facies sequences, and interpreting these in terms of palaeography within a framework of syn-depositional tectonism (Bryhni 1964, 1975, 1978, Nilsen 1968, Steel 1976, Steel & Gloppen 1980). In consequence of this, sedimentological studies have prevailed and comparatively little work has been done on aspects of post-depositional tectonic deformation. The Downtonian to Middle Devonian sediments do, in fact, display variable effects of what is generally termed late-Caledonian tectonism (Roberts & Sturt 1980), which is an important element in the overall development history of the mountain belt. In this synthesis, the most significant traits of this tectonic deformation are outlined, and discussed briefly in a wider context of the waning phases of Caledonide orogenic evolution. Details of stratigraphic sequences and lithologies are contained in the cited publications, including a recent review article (Steel et al. 1983), and will therefore not be repeated here.

Tectonic structures in the different ORS basins

A feature common to the 8 or 9 ORS basinal sequences is their deformation and dissection by open folds and faults. Many of the sequences have also been detached along thrust surfaces, mostly with evidence of east to southeast translations, although the extent and magnitude of the thrusting is generally not known in any detail.

The Lower to Middle Devonian sequences of *Vestlandet* (western Norway) occur in 4 main areas (Fig. 1). The axial trends of the folds which deform the ORS rocks vary from NE–SW at Solund (Nilsen 1968) through ENE–WSW in the Hornelen and Kvamshesten basins (Bryhni 1964, Höisæter 1971, Bryhni & Skjerlie 1975) to E–W at Haasteinen (Kolderup 1925), with axial

Fig. 1. Locations of the main areas of Old Red Sandstone sediments in southern Norway. The ruled area represents the approximate extent of the Caledonian nappes and autochthon. The remaining, white areas are Precambrian crystalline basement, probably partly allochthonous (Caledonian) in the northwest.

plunges consistently towards the E to NE quadrant (Fig. 2); axial surfaces are either vertical or dip steeply towards NW–N. Major cataclastic faults displacing the ORS rocks show main trends of NE–SW and c. NW–SE, while locally N–S fractures are of importance. Some basins are also fault-bounded, e.g. Hornelen and Haasteinen.

Thrust displacements of the ORS complexes have been reported from all the Vestlandet Devonian basinal areas, although the westernmost contacts of these rocks with their crystalline substrates generally show undisturbed angular–unconformable relationships. At *Solund* a 15° thrust contact delimits the southeastern margin of the ORS rocks. Above a thin basal mylonite with NW–SE quartz grain elongation lineation, conglomerate pebbles are strongly deformed within a zone of 2.5 km mappable width (Nilsen 1968). Pebble lineation is NW–SE, with NW plunge, contrasting strongly with the SE-dipping bedding in the conglomerate. A comparable, pervasive, NW–SE pebble lineation has also been reported from the outermost islands in the Solund district (Indrevær & Steel 1975). This may well relate to a subsurface thrust zone, and quite possibly an extension of the basal Solund Thrust. Dip-slip extensional faults parallel to the Solund Thrust, bedding-plane slippage faults and local, NE–SW, strike-slip faults are other deformational structures present in the Solund ORS district (Nilsen 1968). Minor thrusts occur in other parts of the Solund area, some with thin ultramylonites (Furnes & Lippard, in press); these same authors describe a lowest greenschist facies metamorphic paragenesis and schistosity in Devonian trachytic lavas. A thrust origin was also advocated by Kolderup (1926) for lensoid gabbro bodies occurring within Devonian conglomerates in the Hersvik area, although other interpretations have since been proposed – Nilsen (1968) favoured an intrusive origin, as gabbro sills, whereas Bryhni (1976) considered the gabbro as possibly representing a debris flow deposit. Recent investigations (D. M. Ramsay, pers.comm. 1982), including the recognition of a basal mylonite, favour a thrust emplacement for this gabbro. Cumulative thrust displacement of the Solund ORS sequences as a whole may be several tens of kilometres (B. A. Sturt, pers.comm. 1982).

At *Kvamshesten* (Fig. 1) the folded Devonian rocks are truncated by a flat-lying thrust delimiting the southern, eastern and northeastern boundaries of the ORS outcrop (Bryhni & Skjerlie 1975), with a "strongly mylonitized zone some tens of metres in thickness" (Höisæter 1971) marking the thrust contact with subjacent metasediments. Few details are yet available on this thrusting; those reported are ambiguous with Skjerlie (1971) noting both northward and northeastward directed thrusting, whereas eastward displacement is favoured by Nilsen (1968). An eastward translation seems the more likely, at least for the main fairly ductile detachment and transport of the ORS rocks from their basinal site, and strain reduction to mylonite at the base of the sequence. The thrust surface in the eastern part of the district has been referred to as a probable sole thrust by Bryhni & Skjerlie (1975). Judging from the map-picture (Höysæter 1971, Bryhni & Skjerlie 1975) which shows the thrust, towards the west, also forming the floor to a sheet of Precambrian plutonic rocks and Lower Palaeozoic schists below the unconformably overlying ORS sediments, this characterization would appear valid. Another feature of interest is that some of the faults cutting the ORS rocks do not transect the mylonitic thrust zone, whereas other, younger fractures penetrate across the thrust into the basement. The small *Haasteinen* basin just north of Kvamshesten (Fig. 1) also exposes a thrust contact along its southeastern margin (Kolderup 1925); no structural details are available from this area at the present time.

ORS sediments of the *Hornelen* basin are fault-bounded along their northern and southern margins, and displaced by thrusting in the east. Syndepositional movement along the faults, partly dextral strike-slip, has been considered an important factor in the sedimentation history of this basin (Bryhni 1964, 1978, Steel 1976, Steel et al. 1977, Steel & Gloppen 1980). Structural data from the sole thrust (Bryhni 1978) contact zone are few, although mylonitization features and neocrystallization of quartz, calcite and epidote have been reported by Bryhni (1964). In addition, mudstones adjacent to the marginal faults are strongly cleaved. In contrast to the situation at Kvamshesten, some of the folds in the southern part of this basin appear to deform the basal thrust-fault (Bryhni 1978, fig. 1), indicating at least a local continuation of folding after thrust detachment and movement.

Further north, in the coastal region of Trøndelag and Nordmøre, ORS sediments occur primarily on *Hitra* and *Fosen,* and on islands adjacent to *Smøla* (Fig. 1). The structural trait common to all these sequences is their deformation in a NE–SW to ENE–WSW trending syncline (Siedlecka & Siedlecki 1972, Siedlecka 1975, Fediuk & Siedlecki 1977) (Fig. 2). It is possible that the synclines recorded in the different areas were once part of a single structure, now transected and displaced by faulting (Steel et al. 1983). Principal fault trends in this district are NW–SE, N–S and NE–SW (Fig. 2). On Hitra a high-angle reverse fault cuts out part of the southeastern limb of the Hitra syncline against the pre-ORS basement. Apart from this example, no other cases of post ORS tectonic inversion are known from this district. Also on Hitra, a spaced cleavage paralleling the axial surfaces of mesoscopic NE–SW folds is present in some mudstones, with carbonate concretions showing rotation into the plane of anisotropy (Roberts 1981). N–S trending kink-bands of westerly downstep and a north-dipping crenulation cleavage are minor tectonic structures occurring in the ORS on this same island.

In the southeastern parts of the Trondheim region the Devonian deposits at *Røragen* were deformed during two phases of folding, the main structure being a syncline of E–W trend and easterly plunge. Minor NW-facing folds and local NW-directed thrusts constitute secondary structural elements; these deform a very low grade lepidoblastic fabric which is present in some of the pelitic lithologies (Roberts 1974). Faults show NW–SE and NE–SW trends and probably represent rejuvenated syn-depositional structures (Holmsen 1963, Jakobsson 1978).

The *Oslo basin* of southeastern Norway exposes Ludlow–Downtonian red-

Area/basin	Main macrofold	Fold trend (& plunge)	Faults		Thrusts	Foliate and linear elements
			Syn-sedimentary	Main post-sed. trends		
Solund	Anticline	NE-SW (NE)	E-W in north, ?dip-slip. NE-SW in SE, ?dip-slip	NE-SW NNW-SSE	In SE. Thrusting towards SE. Basal mylonite.	Local schistosity in lava and pelite. Later crenulation cleavage. NW-SE pebble & min. lineation.
Kvamshesten	Syncline	ENE-WSW (ENE)	E-W in north, oblique-slip. ?ENE-WSW in south	NE-SW NW-SE	In S, E & NE. Thrusting towards SE-?E. ?Late movem. towards NE-N. Basal mylonite	None reported
Hasteinen	Syncline	E-W (E)	Not known, but assumed	NE-SW NW-SE N-S	In SE-E & NE. Direction not known, assumed SE-E.	None reported
Hornelen	Syncline	ENE-WSW (ENE)	E-W in north, strike/oblique-slip. E-W/NE-SW in S, dip-slip	NE-SW NW-SE N-S	In E & SE. Direction uncertain. Probably SE-E. Mylonite	Cleavage in mudstones. Nothing else reported.
Smola Islands	Syncline	NE-SW (NE)	?NE-SW in north, slip unknown. SE contact unknown	NE-SW NW-SE	None exposed	None reported
Hitra	Syncline	ENE-WSW (ENE)	NE-SW in north, slip unknown. NE-SW in south	NW-SE NE-SW E-W N-S	None, but high-angle reverse fault in south.	Local spaced cleavage ax.-surf. to folds. Later cren. cleavage in pelites.
Fosen	Syncline	NE-SW (NE)	Probable, but no data available.	NE-SW NW-SE N-S	None exposed	None reported. Conjugate fractures in pebbles.
Roragen	Syncline	E-W (E)	NW-SE in NE, ?dip-slip. NW-SE to E-W in SW, ?dip-slip.	NE-SW NW-SE	No major thrust exposed. Minor NW-directed thrusts.	Local schistosity (phyllites). Late NE-SW minor folds. Minor shears.

Fig. 2. Summary of structural data from the main ORS basins and areas in Norway. The information is culled from the papers cited in the text.

-bed sequences which are folded in Jura-style structures along E–W to NE–SW trends (Strand 1972, Steel et al. 1983). Folds are tighter towards the north where reverse faults transect the succession; these faults become shallower southwards (Bockelie & Nystuen 1983). In general, this fold- and thrust-shortening in the Oslo area is taken up by décollement along pelitic horizons – a characteristic feature along other parts of the Caledonian front.

Outside of the presently exposed ORS basin areas, biostratigraphic control on the age of the Caledonoid tectonic structures is of course lacking. There is no reason, however, to doubt Vogt's (1928) contention that structures belonging to this late-Caledonian deformation phase are to be found throughout the length of the Norwegian Caledonides. Open folds and fault sets which deform the ORS rocks in west Norway, in many cases can be traced inland for considerable distances, and it is considered likely that the synformal 'Faltungsgraben' and Bergen arc structure also belong to this orogenic phase. Similarly, the NE–SW Tyin–Gjende Fault of the Jotun region (Battey & McRitchie 1973, Emmett 1982) is also accorded a Devonian age (M. Heim, pers.comm. 1981), while in the Hardangerfjord region late, open, NE–SW macrofolds and normal extension faults (Naterstad et al. 1973) may relate to this late-Caledonian event. Further north in the Caledonides, several of the orogen-transverse basement culminations, e.g. in the Grong district, are also probably Devonian in age.

Timing of deformation

Traditionally, the folds, related faults and slightly later thrusts which affect the ORS sequences of southern Norway have been ascribed to the *Svalbardian* orogenic phase of Vogt (1928). This lowermost Upper Devonian (Frasnian) event was established from investigations on Spitsbergen and Bjørnøya where faunal control is more than adequate; on Bjørnøya, Upper Devonian sediments lie unconformably above folded Middle Devonian and pass conformably up into the Carboniferous. Vogt (1928, 1933) noted the strong similarity of the folding, in its style and character, to that seen in the Norwegian ORS successions, as well as in parts of the Scottish ORS, and introduced the term 'Svalbard folding' or 'Svalbard orogenesis' for this widespread event.

Constraints on the age of the deformation on the Norwegian mainland are less precise and more indirect than on Svalbard. Faunas in Upper Palaeozoic sediments in the Oslo region, overlying folded Siluro-Downtonian, indicate a Middle Carboniferous age for the oldest fossiliferous rocks (Olaussen 1981). Much further afield, undeformed mafic dykes on Ytterøy, Trondheimsfjorden (Priem et al. 1968) and Varangerhalvøya, Finnmark (Beckinsale et al. 1975), have also yielded uppermost Devonian intrusive ages. Elsewhere in northern Norway, where Devonian sediments are nowhere exposed, K-Ar studies have revealed a Lower Carboniferous weathering profile on the island of Andøya (Sturt et al. 1979). Thus, for mainland Norway, Vogt's Svalbardian event can be placed somewhere in the late Devonian. Although a Frasnian age is likely, the event need not have been perfectly synchronous everywhere; diachronism of orogenic deformation is indeed a hallmark of the Caledonian mountain belt. In recognition of this it may therefore be justifiable to adopt local names for this post-Middle Devonian deformation, e.g. *Røragenian* event for eastern Norway (Roberts 1974).

An exception to this general pattern of late Devonian deformation appears to obtain in the Oslo area. There, the tectonism affects the marine Silurian succession as well as the conformably overlying red-beds, and is essentially the diachronous end-stage of the major Scandinavian or Scandian (mid to late Silurian) orogenic event. This deformation, in the Oslo region, could well be Gedinnian or Siegennian in age rather than late Devonian (cf. Roberts 1974). If so, it would correlate in time with ORS molasse sedimentation in the several intermontane basin areas further to the northwest within the already deformed and uplifted mountain chain.

Basin development and tectonic deformation in a regional context

GENERAL ASPECTS

Recent studies of the west Norwegian ORS basins have confirmed Bryhni's (1964) thesis of a close relationship between ORS sedimentation and tectonism, and led to refinements in the concept of basin development in a tectonic regime involving a combination of strike-slip and dip-slip motions (Nilsen 1968, Steel 1976, Steel & Gloppen 1980). In this general model, most of the basins provide evidence of growth by lateral accretion with the depocentre migrating east- to southeastwards through time. Sediment transport of the basinal axial deposits was mainly directed towards west or northwest. Similar evidence of tectonic control over sedimentation has emerged from work in coastal Trøndelag and at Røragen (Siedlecka & Siedlecki 1972, Siedlecka 1975, Roberts 1974, Jakobsson 1978).

Prior to the establishment of this broadly extensional or transtensional (Harland 1971) regime in Devonian time, the Brito-Scandinavian Caledonides and their East Greenland counterpart evolved from the final cratonic suturing of the Laurentian and Baltoscandian plates in a major phase of crustal contraction. This Scandian crustal shortening involved southeastward nappe translations of hundreds of kilometres in the Norwegian part of the belt, with the shortening vector swinging towards south in the Oslo district. Nappe stretching lineations, c. NW–SE, and internal foliations were then deformed by major NE–SW folds, with antiforms providing the loci of many of the tectonic windows which occur throughout the orogen; and subsequent to this a period of crustal distension was initiated, aided by gravitational sagging of the tectonostratigraphically thickened uppermost crust (Roberts & Sturt 1980). It is important to remember, however, that at the same time as this NW–SE extensional regime was coming into play in tectonically higher and internal parts of the orogen, in the southeastern frontal districts the red-beds of the Oslo region were still being deposited or were just about to suffer their Scandian tectonic contraction.

REGIONAL AND OROGENIC FACTORS

The surface-crustal extension which led to the production of the ORS basins was itself influenced by two main factors, one of orogenic proportions and the other of more regional character. The regional factor pertains to the existing Scandian, post-thrusting, macrofold pattern and associated dislocations in southern Norway which, it is submitted, would be expected to have controlled the generation of the extension faults bordering the ORS molasse basins. Support for this is seen in minor dip-slip faults outside of the ORS basins, associated with the gravitational vertical-shortening stage of the Scandian orogeny: in most cases these faults strike parallel to the c. NE–SW axial trend of the macrofolds. The second factor, of a more orogen-wide magnitude, relates to the imposition of a palaeomagnetically and geologically defined wrench fault regime throughout the Caledonian–Appalachian belt (Harland 1973, Storetvedt 1973, Morris 1976, Kent & Opdyke 1979) following the final, late Silurian, cratonic suturing. The principal strike-slip megafracture here is considered to have extended from Svalbard south-southwestwards to northern Britain and into the northern Appalachians, and involved sinistral displacement of some 1500–2500 km during the Devonian period, mostly during early to mid Devonian time (Morris 1976, Ziegler 1978).

The faults bounding the west Norwegian ORS basins relate well, kinematically, to the above, first-order, sinistral mega-shear. On a theoretical basis, the E–W to ENE–WSW fractures can be viewed as second-order, right-lateral, strike-slip or oblique-slip faults (Moody & Hill 1956, Wilcox et al. 1973) splaying and propagating from the first-order structure (Fig. 3). This picture is not quite so simple, however, since the strike-slip motions were themselves secondary or subsidiary to the fundamental, intra-plate, NW–SE to NNW–SSE crustal extension; the overall kinematic regime was thus one of transtension. In this setting the individual basins would have gradually opened by combined, componental, dextral strike-slip (or oblique-slip) and extensile motions (cf. Steel & Gloppen 1980). The dip-slip extensional faults would have tended to splay, or hinge, from the master fault in the manner of listric structures as the pull-apart basin gradually widened, with the migrating depocentre producing successive E–SE onlap of cyclothems across the buried listric splays.

The southern or southeastern basin-margin faults, which are now essentially thrust contacts, were therefore extension fractures during Lower to Middle Devonian time. The subsequent tectonic inversion led to a reversal of the slip vector along these fractures during the Svalbardian contractile phase (Steel et al. 1978). The master extension faults, in some cases, could also have had a listric form judging from observations of regional structure (Bryhni & Skjerlie 1975, Bryhni 1978). This feature would have facilitated their ultimate conversion into sole thrusts, which may have interconnected across the region. Another feature of the transtensional field is that N–S extensional faults may well have developed, under favourable circumstances. The overall slip motion in these ORS basins during their growth would thus be expected to have varied between northwest and west. An interesting exercise would be to attempt to assess the amount of extensional slip involved, i.e. prior to the reversal of motion and initiation of compressional tectonics. In other mountain belts extensional displacements of fault hanging-walls over distances of several kilometres have been recorded (cf. Wernicke 1981).

OTHER DEFORMATION FEATURES

As well as the major, combined dextral-slip and extensile E–W to NE-SW faults arising from the imposed orogen-parallel sinistral mega-shear, NW–SE to NNW–SSE second-order sinistral strike-slip faults of Devonian age are theoretically feasible products of this wrench regime (cf. Moody & Hill 1956). Unlike the dextral fractures, however, these would normally not be expected

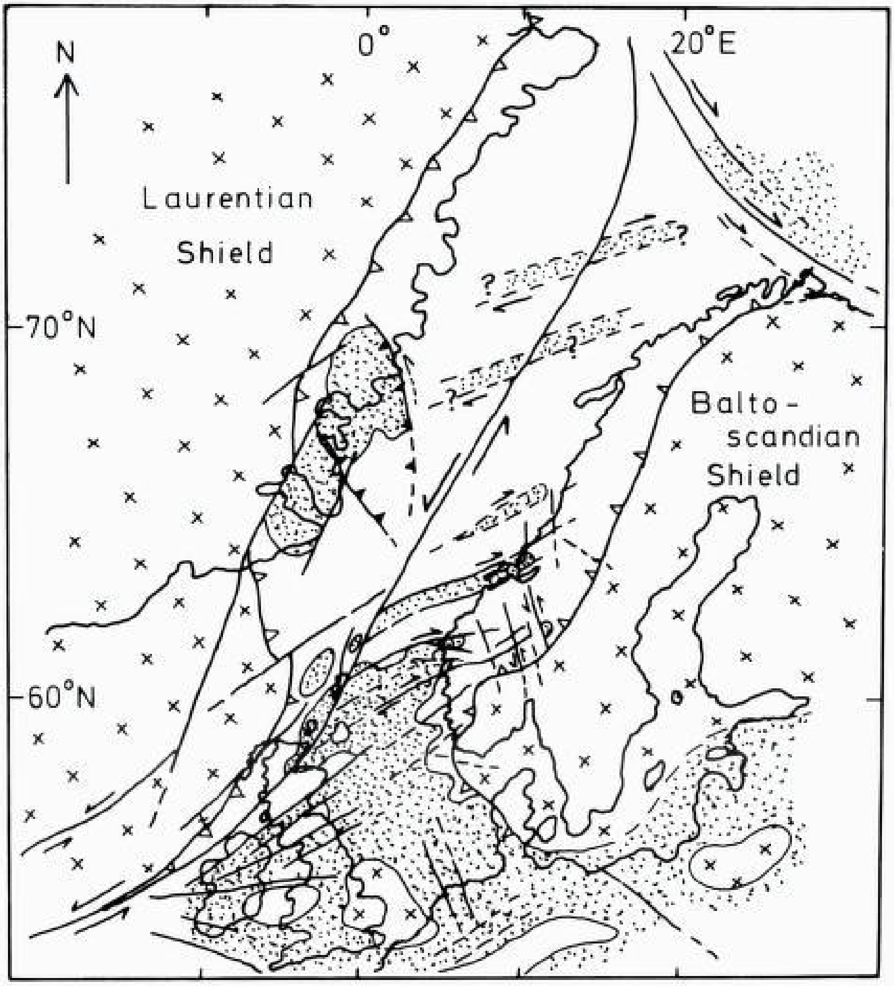

Fig. 3. Simplified map depicting the palaeographic situation and major fractures in the Balto-scandian–Laurentian–British/Irish area in approximately Middle Devonian time: modified from Ziegler (1978, fig. 3).

Ornament: *stippled* – areas of Devonian sedimentation; *crosses* – Precambrian crystalline basement; *white areas* – Caledonian fold belts. Faults are shown by thicker, full lines, with known or deduced strike-slip movement indicated by semi-arrows. Main thrusts are shown by the traditional ticked (triangled) lines. Drillcore data from the North Sea indicate that ORS sediments can be traced from the Scottish basins across to Vestlandet (Ziegler 1978).

to have opened up to produce divergent transtensional basins since they strike sub-parallel to the principal crustal extension trend. Only where fault surfaces were slightly curved, or deflected by pre-existing structures, would such divergence have been possible. Fractures of this trend do occur over large parts of southern Norway, and many carry post-Silurian displacements denoting left-lateral movement (Guezou 1981). However, some of these faults may also have been reactivated during the late Palaeozoic (Hercynian) to early Mesozoic rifting events. The only Devonian pull-apart basin associated with these NW–SE trending faults recorded in Norway is that at Røragen. Others may well have existed, the ORS sediments having since been removed by post-Caledonian erosion.

ORS sedimentation, as noted earlier, was followed by the Svalbardian deformation phase which produced major folds, local very low grade metamorphic fabrics, and inverted the basin-margin dip-slip or oblique-slip dislocations into contractile thrust-faults (Steel et al. 1978). By analogy with the early Devonian complex transtensional regime, this deformation was most likely partly transpressive rather than a simple strike-normal contraction. This is also supported by the fact that although most of the observed thrusts are younger than the fold structures (e.g. Höisæter 1971), there are cases where the thrust surfaces and mylonite zones are also deformed by folds of this same broad generation.

Conclusions

The Old Red Sandstone basinal sediments of southern Norway display a variable deformation by folds, faults, cleavages and thrusts which are part of a late-Caledonian orogenic pulse. Stratigraphic constraints within Norway indicate an Upper Devonian age for these structures; orogen-wide correlations favour a Frasnian age. However, perfect synchroneity of deformation along and across the Caledonian mountain belt is thought to have been unlikely.

Basinal development in Lower to Middle Devonian time (or slightly earlier in the northwest) occurred during the imposition of a fundamental c. NW–SE crustal extension following the major Scandian (Silurian) tectonometamorphic event. This was strongly influenced, however, by a broadly simultaneous orogen-parallel, sinistral strike-slip mega-shear from the Svalbard area south–southwestwards to northern Britain and further into the northern Appalachians. In combination with the fundamental extensile field this produced a transtensional regime with second-order dextral strike-slip faulting acting in unity with the extensional fault motions. Subsequently, in early Upper Devonian time, the transtensional field was replaced by a SE-directed compression, probably in part or even largely transpression. During this contractile stage the NE–SW to E–W trending fold structures were developed, and many of the earlier, syndepositional extension faults (either listric or planar) were rejuvenated as reverse-faults and thrusts. Mylonites were developed along some of the most ductile movement zones, and the ORS rocks and their crystalline substrates were translated east- to southeastwards as thrust sheets over distances of several kilometres and, in some cases, perhaps tens of kilometres (B. A. Sturt pers.comm.). Later movements in some areas produced rather more brittle, NW- to NNE-directed thrusting or reverse faulting, and crenulation cleavages.

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REFERENCES

Battey, M. H. & McRitchie, W. D. 1973: A geological traverse across the pyroxene-granulites of the Jotunheimen in the Norwegian Caledonides. *Norsk geol. Tidsskr. 53,* 237–265.

Beckinsale, R. D., Reading, H. G. & Rex, D. C. 1875: Pottasium-argon ages for basic dykes from East Finnmark: stratigraphical and structural implications. *Scott. J. Geol. 12,* 51–65.

Bockelie, J. F. & Nystuen, J. P. 1983: The southeastern part of the Scandinavian Caledonides. *In* The Caledonian orogen – Scandinavia and related areas (eds. Gee, D. & Sturt, B. A.) John Wiley, London, (in press.)

Bryhni, I. 1964: Migrating basins on the Old Red continent. *Nature, 202,* 384–385.

Bryhni, I. 1975: The west Norwegian basins of Old Red Sandstone. *IX. Intern. Geol. Congr. Sedimentology, Nice, Theme 5,* 111–115.

Bryhni, I. 1976: 'Gabbro' in the Solund conglomerates – a Devonian debris flow deposit? *Norsk geol. Tidsskr. 56,* 95–102.

Bryhni, I. 1978: Flood deposits in the Hornelen Basin, west Norway. *Norsk geol. Tidsskr. 58,* 273–300.

Bryhni, I. 1982: Mekanismen bak devonbassengenes mektighet? (Abstract) *Geolognytt, 17,* 16–17.

Bryhni, I. & Skjerlie, F. J. 1975: Syndepositional tectonism in the Kvamshesten district (Old Red Sandstone), western Norway. *Geol. Mag. 112,* 593–600.

Emmett, T. F. 1982: Structure and petrology of the Bergen–Jotun kindred rocks from the Gjendebu region, Jotunheimen, central southern Norway. *Norges geol. Unders. 373,* 1–32.

Fediuk, F. & Siedlecki, S. 1977: Smøla. Beskrivelse til det berggrunnsgeologiske kart 1321 I – M 1:50,000. *Norges geol. Unders. 330,* 1–23.

Furnes, H. & Lippard, J. S. 1983: Devonian lavas from Solund, west Norway – field relationships and geochemistry. *Norges geol. Unders.* (in press).

Guezou, J. C. 1981: Røros 1:250,000. Preliminary bedrock map. *Norges geol. Unders.*

Harland, W. B. 1971: Tectonic transpression in Caledonian Spitzbergen. *Geol. Mag. 108,* 27–42.

Harland, W. B. 1973: Tectonic evolution of the Barents Shelf and related plates. *Am. Assoc. Pet. Geol. Mem. 9,* 599–608.

Höisæter, T. 1971: Thrust Devonian sediments in the Kvamshesten area, western Norway. *Geol. Mag. 108,* 287–292.

Holmsen, P. 1963: On the tectonic relations of the Devonian complex of the Røragen area, east central Norway. *Norges geol. Unders. 223,* 127–238.

Indrevær, G. & Steel, R. J. 1975: Some aspects of the sedimentary and structural history of the Ordovician and Devonian rocks of the westernmost Solund islands, West Norway. *Norges geol. Unders. 317,* 25–32.

Jakobsson, K. H. 1978: *Sedimentology of the Røragen Group (Lower Devonian) in Eastern Norway.* Unpubl. thesis, Univ. of Bergen, 113 pp.

Kent, D. V. & Opdyke, N. D. 1979: The early Carboniferous palaeomagnetic field of North America and ist bearing on tectonics of the Northern Appalachians. *Earth Planet. Sci. Lett. 44,* 365–372.

Kolderup, C. F. 1925: Haasteinens devonfelt (Engl. summary). *Bergens Mus. Aarbok 1923–24, Naturv. Række 11,* 1–23.

Kolderup, C. F. 1926: Solunds devonfelt (Engl. summary). *Bergens Mus. Aarbok 1924–25, Naturv. Række 8,* 1–73.

Moody, J. D. & Hill, M. J. 1956: Wrench-fault tectonics. *Geol. Soc. Am. Bull. 67,* 1207–1246.

Morris, W. A. 1976: Transcurrent motion determined palaeomagnetically in the Northern Appalachians and Caledonides and the Acadian orogeny. *Can. Jour. Earth Sci. 13,* 1236–1243.

Naterstad, J., Andresen, A. & Jorde, K. 1973: Tectonic succession of the Caledonian nappe front in the Haukelisæter-Røldal area, Southwest Norway. *Norges geol. Unders. 292,* 20 pp.

Nilsen, T. H. 1968: The relationship of sedimentation to tectonics in the Solund Devonian district of Southwestern Norway. *Norges geol. Unders. 259,* 10 pp.

Olaussen, S. 1981: Marine incursion in Upper Palaeozoic sedimentary rocks of the Oslo region, southern Norway. *Geol. Mag. 118,* 281–288.

Priem, H. N. A., Verschure, R. H., Boelrijk, N. A. I. M. & Hebeda, E. H. 1968: Rb–Sr and K-Ar age measurements of phlogopitic biotite from the ultrabasic lamprophyre dyke on the island of Ytterøy, Trondheimsfjord, Norway. *Norsk geol. Tidsskr. 48,* 319–321.

Roberts, D. 1974: Sedimentary, tectonic and metamorphic features of the Devonian of Røragen, Sør-Trøndelag. *Norges geol. Unders. 311,* 89–108.

Roberts, D. 1981: Some features of tecronic deformation of Old Red Sandstone sediments on Hitra, west central Norway. *Norges geol. Unders. 370,* 45–48.

Roberts, D. & Sturt, B. A. 1980: Caledonian deformation in Norway. *Jour. Geol. Soc. Lond. 137,* 241–250.

Siedlecka, A. 1975: Old Red Sandstone lithostratigraphy and sedimentation of the outer Fosen area (Trondheim region). *Norges geol. Unders. 321,* 1–28.

Siedlecka, A. & Siedlecki, S. 1972: A contribution to the geology of the Downtonian sedimentary rocks of Hitra. *Norges geol. Unders. 275,* 1–28.

Skjerlie, F. J. 1971: Sedimentasjon og tektonisk utvikling i Kvamshestens devonfelt, Vest-Norge. *Norges geol. Unders. 270,* 77–108.

Steel, R. J. 1976: Devonian basins of western Norway – sedimentary response to tectonism and to varying tectonic context. *Tectonophysics, 36,* 207–224.

Steel, R. J. & Gloppen, T. G. 1980: Late Caledonian (Devonian) basin formation, western Norway: signs of strike-slip tectonics during infilling. *Spec. Publ. inr. Assoc. Sediment. 4,* 79–103.

Steel, R. J., Mæhle, S., Røe, S. L., Spinnanger, A. & Nilsen, H. R. 1977: Coarsening-upwards cycles in the alluvium of Hornelen Basin (Devonian, Norway) – sedimentary response to tectonic events. *Geol. Soc. Am. Bull. 88,* 1124–1134.

Steel, R. J., Siedlecka, A. & Roberts, D. 1978: Sedimentation in the Devonian basins of Norway – some controlling factors. (Abstract) *II Meet. Eur. Geol. Societies, Amsterdam,* 65.

Steel, R. J., Siedlecka, A. & Roberts, D. 1983: Devonian basins of Norway and their deformation. *In* The Caledonian orogen – Scandinavia and related areas. (eds. Gee, D. & Sturt, B. A.) John Wiley, London, (in press).

Storetvedt, K. M. 1973: A possible large-scale sinistral displacement along the Grest Glen Fault in Scotland. *Geol. Mag. 111,* 23–30.

Strand, T. 1972: The Norwegian Caledonides. In Scandinavian Caledonides (eds. T. Strand & O. Kulling). Wiley Interscience, London.

Sturt, B. A., Dalland, A. & Mitchell, J. L. 1979: The age of the sub-Jurassic weathering profile of Andøya, northern Norway, and the implications for the Late Palaeozoic palaeography in the North Atlantic region. *Geol. Rundschau 68,* 523–542.

Vogt, T. 1928: Den norske fjellkjedens revolusjonshistorie. *Norsk. geol. Tidsskr. 10, 97–115.*
Vogt, T. 1933: Orogenesis in the region of Palaeozoic folding of Scandinavia and Spitsbergen. *XVI Int. geol. Congr. Report,* 953–955.

Wernicke, B. 1981: Low-angle normal faults in the Basin and Range Province: nappe tectonics in an extending orogen. *Nature, 291,* 645–648.

Wilcox, R. E., Harding, T. P. & Seely, D. R. 1973: Basic wrench tectonics. *Bull. Am. Assoc. Pet. Geol. 57,* 74–96.

Ziegler, P. A. 1978: North-western Europe: tectonics and basin development. *Geol. Mijnb. 57,* 589–626.

Precambrian Stratigraphy in the Masi Area, Southwestern Finnmark, Norway

ARNE SOLLI

Solli, A. 1983: Precambrian stratigraphy in the Masi area, southwestern Finnmark, Norway. *Norges geol. Unders. 380*, 97–105.

The rocks in the Masi area consists of a comformable stratigraphic sequence containing three formations. The lowest is the Gål'denvarri formation, mainly containing metamorphic basic volcanics. Above this is the Masi Quartzite with a conglomerate at its base. The upper formation in the Masi area is the Suoluvuobmi formation containing metamorphic basic volcanics, metagabbros, mica schist, graphitic schist and albite fels. The eastern part of the area is dominated by granites which have an intrusive relationship to all three formations. Remnants of the Archean basement situated further to the east probably occur within the younger granites. Nothing conclusive can be said about the ages of the rocks, but for the Gål'denvarri formation an Archean age is considered most probable. The ages of the Masi Quartzite and Suoluvuobmi formation may be Svecokarelian, but based on correlation with rocks in Finland, Archean ages also seem likely for these units.

A. Solli, Norges geologiske undersokelse, P.O.Box 3006, N-7001 Trondheim, Norway.

Introduction

The central part of Finnmarksvidda is occupied by a dome structure of Archean granitic gneisses (Fig. 1). On each side of the dome are supracrustal rocks. To the east is the Karasjok region, and to the west is the Kautokeino–Masi region with the same types of rocks even though no direct correlation has yet been established between the two regions. Greenstones, quartzites and mica schist are the dominating rock types. A brief summary of the geology of Finnmarksvidda is given by Skålvoll (1978), and Fig. 1 mainly follows his map.

The Archean age of the central gneiss dome is indicated by a U/Pb wholerock age determination of 2800 m.y. (Skålvoll 1972, Meriläinen 1976). Just to the south of Skoganvarre (Fig. 1) the gneiss is overlain by a quartzite unit which has a conglomerate at the base with pebbles similar to the underlying gneiss. Skålvoll (1978) interprets this conglomerate as the lowest part of the Proterozoic on Finnmarksvidda. The greenstones in western Finnmark are commonly said to be of Karelian age, with ages of 1800–2000 m.y. (e.g. Oftedahl 1980), but this must be reconsidered.

The geology of the Kautokeino–Masi region is mainly known due to the work of Holmsen et al. (1957). In 1980 remapping in the Masi area started as a part of a resource investigation for the Finnmark county. The work is still in progress, but some of the new results, mainly those concerning stratigraphy within the map-sheet Masi 1:50,000 (Fig. 2), are presented here.

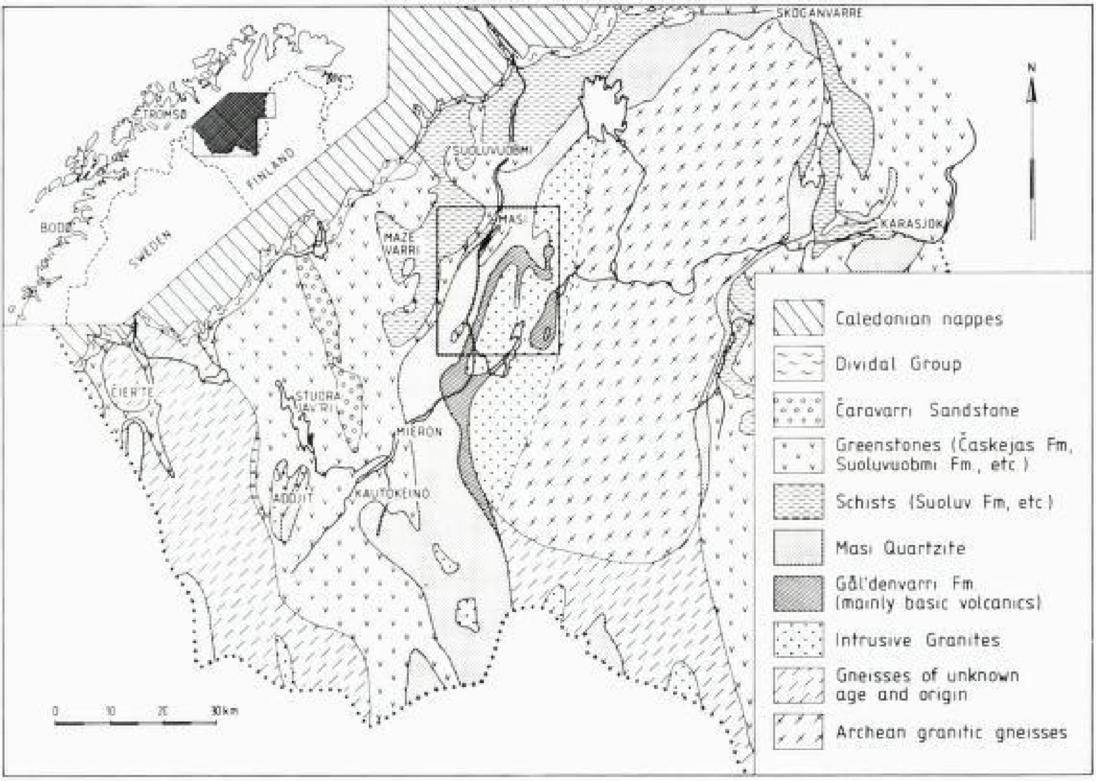

Fig. 1: Geological map of Finnmarksvidda. Location of Masi area shown by frame.

Geological setting

The lowest supracrustal formation found in the Masi area is a series of metavolcanics which is informally called the Gål'denvarri formation. It is succeeded upwards by the Masi Quartzite. This quartzite, the name of which was introduced by Holmsen et al. (1957), has a conglomerate at its base that seems to rest unconformably on the Gål'denvarri formation. The conglomerate has pebbles of granitic gneiss and has a strong resemblance to the conglomerate at Skoganvarre. According to new field work in the area between Skoganvarre and Masi there can be little doubt on the correlation of these two conglomerates. To the west and stratigraphically above the Masi Quartzite is the Suoluvuobmi formation containing mainly basic metavolcanics and mica schists. The Suoluvuobmi formation is probably an equivalent to the Časkejas Group of Holmsen et al. (1957). The latter term has been used as a loose designation for all basic metavolcanics on western Finnmarksvidda, but its type area is around Bied'djuvaggi Mine (Hagen 1982), west of the Čarravarri Sandstone (Fig. 1). Since the correlation between rocks there and in the Masi area is not obvious, the new term Suoluvuobmi formation is introduced. As may be seen from the map, Fig. 1, no attempt is made to differentiate between the regional distribution of these two units.

The eastern part of the Masi area is occupied by granite rocks. Formerly, these were thought to belong to the Archean basement, but the new mapping has shown that they are younger intrusives. Remnants of the basement may, however, be preserved in the granites.

Gål'denvarri formation

The rocks of the Gål'denvarri formation were included in the Časkejas Group by Holmsen et al. (1957), but more detailed mapping has clearly shown its position below the Masi Quartzite. It is mainly composed of amphibolites which are interpreted as metabasalts. A faint layering may be seen in the amphibolites, mainly due to differences in grain size and mineral composition, but also some porphyritic layers and schist horizons occur. Primary volcanic structures are difficult to recognize; most convincing is a fragmental amphibolite, interpreted as a volcanic breccia. It is found at several localities, and seems to occur at different levels in the stratigraphy. Some ultrabasic rocks occur in very elongated lenses, up to 30–40 m thick and 2–300 m long. The main minerals are amphibole and chlorite, but also olivine is found. Total thickness of the formation is about 1–1,5 km.

Chemically the amphibolites are classified as tholeiitic basalts. Characteristic is a low TiO_2 content (mean 0.5% for 16 samples) compared to the amphibolites of the Suoluvuobmi formation (mean 1.5%, 21 samples) which are also tholeiitic. Also characteristic is a relatively high MgO content (mean 12.3%) in relation to the SiO_2 content (mean 52.5%). In the Suoluvuobmi formation these concentrations are 8.5% and 47.3%, respectively. There are also significant differences in trace elements between the two formations, especially in Cr and Ni contents which are much higher in the Gål'denvarri formation.

Schists are a minor component of the Gål'denvarri formation, but in some of these layers the minerals garnet, staurolite and sillimanite are found. The paragenesis cordierite–anthophyllite is found both in basic rocks and in quartz-rich schist. Except for garnet, none of the other minerals is found elsewhere in the Masi area. Many of the minerals listed above are typical for thermal metamorphism, and the apparent higher grade of metamorphism in the Gål'denvarri formation could be ascribed to contact metamorphism from the intrusive granite to the east. However, except for sillimanite the mineralogical textures indicate that the minerals are formed at an early stage by regional metamorphism.

The lower parts of the Gål'denvarri formation are invaded by younger granites and the base of the formation is not found. At one single locality east of Gål'denvarri (Fig. 2) a poorly sorted conglomerate occurs as the lowest exposed member of the formation. Well rounded pebbles of granitic gneiss ranging in size up to 30 cm occur rather scattered in an arkosic matrix. This conglomerate may represent a basal layer on an old gneiss basement, even if there is little evidence for such an assertion.

Masi Quartzite

The Masi Quartzite is a relatively thick and homogeneous formation of widespread distribution, and may represent one of the most important formations regarding correlation of geological units in different parts of Finnmarksvidda.

The quartzite has a white, grey-white or pink colour and commonly shows a fine lamination caused by differences in feldspar content or heavy mineral dust. It is recrystallized so that primary sedimentary grains are usually not seen. Cross-bedding is observed at several localities. Much of the Masi Quartzite could be classified as a feldspathic quartzite. Total thickness is about 500–1000 m.

At the base of the Masi Quartzite there is usually a conglomerate with pebbles of granitic gneiss and quartz. Other types of pebbles are rarely found, even where the conglomerate rests directly on the amphibolites of the Gål'denvarri formation. The conglomerate has a calcareous arkosic matrix. The pebbles are normally less than 5 cm in diameter, but at several localities, especially in the northern areas, pebbles up to 30 cm in diameter are observed.

Another conspicuous feature of the Masi Quartzite is the occurrence of the chrome-bearing mica fuchsite. It is especially abundant in impure zones, but it may also be evenly distributed in the quartzite, giving it a green colour. Fuchsite is a common mineral in sandstones and quartzites of Archean age, and may be formed as an alteration product after detrial grains of chromite (Schreyer 1982).

Suoluvuobmi formation

The Suoluvuobmi formation is dominated by amphibolites and mica schists. The amphibolites appear to be of two types (Fig. 2), which may be of different origin. One type is relatively fine grained, dark green and foliated. Although no primary structure are preserved, this rock is thought to represent metabasalt. Another rock variety that probably originally was a lava, is an ultrabasic rock, now consisting of amphibole and chlorite.

The other type of amphibolite has a gabbroic texture. In the field the boundaries of this type seem to be parallel to the layering of other rocks, but on a regional scale they appear to be discordant. These amphibolites are therefore thought to represent metadolerites or gabbros. The geochemistry of both kinds of amphibolite, however, is identical with regard to both major and trace elements, so they may be closely related. This type of gabbro seems to be limited to the Suoluvuobmi formation.

Mica schist is the other dominant rock type. It is often characterized by a porphyroblastic growth of biotite, the dominating mica. The total thickness of the formation is difficult to estimate, but is certainly more than 1 km.

Two other characteristic rocks of the Suoluvuobmi formation are graphitic schists and a rock that in Finnmark has traditionally been called albite fels or quartz-albite rock. These two rock types are often connected and seem to

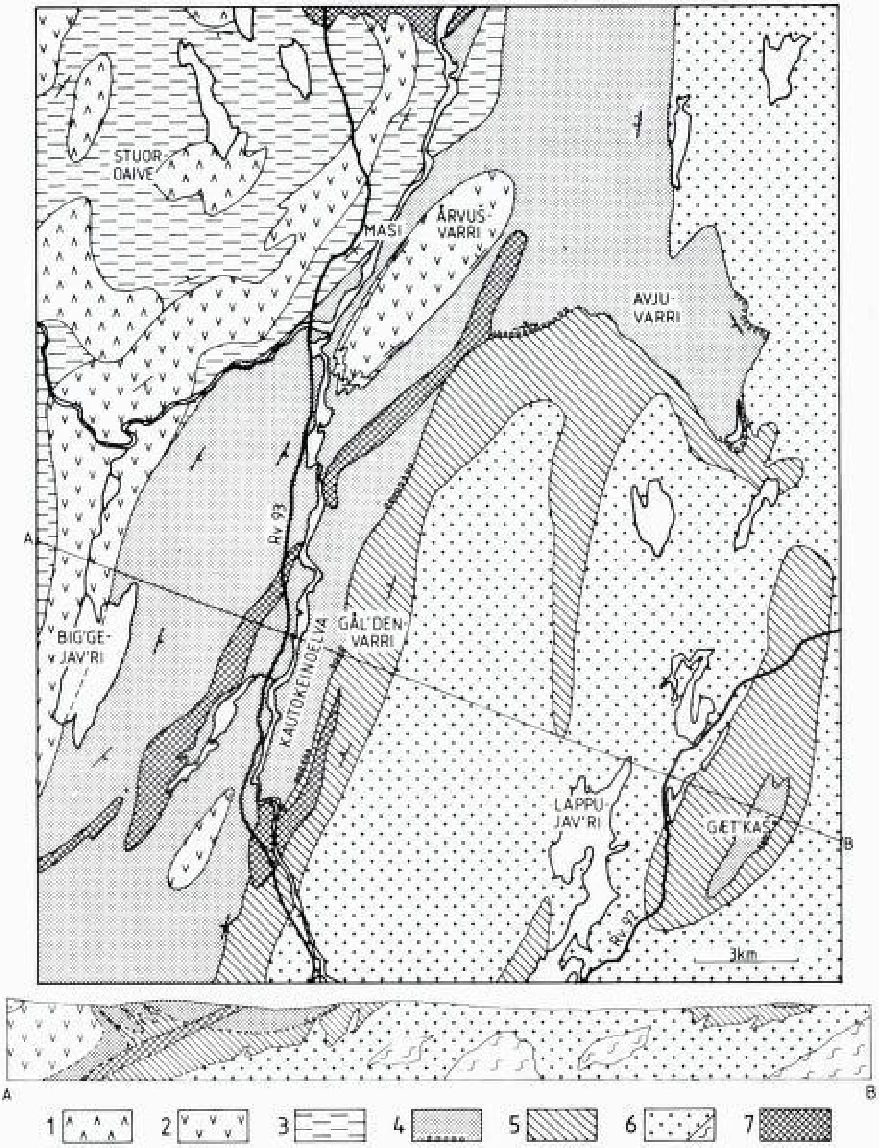

Fig. 2: Simplified geological map of the Masi area.
1–3: Suoluvuobmi formation; 1. Metagabbro. 2. Metabasalt. 3. Mica schist. 4. Masi Quartzite. 5. Gål'denvarri formation; Metabasalt. 6. Intrusive granite. Remnants of Archean gneisses are indicated. 7. Albite diabase.

have a preferred occurrence in the contact zone between amphibolite and mica schists. They are of special interest because there are ore minerals associated with them. On Finnmarksvidda they are particular well known from the Bied'djuvaggi Mines (Hagen 1982), but they are also known from mines in northern Sweden and Finland (Inkinen 1979). The origin of these rocks has been discussed for a long time, and both a metasomatic and a volcanic origin has been proposed; the latter is favoured by the present author.

The contact relations between the Suoluvuobmi formation and the underlying Masi Quartzite are generally very complicated. The Masi Quartzite is locally inverted and thrust above the Suoluvuobmi formation, especially in the area south of Masi, and it is easy to get the impression of an important break between these two formations. At other localities like the Årvusvarri syncline east of Masi (Fig. 2) and east of Mazevarri (Fig. 1) where the Masi Quartzite is deformed in an antiform beneath the Suoluvuobmi formation, there seems to be a normal stratigraphic sequence from the Masi Quartzite up to the Suoluvuobmi Formation. This relationship is further supported from evidence in the Addjit area to the south (Fig. 1) where a probable equivalent to the Masi Quartzite underlies the geenstones of the Časkejas Group (Holmsen et al. 1957).

Basement and granite intrusions

The geological map, Fig. 2, gives the impression of a concordant boundary between the Gål'denvarri formation and the granite to the east. However, at several localities it can be demonstrated that the granite has an intrusive relationship to both the Gål'denvarri formation and the Masi Quartzite. Granitic intrusions are not common in the Suoluvuobmi formation, but at least two larger massifs occur (Holmsen et al. 1957; and Fig. 1), and veins are observed at several localities. There are also many xenoliths, especially of the Gål'denvarri formation, in the granite. These relations seems to be even more abundant to the south of the Masi area, and were also recognized by Holmsen et al. (1957).

Large parts of the granite are a rather homogeneous red to white biotite granite; but within the granite there are also areas of strongly foliated rocks of granitic, granodioritic and tonalitic compostion. These gneisses are thought to be remnants of the Archean basement; and it seems that in the Masi area there was originally a gneissic basement which was later intruded by granite. Conclusive proof that the gneisses are xenoliths is, however, difficult to find, both due to the general lack of exposures and because the intruding granite also has a marked foliation. An important reason for the suggestion of the gneisses being basement is therefore the situation in Skoganvarre, where a gneissic basement for the supracrustals is present. Since the intrusive granite is so dominant in the Masi area, only this is shown on the map (Fig. 2). The boundary between the granite and real Archean basement shown in Fig. 1 is not known in detail.

Albite diabase

Albite diabase is an intrusive rock commonly occurring in the Masi area as well as in the Precambrian of northern Finland and Sweden. Meriläinen (1961) gave a complete description of the rock type. It is characterized by preserved igneous texture and a high content of albite (up to 60%) that seems to be of primary origin. Dark minerals are generally biotite and hornblende, but also clinopyroxene may occur. Magnitite, up to 10%, is present; this make the rock easy to recognize from aeromagnetic maps. The high albite content also produces a high Na_2O content (mean nearly 7%), and this is probably why the albite diabases have been confused with the albite felses that also commonly occur on Finnmarksvidda, but these rocks seeem to have a totally different origin.

Discussion and regional correlations

Mapping in the Masi area has shown that the basic metavolcanics are not restricted to one single unit in the stratigraphy on Finnmarksvidda. They occur in the Gål'denvarri formation below the Masi Quartzite and in the Suoluvuobmi formation above. Recent mapping to the west of Mazevarri (Fig. 1) has also revealed a low-metamorphic greenstone sequence where primary volcanic structures are well preserved. This greenstone has a distribution east of and parallel to the Čaravarri Sandstone, and could represent an even younger sequence than the Suoluvuobmi formation. Low-metamorphic greenstones are also found to the west of and below the Čaravarri Sandstone, which is thought to represent the youngest rock unit of the Precambrian on Finnmarksvidda.

The correlation between the Kautokeino–Masi region and the tectonic windows of Alta–Kvænangen and Komagfjord about 50–100 km to the north is based on both geophysical (Åm, 1975) and lithological evidence. The section from the Kvænvik Greenstone up to the Skoadduvarri Sandstone at Alta (Zwaan & Gautier 1980) corresponds very well with the section from the low-metamorphic greenstone up to the Čaravarri Sandstone on Finnmarksvidda.

Even if no direct correlations can be made between Finnmarksvidda and northern Finland, very similar rocks occur there. It is mainly based on this long distance correlation that the basic metavolcanics in western Finnmark are considreed to be of Karelian age (e.g. Oftedahl 1980, Zwaan & Gautier 1980). However, the volcanics in Finland vary in age from Archean to Proterozoic (e.g. Silvennoinen et al. 1980), and since it is now apparent that the volcanics within the Masi area are not restricted to one single unit, differences in age are considered to exist here too.

A short comment should here be given on the term Karelian, which has been used for the rock sequences in western Finnmark. In Norway, Karelian until recently has been used for the period from 1800 m.y. to 2000 m.y.; see for instance Oftedahl (1980) and Zwaan & Gautier (1980). This is not in accordance with modern Finnish use. Originally, Karelian was used for rocks in the eastern and northern part of Finland which were thought to represent a younger orogenic belt than the Svecofennian in western Finland. Age determinations and later geological investigations have shown that rocks of the Karelian and Svecofennian orogenic belts are of the same age, and they have been renamed the Svecokarelian orogeny (Gaal 1982). The most important metamorphic episode in this orogeny occurred about 1800–1900 m.y. ago (Simonen 1980), but the depositional age of rocks involved seems to have been from 2600 to 1800 m.y. (Simonen 1980, Silvennoinen et al. 1980).

Even if a direct correlation with the rock sequences in the Masi area and Finland should be made with great care, there seem to be two reasonable possibilities for correlation. The key horizon for the first possibility is the conglomerate at the base of the Masi Quartzite. In Finland it is well documented that the beginning of the Proterozoic was a period of denudation and peneplanation of the Pre-svecokarelian basement (Meriläinen 1980, Silvennoinen et al. 1980, Simonen 1980). This has resulted in widespread occurrences of conglomerates resting on the basement, which contained rocks like granitic gneisses, greenstones and sediments. The most obvious unit to associate with such an unconformity on Finnmarksvidda would be the conglomerate at the base of the Masi Quartzite, resting partly on the Gål'denvarri formation as in the Masi area, and partly on a gneissic basement as at Skoganvarre. This correlation will be in accordance with that of Skålvoll (1976) and imply that the Gål'denvarri formation is of Archean age. The Masi Quartzite and Suoluvuobmi formation will thus have a depositional age from about 2500 m.y. and younger, and have been deformed in the Svecokarelian orogeny. K-Ar determinations on basic metavolcanics in the Komagfjord tectonic window confirm that there has been a metamorphic event at 1800 m.y. in Finnmark (Pharaoh et al. 1982). As will be seen from stratigraphic sections from Finland (e.g. Meriläinen 1980, Silvennoinen et al. 1980) there are quartzites, greenstones and schists of the same depositional age that could be correlated with the Masi Quartzite, Suoluvuobmi formation and other rocks in western Finnmark.

If the correlations suggested above are accepted, the bulk of the rocks on Finnmarksvidda will be of Proterozoic age. The only Archean rocks, except for the central gneiss dome, would be the Gål'denvarri formation, which seems to have a very limited distribution (Fig. 1). Recently, however, the Proterozoic age of the supracrustal rocks on Finnmarksvidda has been questioned by several workers in the Finnish Precambrian, because radiometric age determinations have proven an Archean age for the Kittilä greenstone (Gaal et al. 1978, Barbey et al. 1981, Gaal 1982). A correlation of the Kittilä greenstone with the greenstones of the Karasjok region, and possibly also those of the Kautokeino–Masi region, seems obvious from existing geological maps. It is also easy to see many similarities between the stratigraphy of the Masi area and the Archean stratigraphy in this part of Finland. According to Rastas (1980) the lowest supracrustal unit above the gneissic basement in the Kittilä area is a volcanic formation, mainly comprising lavas. Above this is a sericite quartzite, occasionally fuchsite-bearing, and this is followed by the Kittilä greenstones with mica schist and graphitic schists. The lithological correlation with the sequence gneissic basement, Gål'denvarri formation, Masi Quartzite and Suoluvuobmi formation is a good one, and if this is correct it will mean that the only Proterozoic rock unit on Finnmarksvidda is the Čaravarri Sandstone.

The question as to which of the two correlations outlined above is correct, will remain open until age determinations are carried out and more mapping is done. The possibility of an Archean age for the greater part of the rocks on Finnmarksvidda should, however, be kept open.

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REFERENCES

Åm, K. 1975: Aeromagnetic basement complex mapping north of latitude 62 °N, Norway. *Norges geol. Unders. 316*, 351–374.

Barbey, P., Convert, J., Martin, H., Moreau, B., Capdevila, R. & Hameurt, J. 1980: Relationships between granitegneiss terrains, greenstone belts and granulite belts in the Archean crust of Lapland (Fennoscandia). *Geol. Rundsch. 69*, 648–658.

Gaal, G., Mikkola, A. & Søderholm, B. 1978: Evolution of the Archean crust in Finland. *Precambrian Res. 6*, 199–215.

Gaal, G. 1982: Proterozoic tectonic evolution and late Svecokarelian plate deformation of the Central Baltic Shield. *Geol. Rundsch. 71*, 158–170.

Hagen, R. 1982: The Bidjovagge Copper–Gold Deposits of western Finnmark, Norway. *Geol. Rundsch. 71*, 94–103.

Holmsen, P., Padget, P., Pehkonen, E. 1957: The Precambrian geology of Vest-Finnmark, Northern Norway. *Norges geol. Unders. 201*, 106 pp.

Inkinen, O. 1979: Cu, Zn, and U occurrences at Pahtavouma in the Kittilä Greenstone Complex, Northern Finland. *Econ. geol. 74*, 1153–1165.

Meriläinen, K. 1961: Albite diabases and albitites in Enontekiö and Kittilä, Finland. *Bull. Comm. géol. Finlande 195*, 77 pp.

Meriläinen, K. 1976: The granulite complex and adjacent rocks in Lapland, northern Finland. *Geol. Surv. Finland, Bull. 281*, 129 pp.

Meriläinen, K. 1980: Stratigraphy of the Precambrian in Finland. *Geol. Fören. Stockh. Förh. 102*, 177–180.

Oftedahl, Chr. 1980: Geology of Norway. *Norges geol. Unders. 356*, 3–114.

Pharaoh, T. C., Macintyre, R. M. & Ramsay, D. M. 1982: K-Ar age determination on the Raipas suite in the Komagfjord Window, northern Norway. *Norsk geol. Tidsskr. 62*, 51–57.

Rastas, P. 1980: Stratigraphy of the Kittilä area. In Silvennoinen, A. (ed.): Jatulian geology in the eastern part of the Baltic Shield. (Proceedings). The Committee for Scientific and Technical Cooperation between Finland and the Soviet Union, Rovaniemi, 145–151.

Schreyer, W. 1982: Fuchsite-aluminium silicate rocks in Archean greenstone belts: Are they metamorphosed alunite deposits? *Geol. Rundsch. 71*, 347–360.

Silvennoinen, A., Honkamo, M., Juopperi, H., Lehtonen, M., Mielikäninen, P., Perttunen, V., Rastas, P., Räsänen, J. & Väänänen, J. 1980: Main features of the stratigraphy of North Finland. In Silvennoinen, A. (ed.): Jatulian geology in the eastern part of the Baltic Schield. (Proceedings) The Committee for Scientific and Technical Cooperation between Finland and the Soviet Union, Rovaniemi, 153–162.

Simonen, A. 1980: The Precambrian in Finland. *Geol. Surv. Finland, Bull. 304*, 58 pp.

Skålvoll, H. 1972: Geologisk kart over Norge – berggrunnskart Karasjok – 1:250000. *Norges geol. Unders.*

Skålvoll, H. 1978: Berggrunnsgeologi (with map enclosure). In Finnmarksvidda – natur – kultur. Norges off. utredn. 18A, Universitetsforlaget, Oslo. 35–39.

Zwaan, K. B. & Gautier, A. M. 1980: Alta og Gargia. Beskrivelse til de berggrunnsgeologiske kart 1834 I og 1934 IV, M 1:50000. *Norges geol. Unders. 357*, 47 pp.

A Fluidization Breccia in Granite at Skaget, Svellingen, Frøya

TORE TORSKE

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The breccia consists of fragmented and comminuted host rock material and small amounts of hydrothermal quartz, epidote, laumontite, ?stilbite, calcite and montmorillonite. At least three brecciation and mineralization events have occurred. The breccia is considered to have formed by hydraulic fracturing by a mixed, high-pressure CO_2–H_2O fluid. Upon explosive venting to the surface, the fluid is thought to have fluidized and fragmented the fractured rock. The carbon dioxide of the fluid may have originated in the mantle, and could have ascended through the crust by hydraulic fracture propagation. It is suggested that the occurrence within the same region of the Skaget Breccia, Tertiary olivine-nephelinite plugs, and scattered zeolite mineralization may possibly define a palaeo-geothermal province.

Tore Torske, Institutt for Biologi og Geologi, Universitetet i Tromsø, Postboks 3085, N-9001 Tromsø, Norway.

Introduction

A small breccia body cutting through granite at Skaget, Svellingen, Frøya (UTM co-ordinates NR 912752; Fig. 1) was described by Keilhau (1850) and mentioned by Reusch (1914), but has not been treated in newer literature. It occupies the tip of a small, pointed headland called Skaget (Figs 2 & 3). On land, the Skaget Breccia covers about 5000 m², but is surrounded by the sea on three sides; accordingly, its full extent and shape are unknown. In addition to the main breccia outcrop, there are a few, scattered, small occurrences of breccia over a distance of a couple of hundred metres southward along the coast.

Keilhau (1850) regarded the breccia as an epiclastic sedimentary rock, and interpreted its gradational boundary with the surrounding granite in terms of his neptunistic ideas: as evidence for the transformation of a sedimentary rock into granite. Tre present study indicates that the rock was formed by fragmentation of the host granite and fluidization of the fragments by streaming gas.

Regional and field geology

Host rock to the Skaget Breccia is the Frøya Granite, a large, homogeneous pluton of granitic to granodioritic composition. It covers the northern part of Frøya and the adjoining archipelago, including the Sula islands (Fig. 2). The granite is intrusive into migmatites and gneisses of the Northwestern Gneiss

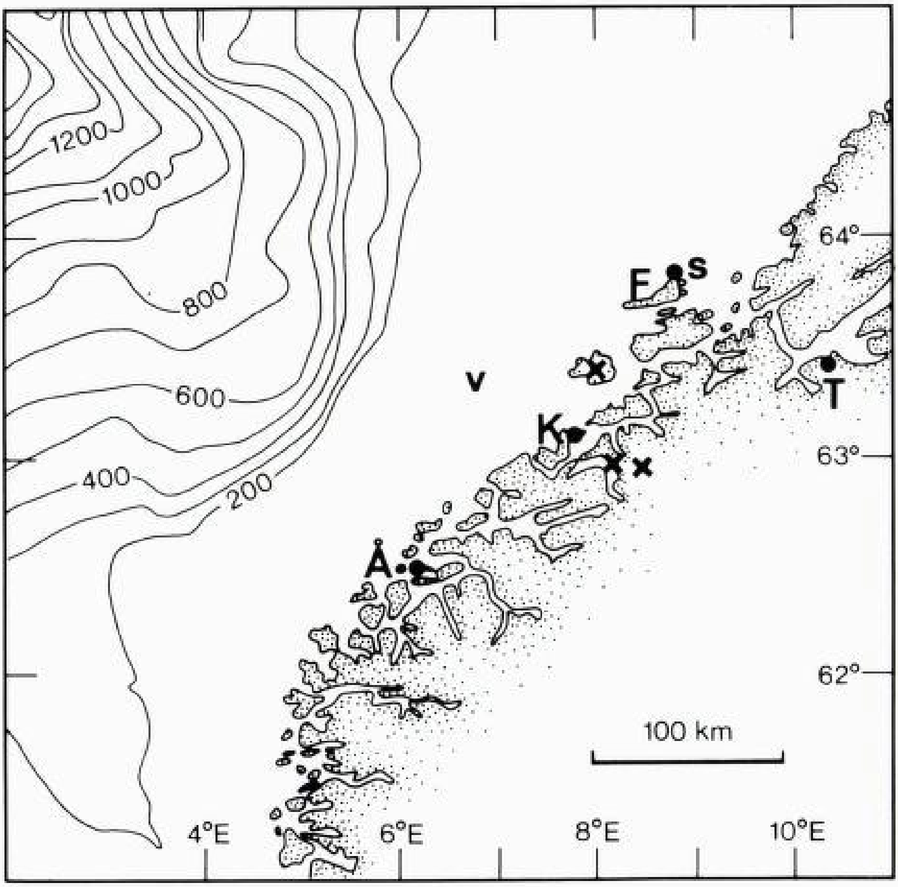

Fig. 1. Location map of west-central Norway. S: Skaget; F: Frøya; V: Olivine nephelinite plugs; crosses: known zeolite occurrences. K: Kristiansund; T: Trondheim; Å: Ålesund. Isobaths in fathoms.

complex of the Norwegian Caledonides (Strand 1960, Oftedahl 1981). In places, the gneisses contain small bodies of partly silicified marble (Reusch 1914). A provisional, model Rb.-Sr dating gave a Caledonian minimum age for the Frøya Granite (B. Sundvoll, pers. comm.).

The Skaget Breccia occurs well inside the pluton, about four kilometres from the nearest, steeply dipping contact between the Frøya Granite and the gneisses adjoining it to the south. It consists of fragments and matrix of comminuted host-rock material and secondary hydrothermal minerals; predominantly quartz, minor epidote, and small amounts of Ca-zeolites, montmorillonite and calcite.

There are two, texturally different types of breccia (Fig. 3): a dark, unsorted, granite breccia (Fig. 4) and a light silicic microbreccia (Fig. 5a, b). The latter forms a dyke-like sheet along part of the boundary between the main, granite breccia and the host granite (Fig. 3). Its emplacement post-dated that of the local granite breccia, because it cross-cuts a small portion of the main breccia body. On the other hand, quite similar microbreccia occurs

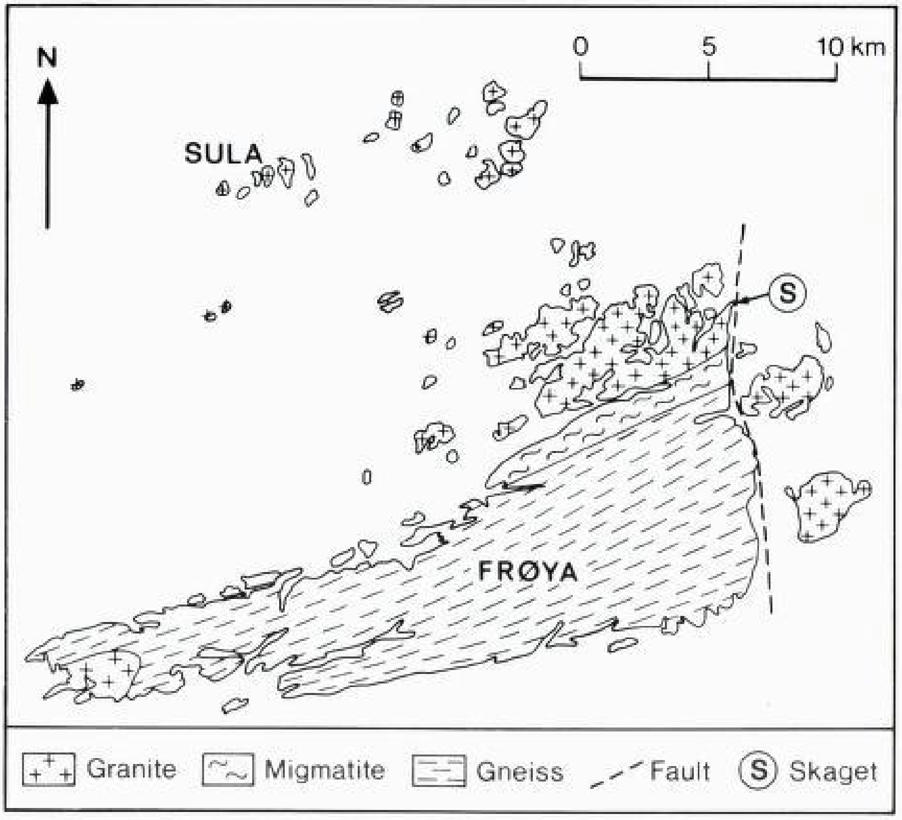

Fig. 2: Bedrock geology of Frøya. Redrawn after Askvik (1979). S: Skaget.

as fragments in the granite breccia near the boundary between the two breccia types. The microbreccia sheet is sharply delimited, up to 12 m thick, and dips about 35° ESE. This orientation parallels that of a joint system in the bordering granite. The limited exposure of the granite breccia precludes an assessment of the overall shape and extent of this variety.

Petrography and mineralogy

The Skaget Breccia and ist surroundings may be described in terms of three main rock types: 1) the host granite; 2) the granite breccia; and 3) the microbreccia.

1) *The Frøya Granite* is a massive, medium to coarse-grained, red, biotite granite to granodiorite. In places it is semi-porphyric, with up to several centimetres long, stubby megacrysts of brick red microcline. The megacrysts may locally have a crudely defined parallel orientation. Sparse and scattered, dark, thin, long, biotite schlieren have the same orientation as the feldspar megacrysts, and are probably strongly flattened and partly digested xenoliths.

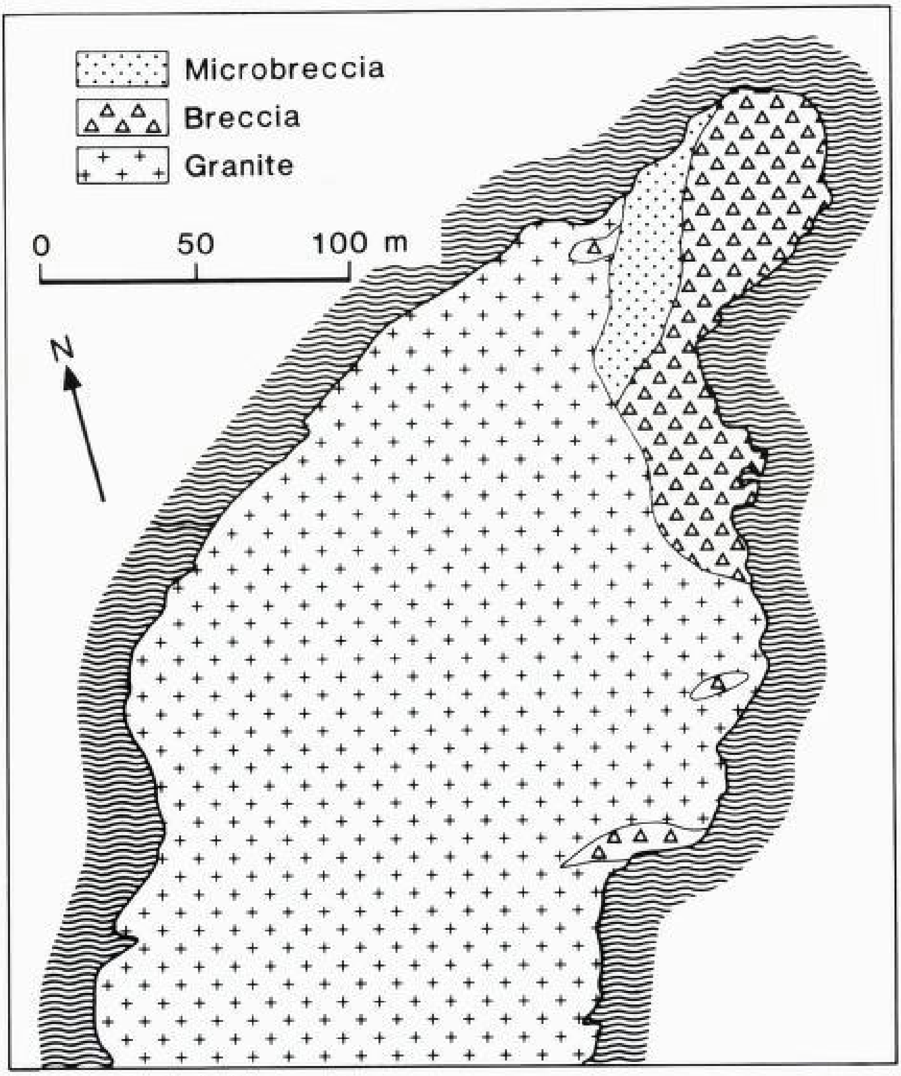

Fig. 3. Geological sketch map of Skaget.

Some granite pegmatite dykes have up to 10 cm long, red, euhedral K-feldspar crystals, much quartz, and subordinate plagioclase. Pegmatites with smaller grain sizes form narrow, straight and parallel-walled dykes, a few centimetres wide and up to some tens of metres in length. The granite has a variably developed parallel jointing, with joints spaced from a few decimetres to a couple of metres apart. They appear to be oriented more or less perpendicular to the indistinct foliation, defined by parallel oriented feldspar megacrysts and biotite schlieren.

Along its external border, against the gneiss complex to the south, the Frøya pluton is associated with a zone of migmatite. There, more or less irregular dykes and veins of massive red granite, muscovite-bearing aplite, and pegmatite invade the older gneisses.

Fig. 4. Granite breccia, Skaget. Subangular to rounded granite fragments in unsorted matrix. Coin is 2.5 cm across.

The texture of the typical Frøya Granite is hypautomorphic-granular (Johannsen 1939); with subhedral to anhedral, predominantly unstrained microcline and albite-oligoclase crystals, unoriented flakes of biotite, and intersertal, granular aggregates of fine- to medium-grained, anhedral, strained quartz. Sphene is the predominant accessory mineral, but apatite is also very common. In addition, there are scattered grains of zircon and allanite. The plagioclase may be slightly sericitized, but has apparently not been epidotized in the typical Frøya Granite. Biotite may show incipient chloritization.

In the nearest several metres to the breccia boundary, the Frøya Granite is strained, fractured and hydrothermally altered. The characteristic deformation features are: strain effects in plagioclase, like undulatory extinction and bent or kinked twin lamellae; and small-scale fracturing and veining by microbreccia and quartz, as well as by laumontite. The effects of hydrothermal alteration are: complete chloritization of biotite, minor adularization as well as partial epidotization of plagioclase, and local dissolution of primary, strained quartz and subsequent deposition within the same space of unstrained, subhedral quartz grains.

In many cases chlorite flakes bordering on plagioclase are accompanied by a K-feldspar rim in the latter, in such a way that the two secondary minerals together form a composite pseudomorph after biotite (Fig. 5 c). From the restricted scale of this type of plagioclase alteration, and from the apparent non-participation of the primary K-feldspar of the granite in these secondary reactions, it may be inferred that potassium for the secondary K-feldspar in

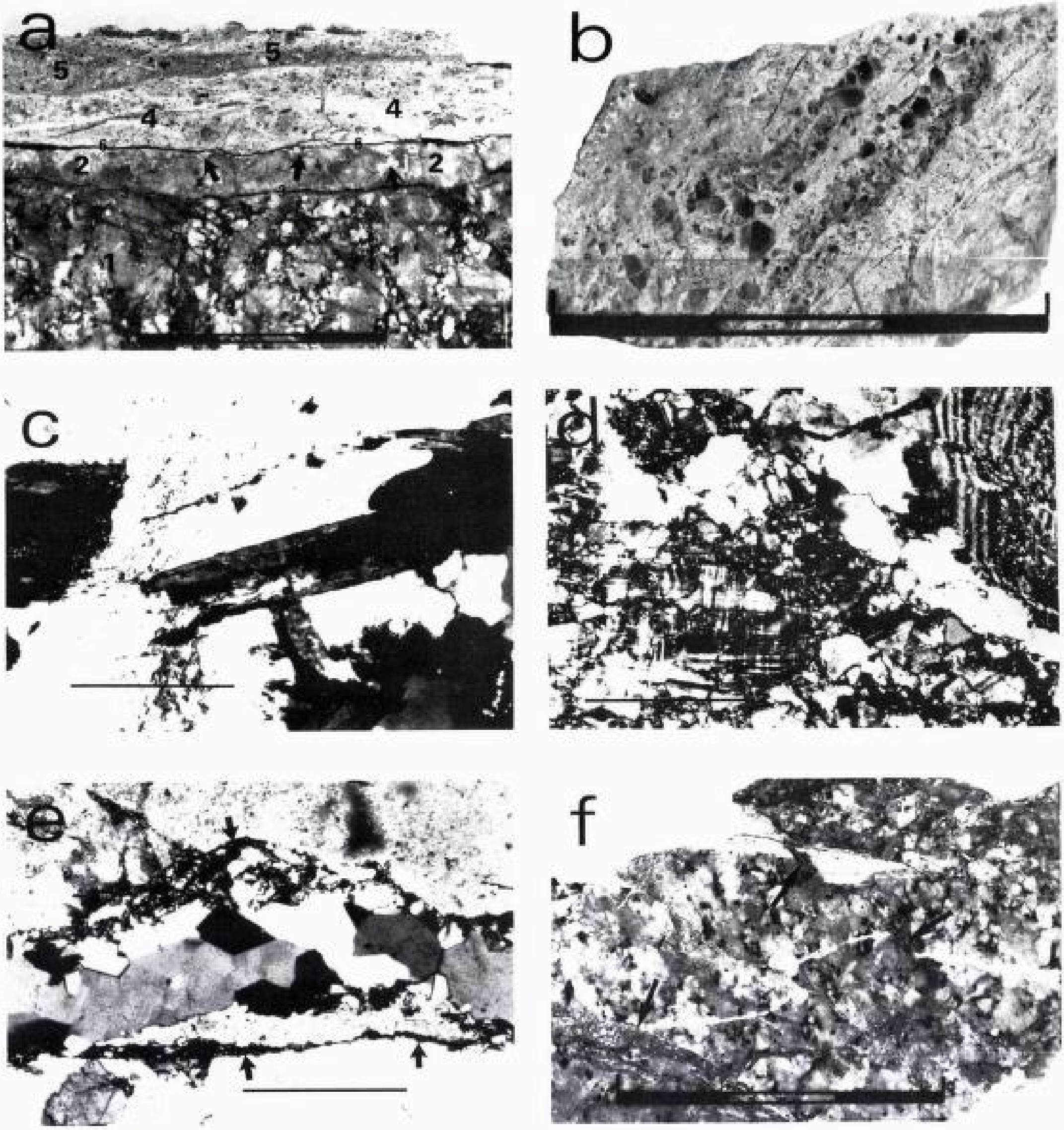

Fig. 5(a): Border between host granite (below) and micro breccia: 1, microfractured granite; 2, quartz-free rim, consisting of highly strained feldspar (cf., Fig. 5 d), and with microbreccia-filled embayments (arrows); 3, younger quartz vein (cf., Fig. 5 e); 4 and 5 are two generations of microbreccia; crack (6) is sampling artifact. Polished slab. Bar scale in cm.

(b) Microbreccia with train of rock crystal fragments. Polished slab. Bar scale in cm.

(c) Adularia (medium grey) and chlorite (dark) pseudomorph after biotite in plagioclase (light). Frøya granite. Planepolarised light. Bar scale = 0.5 mm.

(d) Severely strained plagioclase and microcline in feldspathic rim in granite bordering against microbreccia (cf., Fig. 5a). Crossed nicols. Bar scale = 0.5 mm.

(e) Quarts-filled fracture in granite (cf., Fig. 5a). Note euhedral outlines of some quartz grains, and older microbreccia-filled fractures (arrows). Crossed nicols. Bar scale = 0.5 mm.

(f) Laumontite veins (light) in granite offset by younger microbreccia-filled fractures (arrows). Polished slab. Scale in cm.

plagioclase was derived mainly from the breakdown of local biotite, in general accordance with the process proposed by Chayes (1955). Most K-feldspar patches in plagioclase are turbid; they have very low birefringence and small but indeterminate, apparently variable $2V_X$, indicating adularia; and thus pointing to hydrothermal alteration of the affected plagioclase. Epidote in plagioclase consists mostly of fairly large grains, frequently only one large epidote grain per plagioclase host grain, rather unlike the fine-grained products of typical saussuritization.

The strain effects in plagioclase and quartz give evidence of ductile deformation of these minerals. Microcline appears to be unaffected by this, except at the immediate contact with the microbreccia (Figs. 5a, d). There, a feldspathic rim, about 5 mm thick, shows evidence of intensive ductile as well as brittle strain. This, together with intensive micro-fracturing, indicates that the deformation behaviour of the granite in the contact zone was near the brittle-ductile transition field. The thin rim with highly strained feldspar appears to have lost its primary quartz content by dissolution. It is embayed by microbreccia-filled solution pits and cracks (Fig. 5 a), and invaded by younger, unstrained veins of hydrothermal quartz (Figs 5 a, e).

Four types of microfracture occur in the altered granite and in larger fragments in the breccia: a) healed fractures (Batzle & Simmons 1976), where fractured mineral grains have recrystallized across the fracture; b) sealed fractures (*ibid.*), where new mineral grains have crystallized within the fracture (Figs 5 e, f); c) filled fractures, where minute fragments, forming a microbreccia, occupy the fracture space (Fig. 6 a); and d) closed fractures, where little or no material intervenes between the fracture walls; these are in immediate contact, and only small offsets along the crack or, for instance, a thin coating of iron oxide delineate the fracture. Any particular fracture may change type one or several times along its extent. The fracture-sealing mineral is predominantly quartz (Fig. 5 e), often with small amounts of epidote, but laumontite veins have also been identified (Fig. 5 f). Some laumontite veins are cut by thin, microbreccia-filled fractures (Fig. 5 f). Calcite is a very subordinate vein mineral.

2) *The granite breccia* forms the major part of the Skaget occurrence. It consists of more or less rounded, unsorted, granite clasts and mineral fragments, interspersed with scattered fragments of light-coloured microbreccia and vein quartz; all set in a fine-grained to dense, greyish brown to green groundmass. The granite clasts are mostly less than 10 cm, but may reach 50 cm in size. Clasts of microbreccia are mostly less than 2 cm across.

In a few places, the granite breccia has irregular, locally anastomosing, steep, tubular or podiform cavities, up to 2 cm in diameter and possibly up to several tens of centimetres long. They are either empty, or filled with pasty, white, pure montmorillonite. Some are lined with a thin layer of zeolite or quartz and calcite.

The border between granite breccia and host granite is diffuse and irregular: there is a gradational transition over one to several metres from large blocks of granite in situ, partially surrounded by thin veins of granite breccia, to matrix-supported breccia with rounded, smaller fragments. The boundary between the granite breccia and the microbreccia, on the other hand, is a well-defined, sharp, straight to smoothly undulating contact.

The fragments in the granite breccia vary in size from around 10 cm down to about 10 μm. The matrix is composed essentially of fine-grained to dust-sized, fragmented rock material (Fig. 6 b). Fragments of quartz and feldspar may show internal, pre-brecciation strain phenomena; whereas primary quartz from the granite is strained and polygonized, hydrothermal vein quartz fragments are nearly always unstrained. Flakes of chlorite may show externally induced deformation from being pinched between other mineral fragments or between the walls of closed fractures. The breccia matrix is cemented by secondary quartz and smaller amounts of epidote-clinozoisite; the grain size of the latter is only about 5 μm. In addition, very small but irregularly distributed amounts of zeolite minerals form thin veins and replacement textures. Zeolites partially replacing quartz and other minerals (Fig. 6 c) have been tentatively identified under the microscope as ?stilbite; and vein-forming laumontite was identified by X-ray diffraction methods, as was the montmorillonite mentioned above. Calcite occurs in trace amounts.

The bulk chemical composition of the granite breccia is very similar to that of the host granite (Table 1), the most notable difference being a higher content of Na_2O and slightly lower CaO and SiO_2.

3) *The microbreccia* has well-defined and sharp boundaries against both the granite breccia and the granite. It sends a few small, thin veinlets into the latter. The rock is fine-grained to dense, and breaks with a rough, irregular to flinty fracture. It is light grey with varying shades of pale green, caused by finely dispersed epidote. Some microbreccia varieties have small, reddish stains of hematite. Thin, irregular quartz veins occur locally.

The microbreccia commonly contains small (less than 2 cm), more or less rounded fragments of similar but older microbreccia. Nearly all these fragments are transected by older quartz veins, clearly predating the latest fragmentation. Scattered, discrete quartz vein fragments are also macroscopically visible. In a few places, small concentrations of fragmented, euhedral low-quartz crystals are embedded as clasts in the microbreccia (Fig. 5 b).

Under the microscope, the microbreccia is seen to consist of a fine-grained, varied mixture of mineral fragments, mostly between 0.01 and 0.50 mm in size (Fig. 6 d). The fragments are mainly of feldspar and quartz, with trace amounts of chlorite. The feldspar and quartz fragments derived from granite may show severe, internal, pre-brecciation strain. The near-ubiquitous fragments of older, quartz-veined microbreccia are commonly richer in secondary epidote than the younger, surrounding microbreccia matrix. In other cases, however, the matrix of these older fragments of microbreccia is so similar to the younger host matrix that such fragments can be identified only by their old quarts veins, abutting against the fragment borders (Fig. 6 e). Even though the breccia matrices in these instances are almost identical, this

Table 1: Chemical analyses and CIPW norms of rock types at Skaget.

	Frøya Granite wt %	Granite breccia wt %	Microbreccia wt %
SiO_2	69.77	68.47	91.37
Al_2O_3	15.58	15.92	3.68
CaO	2.19	2.05	0.97
MgO	0.91	0.79	0.10
Fe_2O_3	0.72	1.40	0.96
FeO	1.58	1.28	0.21
Na_2O	4.00	5.29	1.23
K_2O	4.05	3.24	0.81
TiO_2	0.40	0.46	0.09
MnO	0.05	0.03	0.00
P_2O_5	0.08	0.11	0.02
CO_2	0.09	0.11	0.04
H_2O^+	0.68	1.04	0.23
Total	100.10	100.19	99.71
CIPW norms			
Q	24.6	20.1	79.5
C	1.0	0.5	–
Or	23.9	19.1	4.8
Ab	33.8	44.8	10.4
An	9.8	8.8	2.1
Wo	–	–	0.7
Di	–	–	0.5
Hy	4.0	2.5	–
Mt	1.0	2.0	0.4
Hm	–	–	0.7
Il	0.8	0.9	0.2
Ap	0.2	0.3	–
Cc	0.2	0.3	0.1
H_2O	0.7	1.0	0.2
Total	100.1	100.2	99.7

Analyst: Per-Reidar Graff.

mode of occurrence of vein quartz proves that the microbreccia fragments did not form by agglomeration in the present host matrix, but represent one or more, separate, older microbreccia generations. Some microbreccia fragments carry even older, rounded, fragments of microbreccia (Fig. 6 f); thus, there are at least three generations of microbreccia.

The secondary minerals in the microbreccia are quartz and minute (about 5 μm) grains of epidote; in addition there are trace amounts of calcite. Unlike the granite breccia, the microbreccia does not appear to have been subject to zeolite and montmorillonite deposition.

Chemically, the microbreccia is highly siliceous, and is enriched in calcium relative to alkalies and alumina, as compared with both the Frøya Granite and the granite breccia (Table 1). The microbreccia is diopside and wollastonite normative, whereas the two other lithologies are hypersthene and slightly corundum normative.

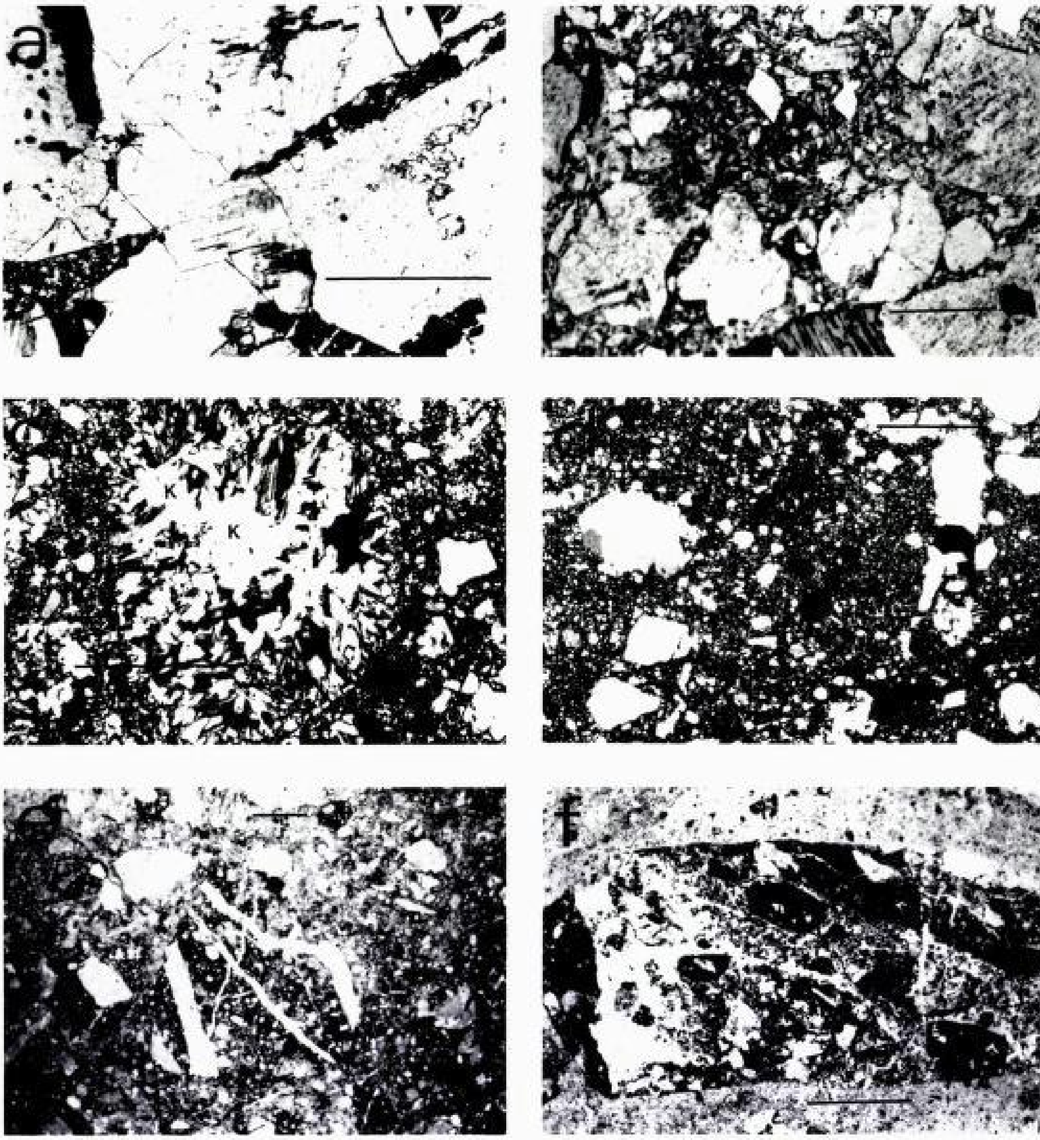

Fig. 6(a): Epidotized microbreccia-filled fractures in granite. Plane polarised light. Bar scale = 0.5 mm.

(b) Granite breccia matrix; unsorted rock and mineral fragments. Plane polarised light. Bar scale = 0.5 mm.

(c) Zeolitized fragment in granite breccia. ?Stilbite, replacing calcite (K), has grown more or less perpendicularly inward from the rim of the fragment. Crossed nicols. Bar scale = 0.5 mm.

(d) Microbreccia with fragments of vein quartz. Crossed nicols. Bar scale = 0.5 mm.

(e) Quartz-veined microbreccia fragment in younger, similar microbreccia. Plane polarised light. Bar scale = 1 mm.

(f) Microbreccia fragment with older microbreccia fragments in microbreccia. The older fragments (dark) are more epidote-rich than younger matrix. Plane polarised light. Bar scale = 0.5 mm.

Discussion

States of stress in the Earth's crust producing localized brittle fragmentation of rocks can arise from a variety of geological causes, such as: igneous intrusion and extrusion, tectonic activity, meteorite impact, and the release of thermal and potential energy from subsurface fluids.

The following points are considered important for an understanding of the formation of the Skaget Breccia. Brecciation by magmatic processes can be dismissed because no igneous material is present in or associated with the breccia; except the host granite, which clearly had crystallized completely prior to the breccia formation. Tectonic fragmentation appears unlikely because of the irregular and gradational boundary between the host granite and the granite breccia, and also because of the virtual lack of strain effects in fragments of vein quartz and other secondary minerals even in the most intensively comminuted fractions. Impact brecciation can be excluded as a mechanism because of the recurrence of brecciation events, separated by episodes of mineral deposition and veining; besides, no evidence of shock phenomena indicative of impact have been found.

On the other hand, brecciation of the granite by volatile fluid overpressure with subsequent gas streaming and fragment comminution appears to be consistent with all the known facts about the Skaget Breccia. The following discussion presents a short examination of this interpretation.

By combining the observations presented above, it is possible to identify some of the effects and properties of the fluid medium:

1) Initially, the fluid exerted a pressure sufficient to break up the Frøya Granite, probably by hydraulic fracturing (Fyfe et al. 1978), and to cause incipient ductile deformation of feldspar as well as dissolution of quartz in the wall-rock margin.

2) Through sudden pressure release, probably by explosive venting to the surface, the fluid thereupon expanded greatly and formed a fluidizing gas, which enabled inter-particle collisions among the granite fragments to abrade and comminute the material during vigourous, bubbling fluidization.

3) By entrainment and elutriation of the smallest particles, the streaming gas separated a fraction of fines from the main and unsorted part of the fluidized system to form the microbreccia in a marginal portion of the system (Woolsey et al. 1975).

4) When the fluidized material had settled after the eventual escape of the gas a hydrothermal, liquid solution following in its wake rapidly deposited minute grains of epidote/clinozoisite and, much more slowly, large amounts of quartz – some of it as euhedral, inclusion-free crystals. These have obviously grown in open vugs, cavities and fractures in the breccia. Subsequently, Ca-zeolites were formed along veins and fractures; some of them probably by reaction between the solution and unstable minerals (quartz, calcite), and montmorillonite was deposited in small cavities, which were formed as fluidization bubbles (Reynolds 1954).

The material deposited from solution was probably derived, at least in part, from quartz and plagioclase (anorthite) in the fragmented granite.

5) This sequence of events: fracturing, break-through to the surface, gas fluidization and hydrothermal mineral deposition was repeated at least twice (Fig. 6 f), in the same order and with apparantly identical effects.

6) The temperature was probably never very high, since microbreccia-filled fractures cross-cutting and offsetting veins with unaltered laumontite in the granite and in breccia fragments indicate that the upper stability temperature of this zeolite (ab. 220 °C; Liou 1971, Bird & Helgeson 1981) was not overstepped during the fracturing and fluidization events. This is well below the critical temperature for pure water (374 °C). At a temperature of, say, 200 °C the vapour pressure of pure water is only 15.5 bars (Fyfe et al. 1978). This would probably be insufficient to break through the granitic overburden at a depth where the rock did not already contain shallow joints more or less open to the surface.

The most probable fracturing and gas fluidization medium at such maderate temperatures is, therefore, considered to have been a fluid composed of carbon dioxide and water, and separated into a H_2O-rich, liquid phase and a CO_2-rich, incondensible fluid phase (Takenouchi & Kennedy 1964). Such mixtures have thermal expansivities, compressibilities and specific heats much higher than those of pure water (Schubert & Straus 1981), and would expand much more than water under the same conditions. The fluidizing agent would then consist of the CO_2-rich gas phase.

Chemically, the CO_2 content of the initial fluid mixture would induce a relatively high solubility for anorthite as compared to albite and K-feldspar (Thompson 1974.) This is consistent with the deposition of Ca minerals upon loss of CO_2, whereas there is no mineralogical evidence that the alkali feldspars have been affected by the fluid. While calcium ion activities could be lowered due to complexing with carbonate species in solution (Wigley & Plummer 1976, Blatt et al. 1980, Anderson 1981); alkali feldspar solubilities, on the other hand, could be reduced by the common-species effect if the fluid were at or close to silica saturation (Fyfe et al. 1978). The sudden escape of CO_2 during fluidization would lead to rapid deposition of epidote/clinozoisite and probably calcite, which would redissolve during falling temperatures as a consequence of its retrograde solubility (Fyfe et al. 1978). Calcium zeolites may in part have formed at the expense of calcite.

The many complexities in this system, such as disequilibrium conditions, alternation between open and closed-system behaviour, and spatial variations in the relative importance of fluid flow versus diffusion transport, discourage further detailed discussions based on the available mineralogical and chemical data for the Skaget Breccia.

With only moderate temperatures prevailing in and near the system during brecciation activity, there is no evidence that the fluid was significantly heated in or by the host rock. On the contrary, the onset of incipient ductility of the granite material at the very contact may indicate that the rock margin was heated by the fluid (Paterson 1978, Handin & Carter 1979).

Therefore, in situ thermal expansion does not seem to afford a viable explanation for the dynamics of hydraulic fracturing, explosion and fluidization in the present case. This would also seem to disqualify circulating groundwater as an immediate source for the fluids; such waters provide only hydrostatic pressures (Henley & McNabb 1978). The fluids must have come from deeper levels.

A deep-lying fluid and energy source within the granite pluton itself would entail the involvement in the brecciation events of a magmatic fluid (Henley & McNabb 1978), equilibrated at high temperatures with crystallizing or hot crystalline granite in the interior of the pluton, below a thick, cool granite shell, and perhaps driven out and upward by retrograde boiling (*ibid.*). However, the chemical effects of the fluid on the brecciated granite do not appear to be compatible with the expected composition of a 'granitic' magmatic fluid (Burnham 1967, Ellis & Mahon 1967, Hibbard 1980). On the other hand, the magmatic fluids could have had an unusual composition; perhaps in part resulting from decomposition of marble inclusions derived from the older gneisses. This is considered less likely, however, because the Frøya Granite appears to be notably free from inclusions of the adjacent gneisses. Moreover, during the late crystallization history of the accessible parts of the Frøya Granite pegmatitic melts, in all probability saturated with volatiles, were retained within the pluton or its migmatitic aureole even at the stage where the solidified granite could sustain large-scale fracturing. This indicates that the Frøya Granite was not emplaced at shallow levels in the crust, where late to post-magmatic fluids would tend to actively force their way upwards. Although there are a number of similarities between the Skaget Breccia and syn-plutonic breccias in granitoid and other igneous rocks, for instance in the British Caledonides (Brindley et al. 1976) and in and near Variscan granites (Goode & Taylor 1980), these plutons seem to have intruded at higher crustal levels than did the Frøya Granite; and thus to have been able to interact, as hot rocks, with water of meteoric origin. The same appears to be the case with breccias associated with intrusive stocks of porphyry copper systems (Phillips 1973, Henley & McNabb 1978). Diatremes and breccia pipes in appinitic plutonic rocks in the Caledonides in Scotland and Ireland differ from the Skaget Breccia in being intimately associated with magmatic products and clearly connected with intrusive processes (e.g. Pitcher & Berger 1972). They are considered to have been produced by volatile escape from water-rich, mantle-derived magma (Wright & Bowes 1979).

The brecciating fluid at Skaget and its energy content may possibly have been independent of and, by implication, younger than the hosting, Frøya Granite.

One alternative derivation for the fluid could have been prograde, devolatilizing metamorphism at deep levels in the crust of metasedimentary rocks associated with deep thrusting. Such a proposition is without supporting evidence at the present time.

A possible mantle origin for the Skaget fluids is an interesting alternative in view of the present, widely held opinion that carbon dioxide is widespread and common in the mantle (e.g., Eggler 1975, Mysen & Boettcher 1975, a, b, Wyllie 1978, Barnes & McCoy 1979, Wendlandt & Harrison 1979, Newton et al. 1980). Newton et al. (1980) consider that such fluids can enter the overlying crust, and scavenge water from crustal rocks on their way upwards. For carbon dioxide to penetrate into the crust from below would probably require anomalously high temperatures. Appropriate conditions would be expected to occur in the thermal regime of a hot spot or mantle plume.

The heat energy and intrinsic buoyancy of such fluids could probably not alone provide a mechanism for their upwards penetration from the mantle to shallow crustal levels. Under certain conditions, however, release of potential energy could possibly bring this about, as shown theoretically by Secor & Pollard (1975), who applied their model specifically to water at high crustal levels. In this model, the fluid acts as a pressure medium, conveying high static pressures upwards from deeper levels (Fig. 7). In the situation depicted in Fig. 7, a horizontal least-principal rock stress has a vertical gradient (dS_x/dZ) which is moderately higher than the vertical pressure gradient in the fluid (dP_f/dZ). The fluid holds open a long vertical hydraulic fracture by means of the enhanced pressure induced on it by the rock stress at the lower closure of the fracture. If the fluid pressure at the top of the fracture exceeds the sum of the local least principal stress and the tensile strenght of the rock, the crack may propagate upwards by hydraulic fracturing; the lower end of the fracture would close, in step with the upward advance. Thus, the length of the fracture will remain constant as long as the fluid ascends under unchanged conditions, assuming that the fluid is incompressible and that the process occurs under

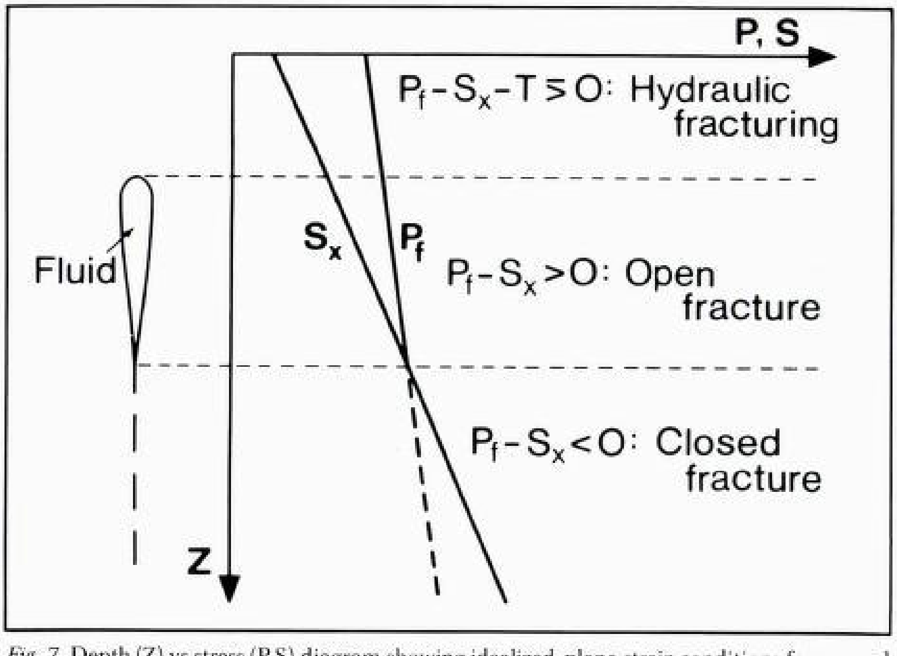

Fig. 7. Depth (Z) vs stress (P,S) diagram showing idealized, plane-strain conditions for upward propagation of vertical, fluid-filled, hydraulic fracture. Adapted from Secor & Pollard (1975). P_f: fluid pressure; S_x: horizontal, least principal stress; T: Tensile strength of rock.

plane-strain conditions (Secor & Pollard 1975). For the present application, these last assumptions make a rather crude approximation. Energy dissipation associated with the work of fracture, etc., would probably halt further ascent of an isolated batch of fluid such as this after a finite distance. However, since the length of the stable, open fracture is determined by the *local* stress and pressure variables and gradients, it should be entirely possible for new batches of fluid to follow in the hydro-fractured wake of the first, an to overtake it. The merging of sequential, open fractures and their contained fluids would probably add renewed impetus to hydraulic fracturing, and so continue a stepwise or pulsed ascent of the fluid. A possible consequence of this latter mechanism could be cyclic brecciation with intervening quiet periods, similar to the processes inferred for the Skaget Breccia, where self-sealing (Keith et al. 1978) by quartz deposition would probably result from hydrothermal activity following each explosive outburst of brecciation and fluidization triggered by the arrival of new pulses of mantle fluid.

High-pressure melting experiments with synthetic and natural mantle material have indicated that melts yielding, for instance, larnite-normative olivine nephelinite can form under realistic conditions from mantle material only in systems with carbon dioxide and water (Mysen & Boettcher 1975 a, b, Eggler & Holloway 1977, Burnham 1979). It is therefore particularly pertinent to the present discussion about possible deep sources for carbon dioxide that plugs of larnite-normative olivine nephelinite, dated at about 56 Ma, occur on the continental shelf only about 120 km from Skaget (Fig. 1; Bugge et al. 1980). In this context it is also of interest to note that this part of the coast has been recorded by previous authors as a region of widespread zeolite mineralization on joints and fractures in basement gneisses and other rocks (Fig. 1; Sæther 1950, Strand 1952, Fediuk & Siedlecki 1977). Based on other considerations, a mantle-plume generated uplift has been tentatively proposed for this region (Torske 1975).

Conclusions

Based on the foregoing discussion on the Skaget Breccia and on the occurrence of young(?) zeolite mineralization as well as Tertiary volcanic activity, it is suggested that the coastal and offshore Nordmøre region may constitute a palaeo-geothermal province. This could conceivably have a bearing on the hydrocarbon potential in this part of the continental shelf in terms of palaeogeotherms and permeability effects of mineralizations.

A critical topic in the assessment of the possible genetic affiliations of the Skaget Breccia would be its mineralization age. This is unknown at the present time.

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REFERENCES

Anderson, G. M. 1981: Aqueous silicate and carbonate solute species at high P and T. *Geological and Mineralogical Associations of Canada Joint Annual Meeting, Program with Abstracts 6*, 1. (Calgary).

Askvik, H. 1979: *Berggrunnskart Kristiansund 1:250000;* foreløpig utgave, Norges geologiske undersøkelse, Trondheim.

Barnes, I. & McCoy, G. A. 1979: Possible role of mantle-derived CO_2 in causing two «phreatic» explosions in Alaska. *Geology 7*, 434–35.

Batzle, M. L. & Simmons, G. 1976: Microfractures in rocks from two geothermal areas. *Earth planet. Sci. Let. 30*, 71–93.

Bird, D. K. & Helgeson, H. C. 1981: Chemical interaction of aqueous solutions with epidote-feldspar mineral assemblages in geologic systems. II. Equilibrium constraints in metamorphic/geothermal processes. *Am. J. Sci. 281*, 576–614.

Blatt, H., Middleton, G. & Murray, R. 1980: *Origin of sedimentary rocks*. 2nd ed., 782 pp. Prentice-Hall, Inc., Englewood Cliffs, N.J.

Brindley, J. C., Gupta, L. N. & Kennan, P. S. 1976: Explosion breccias and related features in the Leinster Caledonian Massif, South-East Ireland. *Proc. R. Ir. Acad. 76, Sect. B, 22*, 337–48.

Bugge, T., Prestvik, T. & Rokoengen, K. 1980: Lower Tertiary volcanic rocks off Kristiansund, Mid-Norway. *Mar. Geol. 35*, 277–86.

Burnham, C. Wayne 1967: Hydrothermal fluids at the magmatic stage. *In* Barnes, H. L. (ed.): *Geochemistry of hydrothermal ore deposits*, 34–76. Holt, Rinehart and Winston, Inc., N.Y., 670 pp.

Burnham, C. Wayne 1979: The importance of volatile constituents. *In* Yoder, H. S. (ed.): *The evolution of the igneous rocks – fiftieth anniversary perspectives*, 439–82. Princeton University Press, Princeton, N.J., 588 pp.

Chayes, F. 1955: Potash feldspar as a by-product of the biotite–chlorite transformation. *J. Geol. 63*, 75–82.

Eggler, D. H. 1975: CO_2 as a volatile component of the mantle: the system Mg_2SiO_4–SiO_2–H_2O–CO_2. *In* Ahrens, L. H., Dawson, J. B., Duncan, A. R. & Erlank, A. J. (eds.): *Physics and chemistry of the Earth 9*, 869–81. Pergamon Press, Oxf., 940 pp.

Eggler, D. H. & Holloway, J. R. 1977: Partial melting of peridotite in the presence of H_2O and CO_2: principles and review. *Ore. Dep. Geol. Miner. Ind. Bull. 96*, 15–36.

Ellis, A. J. & Mahon, W. A. J. 1967: Natural hydrothermal systems and experimental hot water/rock interactions (Part II). *Geochim. cosmochim. Acta 31*, 519–38.

Fediuk, F. & Siedlecki, S. 1977: Smøla. Beskrivelse til det berggrunnsgeologiske kart 1321–1:50000. *Norges geol. Unders. 330*, 1–23.

Fyfe, W. S., Price, N. J. & Thompson, A. B. 1978: *Fluids in the Earth's crust.* Elsevier, Amst., 383 pp.

Goode, A. J. J. & Taylor, R. T. 1980: Intrusive and pneumatolytic breccias in south-west England. *Inst. geol. Sci. Rep. 80/2*, 1–23. London.

Handin, J. & Carter, N. 1979: Rheological properties of rocks at high temperatures. *4th Int. Soc. Rock Mech. Proc. 3*, 97–106, Montreux.

Henley, R. W. & McNabb, A. 1978: Magmatic vapor plumes and ground-water interaction in porphyry copper emplacement. *Econ. Geol. 73*, 1–20.

Hibbard, M. J. 1980: Indigenous sources of late-stage dikes and veins in granite plutons. *Econ. geol. 75*, 410–23.

Irwin, W. P. & Barnes, I. 1980: Tectonic relations of carbon dioxide discharges and earthquakes. *J. geophys. Res. 85, (B 6)*, 3115–21.

Johannsen, A. 1939: *A descriptive petrography of the igneous rocks 1*, Univ. Chicago Press, Chicago. 318 pp.

Keilhau, B. M. 1850: *Gæa Norvegica 3*, 342–516, Chr.a.

Keith, T. E. C., White, D. E. & Beeson, M. H. 1978: Hydrothermal alteration and self-sealing in Y-7 and Y-8 drillholes in the northern part of Upper Geyser Basin, Yellowstone National Park, Wyoming. *U.S. geol. Surv. prof. Pap. 1054 A*, 1–26.

Liou, J. G. 1971: Stilbite–laumonite equilibrium. *Contr. Miner. Petrology 31*, 171–77.

Mysen, B. O. & Boettcher, A. L. 1975 a: Melting of a hydrous mantle. I. Phase relations of a natural peridotite at high pressures and high temperatures with controlled activities of water, carbon dioxide and hydrogen. *J. Petrology 16*, 520–48.

Mysen, B. O. & Boettcher, A. L. 1975b: Melting of a hydrous mantle. II. Geochemistry of crystals and liquids formed by anatexis of mantle peridotite at high pressures and high temperatures as a function of controlled activities of water, hydrogen and carbon dioxide. *J. Petrology 16*, 549–90.

Newton, R. C., Smith, J. V. & Windley, B. F. 1980: Carbonic metamorphism, granulites and crustal growth. *Nature 228*, 45–50.

Oftedahl, C. 1981: *Norges geologi.* Tapir forlag, Trondheim, 207 pp.

Paterson, M. S. 1978: *Experimental rock deformation – the brittle field.* Springer-Verlag, Berlin. 254 pp.

Phillips, W. J. 1973: Mechanical effects of retrograde boiling and its probable importance in the formation of some porphyry deposits. *Inst. Min. Metall. Trans. B 82*, 90–98.

Pitcher, W. S. & Berger, A. R. 1972: *The geology of Donegal.* Wiley-Interscience, New York. 435 pp.

Reusch, H. 1914: Nogen bidrag til Hitterens og Smølens geologi. *Norges geol. Unders. 69 (4)*, 1–50.

Reynolds, D. L. 1954: Fluidization as a geological process and its bearing on the problem of intrusive granites. *Am. J. Sci. 252*, 577–614.

Sæther, E. 1950: Funn av laumontitt i gneisfeltet på Møre. *Norsk geol. Tidsskr. 28*, 50–51.

Schubert, G. & Straus, J. M. 1981: Thermodynamic properties for the convection of steam-water-CO_2 mixtures. *Am. J. Sci. 281*, 318–34.

Secor, D. T. Jr. & Pollard, D. D. 1975: On the stability of open hydraulic fractures in the Earth's crust. *Geophys. Res. Letters 2 (11)*, 510–13.

Strand, T. 1952: Occurrences of zeolites in Nordmøre. *Norsk geol. Tidsskr. 30*, 210–12.

Strand, T. 1960: The pre-Devonian rocks and structures in the region of Caledonian deformation. pp. 170–284 *in* Holtedahl, O. (ed.): Geology of Norway. *Norges geol. Unders.* 208, 540 pp.

Takenouchi, S. & Kennedy, G. C. 1964: The binary system H_2O-CO_2 at high temperatures and pressures. *Am. J. Sci. 262*, 1055–74.

Thompson, A. B. 1974: The instability of feldspar in metamorphism. *In* MacKenzie, W. S. & Zussmann, J. (eds.): *The feldspars.* pp. 645–72. Manchr. University Press, Manchr. 717 pp.

Torske, T. 1975: Possible Mesozoic mantle plume activity beneath the continental margin of Norway. *Norges geol. Unders. 322*, 73–90.

Wendlandt, R. F. & Harrison, W. J. 1979: Rare earth partitioning between immiscible carbonate and silicate liquids and CO_2 vapor: results and implications for the formation of light rare earth-enriched rocks. *Contr. Miner. Petrology 69*, 409–19.

Wigley, T. M. L. & Plummer, L. N. 1976: Mixing of carbonate waters. *Geochim. cosmochim. Acta 40*, 989–95.

Woolsey, T. S., McCallum, M. E. & Schumm, S. A. 1975: Modeling of diatreme emplacement by fluidization. *In* Ahrens, L. H., Dawson, J. B., Duncan, A. R. & Erlank, A. J. (eds.): *Physics and chemistry of the Earth 9*, 29–42. Pergamon Press, Oxford. 940 pp.

Wright, A. E. & Bowes, D. R. 1979: Geochemistry of the appinite suite. pp 699–704, *in* Harris, A. L., Holland, C. H. & Leake, B. E. (eds.): *The Caledonides of the British Isles – reviewed.* Scottish Academic Press, Edinb., 768 pp.

Wyllie, P. J. 1978: Peridotite-CO_2-H_2O and the Low-Velocity Zone. *Bull. volcanol. 41 (4)*, 670–83.

Classification of Uranium Mineralization in Norway

INGVAR LINDAHL

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NGU has carried out exploration for uranium on a small but varying scale since the second World War. An attempt is made to classify types of uranium mineralization occurring in Norway, following a system based on genetic types. Although difficulties are encountered in applying the classification, it has proved possible to categorize many of the occurrences in this way. Information on some occurrences is limited, and future revisions must be expected. The uranium mineralizations described are from all parts of the country, and range from Proterozoic to Lower Paleozoic in age. The most prominent and promising occurrences are largely of sedimentary, intrusive, metamorphic and supergene origin, the most promising of which are found in the region of northern Nordland. As yet no uranium deposits of importance have been found, based on prevailing metal prices, but insufficient work has been done on a number of known occurrences and the potential for the existence of economic deposits in Norway has been assessed as good.

I. Lindadl, Geological Survey of Norway, P.O. Box 3006, N-7001 Trondheim, Norway.

Introduction

Early in the last century geologists and mineralogists became interested in the pegmatites in the Precambrian shield of southern Norway because of their many rare minerals, often developed as large idiomorphic crystals. Several new minerals were identified and described from these deposits, some of them containing uranium and thorium. Minerals with uranium and thorium as main constituents were also recognized. Quartz and feldspar were produced from small quarries in the pegmatites. Uranium was a by-product at a few of them and used mostly as a glass-colouring agent.

After the World War II, previously known occurrences of uranium-bearing minerals in southern Norway were reexamined. Drilling programmes were conducted on the Einerkilen and Vats pegmatites in Telemark and Rogaland, and on the alum shale in the Oslo region (Rosenquist 1948). The Norwegian Defence Research Establishment (FFI), the Joint Establishment for Nuclear Energy Research (IFA), and the Geological Survey of Norway (NGU) were involved in this work. This intensive prospecting campaign culminated in 1951 when IFA contracted for supply of uranium to their experimental reactor (Sverdrup et al. 1967).

In the 1950's there was a boom for uranium among amateur prospectors. A minor programme was initiated at NGU in 1954 (Sverdrup et al. 1967), and systematic work was done throughout the country. The activity slowed down towards the end of the 1950's, and by the end of the 1960's prospecting activity was very limited. An attempt to evaluate the possible and speculative resources of uranium in Norway was made in 1975 (NOU 18:1972). The potential for economic uranium mineralization in Norway was regarded as small. However, NGU pointed out in comments to this publication that the conclusions had been based on very limited background information.

The present uranium exploration programme at NGU commenced in 1975. Geological reasoning indicated the possibility of finding resources of uranium to be good. Since 1975 a small prospecting effort has been in progress. In addition, NGU has evaluated specific occurrences on behalf of the USB programme (Investigation of state owned mining claims).

One of the aims of the Ore Section at NGU is to give an estimate of available resources of metallic elements in Norway, among them uranium and thorium. No mining company or other institution has prospected for uranium, and therefore most of the information on uranium and thorium in Norway has been collected by, and is recorded at, the Survey. There has been close contact between Norway and the UN-body IAEA (International Atomic Energy Agency), the IEA (international Energy Agency), and the OECD/NEA (Nuclear Energy Agency). The problems with uranium supply in the 1970's resulted in the creation of IUREP (International Uranium Resources Evaluation Project).

The IUREP-report from the 'orientation phase' was completed in 1978, a report signed by a joint steering group on uranium resources from OECD/NEA and IAEA. Countries with different potentials for uranium deposits were grouped; Norway was assigned to the group 'Countries with good potential where exploration is to be encouraged'. The IUREP programme offered expert missions to countries falling in this classification. Norway welcomed experts financed by the IUREP-programme, who visited the Survey early in 1980 and participated in a two week field trip in the summer of 1980. The IUREP-mission resulted in a report by the three experts, M. Cuney from CREGU (Centre de Recherches sur la Géologie de l'Uranium), France, D. M. Taylor, secretary to OECD/NEA, and M. Wilson, head of the uranium prospecting group at SGU (Geological Survey of Sweden), Sweden (Cuney et al. 1981).

The purpose of this paper is to classify the types of uranium mineralizations in Norway according to one of the proposed international systems (Dahlkamp 1978).

Geological Models

The most important factors in selecting initial targets for uranium prospecting in Norway have been geological environments and age of formation of the unit. These have been compared with the milieu and age of known uranium deposits on a global scale. Uranium mineralizations in neighbouring countries, especially Sweden, were particularly considered. The philosophy of this programme was presented by Lindahl & Heier (1977).

One possible position for uranium is along unconformities and major tectonic discontinuities, either in the weathered zone just beneath the peneplain, or in the basal parts of the overlying sediments in the case of unconformities. Examples of this type of deposit on a global scale include quartz pebble conglomerates and uranium in basal sandstones and more fine-grained sediments. Certain geological discontinuites in Norway could be favourable. One is on the Baltic Shield in Finnmark between the dome of supposed Archaean age (between Kautokeino and Karasjok), and the overlying Proterozoic rocks. Conglomerates, but not of the quartz pebble type, have been found on the northern margin by Skålvoll (1971). The border of the basal sequence around this dome lies in terrain heavily covered by till, and possibility for uranium concentrations along the zone cannot be excluded. Other domes in Finnmark of possible Archaean age could have similar potential. The Lofoten–Vesterålen province contains important geological discontinuities (Tveten 1978), but the rock units in this region have very low radio-element content.

Another major age discontinuity is at the boundary between Precambrian basement of varying ages and overlying late Precambrian and Cambrian rocks along or east of the Caledonian front. In some places nappes directly overlie the contact, but in most cases thin sequences of autochthonous sediments are found. A number of uranium anomalies, mineralizations and deposits are located in a relatively narrow zone along this discontinuity, both in the basement and in the sediments above. Deposits of this type have been described by several authors (Gee 1972, IAEA symposium in Athens 1974, Barbier 1974, Lindahl & Heier 1977, Cuney et al. 1981).

Uranium in its oxidized state is one of the most mobile of elements, and is concentrated under favourable geochemical conditions during peneplanation, either in the sediments above or in the basement (Barbier 1974). Other locations for concentration may be along tectonic zones, thrust planes, nappe structures, breccias and geotectonic lineaments. The potential should be even better where two or more of these favourable conditions coincide (Lindahl & Heier 1977). However, a favourable geological structure for deposition of uranium is not sufficient alone; a uranium source is needed to form a deposit. It is therefore more likely that basement with high uranium concentration would produce deposits in younger overlying rocks than basement with low uranium contents. The same suggestions have been put forward for the lead deposits along the Caledonian front (Bjørlykke 1977).

The investigations have included follow-up studies in previously recognized anomalous areas and provinces. Killeen & Heier (1975) described a belt from Båhus–Iddefjord to Flå containing granites with anomalously high radio-element contents. In the uranium programme a study was made of the Flå granite area, with special interest paid to the overlying rocks of Late Precambrian and Cambro–Silurian age.

The study of several regions with intrusive granites and granitoids showed that the basement granitoids (1,700–1,800 m.a.) in the region from Rana to north of Rombak in Nordland are anomalously high in uranium and thorium. This may be an extension of the uranium province in northern Sweden described by Adamek & Wilson (1979). Further, we have examined areas with Devonian and younger sediments on the western coast of Norway between Sognefjorden and Nordfjorden, on Ørlandet and on Andøya (Fig. 1). A parallel has been drawn with the uranium mineralization in the Old Red Sandstone in Scotland, including the Orkneys (Gallagher et al. 1971, Michie 1972). So far the results have been negative.

Fig. 1: Main geological features of Norway with uranium occurrences plotted.

Types of Mineralization

Several authors have made an attempt at classification of uranium deposits. Dahlkamp (1978) proposed a relatively extensive but uncomplicated classification table based on genetic types which are not time related. It embraces the following groups:

1) Sedimentary – Deposits in conglomerates. Potentials in black shale and phosphatic sediments.
2) Effusive – Mineralizations in acidic volcanics.
3) Intrusive – Deposits in alaskites and acid intrusives including related hydrothermal phases (veins).
4) Contact metasomatic – Deposits in calc-silicate rocks.
5) Metamorphic – Remobilized deposits in phyllite, schist, etc. (veins).
6) Supergene – Deposits in sandstone (rolls), calcrete etc.

It is difficult to classify many Norwegian occurrences, partly because insufficient work has yet been done. The classification system of Dahlkamp (1978) does not seem to be fully applicable, nor does the classical system (Cuney et al. 1982). In the following, some selected occurrences and areas of uranium potential are concidered in the light of the Dahlkamp classification.

SEDIMENTARY DEPOSITS

Dahlkamp's (1978) classification includes in this group detrial weathering products more than 2,200 m.a. in age. We have as yet found no example of this type of deposit. However, the younger alum shales, of varying metamorphic grade, contain uranium. Two occurrences of this type will be described briefly. Pelitic and psammitic sediments may host uranium mineralization, but no examples are known, neither in the late Precambrian arkosic sediments along the Caledonian front nor in the Devonian sandstones in western Norway.

Black shale and graphitic schist of various, in some cases uncertain, age occur in or along the Caledonian mountain belt. These rocks show wide ranges of uranium and other trace element content. The only graphite deposit in operation, at Skaland in Troms county, is low in uranium as are other smaller deposits in the same region, all of presumed Precambrian age.

Rendalsvik in Nordland county is another coarse-grained, high-metamorphic graphite deposit in strongly tectonized and partly granitized terrain. The graphite schist is thought to be of Cambro-Silurian age (Skjeseth & Sørensen 1958). The Rendalsvik graphite schist is a mica schist with up to 10% graphite and 8% ore minerals (Sverdrup et al. 1967). Uraninite was identified as occurring in well defined crystals. In the heavy mineral fraction obtained from systematic resampling uraninite could not be identified, but the presence of the minerals rutile, sphene, apatite, sphalerite, uvarovite, clinozoisite, and tourmaline was determined by X-ray (Gust & Thoresen 1981).

The resampling of the Rendalsvik graphite schist (75 profile samples) gave an average of 45 ppm U (range 8–183 ppm) as analysed by gamma-spectrometer. A typical profile is shown in Fig. 2 with analytical results for the samples and radiometric measurements in situ. Analytical results for other elements are on average:

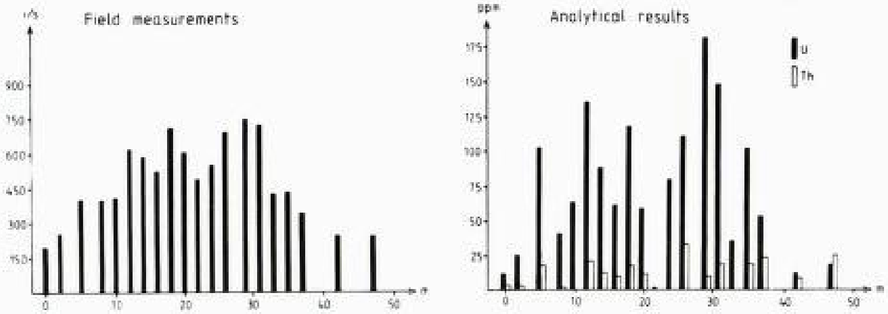

Fig. 2: An example of a profile across the graphite schist in Rendalsvik, Nordland county, showing the results of scintillometer registration (Knirps instrument) compared with U/Th analyses of samples from the same points.

Cu – 85 ppm (AAS), Zn – 105 ppm (AAS), Pb – 40 ppm (AAS), Mo – 85 ppm (AAS), V – 450 ppm (XRF) (max. 0.3%). On the basis of drilling, geophysics and geological interpretation the potential of crystalline graphite ore was estimated to be 3 mill. tons. Assuming the tonnage to average 45 ppm uranium, this amounts to a total of about 100 tons of uranium.

The Cambrian–Ordovician alum shales of the Caledonian front are enriched in uranium. The unit cannot be followed continuously, but has been registered in several places. The deposits in the Tåsjø area in Sweden have been described by Gee (1972). The alum shales near Østersund have been studied in recent years (Gee & Zachrisson 1979). Both these deposits are phosphorus-bearing (Armands 1970). The alum shale in the Østersund area shows an average content of 150–200 ppm U, with subordinate layers of up to 250 ppm. Typical for the shale is a high vanadium content, reaching 0.34%, and molybdenum averaging 0.04–0.05%.

Remnants of alum shale units enriched in uranium have been registered along parts of the borders around Precambrian windows in the Caledonian mountain belt, e.g. the Olden window, the Tømmerås window, the Nasafjell window, and parts of the Rombak window. On the southern part of the Baltic Shield the alum shales, of Upper Cambrian and Lower Ordovician age, are less disturbed by tectonism and metamorphism. The alum shales originally covered an extensive area, with remnants now in central Sweden and along the margins of the Oslo region. The uranium content in the alum shales depends on the availability of uranium and on conditions during sedimentation. The highest contents are found in the Ranstad deposit as described by several authors (Edling 1974), with an average of 300 ppm U.

The alum shale in the Oslo region, 40–50 m thick and deposited during a period of 40–50 mill. years, has been described by Rosenquist (1948), Siggerud (1956) and Skjeseth (1958). The uranium content was found to be 50–100 ppm, with a maximum of 170 ppm in layers of up to 10 cm thick. The highest values are found in the Peltura–Leptoplastus beds (stage 2c–2d) of

Upper Cambrian age. Random samples taken during car-borne surveys, however, gave higher uranium values, and reanalyses of old samples showed the previous analyses to be too low.

Skarn deposits of zinc in the Elsjø and Kirkeby areas have been studied by Olerud (1982). In the Elsjø area contact-metamorphosed Cambro–Ordovician sediments are surrounded by Permian intrusives. In the Kirkeby area they rest on Precambrian gneisses, but border Permian intrusives on the west. Fig. 3 shows Olerud's map of the area. The zinc deposits occur in skarn-

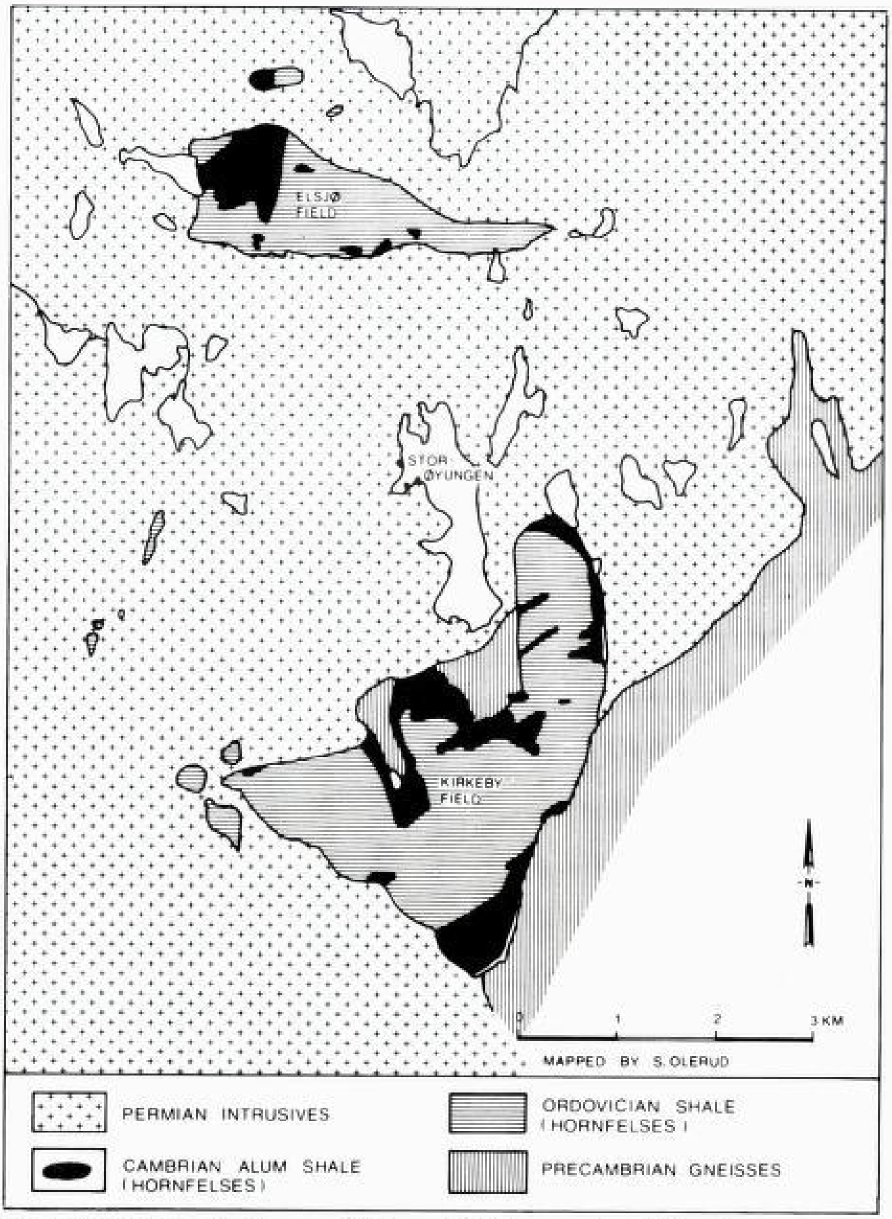

Fig. 3: Simplified geological map of Elsjø and Kirkeby areas; by S. Olerud.

altered limestone lenses and layers at various levels in the sediments, but special interest was paid to the uranium-rich Dictyonema to Olenid zones. The alum shales in the Elsjø area are intensely folded along ENE–WSW axes, and also thrusted (Olerud 1982). The total tectonic thickness of the shales is locally more than 150 m, and thus more favourable than in many other parts of the Oslo region. Limestone layers and lenses are also more abundant. The zinc-bearing skarn was formed locally in limestone lenses through metasomatism. There is no sign of remobilization of uranium in the black shales. In 1978 three holes were drilled in the Elsjø area (Olsrud 1982). The maximum uranium value in drill-core samples was 240 ppm, but the average value of certain 10–20 m sections was 150 ppm U. The average grades for other metallic elements are: 200 ppm Mo, 800 ppm V (max. 0.2%), 150 ppm Ni and 150 ppm Co. A positive correlation is found between uranium and molybdenum, and between vanadium and molybdenum.

EFFUSIVE DEPOSITS

No significant uranium mineralizations have so far been found in rocks which can be definitely classified as effusive. It is, however, difficult to identify acid volcanics in Precambrian basement areas, where the rock units have been subjected to several episodes of tectonism and metamorphism.

An albite fels associated with greenstone in the western part of Finnmarksvidda is thought to be of acidic volcanic origin. Mathiesen (1970) identified a local occurrence of a uranium-bearing mineral in this rock at Bidjovagge. In the uranium exploration programme anomalies were found in boulders of a similar rock near Biggeluobbal west of Masi. Bidjovagge was developed for its deposits of copper and associated gold. There seems to exist a correlation between the gold and higher than background values of uranium. In the Biggeluobbal locality molybdenum is present.

The Duobblon uranium deposit in northern Sweden is located within rhyolitic ignimbrites of Middle Precambrian age (between 1,725 and 1,785 m.a.) (Lindroos & Smellie 1979). In this area, clear primary structures can be found confirming association with the effusive volcanic rocks. It is possible that a grey, medium-grained, granitic gneiss with unusual chemistry, outcropping in the Høgtuva Precambrian window just west of Mo i Rana in Norway could be an acidic volcanic rock. This rock has anomalously high uranium contents. Acidic volcanics have been mapped in the Nasafjell window in basement of the same age (M. Wilson pers. comm.). The Høgtuva area has not yet been studied in detail, but work is in progress.

The Permian rocks of the Oslo region are thorium-enriched and constitute a thorium province. At Sæteråsen west of Holmestrand in Vestfold county, a radiometric anomaly was found by airborne surveys in the 1960's. Later work showed the anomaly to be related to two fine-grained trachytic lava flows (Ihlen 1982) with average uranium contents of 40–50 ppm and thorium contents of 400 ppm. In addition to thorium and uranium, approximately 1% REE and 0.25% Nb are present.

INTRUSIVE DEPOSITS

According to Dahlkamp (1978) deposits of this group are associated with peralkaline syenites, carbonatites, alaskites, alkali granites, granites, pegmatites and hydrothermal veins derived from such intrusions. The most differentiated intrusives, i.e. the youngest ones, are thought to be the best prospecting targets.

The oldest granites in Norway that we know of as highly radioactive are the granites and granitoids in the northern Nordland region from Mo i Rana to the Rombak window, as well as some of the granitoids in the Olden window. The age of this basement is 1,700–1,800 m.a. The granites are usually coarse-grained, but also medium-grained and porphyritic types are found. In the Rombak and Tysfjord region fluorite is a common accessory mineral. The windows in the Meløy–Gildeskål area have typically magnetite crystals up to 5 mm across in the granitoids, and even 3–4 mm crystals of uraninite are found. This province, showing high background values for uranium and thorium, has been subjected to car-borne surveys, field work and a large analytical programme. The northern part of the province shows an approximately normal crustal U/Th ratio, but in the southernmost parts, in the Rana region, the U/Th ratio is close to 1.

So far, no occurrence of economic significance has been discovered, but the province contains large areas of granitoid rocks with 10–50 ppm U. Hydrothermal activity and alteration has only been effective on a small scale, and only limited mobilization of uranium has been recognized. Local enrichments have been found in pegmatites, thin veins, and along the borders of amphibolite bodies. Significant mineralizations are found within a zone along the basement/cover contact, in foliated granitoids or gneisses of uncertain origin, and occasionally in arkoses. The mineralizations are probably in most cases of younger age, and could be derived from sources other than intrusives. They should therefore be classified as sedimentary or metamorphic.

Leucogranites, often pegmatitic, including Orrefjell, intrude several windows(?) of basement gneiss along a N–S, 15 km long trend in the Salangen valley in Troms county. In one intrusion at Orrefjell a uranium deposit was discovered in the 1960's and described by Sverdrup et al. (1967). In this region the Caledonian nappes are thin and authochthonous basement outcrops in some of the valleys (Gustavson 1974). Slices of basement rocks are also included in the nappes. The Orrefjell basement could be such a slice; it occurs on top of a hill approximately 300 m above the valley floor of presumed autochthonous basement.

Fig. 4 presents a simplified geological map of the Orrefjell area by Rundberg and Rindstad (Rindstad 1982). The uranium mineralization occurs irregularly over a length of 1.5 km along the western margin of the window(?), with a thickness up to 20 m. The host rock is the leucogranite pegmatite, and the contact with the Caledonian rocks dips at 45° towards the west. The host rock consists of coarse-grained white microcline, albite and quartz, with small but varying amounts of biotite, muscovite and chlorite. Locally in the mine-

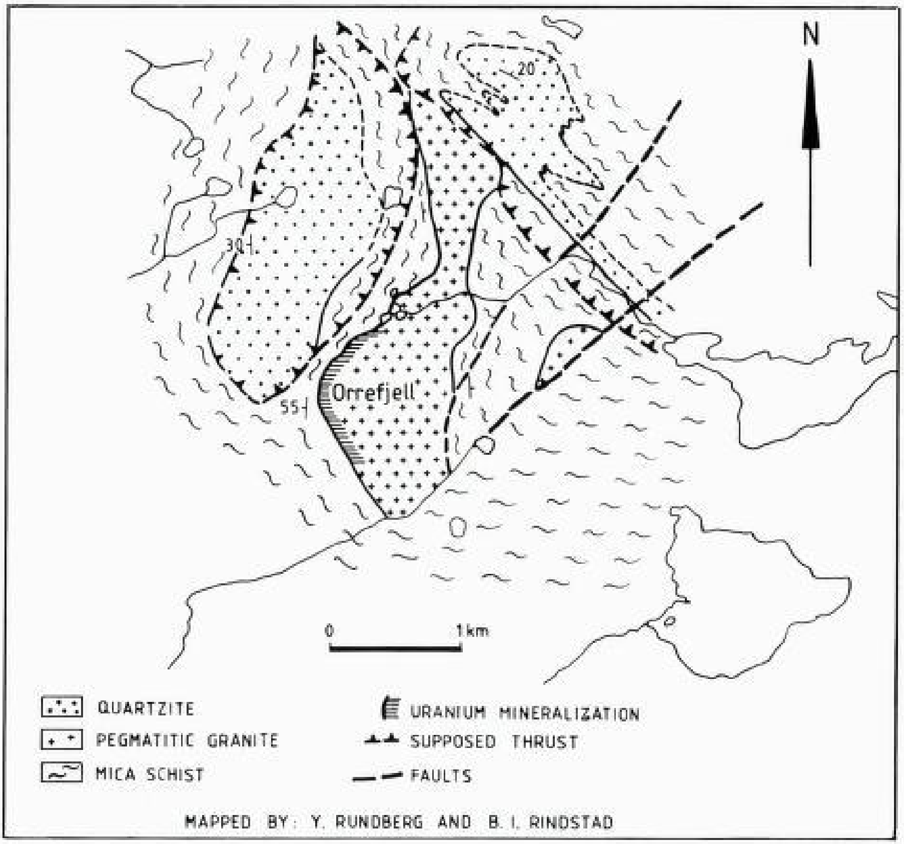

Fig. 4: Simplified geological map of the Orrefjell area; by B. I. Rindstad.

ralized zone, magnetite and iron sulphides occur (Rindstad 1982). Lenses and bands of Precambrian intermediate gneiss and amphibolite are also found. Evidence of propyllitic alteration in the amphibolite has been observed. In the Caledonian mica schist several percent of iron sulphides occurs, and also some graphite. The main uranium mineral seems to be uraninite, which can occur in crystals up to 2–3 mm in size. So far, however, no mineralogical study has been done. The secondary mineral uranotile was identified by Thorkildsen (Sverdrup et al. 1967). Molybdenite is observed locally within the mineralized zone.

The deposit was drilled during the years 1979–81, and split drill-core from the pegmatite was analysed in lengths of 1–2 m. The results of a typical section are given in Table 1. The mineralogy of the leucogranite pegmatite corresponds to the definition of alaskite (Spurr 1900), and the analyses are nearly identical to those of the alaskites in the Rössing uranium deposit in Southwest Africa (Berning et al. 1976, M. Cuney pers. comm.).

Age determinations were made on the Orrefjell pegmatite by the Rb–Sr method and on the uraninite from the mineralization by the U-Pb method. The sampling and Rb–Sr determinations were made by A. Andresen. Urani-

Table 1: Analyses of 1–2 m long samples of drill-core from Orrefjell. Analysed at NGU by XRF-method (All values in wt. percent. L.O.I. = loss of ignition. – = less than).

Sample nr. Element	114	115	116	117	118	119	Average
SiO_2	75.23	73.39	73.62	75.04	77.50	74.55	73.9
Al_2O_3	13.03	14.37	14.46	13.36	12.25	13.78	13.5
Fe_2O_3	0.13	0.23	0.27	0.23	0.27	0.60	0.3
TiO_2	0.03	0.01	0.01	0.03	0.05	0.04	0.03
MgO	−0.01	−0.01	−0.01	−0.01	−0.01	−0.01	−0.01
CaO	0.34	0.54	0.83	0.23	0.67	0.58	0.5
Na_2O	2.70	3.10	3.60	2.40	3.20	4.40	3.2
K_2O	7.17	7.42	6.38	9.17	5.64	4.73	6.8
MnO	−0.01	−0.01	−0.01	−0.01	−0.01	0.22	−0.01
P_2O_5	0.02	0.03	0.03	0.05	0.03	0.02	0.03
L.O.I.	0.39	0.50	0.61	0.21	0.38	0.29	0.4

nite concentrates for U-Pb dating were separated at NGU and the isotope determination was made at IGS (Institute of Geological Science) in London by I. G. Swainbank. The data are in preparation for publication (Andresen, Lindahl et al.). A Rb/Sr isochron age of 1,600 m.a. is indicated. There is a large scatter around the isochron due presumably to the sericitization of the feldspar. The U-Pb dating gives an age of 1,745 m.a. The mineralization crystallized at that time, but either lost lead or gained uranium during a late Caledonian episode (360 m.a.). The results indicate that alaskite intrusion in the Precambrian basement took place at around 1,750 m.a. ago and that the uranium mineralization belongs to the same intrusive event.

Several granites in southern Norway were emplaced during the last stages of the Sveconorwegian orogeny. Some of these have an anomalously high radio-element content (Killeen & Heier 1975a), e.g. the Båhus–Iddefjord–Flå belt. Killeen & Heier (1975b) have studied ten granites in southern Norway, some of them enriched in uranium and thorium. The Bessefjell granite analysed by Killeen & Heier (1975b, 13 samples) gave an average of 13.3 ppm U and 54.7 ppm Th. Grid-sampling of the granite (87 samples) by NGU gave 10 ppm U and 51 ppm Th (gamma-spectrometer). The Homme granite, which has been studied by Falkum & Rose-Hansen (1978), also gave anomalously high uranium and thorium values.

The enriched granites described from southern Norway have lower radio-element contents than those in northern Nordland. Economically significant grades have not been registered. Late-stage hydrothermal alteration connected with the granites is limited, but a number of pegmatites with uranium-bearing minerals have been studied, in some cases in detail. These are the pegmatites in the counties of Øst- and Vest-Agder, e.g. Einerkilen and Rogaland.

A pegmatitic uranium mineralization of this type has been found at Bagn in Valdres, and belongs to the late stage of the Flå granite (D. van der Wel, pers. comm.). One uraninite sample gave an isotopic composition indicating crystallization at about 900 m.a., influenced by the late-stage Caledonian event. Uranium occurs in the pegmatite as primary uraninite, but is also found in fractures. The uranium content exceeds that of thorium.

Parts of the Fen carbonatite complex are enriched in uranium and should genetically be classified as intrusive deposits. The complex is thorium dominated, and niobium and REE occur in significant amounts.

CONTACT METASOMATIC DEPOSITS

Typical contact metasomatic mineralization in calc-silicate rocks has not yet been identified in Norway.

METAMORPHIC DEPOSITS

The Proterozoic rocks in Norway, containing most of the uranium occurrences, are always metamorphosed to some extent. Depending on geochemical conditions, uranium may or may not be leached and moved by metamorphic fluids. Examples in which the uranium has not moved include the alum shale of the Oslo region and graphitic schists of the Caledonides. In the Elsjø area (Olerud 1982) it has not been possible to register mobilization of uranium even where the host rock is completely hornfelsed. Fluids have passed through the alum shale without moving the uranium.

Under oxidizing conditions fluids can leach uranium, and precipitate it elsewhere under reducing conditions. If uranium is available and the geochemical trap is effective, uranium deposits can be formed. When the rock units are overprinted by several metamorphic events it is difficult to classify, for example, the vein deposits. Fluids may have been generated through metamorphic processes, as well as from intrusives.

Rock may be depleted in uranium during metamorphism. Some investigators regard depleted rocks with less than Clark values as a guide to ore, assuming the uranium to have been mobilized and deposited elsewhere. Other geologists look for regions with anomalously high uranium contents, as such rock types may be a protore to deposits if hydrothermal processes have been active. Both lines of reasoning may have merit.

Krause (1980) discovered a vein uranium mineralization in the Porsa sulphide deposits of the Komagfjord window in west Finnmark. The mineralized veins cut the sulphide layers, and the highest uranium contents are found at intersections with the sulphides, which act as a reducing agent on the fluids. Most likely the fluids were of metamorphic origin. Krause (1980) believes that black shales in the area were the source rock for the uranium. Experience from other areas shows that leaching of uranium from the black shales is an ineffective mechanism.

On Kvaløya west of Tromsø a uranium mineralization has been found in a shear zone in granitized basement (Sverdrup et al. 1967). The background radiation in the basement is moderate, but with several anomalous points and narrow veins. The uranium was probably mobilized by the process of granitization.

The Berg copper mine, located at Borkenes in Kvæfjord west of Harstad, has been studied by Often (1982). During the sulphide exploration, localli high radiation was discovered. The Precambrian country rock comprises gneisses of mafic to intermediate composition, with layers interpreted as agglomerates. The sulphide deposit, located where a granite intrudes the gneisses, contains up to 2% Cu, up to 100 ppm Ag and traces of gold. The uranium content is on average low, but some samples gave up to 0.3% U. U-Pb dating on uraninite (3 samples) from the Berg mineralization indicates a Caledonian age. If that is correct, the uranium must be epigenetic and deposited in late Caledonian time. No granites of this age are known in the region, and the fluids are thought to be of metamorphic origin.

Studies of basement/cover relations in northern Nordland have led to the discovery of occurrences of molybdenum and uranium. One promising mineralization is located at Harelifjell SE of Straumen in Sørfold (Fig. 5). The host rock belongs to the Rishaugfjell basement window. The locality is near the basement/cover border zone, and influenced by the Caledonian thrusting. The basal sedimentary sequence occurs as remnants below the thrust-plane on Harelifjell mountain, with meta-arkoses and in one locality a thin conglomerate. The Harelifjell occurrence has not yet been mapped in detail. The mineralization is located in a vein or series of veins steeply dipping to the east. A schematic section is shown in Fig. 6. The mineralization outcrops continuously over a length of 300 m, and the radiometric anomaly, as determined by the helicopter-borne surveys, extends about 5 km further to the south. The mineralization lies in a relatively finely grained gneissic rock of granitic composition, which originally may have been arkose. In the mineralized zone 1–3% sulphides are present, and the main uranium mineral is uraninite. The isotope composition has been determined, indicating the deposition age to be late Caledonian (400 m.a.). No intrusion of that age is known in the region; thus the mineralization is thought to have been deposited by metamorphic fluids, with the sulphides acting as a chemical trap.

The Øksnanuten uranium occurrence has been described (Sverdrup et al. 1967) as a series of thin quartz veins intersecting rocks of several types, partly ultramafic (serpentinite). The contrast in chemistry may also here have caused deposition.

SUPERGENE DEPOSITS

According to Barbier (1974), supergene deposits could be expected to be found along old discontinuities as a result of recent weathering, along breccia zones and in permeable rocks with groundwater perculating through them.

The Njallaav'ži uranium deposit in western Finnmark is thought to belong to the supergene type. The geology in the region has been described by Fareth et al. (1977) and the prospecting by Lindahl et al. (1979). The mineralization was first described by Gjelsvik (1957). The uranium mineralization occurs within brecciated albite dolerite. In addition, uranium concentrations occur in outcrop in brecciated syenite and carbonate breccia (Gjelsvik 1957).

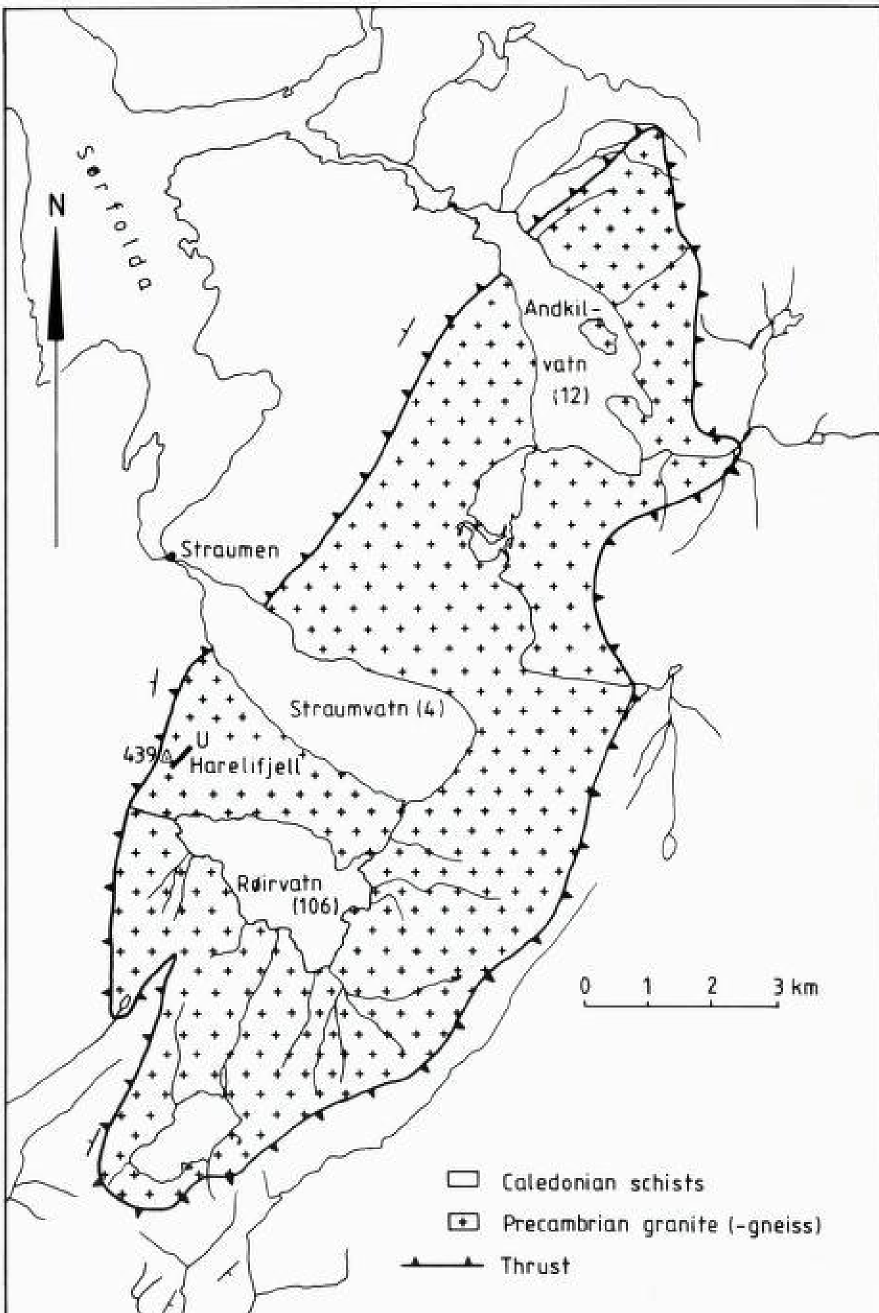

Fig. 5: Location of the Harelifjell uranium occurrence within the Rishaugfjell Precambrian window. Anomalies continue 5 km towards the south.

North of the deposit a circular dome structure mapped by Fareth et al. (1977) consists of granite and acidic gneisses. A quartzite, with or without fuchsite, occurring along the rim, especially towards the east, is overlain by green-

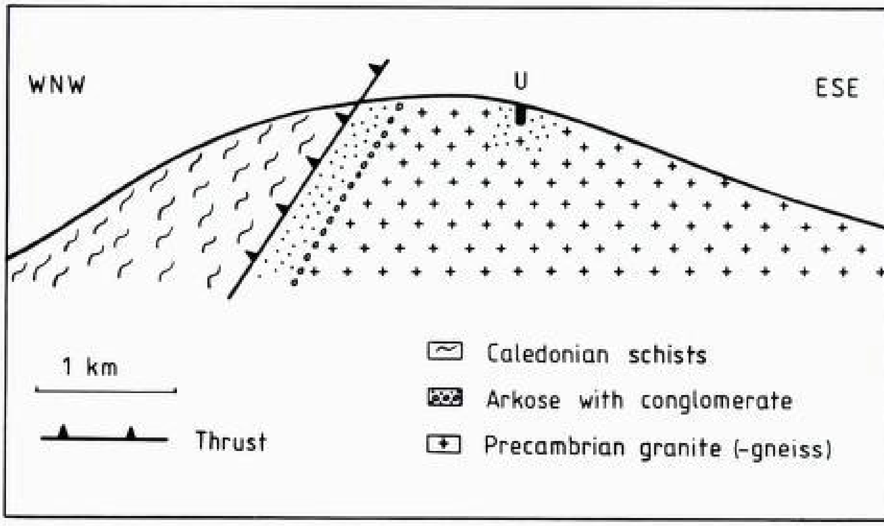

Fig. 6: Schematic profile of the Harelifjell uranium occurrence. The host rock is fine-grained gneiss within coarse-grained granite with weak foliation.

stone. The gneisses and granite could be of Archaean age; the quartzite and greenstone are probably younger.

Two diamond drill-holes intersect the mineralization. The best of these contained 1,800 ppm U over a 1 m thick zone (Th less than 20 ppm). The breccia contains colloidal pitchblende and, as gangue minerals, fine-grained haematite, calcite and chlorite (± biotite). Accessory minerals such as pyrite, chalcopyrite, bornite and galena have been identified locally (Gjelsvik 1957). Other uranium minerals are rare, but Thorkildsen (Sverdrup et al. 1967) identified uranotile and Often (1975) identified liebigite. The intensely brecciated host rock has undergone hydrothermal alteration. Preglacial weathering of sulphide deposits in this part of Finnmarksvidda has been described by Gjelsvik (1956) and Lindahl (1976), indicating that glacial erosion was locally slight. In the greenstone belt of the Råggejav'ri area, north of the supposedly Archaean dome and 12 km NW of Njallaav'ži, there is an iron sulphide deposit outcropping approximately 100 m below the late Precambrian peneplain. From this deposit Lindahl (1976) described slightly metamorphosed weathering textures in pyrrhotite below the preglacial weathering zone. The metamorphism is thought to be Caledonian and the weathering late Precambrian.

The uranium mineralization in Njallaav'ži, where the depth below the peneplain is greater or of similar magnitude to the depth in the Råggejav'ri area, could have the same origin. Dating of the uraninite in Njallaav'ži (two samples) indicates a deposition age of 990 m.a., with influence from either lost lead or gained uranium in late Caledonian time. This age fits with the model described above and a deposition of uranium during late Precambrian peneplanation. We do not know of younger intrusives of this period.

As yet we have no other examples of deposits of supergene origin.

Discussion

Geological modelling is important for selecting target areas for uranium prospecting. Experience will lead to better models, for which high-quality geological maps are very important. Mapping programmes have advanced markedly in recent years, providing better structural and genetic understanding. This is of great help for geological modelling in all prospecting. In Norway coverage by the various exploration methods is uneven, and is areally very limited for the more expensive techniques. Stream sediment coverage is relatively good, but many areas regarded as promising from a uranium resource point of view have not yet been sampled. This paper has attempted to assign the classification system of Dahlkamp (1978) to Norwegian occurrences based on limited information regarding geological and mineralogical data and preliminary age dating using the U-Pb method. The grouping of the deposits is speculative, and future revision must be expected.

The Precambrian granitoid rocks of 1,700–1,800 m.a. age in northern Nordland seem to define a uranium metallogenic province, possibly an extension of the uranium province in the Arjeplog–Arvidsjaur district defined in Sweden by Adamek & Wilson (1979). The province relates to the SSW margin of an early Proterozoic continent which continues towards the west beneath the Caledonian mountain belt.

The present prospecting campaign has been operating for eight years, of which the last 2–3 years have given some success. The potential for uranium deposits in Norway has been assessed by Cuney et al. (1981), and it is regarded as good. They foresee that further prospecting will lead to the discovery of new occurrences, and point to favourable areas. Rock samples collected in the uranium programme and other programmes have been analysed for 15–20 elements. An important spin off is that our results can be used in prospecting for other metals. Correlations have been found to exist between uranium and such metals as Mo, W, REE and Sn in certain provinces, a relationship which will be further studied. Radiometric survey could therefore be important in prospecting for certain metals.

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REFERENCES

Adamek, P. 1975: Geology and mineralogy of the Kopparåsen uraninite–sulphide mineralization, Norbotten county, Sweden, *Sveriges Geol. Unders., ser. C, nr. 712, 69 pp.*

Adamek, P. M. & Wilson, M. R. 1979: The evolution of a uranium province in northern Sweden. *Phil. Trans. R. Soc. Lond. A. 291, 355–368.*

Armands, G. 1970: A uranium-bearing layer from the Lower Ordovician, Sweden. *Geol. För. Forh. 92,* 481–490.

Barbier, M. J. 1974: Continental Weathering as a possible origin of vein-type uranium deposits. *Mineral. Dep. 9,* 271–288.

Berning, I., Coole, R., Hiemstra, S. A. & Hoffmann, U. 1976: The Rössing deposit, South West Africa. *Econ. Geol. 71*, 351–368.

Bjørlykke, A. 1977: Om blyforekomster i Eokambriske/Kambriske sandsteiner på det Baltiske Skjold. In: Bjørlykke, A., Lindahl, I. and Vokes, F. M., eds. Kaledonske Malmforekomster: BVLI–Malmsymposiet, 72–76.

Cuney, M., Taylor, D. M. & Wilson, M. 1981: IUREP–Orientation Phase report, NORWAY. *IUREP-report 1981*, 102 pp.

Dahlkamp, F. J. 1978: Classification of uranium deposits. *Mineral Dep. 13*, 83–104.

Edling, B. 1974: Uranfördelningen i överkambrisk alunskiffer från Ranstad, Billingen, Västergötland. *Publ. Paleont. Inst. Univ. Uppsala, spec. vol. 2*, 118 pp.

Falkum, T. & Rose-Hansen, J. 1978: The application of radioelement studies in solving petrological problems of the Precambrian intrusive Homme granite in the Flekkefjord area, South Norway. *Chemical Geology, 23*, p. 73–86.

Fareth, E., Gjelsvik, T. & Lindahl, I. 1977: Cier'te. Description to map sheet 1933 II – 1:50000. *Norges geol. Unders. 331*, 28 pp.

Gallagher, M. J., Michie, U. M., Smith, R. T. & Haynes, L. 1971: New evidence of uranium and other mineralization in Scotland. *Trans./Sect. B, Inst. Min. Mt. 80*, 150–173.

Gee, D. 1972: The regional geological context of Tåsjø uranium project, Caledonian front, Central Sweden. *Sveriges geol. Unders. C671*, 1–36.

Gee, D. G. & Zachrisson, E. 1979: The Caledonides in Sweden. *Sveriges geol. Unders. C769*, 48 pp.

Gjelsvik, T. 1956: Pre-glacial forvitringsfenomener i kopperforekomster i den syd-vestlige del av Finnmarksvidda. *Geol. För. Förhandl. 78*, 659–665.

Gjelsvik, T. 1957: Pitchblende mineralization in the Precambrian plateau of Finnmarksvidda, Northern Norway. *Geol. För. Förh. 79*, 572–580.

Gust, J. & Thoresen, H. 1981: Radiometriske målinger og prøvetakning i Rendalsvik grafittfelt, Meløy, Nordland. *Unpubl. NGU-rapp. nr. 1650/59A*, 16 pp.

Gustavson, M. 1974: Geologisk kart over Norge, berggrunnskart NARVIK 1:250000. *Norges geol. Unders.*

Ihlen, P. 1982: Sæteråsen niob-forekomst. *Unpubl. NGU-rapp. nr. 1800/76A*, 11 pp.

Killeen, P. G. & Heier, K. S. 1975a: A uranium and thorium enriched province of Fennoscandian shield in Southern Norway. *Geochim. Cosmochim. Acta. 39*, p. 1515–24.

Killeen, P. G. & Heier, K. S. 1975b: Th, U, K and heat production measurements in Ten Precambrian granites of the Telemark area, Norway. *Norges geol. Unders. 319*, 59–82.

Krause, M. 1980: Some uranium mineralizations in the Raipas Suite of the Komagfjord tectonic window, Finnmark, Norway. *Norges geol. Unders. 355*, 49–52.

Lindahl, I. 1976: Prospektering på kartblad Cier'te 1974. *Unpubl. NGU-rapp. nr. 1243/1A-I*, 18 pp.

Lindahl, I. & Heier, K. S. 1977: Possible uranium mineralizations in the light of recognized geological and geotectonic structures in Norway. In: «Recognition and evaluation of uraniferous areas» – IAEA p. 123–129.

Lindahl, I., Bergstrom, B., Fareth, E. & Often, M. 1979: Heavy-metal exploration in the western part of Finnmarksvidda, northern Norway. In: Prospecting in areas of glaciated terrain 1979. *Inst. Min. Met.*, 59–66.

Lindroos, H. & Smellie, J. 1979: A stratabound uranium occurrence within Middle Precambrian ignimbrites at Duobblon, northern Sweden. *Econ. Geol. 74*, 1118–1130.

Mathiesen, C. O. 1970: An occurence of unusual minerals at Bidjovagge, Northern Norway. *Norges geol. Unders. 266*, 85–104.

Michie, U. M. 1972: Further evidence of uranium mineralization in Orkney. *Trans./Sect. B. Inst. Min. Met. 81*, 53–54.

Often, M. 1975: Geokjemisk U-prospektering i Njallaav'ži, Nordreisa, Troms. Unpubl. thesis, Technical Univ. of Norway, Trondheim, 65 pp.

Often, M. 1982: Geologiske undersøkelser og diamantboring ved Berg Gruve, Borkenes. *Unpubl. NGU-rapp. nr. 1575/1B*, 19 pp.

Olerud, S. 1982: Elsjøfeltets geologi, tektonikk og innhold av sporeelementer i Kambro–Ordoviciske skifre. *Unpubl. NGU-rapp. nr. 1650/16D*, 49 pp.

Rindstad, B. I. 1982: Geologiske undersøkelser på Orrefjell, Salangen, Troms. *Unpubl. NGU-rapp. nr. 1800/48D, 12 pp.*

Rosenquist, I. 1948: Uranforekomster i Norge. *Tidsskr. Kjemi, Bergvesen og met.*, 7, 116–119.

Siggerud, T. (1956): The occurence of Uranium and Thorium in Norway. Proc. Int. Conf. on Peaceful use of Atomic Energy, Geneve, 8.-20. aug. 1955, 6 UN, 178-182.

Skjeseth, S. & Sørensen, H. 1953: An example of granitization in the Central zone of the Caledonides of Northern Norway. *Norges geol. Unders. 184,* 154-187.

Skjeseth, S. 1958: Uran i kambrisk alunskifer i Oslofeltet og tilgrensende områder. *Norges geol. Unders. 203,* 100-111.

Skålvoll, H. 1971: Geologisk kart over Norge, berggrunnskart KARASJOK 1:250000. *Norges geol. Unders.*

Spurr, J. E. 1900: Classification of igneous rock according to composition. *American Geol. 25,* 229-230.

Sverdrup, T. L., Thorkildsen, C. D. & Bjørlykke, H. 1967: Uran og thorium i Norge. *Norges geol. Unders. 250A,* 31 pp.

Tveten, E. 1978: Geologisk kart over Norge, berggrunnskart SVOLVÆR 1:250000. *Norges geol. Unders.*

Sulphur Isotope Composition of the Sandstone–lead Deposits in Southern Norway

ARNE BJØRLYKKE

Bjørlykke, A. 1983: Sulphur isotope composition of the sandstone-lead deposits in southern Norway. *Norges geol. Unders. 380*, 143–158.

Samples of galena, pyrite and barite from the Norwegian sandstone-lead occurences in the Gjøvik area and at Osen and Galåa have been analyzed for their sulphur isotope composition. Samples from the Vassbo and Laisvall deposits in Sweden were also analyzed. The galena samples are characterized by a large variation in the sulphur isotopic composition (+10 to +26 ‰) and they have an average $\delta^{34}S$ composition of +18.56 ‰. The sulphur source has earlier been interpreted to be oil-field waters, and these data are not inconsistent with this interpretation. However, the data also permit, as an alternative source for the sulphur, sea-water sulphate locally reduced by bacteria.

A. Bjørlykke, Norges geologiske undersøkelse, Postboks 3006, – 7001 N-Trondheim, Norway

Introduction

Several deposits and occurrences of lead are known in Late Precambrian to Early Cambrian sandstones of the Baltic Shield. They occur mainly along the eastern and southeastern frontal zone of the Scandinavian Caledonides, over a distance of approximately 2000 km (Fig. 1). Galena is the main base-metal sulphide in the deposits and the Pb/Zn ratio varies from 5–15, but locally within each deposit Zn may be the dominant base metal.

Laisvall is the largest deposit with 80 million metric tons of ore at an average grade of 4% Pb (Rickard et al. 1979). Two different genetic models for these deposits have recently been published. Rickard et al. (1979) proposed a basinal brine model, where dewatering of sediments is envisaged as the process during which lead was complexed and moved towards the basin margins. Precipitation took place when the metal-bearing brine mixed with a sulphide-bearing brine contained in the Laisvall sandstone aquifer. Bjørlykke & Sangster (1981), on the othe hand, suggested that the lead was released during weathering from the underlying granitic basement either directly or via formation of continental arkoses. Groundwater may then have carried the lead (and zinc) towards the basin and sulphides were precipitated when the groundwater met reducing conditions. Bjørlykke & Sangster (1981) also showed that the sulphur isotope composition is dependent upon the depositional environment: galena in continental sediments has a light sulphur isotope composition whereas in marine sediments the sulphur is heavy. On the Baltic Shield the sulphur isotope composition has earlier been analyzed from Laisvall (Rickard et al. 1979) and from Vassbo (Wallin 1980). Both studies showed a heavy sulphur isotope composition in galena.

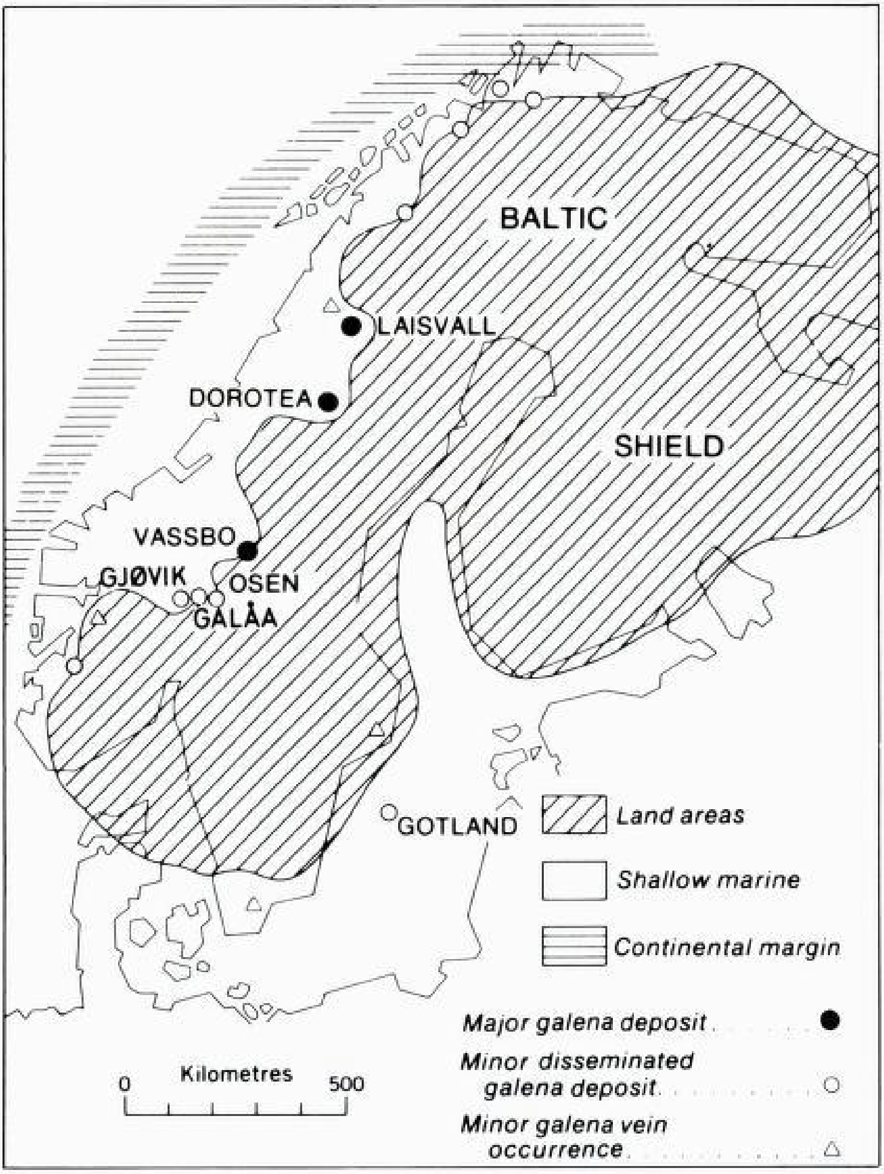
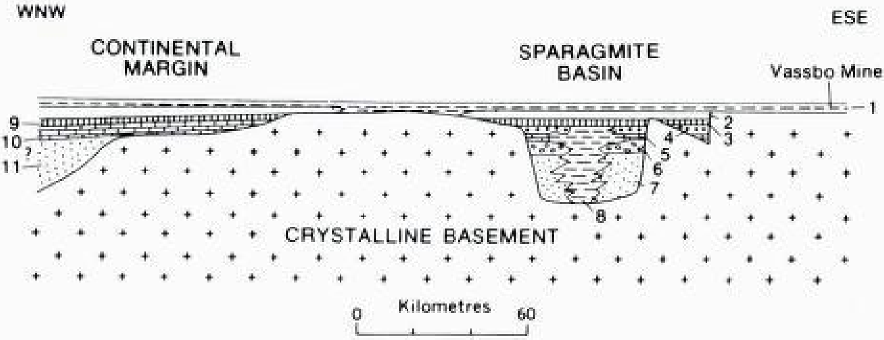


Fig. 1. Stylized paleographic map of the Baltic Shield in Early to Middle Cambrian showing the position of sandstone lead deposits (modified from K. Bjørlykke 1974).

This study compares the sulphur isotope compositions in the Norwegian deposits with those from Laisvall and Vassbo in Sweden, and samples from Norwegian occurrences were therefore analyzed together with a few samples from Laisvall and Vassbo.

GEOLOGICAL SETTING

The Late Precambrian sediments of Scandinavia were deposited along the continental margin and in fault-controlled basins (Sparagmite Basin, Fig. 2).



1. Lower Cambrian shale and sandstone
2. Vangsås Formation (sandstone)
3. Moelv Tillite
4. Ring Formation (arkose)
5. Biri Formation (limestone)
6. Biskopås Conglomerate
7. Brøttum Formation (sandstone)
8. Shale
9. Tillite
10. Carbonate (dolomite)
11. Sandstone

Fig. 2: Reconstructed cross-section through the Baltic Shield in early Paleozoic time showing a rift valley basin (sparagmite basin) and an onlapping continental margin sequence (modified from K. Bjørlykke 1978).

The sediments on the continental margins are widespread and consist mainly of sandstones and dolomites. The Sparagmite Basins contain turbidites (Brøttum Formation), limestones and shales (Biri Formation), and coarse arkoses and conglomerates (Biskopås and Ring Formations).

Deposition of Late Precambrian glacial sediments (Moelv Tillite) was followed by fluvial and/or deltaic sedimentation (lower part of Vangsås Formation). The lowermost Lower Cambrian (Tommotian; Cowie & Cribb 1978) contains quartzites deposited in a tidal or beach environment, and containing *Scolithus* burrows (Skjeseth 1963) (upper part of Vangsås Formation). In most areas fossiliferous Lower Cambrian strata begin with sandstones fining upwards to green shales. During Middle and Late Cambrian times most of the Baltic Shield was covered by a shallow and stagnant epicontinental sea in which mainly black shales were deposited (K. Bjørlykke 1974).

The Caledonian nappe movements occurred in an early Finnmarkian phase (Upper Cambrian–Lower Ordovician; Sturt 1978) and in a Silurian to lowermost Devonian phase. During the later phase, sediments deposited on the crystalline basement or along the continental margin were thrust eastward or southeastward. In the study area the Caledonides are divided into three major tectonostratigraphic units (Gee 1975, K. Bjørlykke 1978):

- Autochthonous sediments lying upon Precambrian basement;
- Lower nappe units with epicontinental, Late Precambrian to Ordovician sediments of low metamorphic grade and moderate deformation;
- Upper nappe units, long-distance transported and strongly deformed and generally comprising fragments of basement (e.g. anorthosites, augen gneisses) and sediments (sandstones and dolomites) deposited along the continental margin.

Lead mineralization occurs in the autochthonous sequences and in the lowermost nappes along the eastern and southeastern border of the Caledonides, suggesting that the lead was restricted to sediments deposited in topographic lows along the marginal areas of the Early Cambrian epicontinental sea (Fig. 1).

Outline of the Geology of the Sampled Areas

In this study sandstone lead occurrences in the Gjøvik area and at Galåa and Osen in Southern Norway were sampled and analyzed together with a few samples from Laisvall and Vassbo in Sweden (fig. 1).

GJØVIK AREA

In the Gjøvik area several occurrences of galene are known to occur in the Vangsås Formation (Fig. 3 and Bjørlykke 1979). The formation is divided into the Vardal Sandstone Member and the overlying Ringsaker Quartzite Member. The Vardal Sandstone is 160 metres thick and consists of arkoses and feldspathic sandstones (Fig. 4), with microcline dominating over plagioclase. The degree of maturity increases upwards and there is a gradual transition into the Ringsaker Quartzite. Conglomerates occur in the southern part of the area (Fig. 3) and thin out rapidly northwards. Red beds are observed

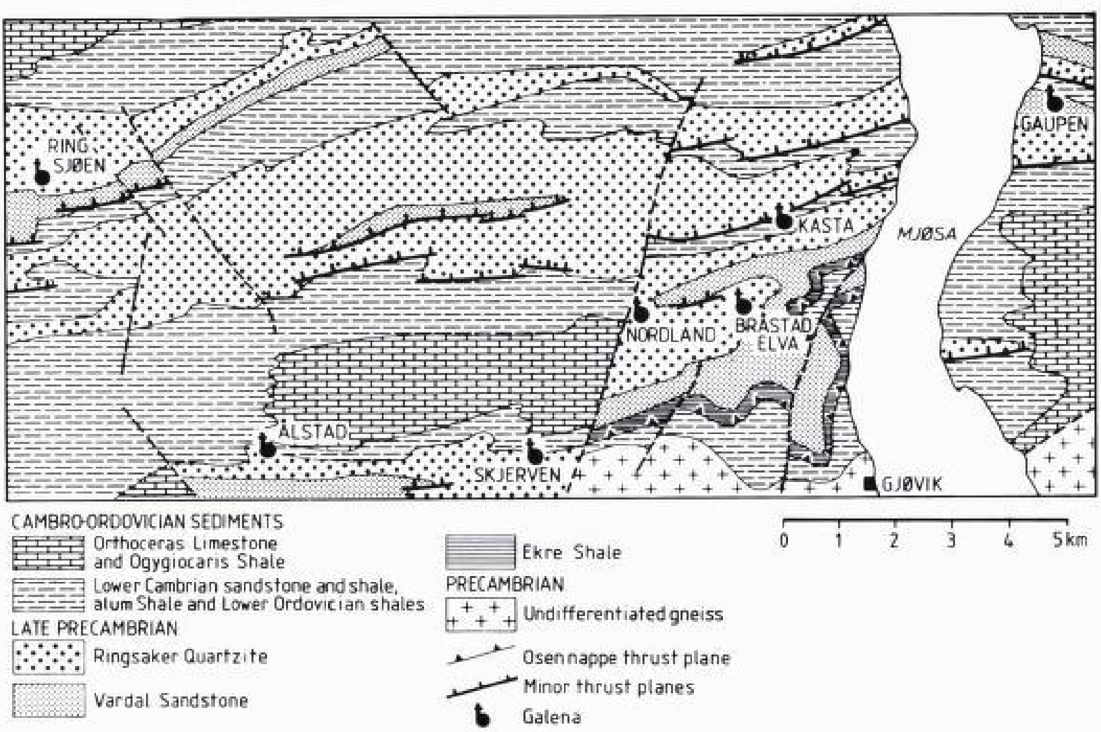


Fig. 3: Geological map of the Gjøvik Area.

in a few places in the lowermost part of the Vardal Sandstones. Thin beds of phosphorites (A. Bjørlykke 1979) indicate a marine origin for at least parts of the Vardal sandstone, which earlier had been interpreted as entirely fluvial (K. Bjørlykke et al. 1976). The Ringsaker Quartzite is 50–80 metres thick. The lower half is a fine- to medium-grained quartz sandstone which is locally rich in altered feldspar and with a sericitic matrix. The upper half is a medium- to coarse-grained quartzite, often with a blue colour caused by a thin film of carbonaceous material enveloping quartz grains. The uppermost part of the Ringsaker Quartzite contains *Scolithus* burrows (Skjeseth 1963), which indicate a shallow-marine environment considered typical for lowermost Lower Cambrian strata (Vidal 1981).

In the Gjøvik area the galena occurs only in the orthoquartzite and mainly as cement often with poikiloblastic texture. Some galena was remobilized during the Silurian nappe movement and fills small cracks and joints in the orthoquartzite. An erosional hiatus beneath the fossiliferous Lower Cambrian strata is marked by a 0.2–0.5 m thick conglomerate consisting of rounded fragments of the underlying Ringsaker Quartzite. The conglomerate is overlain in places by impure limestone followed by sandstones and green shales. The sub-Middle Cambrian break is marked by a conglomerate with pebbles of dark quartzites and phosphorites, overlain by black shales (Alum shale).

During the Silurian the Vangsås Formation and the overlying Cambro-Silurian sediments were thrust towards the south, folded and imbricated.

GALÅA AREA

Galena disseminations in the Galåa area south of Rena were found in the course of a regional geochemistry program (A. Bjørlykke et al. 1973). Middle Cambrian black shales were deposited directly upon weathered Proterozoic basement gneisses. The Osen Nappe, which truncates the autochthonous sequence 10 metres above the basement, consists of the Vangsås Formation and Cambro-Ordovician sediments (Fig. 4).

Because of the imbricated structure of the Osen Nappe it is difficult to find a continuous complete section through the Vangsås Formation. The Vardal Sandstone is approximately 150 m thick. It consists in its lower part of grey and red feldspathic sandstones and arkoses of assumed fluvial origin. The sandstones become more quartzitic in the uppermost part and exhibit a blue-grey colour resulting from an increased organic content. The boundary with the overlying Ringsaker Quertzite is defined by the feldspar content. The Ringsaker Quartzite is approximately 40 m thick and is a white, clean, quartz-cemented quartz-arenite. The quartzite is overlain by shallow-marine fossiliferous fine-grained sandstones and green shales of Early Cambrian age.

The lead mineralization occurs in the blue-grey quartz sandstone of the uppermost part of the Vardal Sandstone both as disseminations and as mobilizations in cracks and joints. Disseminated pyrite occurs frequently also in the grey-to-blue beds of the Vardal Sandstone.

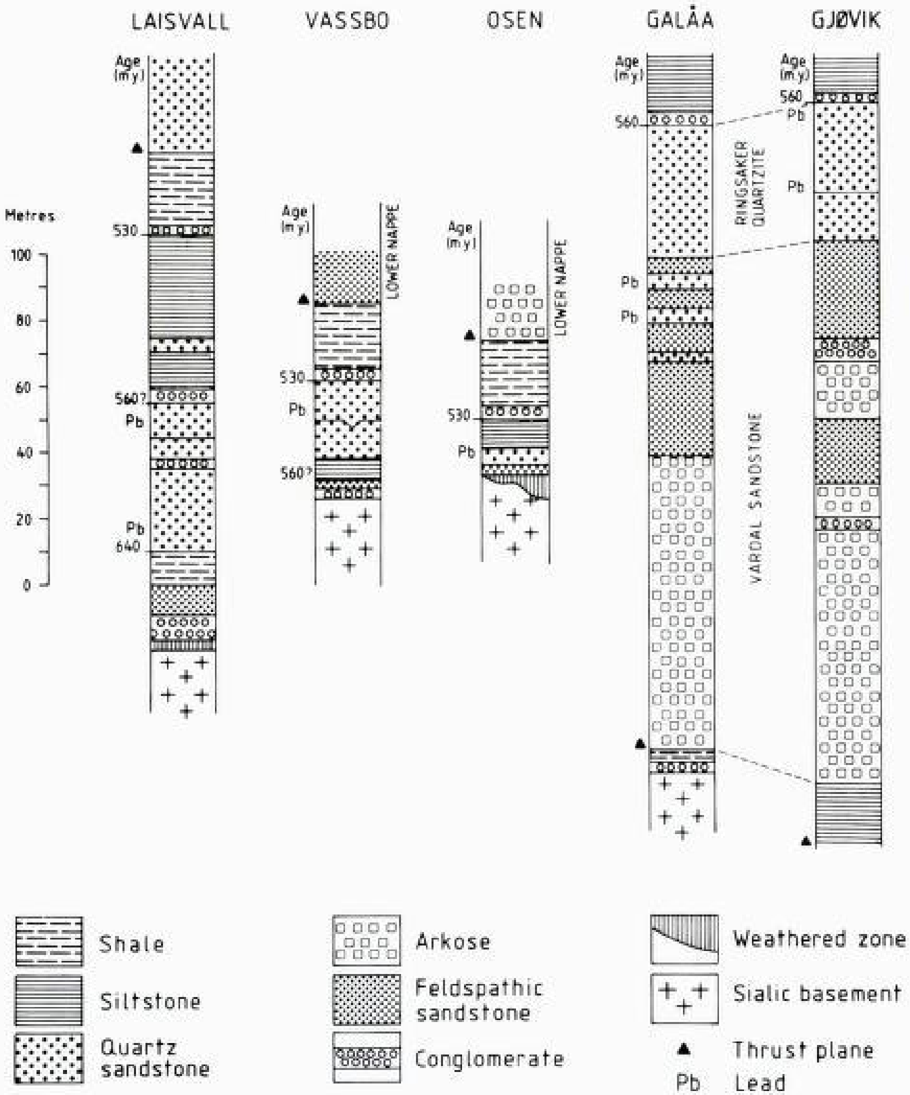


Fig. 4: Lithostratigraphy of five areas of the Baltic Shield containing sandstonelead deposits. Descriptions and bibliographic sources are given in the text.

OSEN AREA

The basement in the Osen area consists mainly of granite with subordinate gneisses of Middle Proterozoic age. The basement forms a peneplain upon which a weathered zone up to ten metres thick was developed. The weathered basement grades upward into a feldspathic sandstone followed by up to 4 m of coarse-grained mineralized quartzite of Early Cambrian age, probably deposited in a beach environment (Høy 1977). It is overlain by 2–3 m of fine-grained, slightly mineralized, dark sandstone of possible tidal origin, which grades into grey, strongly bioturbated siltstones (Nystuen 1969) (Fig. 4).

The Middle Cambrian strata begin with a 0.2 m thick conglomerate containing fragments of shale, quartzite and phosphorite. It is followed by up to 20 m of black shale before the autochthonous sequence is truncated by Caledonian nappes.

LAISVALL

The following data (Fig. 4) are compiled mainly from Rickard et al. (1979), Grip (1967, 1973) and Willdén (1980). The basement in the Laisvall area consists of weathered granites of Proterozoic age. A 1–9 m thick mixtite rests directly on the basement. It is followed by 7–9 m of feldspathic sandstone, with sporadic galena mineralization, overlain by a green-to-grey pebbly shale. This lower part of the sequence is interpreted to be of glacial origin (Rickard et al. 1979) and to have an age of about 654 ± 7 m.y. (Pringle 1973, Sturt et al. 1975).

The glacial sediments are followed by a 40–45 m thick sandstone sequence, which is divided into lower, middle and upper units. The lower unit is up to 25 m thick and consists of a well sorted, medium-grained quartzite with intercalated green shale beds. These are often brecciated or show liquefaction structures. The middle sandstone starts with a conglomerate that contains fragments of the underlying sandstone and truncates the liquefaction structures in the underlying beds. The middle sandstone is 7 m thick and has a more clayey matrix than the lower sandstone; channel structures are also common. The upper sandstone, which starts with a quartzite conglomerate, is 11 m thick with well developed cross-bedding. Fragments of phosphorite occur in the upper part.

The whole sandstone sequence up to the fossiliferous Lower Cambrian (560 Ma) represents a period of approximately 80 million years (Cowie & Cribb 1978); the conglomerate may represent a long period of subaerial exposure. Willdén (1980) has interpreted the lower sandstone, which hosts the main mineralization, to have been deposited in a lagoonal environment. The poorly mineralized middle sandstone was interpreted as a tidal deposit and the well mineralized upper sandstone as deposits formed in tidal channels and on beaches.

Above a thin conglomerate containing fragments of quartzite and phosphorite occurs Lower Cambrian strata, consisting of green shales and siltstones commonly bioturbated and with fossiliferous limestone beds. After a new break in sedimentation between Lower and Middle Cambrian, black shales (alum shale) were deposited in a shallow-marine environment. The sedimentology shows that the Laisvall area was located within a shallow and broad depression surrounded by gently rising areas to the north and south (Willdén 1980). The autochthonous sequence is then truncated by Caledonian thrusts. In the lower nappe, minor lead and zinc mineralization occurs in a blue quartzite which is correlated with the upper sandstone in the autochthonous strata (Willdén 1980).

VASSBO

The following data (Fig. 4) are mainly compiled from Tegengren (1962), Grip (1973), Christoffersen et al. (1979) and Wallin (1982). Basement in the Vassbo area consists of quartz porphyry dated at 1669 ± 38 M.a. (Welin & Lundqvist 1975, Welin 1980), sandstone and diabase dykes. A thin basal conglomerate is overlain by a 1–6 m thick shale of Early Cambrian age, locally with mineralized coarse sandstone beds. The shale is followed by a fine-grained calcite-cemented quartz sandstone, 12–14 m thick, with disseminations of galena in the upper part. A marked erosional disconformity exists between this sandstone and the overlying coarser quartz sandstone. The quartz sandstone, with galena, sphalerite and barite is 7–12 m thick. It contains cross-bedding and ripple-marks, and is interpreted to have been deposited in an open-marine beach environment (Christofferson et al. 1979).

The Middle Cambrian starts with a 0.1 m phosphoritic conglomerate with fragments of quartzite. Some of the fragments contain disseminated galene with a matrix rich in pyrite and minor, possibly remobilized, galena. This is followed by 10–20 metres of black shale before the autochthonus sequence is truncated by Caledonian thrusts.

ORE GEOLOGY

A more detailed description of the ore geology has been published by Rickard et al. (1979), Christofferson et al. (1979) and Bjørlykke & Sangster (1981). The ore bodies are blanket-like in shape and often follow the stratification and other sedimentary structures in the host sandstone. In homogenous sandstones galena appears to have nucleated around organic material, forming poikiloblastic spots and heavier concentration around shale fragments.

The ore minerals occur as cement in zones which were the most permeable parts of the quartz sandstone. The original porosity in the Laisvall deposit was probably around 30% (Rickard et al. 1979) but in the most intensely mineralized parts corrosion and replacement of the detrial grains can be seen. Galena and sphalerite are the most important sulphides; accessory minerals include framboidal pyrite, marcasite, bravoite, pyrrhotite and chalcopyrite (Ncube et al. 1978). Although the galena contains small inclusions of native silver, the silver content of the ore is low, ranging between 7 and 21 g/t (Grip 1973). The sphalerite is mostly light-coloured and has an iron content between 1% and 2% and a Cd content of 0.3% (Rickard et al. 1979). Bravoite was found at four different places in Laisvall, but the nickel content in the bravoite-bearing rocks was found to be only 17–41 ppm (Ncube et al. 1978). The non-sulphide cement consists dominantly of quartz, with some calcite, fluorite and barite. The non-quartz cement, including the sulphides, occupies on average about 6% of the rock volume (Rickard et al. 1979).

Isotope Results

The samples were crushed and the sulphides and sulphates separated at the Geological Survey of Norway. The analytical work was performed at

Geochemisches Institut der Universität Göttingen. The error limit is generally within 0.1‰ for the sulphides and 0.3 ‰ for the sulpates (Nielsen & Sperling 1973).

The results of the sulphur isotope analysis of galena, pyrite and barite from the sandstone-lead deposits in Southern Norway and from Laisvall and Vassbo in Sweden are listed in Table 1. The samples are characterized by a large variation in the sulphur isotope composition of the galena and most of them are isotopically heavy.

Table 1. Sulphur isotope analyses.

No.	Deposit	$\delta^{34}S$
	GALENA SAMPLES	
1	Ringsjøen, Gjøvik	22,3
2		− 0,1
3		25,7
4		17,8
5		25,6
24		14,2
25		13,7
26		14,4
27	Ålstad, Gjøvik	12,9
28		12,2
29		11,8
44	Skjerven, Gjøvik	17,7
45		14,2
46		14,5
30	Kasta, Gjøvik	21,0
31		17,7
32		17,1
42	Nordland, Gjøvik	14,2
43		15,3
33	Tjerne, Gjøvik	18,2
34		14,6
35		16,8
38	Vassbo ore zone	18,6
39		18,5
40		10,2
41		13,7
36	Laisvall ore zone	20,1
37		19,9
6	Galåa	22,9
7		19,7
8		22,8
9		20,6
10		19,8
11		20,8
12		19,3
13		33,4
14		10,8
15		37,1
114		21,0
115/119		21,5
16	Osen	22,3
17		20,0
18		20,2
19		19,8
20		20,8
21		18,4
22		18,0
23		20,7
47		18,0
48		16,7
49		17,2
101		18,7
102/105	Middle Cambrian black shale	0,5
103/109		21,4
	PYRITE SAMPLES	
116	Galåa	28,5
117		14,9
118		28,5
115/119		31,6
126	Vassbo	18,4
128		18,0
102/105	Osen–Black shale	51,2
103/104		16,0
106	Sub-middle Cambrian conglomerate	−20,5
107	Lower Cambrian siltstone	−25,6
110	Middle Cambrian black shale	20,8
111		21,5
112		33,5
113	Lower Cambrian siltstone	−22,5
	BARITE SAMPLES	
127	Vassbo	20,7
134		18,9

A frequency distribution of the $\delta^{34}S$ values in 47 galena samples from Southern Norway is shown in Fig. 5. Except for three samples they all fall between +10 and +26‰ and they have an average $\delta^{34}S$ composition of +18,5‰. The sulphur isotope composition of the sulphides in the Galåa deposit and in the Gjøvik area shows a larger spread than in the Osen deposit (Fig. 5). The average composition in the Gjøvik area is +16.0‰, Osen + 18.3‰ and Galåa +22.5‰ (Fig. 5). Only two galena samples from Laisvall and four from Vassbo were analyzed and the results are within the spread in the data from Southern Norway. In Fig. 5 data published by Rickard et al. (1979) are combined with our data and the average $\delta^{34}S$ composition of galena-sulphur is +22.87‰.

One pyrite-galena pair from Osen and two from Galåa were analyzed (Table 1), and from Laisvall two pairs of sphalerite–galena have been published (Rickard et al., 1979). None of the pairs shows isotopic equilibrium during precipitation. Two samples of barite from Vassbo were analyzed and their sulphur isotope compositions are similar to the average composition of

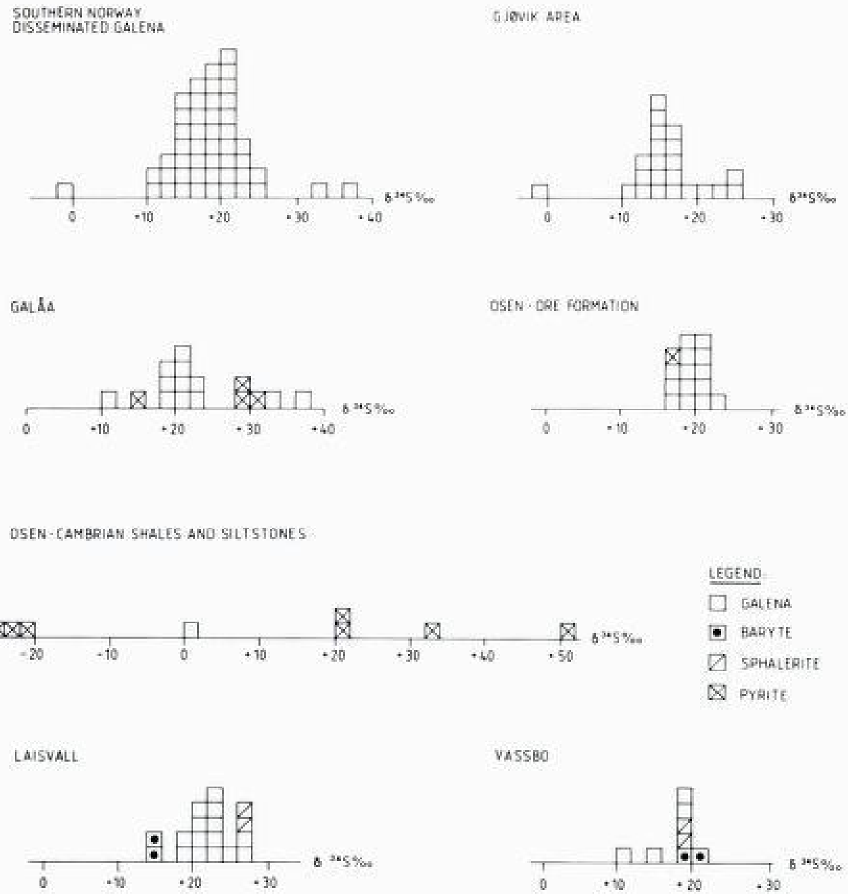


Fig. 5: Sulphur isotopic composition($\delta^{34}S$) in sandstone-lead deposits on the Baltic Shield. The data from Laisvall are partly from Rickard et al. (1979).

the galena. The sulphur isotope composition in barite from Laisvall (Rickard et al. 1979) is 7‰ lighter than the sulphide-sulphur. In Osen. eight samples were taken from above the mineralized zone, from the Lower Cambrian siltstone and from Middle Cambrian black shales. They show a large variation in sulphur isotope composition, from −26 to +51‰ (Fig. 5).

Discussion

SULPHIDE–SULPHUR

The sulphur isotope study reported here shows that sulphides from the Norwegian occurrences have a lighter average and a larger spread in isotopic composition than those from Laisvall published by Rickard et al. (1979). These authors concluded that the Laisvall data are not inconsistent with those expected for sulphide–sulphur in oil-field waters. Solutions low in base metals with hydrocarbon and reduced sulphur could have been supplied during either the late stages of the Caledonian orogeny (as proposed by Rickard et al. 1979) or, alternatively, somewhat earlier from Late Precambrian rift basins.

The new data reported here from Norwegian sandstone-lead occurrences are consistent with the oil-field water model proposed by Rickard et al. (1979) for the Laisvall deposit in Sweden. However, the data also permit, as an alternative source for the sulphide–sulphur, sea-water sulphate reduced locally by bacteria. Data compiled by Bjørlykke & Sangster (1981) indicate that the sulphur isotope composition of galena in sandstone–lead deposits reflects the depositional environment of the host rocks. For example, deposits in fluvial sandstone, such as Maubach, Mechernich and Freihung in Germany (Jensen 1967) have isotopically light sulphur similar to red-bed copper and sandstone–uranium deposits (Sangster 1976). In contrast, lead deposits in shallow-marine sandstones on the Baltic Shield have an isotopically heavy composition, not unlike that in Mississippi Valley-type deposits.

A sea-water source for the sulphur in Mississippi Valley-type deposits was proposed by Sangster (1976). In comparing various deposits which may have sea-water sulphate as their sulphur source, one must take into account the change in isotopic composition in sea-water with time. During the Late Precambrian and Early Cambrian the sea-water contained isotopically heavy sulphur with a $\delta^{34}S$ composition between +30 and +35‰ (Claypool et al. 1980). The fractionation in sulphur isotope composition between marine sea-water sulphate (+30 and +35‰) and galena in the sandstone-lead deposits in southern Norway (av. = +18.56‰) is about 15 per mil. This is the average fractionation value found by Sangster (1976) for deposits in marine sediments and supports the seawater–sulphate source for sulphide–sulphur in the sandstone–lead deposits.

The seawater–sulphate model can be examined in somewhat more detail by consideration of the local depositional environment in a manner described by Schwarcz & Burnie (1973). The sandstone–lead deposits on the Baltic Shield occur in sandstones deposited in a shallow-marine, partly tidal environment. According to Schwarcz & Burnie (1973) this environment is a partly closed sulphur system, both with respect to sea-water sulphate and H_2S gas. In a partly closed system such as this, Δ will range between −3 and + 25‰ ($\Delta = \delta^{34}S_{SO_4{}^{2-}} - \delta^{34}S_{H_2S}$) and is assumed to be a function of both the nature of the reducing organism and the depositional environment. Most galena samples from southern Norway have isotopic compositions ranging between +10 and +26‰ (Fig. 5). The seawater–sulphate in Early Cambrian seas had an isotopic composition between +30 and +35 (Calypool et al. 1980), which gives Δ-values between +4 to +25‰. These are within the range of Δ-values given by Schwarcz & Burnie (1973) for bacterial seawater–sulphate reduction in a tidal environment.

The data from the ore zone and the overlying sediments in Osen fall into three groups corresponding to the lithostratigraphy. 1) The tidal sandstone has heavy sulphur isotopes, indicating a closed system during sulphide deposition. 2) The overlying strongly bioturbated siltstone and the conglomerate (Fig 4) have a light isotope composition (−20.5‰ to −25.6‰). The marked change in composition reflects a transition to an open system both for H_2S and $SO_4{}^{2-}$ possibly due to the bioturbation. The hiatus between Early and Middle Cambrian, which led to the formation of the conglomerate, could also have oxidized previously formed sulphides, changing the $SO_4{}^{2-}$ in the sediments towards a lighter isotope composition. 3) The sulphide samples from the Middle Cambrian black shale (alum shale) have a heavy isotope composition (+0.5‰ to 51.2‰). Black shale environments can produce sulphide–sulphur of both light and heavy composition, Schwarcz & Burnie (1973) and the heavy isotopes at Osen indicate that the alum shale was deposited in a closed system where the influx of marine sulphate was restricted.

The sulphur isotope data from Galåa and Gjøvik show a larger spread than in the Osen deposits. In Galåa and Gjøvik the mineralizations occur in the Osen Nappe and have been more remobilized than in Osen, which could have affected the sulphur isotope composition. There is also a change in the average sulphur isotope composition from area to area. One reason for this change can be that the ore occurs in sediments of different ages but we have too sparse information on the detailed changes in the composition of the ocean water in Upper Precambrian and Early Cambrian to correlate these two factors.

SULPHATE-SULPHUR

Rickard et al. (1979) showed that the barite–sulphur in Laisvall was 7 per mil lighter than the sulphide–sulphur (Fig. 5). In Vassbo, the isotopic composition of the sulphate ($\delta^{34}S$ = 20‰) is 3.6‰ heavier than the composition of sulphide–sulphur (Fig. 5). Because of the large difference between this value and that of Early Cambrian seawater ($\delta^{34}S$ = +30 to +35‰), the latter was considered unlikely to have been the source of sulphate in barite. Rickard et al. (1979) found that the sulphate–sulphur had a composition similar to the pyrite in the Alum shale and proposed, therefore, that the sulphate–sulphur source was oxidized synsedimentary sulphides. The data from Osen show, however, a large spread in the sulphur isotope composition (−25.6‰ to +51.2‰) of synsedimentary sulphides of Early to Middle Cambrian age. It is therefore difficult to establish a genetic relationship between synsedimentary sulphides of the overlying sediments and sulphates in the ore without a more comprehensive investigation.

The present author regards the similarity in composition of sulphate–sulphur and sulphide–sulphur in the Vassbo deposit to indicate that the source of the sulphate could have been supplied by oxidation of the ore–sulphides in the following manner. According to the groundwater model for the formation of these deposits (Samama 1976, Bjørlykke & Sangster 1981), original precipitation of sulphides within the sandstone would have taken place in a zone of mixing between groundwater carrying metals and barium, and H_2S-bearing marine water. Variations in sea-level and groundwater flow would move this mixing zone back and forth in the sandstone such that any one locality in the sandstone could be alternatively oxidizing, then reducing and back again. These changes in redox potential could cause the originally precipitated sulphides at Vassbo (av. $\delta^{34}S = +16.2‰$) to be oxidized to sulphate of the same isotopic composition. This increase in the sulphate content, combined with the barium in the groundwater, would precipitate barite with a sulphur isotope composition close to that of sulphide–sulphur in the ore as shown in Fig. 5.

No simple paragenesis has been found in the sandstone–lead deposits in Scandinavia. However, Rickard et al. (1979) report a marked tendency for sphalerite to be generally older than galena, which in turn usually precedes the fluorite, barite and calcite. However, the minerals are commonly mutually exclusive and several generations of each mineral exists. Rickard et al. (1979) suggest (p. 1266) «that the mineralization was formed by a series of pulses of ore-bearing solutions of roughly similar composition, which were mainly zinc-rich initially and lead-rich during the later main mineralization period. Late-stage solutions are evidenced by corrosion and silicia replacement of galena.»

The groundwater model, which does not involve pulses of different solutions (Fig. 6), shows in an idealized way the transitional zone between groundwater and marine water within a tidal sandstone. Due to a higher organic content in marine sediments than in continental sediments in Cambrian time, the concentration of H_2S will increase seawards and groundwater with lead and zinc will first precipate galena and then sphalerite. During a period of regression the mixing zone will move seawards resulting in a sequence of precipitation with sphalerite, galena and barite, similar to what has been reported by Rickard et al. (1979).

Conclusion

The new data from the mineralizations at Gjøvik, Osen and Galåa reported here are consistent with a sulphure source from oil-field water proposed by Rickard et al. (1979) for the Laisvall deposit in Sweden. However, the data

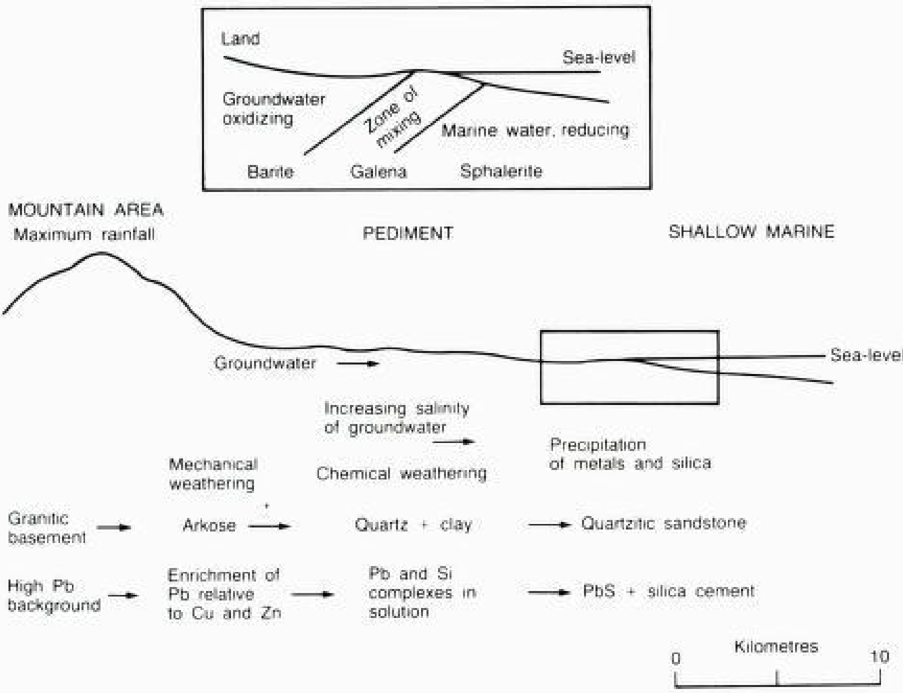


Fig. 6: Schematic representation of the main features in the ground water model for sandstone lead deposits (Modified from Bjørlykke & Sangster 1981).

also permit, as an alternative source for the sulphide-sulphur, sea-water sulphate reduced locally by bacteria. Sulphides from sediments of Early to Middle Cambrian age in Osen show a sulphur isotope stratigraphy reflecting sulphate reduction by bacteria in alternating closed and open systems. In the Vassbo deposit in Sweden sulphate-sulphur has a similar composition to the sulphide, indicating that the sulphur source for the barite could be oxidation of previously formed sulphides.

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REFERENCES

Bjørlykke, A. 1979; Gjøvik og Dokka. Beskrivelse til de berggrunnsgeologiske kart 1816 I og 1816 IV - 1:50.000. *Norges geol. Unders. 344,* 48 pp.

Bjørlykke, A., Bølviken, B. Eidsvig, P., & Svinndal, S. 1973: Exploration for disseminated lead in Southern Norway. In: Prospecting in areas of glacial terrain. IMM, London, 111-126.

Bjørlykke, A. & Sangster, D. F. 1981; An overview of sandstone lead deposits and their relation to red-bed copper and carbonate-hosted lead-zinc deposits. *Econ. Geol.*, 75th Anniversary vol. 179-213.

Bjørlykke, K. 1974; Depositional history and geochemical composition of Lower Paleozoic epicontinental sediments from the Oslo Region. *Norges geol. Unders.* 305, 81 pp.

Bjørlykke, K. 1978: The eastern marginal zone of the Caledonide orogen in Norway. In Caledonian-Appalachian Orogen of the North Atlantic Rwgion; Geological Survey of Canada, Paper 78-13, 49-56.

Bjørlykke, K., Elvsborg, A. & Høy, T. 1976; Late Precambrian sedimentation in the central sparagmite basin of South Norway. *Norsk geol. Tidsskr. 56*, 233-290.

Christofferson, H. C., Wallin, B., Selkman, S., & Rickard, D. T. 1979; Mineralization controls in the sandstone lead-zinc deposit at Vassbo, Sweden. *Econ. Geol.*, 74, 1239-1249.

Claypool, G. E., Holser, W. T., Kaplan, I. R., Sakai, H., & Zak, I. 1980; The age curves of sulfur and oxygen isotopes in marine sulfate and their mutual interpretation. Chem. Geol., 28, 199-260.

Cowie, J. W. & Cribb, S. J. 1978; The Cambrian System. In: Contribution to the Geologic Time Scale. American Association of Petroleum Geologists, Studies in geology, No. 6, 355-362.

Gee, D. G. 1975; A tectonic model for the central part of the Scandinavian Caledonides. Am. Jour. Sci. 275, 468-515.

Grip, E. 1967; The genesis of the lead ores of the eastern border of the Caledonides in Scandinavia. In: Genesis of stratiform lead-zinc-barite-fluorite deposits in carbonate rocks (Brown, J. S. ed), Econ. Geol. Monograph 3, 208-218.

Grip, E. 1973; Sulfidmalm i fjällkedjan och i det baltiska Kambriska flackhavsområdet. In: Grip, E., and Frietsch, R., eds., Malm i Sverige, 2: Stockholm, Almquist and Wiksell, 10-66.

Høy, T. 1977; Forvitring av basement, en kilde for bly/sink mineraliseringer i sparagmittområdet. In: Bjørlykke, A., Lindahl, I., and Vokes, F. M. (eds.), Kaledonske Malmforekomster, BVLI tekniske virksomhet, Trondheim, 77-81.

Jensen, M. L. 1967; Sulphur isotopes and mineral genesis; in Barnes, H. L. (ed.) Geochemistry of hydrothermal ore deposits. Holt, Rinehart, and Winston, Inc., N. Y., 143-165.

Ncube, A., Zweifel, H., & Amstutz, G. C. 1978; On the occurrence of bravoite, framboidal pyrite, marcasite, pyrrhotite and possible plant remains in the Laisvall lead and zinc deposit, Sweden; Neues jahrbuch für mineralogie, Abhandlungen 132, 269-283.

Nielsen, H., & Sperling, H. 1973; Schwefel-Isotopenuntercuchungen an der Blei-Zinc-Ertzlagerstätte Grund (Westhartz, Bundesrepublik Deutschland) Mineral. Deposita, 8, 64-72.

Nystuen, J. P. 1969; Sedimentology of the Lower Cambrian sediments at Ena, Nordre Osen, Southern Norway. *Norges geol. Unders. 258*, 27-43.

Ohmoto, H. & Rye, R. O. 1979; Isotopes of sulfur and carbon, in Barnes, H. L. (ed.) Geochemistry of hydrothermal ore deposits. New York John Wiley and sons, 509-567.

Pringle, I. R. 1973; Rb-Sr age determination on shales associated with the Varanger ice age. *Geol. Mag. 109*, 465-472.

Rickard, D. T., Willdén, M. Y., Marinder, N. E. & Donnelly, T. H., 1979; Studies on the genesis of the Laisvall sandstone lead-zinc deposit, Sweden. Econ. Geol., vol. 74, 1255-1285.

Rickard, D., Coleman, M., & Swainbank, I. 1981; Lead and sulphur isotopic composition of Galena from the Laisvall sandstone lead-zinc deposit, Sweden. *Econ. Geol. 76*, 2042-2046.

Samama, J. C. 1976; Comparative review of the genesis of the copper-lead sandstone-type deposits, in Wolf, K. H. (ed.) Handbook of stratabound and stratiform ore deposits. Vol. 6, 1-20. Amsterdam, Elsevier.

Sangster, D. F. 1976: Sulphur and lead isotopes in strata-bound deposits. In Wolf, K. H. (ed.) Handbook of strata-bound and stratiform ore deposits, Vol. 2, 219-266. Elsvier, Amsterdam.

Schwarcz, H. P. & Burnie, S. W. 1973: Influence of Sedimentary Environments on Sulfur Isotope Ratios in Clastic Rocks; a Review. Mineral. Deposita. 8, 264-277.

Skjeseth, S. 1963: Contributions to the geology of the Mjøsa District and the classical sparagmite area in Southern Norway. *Norges geol. Unders. 220*, 126 pp.

Sturt, B. A. 1978: The Norwegian Caledonides. Introduction. In: Caledonian-Appalachian orogen of the North Atlantic Region. Geol. Survey of Canada, Paper 78-13, 13-16.

Sturt, B. A., Pringle, I. R. & Roberts, D. 1975: Caledonian nappe sequence of Finnmark, northern Norway, and the timing of orogenic deformation and metamorphism. *Bull. Geol. Soc. Am. 86*, 710-718.

Tegengren, F. R. 1962: Vassbo blymalmfyndighet i Idre och dess geologiska inramning. Sveriges geologiska undersökning, ser. C. 586, 62 pp.

Vidal, G. 1981: Aspects of problematic acid-resistant, organic-walled microfossils (Archritarchs) in the Upper Proterozoic of the North Atlantic region. Precambrian Research 15, 9–23.

Wallin, B. G. 1980: Hydrocarbon – derived hydrothermal carbon in carbonates from the ore-bearing sandstone at Vassbo, central Sweden. Annual Report of the Ore Research Group, Stockholm University p. 112–132.

Wallin, B. 1982: Sedimentology of the Lower Cambrian Sequence at Vassbo, Sweden. Stockh. Contr. Geol., 39 (1), 1–111.

Welin, E. 1980: Tabulation of recalculated ages published 1960–1979 for rocks and minerals in Sweden. Geol. Förh. Stockh. Førh., 101, 359–362.

Welin, E. & Lundqvist, T. 1975: K-Ar ages of Jotnian dolorites in Västernorrland County, central Sweden. Geol. För. Stoch. Förh., 97, 83–88.

Willdén, M. Y. 1980: Paleoenvironment of the autochthonous sedimentary rock sequence at Laisvall, Swedish Caledonides; Stockholm contributions in Geology, vol. XXXIII, 100 pp.

Subsidence and Tectonics in Late Precambrian and Palaeozoic Sedimentary Basins of Southern Norway

KNUT BJØRLYKKE

Bjørlykke, K. 1983: Subsidence and tectonics in Late Precambrian sedimentary basins of southern Norway. *Norges geol. Unders. 380,* 159–172.

The assumption that sedimentary basins approach isostatic equilibrium provides a good foundation for modelling basin subsidence based on variables such as cooling rates (thermal contraction), crustal thinning, eustatic sea-level changes and sedimentation. The Sparagmite basin of Central Southern Norway was probably formed by crustal extension during rifting. During Cambrian and Ordovician times the Oslo Region was rather stable part of the Baltic Shield, reflected in slow epicontinental sedimentation. The Bruflat Sandstone (Uppermost Llandovery) represents the first occurrence of a rapid clastic influx, reflecting a pronounced basin subsidence. This change in sedimentation is believed to be related to the emplacement of the first Caledonian nappes in the northern part of the Oslo Region, providing a nearby source for the sediments and resulting in subsidence due to nappe loading. The underlying Palaeozoic sequence was detached along the Cambrian Alum Shale in front of the Osen Nappe. Devonian sedimentation was characterised by vertical tectonics and some of the Devonian basins, such as the Hornelen Basin, may be related to listric faulting rather than strike-slip fractures. The Permian sediments of the Oslo Graben were probably overlain by Triassic and possibly also by Jurassic sediments during post-rift subsidence.

K. Bjørlykke, Geologisk Institutt, Avd. A, Allégt. 41, N-5014 Bergen-Univ. Norway.

Introduction

Sedimentary basins are very sensitive recorders of contemporaneous tectonic movements. Their potential as a key to the understanding of the tectonic history of a region has, however, not always been fully utilized. The present paper will examine some of the Late Precambrian and Palaeozoic sequences in Southern Norway and attempt to relate sedimentation to tectonic movements. The principle of crustal isostasy is old and well established in geological and geophysical literature. It is, however, only recently that Airy's isostatic model has been widely applied to recalculate primary tectonic movements or sea level changes in sedimentary basins (e.g. Kinsman 1975, Watt & Stecker 1979, Watt & Ryan 1976, Watt 1981, 1982).

We may assume that most sedimentary basins are in approximate isostatic equilibrium. This implies that basin subsidence must be due to changes in the isostatic equilibrium (Kinsman 1975). However, at basin margins or along hinge lines in passive margins, flexural rigidity of the crust may play an important role (Watts & Ryan 1976). The requirements that isostatic equilibrium should be maintained, allows us to model the relationship between tectonics and sedimentation and to place important constraints on sedimentological and stratigraphical interpretations. Since several factors influence isostatic equilibrium, this principle rarely gives unique solutions but forces us to consider different mechanisms by which isostatic equilibrium can be maintained.

Primary subsidence of the crust underlying a sedimentary basin may be due to four principal phenomena:

1. Extension and thinning of the continental crust.
2. Cooling, causing thermal contraction and isostatic subsidence due to the increased density.
3. Increased loading, due to the ice or emplacement of tectonic nappes.
4. Eustatic rises in sea level will also cause subsidence in order that isostatic equilibrium can be maintained. Secondary subsidence will rsult from increased load due to deeper water and to the sediments filling the basin.

Isostatic uplift following the melting of glaciers has shown that the crust responds rather quickly to changes in loading in the perspective of the geological time-scale. Nansen (1928) also pointed out the significance of water and sediment loading for isostatic equilibrium.

Stratigraphic information may be used to calculate primary crustal subsidence by removing the effects of loading by sediments and water using a backstripped technique (Stecker & Watts, 1978):

$$Y = \underbrace{S\frac{\rho_m - \rho_s}{\rho_m - \rho_w}}_{\textit{Sediment loading}} - \underbrace{SL\frac{\rho_w}{\rho_m - \rho_w}}_{\textit{Water loading}} + \underbrace{Wd - SL}_{\textit{Water depth}} \quad (1)$$

where Y is the primary basement response, S is the stratigraphice thickness compensated for compaction, ρ_m is the density of the mantle, ρ_s is the density of the sediments, ρ_w is the density of water, SL is the change in sea level and Wd is the water depth.

Late Precambrian and Lower Palaeozoic sedimentation in the Sparagmite Basin and the Oslo Region

In Late precambrian times the Baltic Shield was a mature craton and erosion reached deep into the roots of older orogenic belts (Oftedahl 1980). The Late Precambrian Sparagmite Basin formed as a result of continental rifting (Bjørlykke et al. 1976, Roberts & Gale 1978, Nystuen 1982). Isostatically, the subsidence was probably due to thinning of the continental crust. Contemporaneous faulting and volcanism suggest that rifting was active up to the time of deposition of the Ring Formation (Fig. 1) (Sæther & Nystuen 1981). The Ringsaker Quartzite (Fig. 1) represents an important stratigraphic reference horizon since this extensive, thin, transgressive orthoquartzite suggests a very stable tectonic environment.

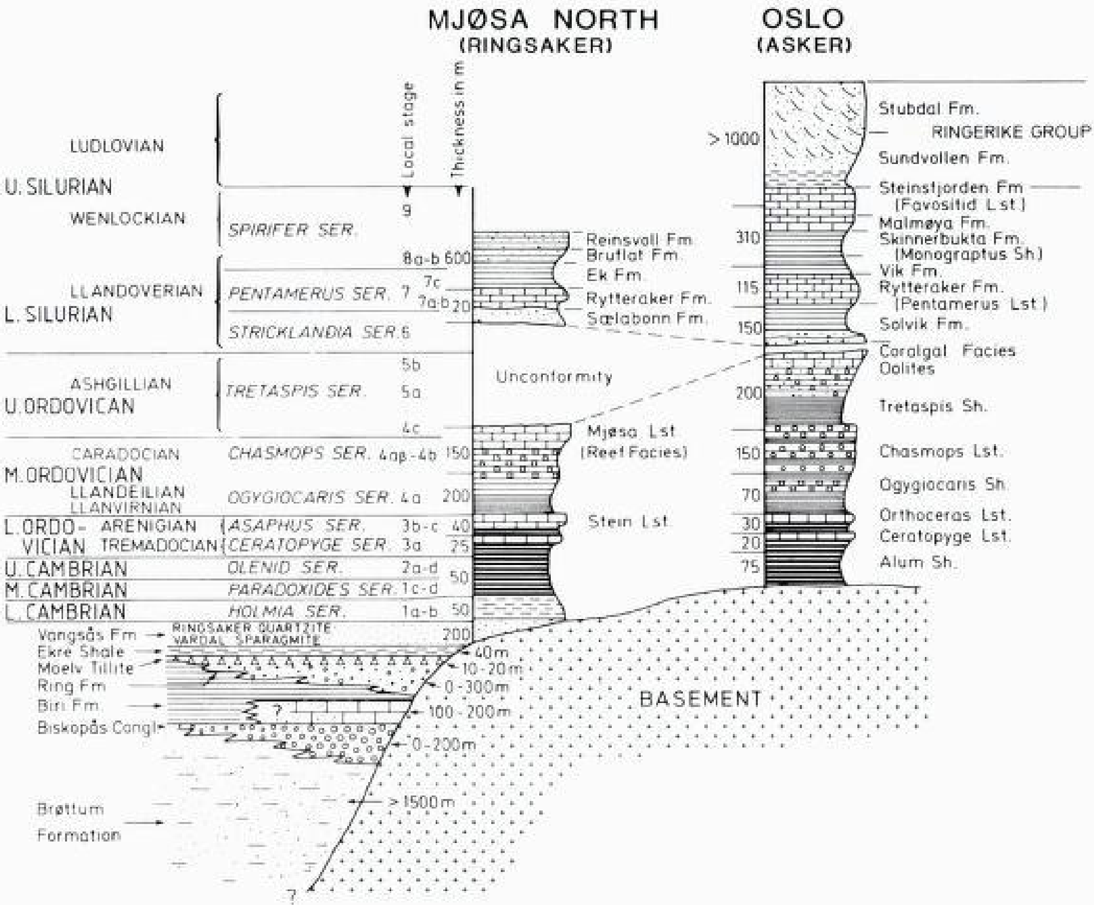

Fig. 1. Stratigraphy and Late Precambrian and Lower Palaeozoic sedimentary sequences in the Lake Mjøsa and Oslo districts. Silurian stratigraphy from Worsley et al. (1982). In the Mjøsa Region the name Helgøy Quartzite is used for the basal quartzite above the Ordovician–Silurian unconformity and Limovnstangen Formation for the Pentamerus Limestone (Høy & Bjørlykke, 1980).

The Vangsås Formation is only 150–200 m thick and the Ekre Shale and the Moelv Tillite are rarely more than 40 and 30 m thick, respectively. The Ekre Shale, however, may locally be thicker and is 150 m thick in Osdalen (Nystuen 1982). These sediments, which may be interpreted as deposited after the cessation of active rifting, are thus only 200–300 m thick in total. Using the back-stripping technique (equation 1), we find that the corresponding primary crustal subsidence due to thermal contraction in the Sparagmite basin was less than 100 m. Assuming that the Moelv Tillite was deposited at 650 Ma (Bjørlykke & Nystuen 1981) and the Ringsaker Quartzite at 580 Ma (lowermost Cambrian) the sequence was deposited over a period of 70 Ma at a low sedimentation rate. By Ringsaker Quartzite times most of the thermal contraction should have been completed, and the Ringsaker Quartzite thus suggests very slow sedimentation in a shallow water marine environment. 100 m is a very modest primary subsidence and it may be pertinent to ask why thermal contraction due to post-rift cooling did not cause more subsidence. One possible explanation is that most of the rifting and crustal heating occurred during the deposition of the Brøttum and Biri Formations, and the Ring Formation may then also have been deposited during the cooling phase. Evidence of contemporaneous volcanicity, however, is found below the Moelv Tillite in Østerdalen (Sæther & Nystuen, 1981). These are the only areas where there is clear evidence of contemporaneous vulcanicity, although basaltic clasts are common in the Biskopåsen Conglomerate (Sæther & Nystuen, 1981). The Sparagmite rift basin would therefore seem to have been characterised by limited volcanic activity and may have had a lower geothermal gradient than other rift basins. This would have resulted in modest post-rift subsidence due to cooling.

The Ringsaker Quartzite is succeeded by Lower Cambrian (Holmia) shales which are transgressive southwards, lapping on to basement in the Mjøsa District. These beds are found not only in other parts of the Baltic Shield (Martinson 1974) but also on several continents (Brasier 1982) suggesting that the transgression was eustatic. Back-stripping this part of the sequence we find that the Lower Cambrian transgression could have been caused by a eustatic sea level rise or a primary basin subsidence of as little as 30–40 m which is considerably less than that suggested by Vail et al. (1977). This may have been related to cambrian sea-floor spreading.

A conglomerate of lower Middle Cambrian age (Paradoxides oelandicus zone) indicates a period of regression and erosion (Skjeseth 1963, Bjørlykke 1974b). The subsequent Middle Cambrian trabsgression gradually covered most of the Baltic Shield. This transgression occurred at the same time in Norway and southern Sweden and probably represents a new eustatic rise in sea-level. The transgression initiated a cycle of black shale sedimentation which continued, interrupted only by episodes of carbonate sedimentation, to Late Tremadocian times when reworking of the upper part of the Ceratopyge limstones developed. Correcting for later compaction and using equation (1), we find that the deposition of the 120 m-thick Middle Cambrian to Late Tremadocian sequence in the Oslo Region could have been caused by a 40 m eustatic change in sea-level.

The Ordovician sequence in the Oslo Region indicates an accelerating rate of subsidence (Fig. 2). Carbonate sediments constitute a shallow-water facies, shales occurring essentially when rates of subsidence or eustatic rises in sea level exceeded the rate of sedimentation. We can recognise the following cycles of deepening and subsequent shallowing:

1. L. Arenig ⟶ U. Arenig (L. Didymograptus Shale (3b) ⟶ Orthoceras Limestone (3L).
2. Llanvirn ⟶ Caradoc (U. Didymograptus Shale (4a) ⟶ Chasmops Limestone (4b).
3. U. Caradoc ⟶ U. Ashgill (Tretapis Shale (4c) ⟶ Limestone (5b).

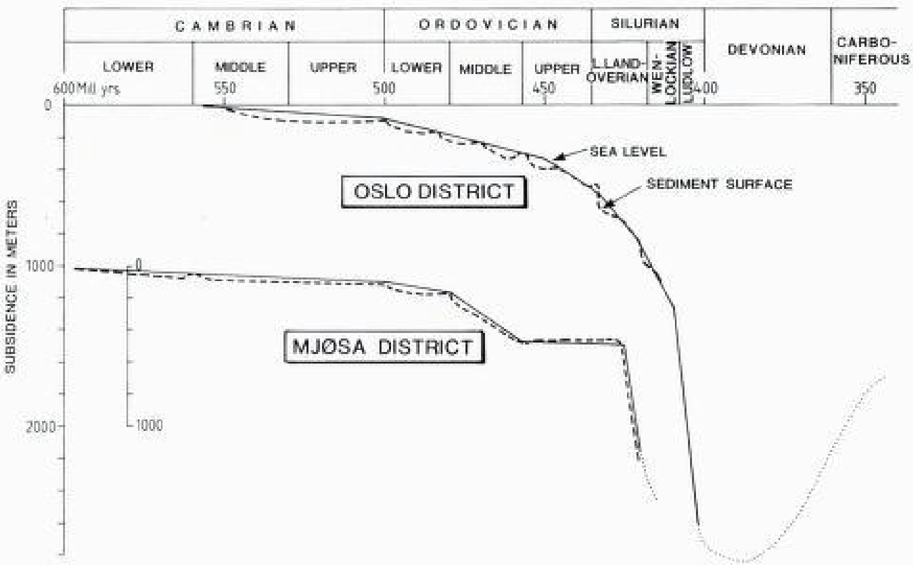

Fig. 2. Basin subsidence curves for the Oslo and Mjøsa districts during Lower Palaeozoic sedimentation.

Within these cycles several minor sea-level oscillations can be recognized. Particularly in the Llanvirn–Caradoc cycle minor oscillations resulted in the alternation of shale and limestone in the Ampyx and the Chasmops series.

The Tretaspis Shale (Uppermost Caradoc to Lowermost Ashgill) is a black shale with a thin phosphate conglomerate near the base, and represents a very significant transgression from the shallow-water environment of the Chasmops limestone. The Tretaspis Shale passes into a shallower water sequence with impure carbonates and coralgal facies at the top. It is difficult to evaluate the effect of eustatic changes in sea level in the Ordivician sequence. The Upper Ordovician glaciation in W. Africa and S. America (Crowell et al. 1981, Biju–Duval et al. 1981, Deynoux & Trompette 1981) should be expected to have cause eustatic sea-level changes that might be detectable in the Oslo Region (Bjørlykke.1974a, p. 18, Brenchley & Newall 1980.

It has proven difficult to establish the age of this glaciation precisely within the time range from Caradoc to Llandovery. Tillites in the Anti-Atlas are, however, better dated and suggested an uppermost Ashgillian (Hirnantian) age (Destombes 1981). The W. African and S. American glaciations of Late Ordivician age may not have been synchronous, but if they were, their extent would suggest a rather significant change in sea-level. Even a much smaller sea-level change than that associated with the Pleistocene glaciation, i.e. 30–50 m, would be expected to have had a quite profound influence on sedimentation in a shallow-marine environment like that in the Oslo Region. It is therefore tempting to correlate the unconformity along the Ordovician/Silurian boundary with the Late Ordovician glaciation, even though the glaciation cannot be dated very precisely. However, the somewhat variable rates of relative subsidence associated with the Lower Silurian transgression in the

Oslo Region suggest a strong tectonic influence (Bjørlykke 1974a). Spjeldnæs (1957) has also suggested that a significant angular unconformity exists at the Ordovician/Silurian boundary in the Oslo Region but it is difficult to find conclusive evidence of this. If the Lower Silurian transgression was only due to a post-glacial eustatic rise in sea-level, it should have been more synchronous and uniform. The U. Ordovician/l. Silurian unconformity in the Oslo Region represents, however, only an episode of relative uplift in a period of accelerating subsidence (Fig. 2), and the unconformity is probably, at least partly, also controlled by tectonic movement. If the erosional relif in the Ordovician sediments at the base of the Silurian sequence (up to 100 m) (Bjørlykke 1974a, p. 21) was entirely due to glacio–eustatic regression, it is surprising that well developed karst structure are rare or absent below the unconformity. On the other hand we might assume that if we had a significant glacio-eustatic sea-level drop in Upper Ordovician time, it could only be placed at the top of the Caradoc or in the Hirnatian, based on the stratigraphic record of the Oslo region.

Relationships between Silurian thrusting and sedimentation (nappe loading)

Sedimentation rates in the Llandovery and Wenlock require significant crustal subsidence, in marked contrast to the stable situation in Middle Cambrian to Middle Ordovician times. A distinct cause of this crustal subsidence should be sought as its magnitude is too large to be explained in terms of eustatic sea-level changes. The Oslo Region must also have been thermally mature and subsidence due to cooling is therefore highly unlikely. The possibility of thinning of the continental crust cannot be ruled out, but it might be expected to have affected the whole of the Oslo Region at the same time. In the author's opinion the cause of the increased rates of subsidence in Late Ordovician and Silurian times may be due to isostatic loading by nappes which were thrust towards the E and SE during Caledonian deformation. A similar relationship is found in the Silurian of North Greenland (Hurst et al. 1983).

There were probably several phases of thrusting, the earliest of which was of Lower to Middle Ordovician age and involved the obduction of ophiolites (Sturt & Thon 1978). The ophiolites were obducted on to Precambrian basement representing a westward extension of the Baltic Shield. Despite the considerable distance to the obduction sites the clastic sediments of the Oslo Region are strongly influenced by debris from basic igneous rocks and distinctive minerals such as chromite can be traced back to ophiolitic source rocks (Bjørlykke 1974a, b, Bjørlykke & Englund 1979). The geochemical influence of ophiolite debris is first felt in the Oslo Region in Arenigian times and reaches a peak in Middle Ordovician (Caradocian) times. Thus the inflow of clastic materials derived from the ophiolites helps to date their emplacement and uplift prior to erosion. A similar pattern of clastic chromite distribution is also found to date the obduction of ophiolites in Newfoundland (Church & Stevens 1971). Early Ordovician thrusting (Sturt & Thon 1976) caused major nappe loading and basin subsidence in regions in proximal positions relative to the nappes, but they were probably too distant to cause major loading effects in the Oslo Region. Middle to Upper Ordovician flysch in Gausdal may however reflect subsidence and sediment supply from advancing Ordovician nappes (e.g. the Jotun Nappe: Nystuen (1981)). The fact that the Upper Ordovician subsidence in the Oslo Region is synchronous with uplift and exposure in the Lake Mjøsa district indicates that relative subsidence at that time was not caused by eustatic sea-level changes or nappe loading but was probably the result of slight adjustments in the part of the craton related to subduction further west.

The Bruflat Sandstone (Fig. 1 & 2) provides the first evidence of relatively rapid subsidence, and the formation of turbidites and deltaic sediments is good evidence of a proximal exposure of a major land mass (Bjørlykke 1974a). This may have been caused by the emplacement of the Osen Nappe or Quartz Sandstone Nappe (Schiötz 1902, Nystuen 1981) towards the Oslo Region. Emplacement was probably the result of gravity sliding producing an imbricate structure (Price 1977) which could cause isostatic subsidence in front of the nappe and at the same time constitute an uplifted source area to the north-west. The Bruflat Sandstone, which is now considered to be of uppermost Llandovery age (Worsley et al. 1982), may thus date the advancement of the Osen Nappe into the Lake Mjøsa Region.

Since the Bruflat Formation and in some areas the Reinsvoll Formation (Worsley et al. 1982) are overlain unconformably by Permian lava and sediments in the Mjøsa district, we are not in a position to detect continued subsidence and sedimentation in the Lake Mjøsa Region through the Upper Silurian. In the Ringerike district carbonate sedimentation was predominant in Wenlockian times, showing that this region was still unaffected by the clastic influx (Worsley et al. 1982). Sedimentation rates however, were relatively high (Fig. 2).

The Ringerike Sandstone (Sundvollen Fm.) of upper Wenlock and Ludlow age (Worsley et al. 1982) represents progradation of fluvial sediments over marine sediments in the Ringerike and Bærum–Asker districts and requires a land area to the north or north-west. The depocentre was displaced to the south in the Upper Llandovery although it appears that the Bruflat Sandstone represents a separate and earlier episode of subsidence. If the Bruflat Sandstone dates the emplacement of the Osen Nappe in the northern part of the Oslo Region, some folding of the Cambro–Silurian sequence might be expected to have occurred in front of the nappe (Fig. 3). The thrusting of the Osen Nappe over the Cambrian Alum Shale requires that the Ordovician and Lower Silurian sequences were detached from the Alum Shale ahead of the Osen Nappe (Fig. 3). Strong deformation of the Alum Shale in the Lake Mjøsa district, due to the movement of the Ordovician–Silurian sequence over the basement, has been documented by Skjeseth (1963). The impor-

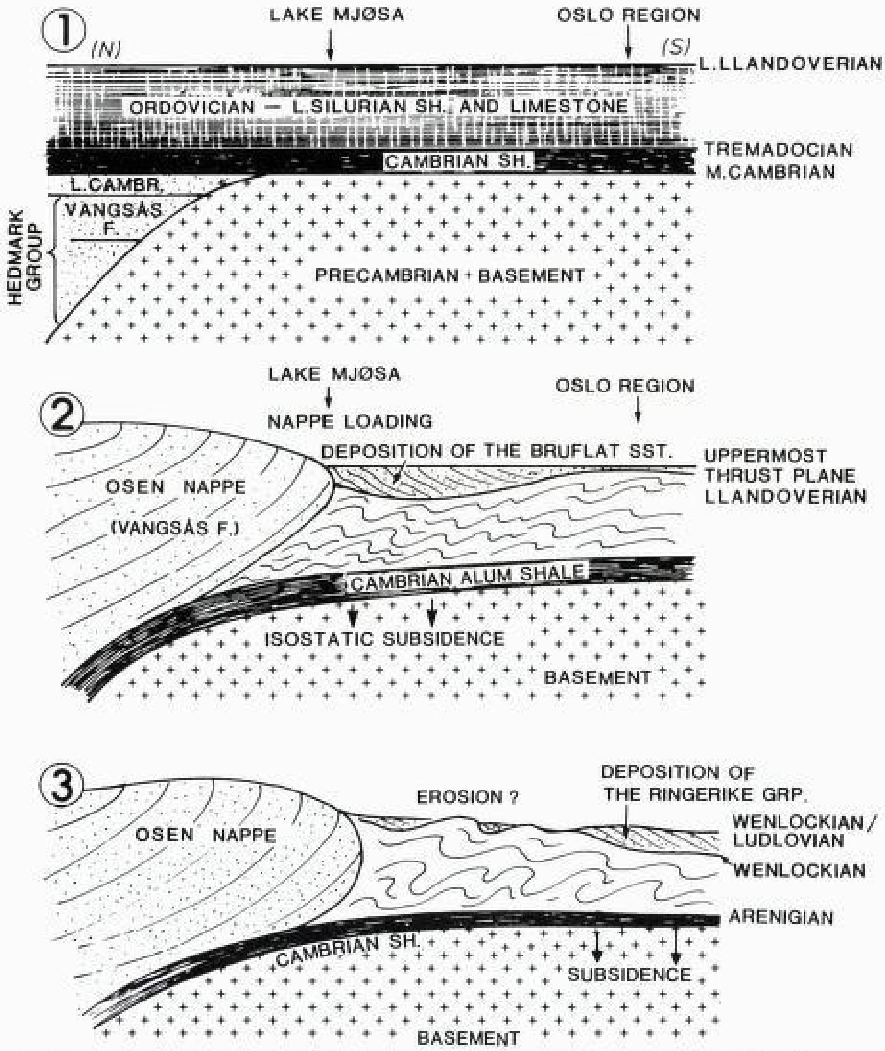

Fig. 3. Hypothetical model for Silurian sedimentation and thrusting in the Oslo Region.

tance of the Alum Shale as a major décollement horizon was first recognised by Brøgger (1882). Nansen (1928, p. 105) stated:

> «Along this slope from the mountain range the sedimentary formations gradually slid downwards towards the Oslo Region, this sliding being facilitated by the loose, and slippery Alum-schist which formed the lowest sedimentary layer and acted as a lubricating material reducing the friction against the underlying surface of Archæan rocks. Thus a lateral pressure arose in the sliding Cambrian and Silurian strata, and they were gradually compressed, folded and crumpled».

The gravity sliding is likely to have been associated with some folding which would then have taken place before the deposition of the Ringerike Sandstone. As pointed out by Worsley et al. (1982) relatively stable areas of carbonate sedimentation existed in the Asker and Ringerike districts through most of Wenlock time. However, the Lower Wenlockian (Sheinwoodian) NW–SE trending facies belt shown on 'time slice' maps (Worsley et al. 1982) may represent surface expressions of gentle early folding.

Before the deposition of the Ringerike Group the underlying Lower Palaeozoic sediments probably consisted of well-cemented limestone interbedded with rather soft muds due to very limited degrees of compaction by the overburden. This sequence could have been folded in a rather ductile manner by differential sliding between the limestone beds.

The Ringerike Sandstone has a more open fold style representing less shortening than for the folds in the underlying sediments. This may possibly be the result of differences in competence but may also be partly explained by some pre-Ringerike Sandstone folding. Further to the south–east the highest rates of subsidence occurred in a basin extending into Scania, Denmark and N. Poland (Størmer 1967). This depression is nearly paralell to the direction of Caledonian thrusting and may be related to extension perpendicular to Caledonian compression along a zone of weakness in the Baltic Shield.

If the Lower Palaeozoic rocks were not deformed prior to the deposition of the Ringerike Sandstone, then the thrusting of the Osen Nappe must have been accomplished during the Silurian. Since the Osen Nappe was emplaced over Cambrian shales, the total shortening of the Ordovician/Silurian sequence must correspond to the magnitude of the Osen Nappe overthrust (Fig. 3).

In the Oslo Region and the Southern part of the Sparagmite Basin the Precambrian basement was not significantly deformed during nappe emplacement. In the southernmost part of the Oslo Region (Skien–Langesund district) there is little evidence of folding or thrusting of the Lower Palaeozoic sequence in relation to the basement. This implies that the shortening of the Ordovician/Silurian sequence related to the emplacement of the Osen Nappe must be represented by folding within the Oslo Region. Based on the occurrence of outliers of the Nappe (Schiøtz 1902, Holtedahl 1915) the minimum distance of translation is 25–40 km. Oftedahl (1943) and Nystuen (1981) have, on various lines of evidence; interpreted the remaining part of the Sparagmite Basin north of Lake Mjøsa to be highly allochthonous (200–400 km). If the 'Sparagmite nappe' had also been sliding on Cambrian shales the shortening of the underlying Cambro–Silurian sequence must have been correspondingly larger. In the present author's opinion the style of the folding in the Cambro–Silurian sequence in the Oslo Region does not suggest shortening of this magnitude.

Late Palaeozoic sedimentation

Subsidence and thrusting during Late Silurian and Early Devonian times resulted in a very significant crustal thickening in the Central Norwegian Caledonides. This resulted in isostatic uplift and the formation of a topographic mountain chain. Since conglomerates and sandstones of Lower to Middle Devonian age i Western Norway (Steel 1978) and in Trøndelag (Siedlecka 1975, Roberts 1974) rest on Lower Palaeozoic sediments or basement, there must have been considerable erosion prior to the formation of the basins. The Devonian basins represent areas of very strong subsidence in the Central Norwegian Caledonides. Such large-scale subsidence requires a thinning of the continental crust, and Steel (1976) suggests for the Hornelen basin that crustal extension and thinning was related to a transcurrent fault system. It is, however, difficult to explain how a major E–W trending strike-slip fault extends and dissipates into the central part of the Baltic Shield.

Bryhni (1964) suggested that the eastward tilting in the Hornelen basin was due to a basin progressively migrating eastwards by successive faulting. Another mode of formation of the sedimentary basin is by detachment faulting between a basement complex and a cover terrain similar to those described from the Cordilleran metamorphic core complexes (Coney 1980). Extensional tectonics related to uplift may produce sedimentary basins by listric faulting (Fig. 4). The sedimentary basins open by movement on a listric gravity fault, producing a basin by extension of the crust in much the same way as one opens a drawer. Sliding of the Hornelen basin with its underlying basement to the west explains the continous infilling of the basin from the

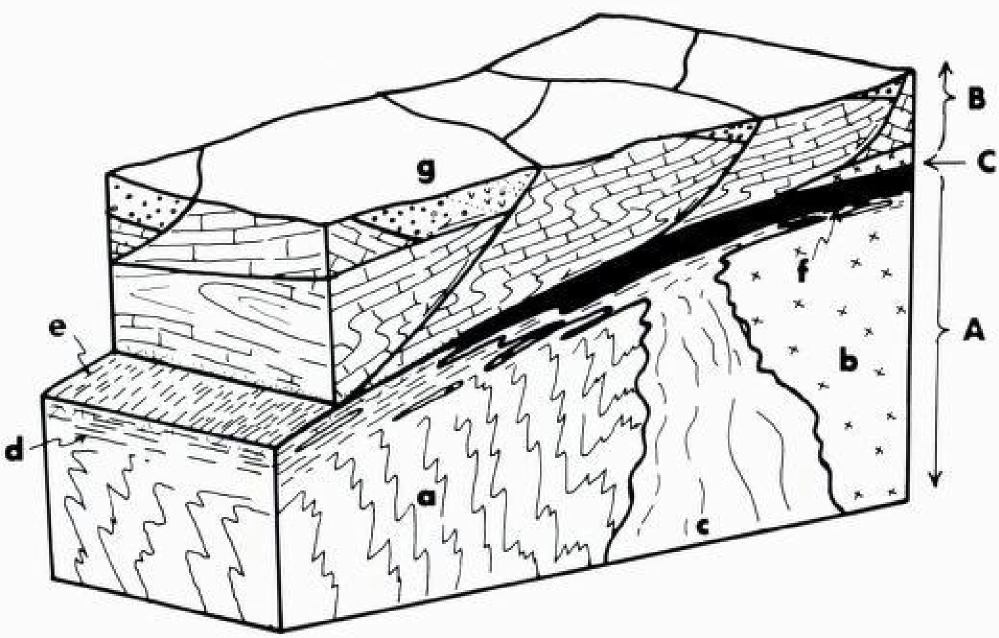

Fig. 4. Formation of sedimentary basins due to listric extensional faulting (from Coney 1980). Note that this model predicts that sedimentary infill occurs asymmetrically as the basin is extended and that the sedimentary sequence will be younging away from the direction of the relative tectonic movement of the fault blocks. The dip of the beds will decrease up section as rotation along the fault-plane takes place during sedimentation.

east, as well as the decrease in dip to the east due to rotation during sedimentation. Detailed tectonic studies are, however, needed before such a model can be substantiated. Basin subsidence related to listric structures and crustal extension has also been suggested by Roberts (1983).

Upper Palaeozoic rifting in the Oslo Region was associated with the production of rather large volumes of igneous rocks (Ramberg & Spjeldnæs 1976) and a sequence of Upper Carboniferous and Permian sediments only a few hundred metres thick was deposited (Olaussen 1981). We would expect that the Upper Palaeozoic rifting was followed by a later phase of subsidence. Post-rifting cooling could theoretically result in a thermal subsidence of 2–3 km, depending upon the degree of thinning of the continental crust and additional subsidence due to sediment loading (Kinsman 1975). It is therefore probable that the Brumunddal Sandstone of the Lake Mjøsa district is only a small part of a once much more extensive and thicker sequence of Post-Permian sediments in the Oslo Region. This assumption is supported by the fact that in the continuation of the Oslo Rift in Skagerak and N. Jylland we find 3–4 km of Triassic sediments (Ziegler 1981, 1982).

In the Oslo Graben, thermal subsidence with the deposition of a 2–3 km thick sequence of sediments would be expected to have followed the Permian igneous activity. It is therefore likely that the Carboniferous–Permian sequence in the Oslo Region was succeeded by approximately 2 km of younger sediments and that sedimentation continued into the Triassic and possibly also the Jurassic.

Conclusions

1. The post-rift subsidence in the Sparagmite Basin is considerably smaller than theoretically expected. This may be explained in two ways:
   a) The rift was characterised by relatively low geothermal gradients.
   b) The main rifting took place during the deposition of the Brøttum Formation and the overlying sequence may be regarded as part of the cooling phase despite the fact that small volumes of volcanic rocks have been found in this overlying sequence.
2. Lower Palaeosoic epicontinental sedimentation indicates a high degree of tectonic stability in Cambrian and Ordovician times. Allowing for basin subsidence due to water loading and sedimentation, the Lower and Middle Cambrian transgression can be explained in terms of much smaller eustatic sea-level changes than those proposed by Vail et al. (1977).
3. The increased rate of subsidence in mid-Silurian times (Late Llandovery and Wenlock) in the Oslo Region may be due to nappe loading, a model which also provides a source for upper Silurian sandstones. If this model is correct it provides a powerful tool for dating thrusting in the Scandinavian Caledonides, as also shown from the Cretaceous of the western United States (Jordan 1983).

4. Devonian sedimentation is characterised by large-scale vertical tectonics and the possibility that Devonian basins in Western Norway may be related to gravity-generated listric faulting rather than strike-slip faulting deserves further attention.
5. In the Oslo Graben, thermal subsidence with the deposition of 2–3 km of sediment would be expected to have followed the Permian magmatism. It is therefore likely that the Carboniferous–Permian sequence in the Oslo Region was succeeded by approximately 2 km of younger sediments and that sedimentation continued into the Triassic and possibly also into the Jurassic.

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REFERENCES

Bjørlykke, K., 1974a: Depositional history and geochemical composition of Lower Paleozoic epicontinental sediments from the Oslo Region. *Norges geol. Unders., 305,* 81 p.

Bjørlykke, K., 1974b: Geochemical and mineralogical influence of Ordovician island areas on epicontinental sediments from the Oslo Region. *Sedimentology, v. 21,* 251–272.

Bjørlykke, K. and Englund, J. O., 1979: Geochemical response to Upper Precambrian rift basin sedimentation and Lower Paleozoic epicontinental sedimentation in South Norway. *Chemical Geology, v. 27,* 271–295.

Bjørlykke, A., 1979: Gjøvik og Dokka. Beskrivelse til de berggrunnsgeologiske kart 1816 I og 1816 IV M 1:50.000. *Norges geol. Unders. 344,* 48 p.

Bjørlykke, K., Elvsborg, A. and Høy, T., 1976: Late Precambrian sedimentation in the central Sparagmite basin of South Norway. *Norsk geol. Tidsskr. 56,* 233–210.

Bjørlykke, K. and Nystuen, J. P., 1981: Late Precambrian Tillites of South Norway. *In:* Hambrey, M. J. and Harland, W. B. (eds.): Earth's pre-Pleistocene glacial Record. Cambridge Univ. Press, p. 624–628.

Biju-Duval, B., Deynoux, M. and Rognon, P., 1981: Late Ordovician Tillites of Sentral Sahara. *In:* Hambrey, M. J. and Harland, W. B. (eds.): Earth's pre-Pleistocene Record. Cambridge Univ. Press, p. 99–107.

Brasier, M. D., 1982: Sea-level changes, facies changes and the late Pre-Cambrian – early Cambrian evolutionary explosion. *Precambrian Research 17,* p. 105–123.

Brenchley, P. J. and Newall, G., 1980: A Facies Analysis of Upper Ordovician Regressive Sequences in the Oslo Region, Norway – A record of glacioeustatic changes. *Palaeogeogr. Palaeoclim. Palaeoecol., 31,* p. 1–38.

Bryhni, I., 1964: Migration basins on the Old Red Continent. Nature, Lond. 202, No. 4930, p. 384–384.

Brøgger, W. C., 1982: Die Silurischen Etagen 2 und 3 in Kristiania Gebiet und auf Eker. Universitätsprogramm für 2. Sem. 1982, 326 p.

Church, W. R. and Stevens, R. K., 1971: Early Paleozoic Ophiolite Complexes of the Newfoundland Appalachian as Mantle–Oceanic Crust Sequences. *Journ. Geophys. Res. 76,* p. 1460–1460.

Coney, P. J., 1980: Cordilleran metamorphic core complexes: An overview. *In:* Crittenden, M. D. (ed.): Cordilleran Metamorphic Core Complexes. *Geol. Soc. Am. Memoir 153,* p. 7–34.

Crowell, J. C., Suarez-Sorucor and Rocha-Campos, A. C., 1981: The Silurian Cancianiri (Zapla) Formation of Bolivia, Argentine and Peru. *In:* Hambrey, M. J. and Harland, W. B. (eds.): Earth's pre-Pleistocene Record. Cambridge Univ. Press, p. 902–907.

Destombes, J., 1981: Hirnantian (Upper Ordovician) Tillites on the North Flank of the Tinodout Anti-Atlas Morocco. *In:* Hambrey, M. J. and Harland, W. B. (eds.): Earth's pre–Pelistocene Record. Cambridge Univ. Press, p. 87–88.

Deynoux, M. and Trompette, R., 1981: Late Ordovician Tillites of the Taodeni Basin of West Africa. *In:* Hambrey, M. J. and Harland, W. B. (eds.): Earth's pre-Pleistocene Record. Cambridge Univ. Press, p. 89–96.

Holtedahl, O., 1915: Iagttagelser over fjeldbygningen omkring Randfjordens nordende. Norges Geol. Unders., No. 75 – Årbok 1915, p. 1–32.

Hurst, J. M., McKerrow, W. S., Soper, N. J. and Surlyk, F., 1983: The relationship between nappe tectonics and Silurian Turbidite deposition in north Greenland. J. Geol. Soc., Lond..

Høy, T. and Bjørlykke, A., 1980: Hamar, Berggrunnskart 1961 IV, m 1:50.000, Norges geol. Unders.

Jordan, T. E., 1981. Thrust Loads and Foreland Basin Evolution, Cretaceous, Western. United States. Am. Ass. Petr. Geol. 65, 2506–2520.

Kinsman, D. J. J., 1975: Rift Valley Sedimentation and Sedimentary History of Trailing Continental Margins. *In:* Fischer, G. and Sheldin, J. (eds.): Petroleum and Global Tectonics, p. 83–129. Princeton Univ. Press, 322 p.

Martinsson, A., 1974: The Cambrian of Norden. *In:* Holland, C. H. (ed.): Cambrian of the British Isles, Norden and Spitsbergen. John Wiley & Son, p. 185–284.

Nansen, F., 1928: The earth's crust, its surface forms and isostatic adjustment. Det Norske Vitenselskaps Skrifter, Mat.-Nat. klasse 1927, nr. 12, 121 p.

Nystuen, J. P., 1981: The Late Precambrian «Sparagmites» of southern Norway: A major Caledonian allochthon – the Osen–Røa Nappe Complex, *Am. J. Sci., v. 281,* p. 69–94.

Nystuen, J. P., 1982: Late Proterozoic basin evolution on the Baltoscandian craton: The Hedmark Group, Southern Norway. *Norges geol. Unders., 375,* p. 1–74.

Oftedahl, C., 1943: Om sparagmitter og deres skyvning innen kartbladet Øvre Rendal. *Norges geol. Unders., No. 161,* 65 p.

Oftedahl, C., 1980: Geology of Norway. *Norges geol. Unders.. No. 356.*

Olaussen, S., 1981: Marine incursion in the Upper Palaeozoic Sedimentary rocks of the Oslo Region, southern Norway. *Geol. Mag. v. 118,* p. 281–288.

Price, N. J., 1977: Aspects of Gravity Tectonics and the Development of Listric Faults. *J. Geol. Soc. Lond., 133,* p. 311–327.

Ramberg, I. B. and Spjeldnæs, N., 1978: The tectonic history of the Oslo Region. *In:* Ramberg, I. B. and Neuman, E. R. (eds.): Tectonics and Geophysics of Continental Rifts. Reidel Publ. Co., Dordrecht, p. 167–197.

Roberts, D., 1974: Sedimentary Tectonics and Metamorphic Features of the Devonian of Røragen, Sør-Trøndelag, *Norges geol. Unders. 311,* p. 88–108.

Roberts, D. and Gale, G. H., 1978: The Caledonian – Appalachian Iapetus Ocean. *In:* Tarling, D. H. (ed.): The Evolution of the Earth Crust, p. 255–342.

Roberts, D., 1983: Devonian tectonic deformation in the Norwegian Caledonides and its regional perspectives. *Norges geol. Unders. 380.*

Schiøtz, O. E., 1902: Den syd-østlige del av Sparagmitt-kvartsfjellet i Norge. *Norges geol. Unders., 35,* p. 135, Map.

Siedlecka, A., 1975: Old Red Sandstone Lithostratigraphy and Sedimentation of the Outer Fosen Area, Trondheim Region. *Norges geol. Unders., 321,* p. 1–36.

Skjeseth, S., 1963: Contributions to Geology of the Mjøsa District and the classical Sparagmite area in Southern Norway. *Norges geol. Unders., 220,* 126 p.

Spjeldnæs, N., 1957: The Silurian/Ordovician Border of the Oslo Area. *Norsk geol. Tidsskr., 37,* p. 355–371.

Steckler, M. S. and Watts, H. B., 1978: Subsidience of the Atlantic-type continental margin off New York. *Earth Planet Sc. Lett., 41,* p. 1–13.

Steel, R. J., 1976: Devonian Basins of Western Norway – Sedimentary Response to Tectonism and to Varying Tectonic Context. *Tectonophysics 36,* p. 207–224.

Størmer, L., 1967: Some aspects of the Caledonian Geosyncline and foreland west of the Baltic Shield. *Quart. Journ. Geol. Soc., v. 123,* p. 183–214.

Sturt, B. A. and Thon, A., 1978: An opholite complex of probable Early Caledonian age discovered on Karmøy. *Nature, Lond. 275,* p. 538–539.

Sæther, T. and Nystuen, J. P., 1981: Tectonic Framework, Stratigraphy, Sedimentation and Vulcanism of the Late Precambrian Hedmark Group, Østerdalen, S. Norway. *Norsk geol. Tidsskr. 61,* p. 193–211.

Vail, P. R., Mitchum, R. M., Todd, R. C., Widmer, J. M. Thompson, S., Sangree, J. B., Bubb, J. N., and Hatlelid, W. G.: Seismic stratigraphy and Global changes of Sea Level, 1977. In Payton, C. E. (Ed.). Seismic Stratigraphy application to hydrocarbon exploration. AAGP Memoir 26. Tulsa, U.S.A., p.49–204.

Watts, A. B., 1981: The U.S. Atlantic Continental margin subsidence history, crustal structure and thermal evolution. *In:* Bally et al. (ed.): Geology of Passive Continental Margins. AAPG Education Course Note Series 19, 2–1 – 2–7–5.

Watts, A. B., 1982: Tectonic subsidence, flexure and global changes at sea level. *Nature, 297,* p. 469–474.

Watts, A. B. and Ryan, W. B. F., 1976: Flexure of the lithosphere and continental margin basins. *Tectonophysics, 34,* p. 25–44.

Watts, A. B. and Steckler, M. S., 1979: Subsidence and eustacy at the continental margin of eastern North America. *In:* Marine Ewine Series, vol. 3. Deep Drilling Results in the Atlantic Ocean; Continental Margins and Paleo-environments, p. 218–239. American Geophysical Union. Washington D.C. 218–234.

Worsley, D., Aarhus, N., Basset, M. G., Howe, P. A., Mørk, A. and Olaussen, S., 1982: The Silurian succession of the Oslo Region. Paleontological contribution from the University of Oslo, No. 276.

Ziegler, P. A., 1981: Evolution of Sedimentary Basins in North-West Europe. *In:* Illing, L. V. and Hobson, G. D. (eds.): Petroleum Geology of the Continental Shelf of North-West Europe. Proceedings of the Second Conference on petroleum Geology of the Continental Shelf of North-West Europe, organised by the Institute of petroleum, London, March 1980, p. 3–42.

Zigler, P. A., 1982: Geological Atlas of Western and Central Europe. Shell Internationale Maatschappij, B. v., 130 p. Elsevier, Holland.

Deglaciation of the Continental Shelf off Southern Troms, North Norway

TORE O. VORREN, MAGNE EDVARDSEN, MORTEN HALD & ELSEBETH THOMSEN

Vorren, T. O., Edvardsen, M., Hald, M. & Thomsen, E. 1983: Deglaciation of the continental shelf off southern Troms, North Norway. Norges geol. Unders. 380, 173–187.

Based on lithostratigraphical and seismic studies, four (possibly five) glacial events are recognized in Andfjorden. The two oldest, the I- and G-events are represented by basal tills. The continental ice sheet probably reached the shelf edge during deposition of these tills. Both the I- and G-events are older than c. 19,000 YBP and possibly younger than c. 36,000 YBP. The uncertain glacial event is represented by a questionable moraine ridge complex 20–25 km from the shelf edge. The Flesen event is represented by end moraines c. 50 km from the shelf edge; an age of 16,000–15,000 YBP is suggested. The D-event is identified as a glaciomarine sedimentary unit with high frequencies of ice-dropped clasts. During D-time the ice sheet crossed the sedimentary/crystalline boundary in Malangsdjupet; in Andfjorden it was probably situated just shoreward of this boundary. The age of the D-event is 14,000–13,000 YBP.

T. O. Vorren, M. Hald & E. Thomsen, Institute of Biology and Geology, University of Tromsø, N-9000 Tromsø, Norway.

M. Edvardsen, Norwegian Petroleum Directorate, N-4000 Stavanger, Norway

Introduction

The purpose of the investigation upon which this article is based is to elucidate the deglaciation history of the continental shelf off the southern part of Troms based on stratigraphical studies. For a review of earlier work on the deglaciation history of this area we refer to Vorren & Elvsborg (1979).

The areas chosen are two troughs, Andfjorden and Malangsdjupet (Fig. 1). There are two main reasons for this choice; 1: The depth of these troughs exceeds 300 m which is a critical depth for iceberg plough marks in this area. Above this depth the stratigraphy of the upper sediment layers was disturbed by Weichselian iceberg ploughing (Lien & Myhre 1977, Vorren et al. 1982). 2: These shelf areas are adjacent to land areas which are under consideration as possible non-glaciated areas during the Weichselian (Ahlmann 1919, Undås 1938, 1967, Grønlie 1941, Dahl 1955, Bergstrøm 1973, Ives 1975).

Physiographic setting

The main bathymetric features of the investigated area (Fig. 1) are two glacial troughs, Malangsdjupet and Andfjorden, with maximum depths of 455 and 505 m, respectively. The area in between comprises a shallow bank, Sveinsgrunnen, with depths less than 100 m.

The bedrock on the shelf comprises Mesozoic and Tertiary sedimentary rocks while older crystalline rocks occur along the coast and on land (Fig. 1).

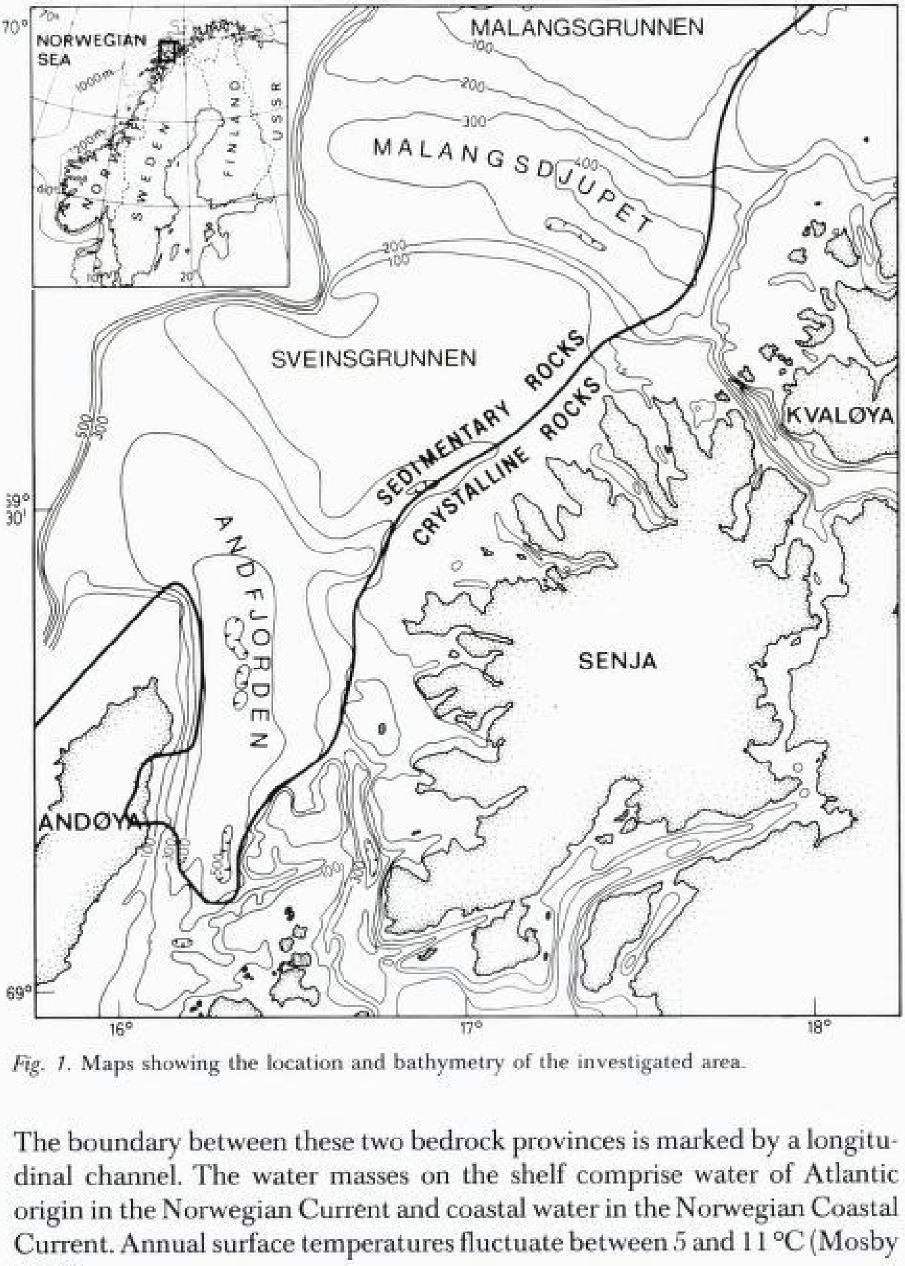

Fig. 1. Maps showing the location and bathymetry of the investigated area.

The boundary between these two bedrock provinces is marked by a longitudinal channel. The water masses on the shelf comprise water of Atlantic origin in the Norwegian Current and coastal water in the Norwegian Coastal Current. Annual surface temperatures fluctuate between 5 and 11 °C (Mosby 1968).

Material and methods

Altogether 120 gravity cores (inner diameter 100 mm) have been sampled and investigated by us (Fig. 2). At several of the sampling stations two or more cores have been recovered in order to achieve good stratigraphic control and partly to get enough material for radiocarbon dating.

The cores were split in two parts at the laboratory. Various geotechnical

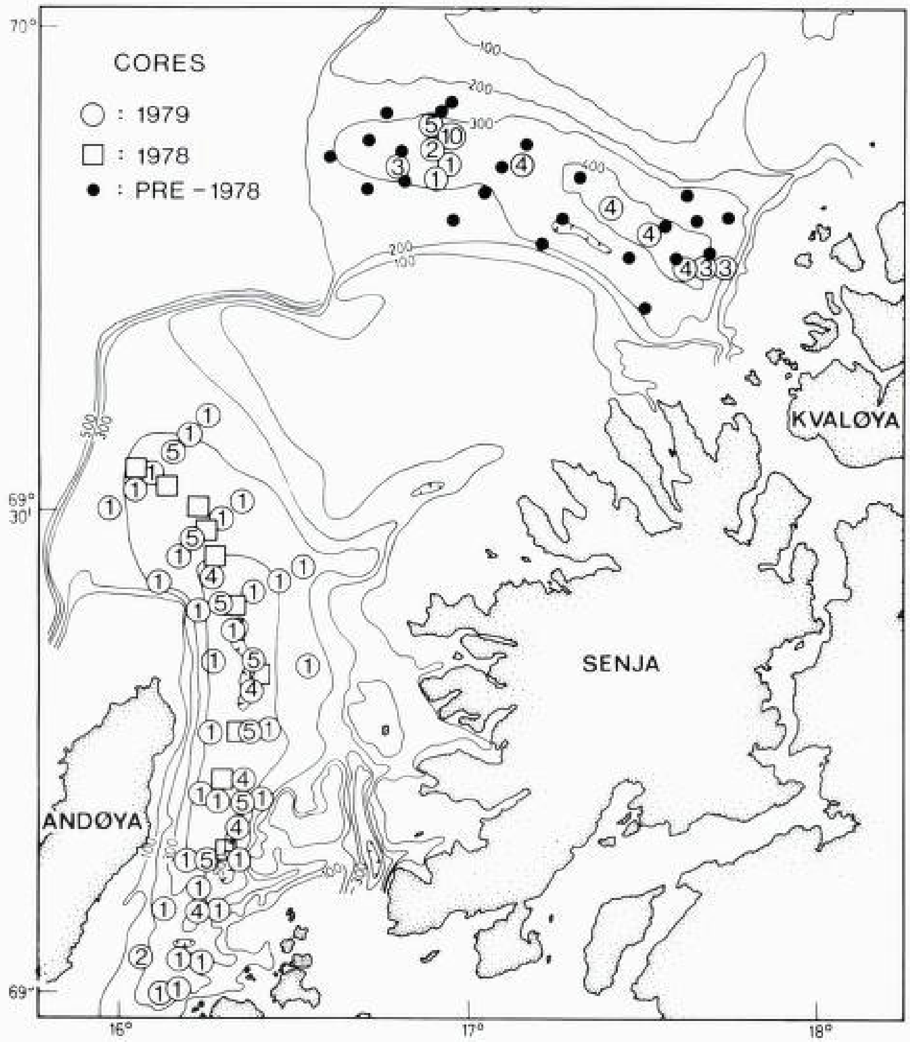

Fig. 2. Station map. Number of cores from each 1979-station is indicated. Only one core was recovered from each 1978 and pre-1978 stations. The pre-1978 cores are described by Elvsborg (1979).

and sediment-petrographical investigations were carried out: colour-determination using Munsell Color Charts; water content; shear strength by fall-cone test (Hansbo 1957); granulometric analysis by wet sieving and pipette analysis; carbonate content by gasometric analysis (Gross 1971); clay mineralogy by XRD and 'pebble' counts on the fraction 1–2 mm. A useful parameter for differentiating between various types glacigenic sediments and to quantify the ice drop activity was the number of lithoclasts in the 1–2 mm grade per 100 g dry sediment (Vorren et al. 1982).

The studies of the seismic stratigraphy are based on sparker profiles kindly put at our disposal by the Norwegian Continental Shelf Institute. The profiles have previously been described by Bugge & Rokoengen (1976).

Seismic stratigraphy

The greatest thickness of Quaternary sediments is found at the mouth of the troughs and at some scattered locations on the inner part of the shelf according to Rokoengen et al. (1979) (Fig. 3A). Based on the seismic studies they infer three glacial units (Fig. 3B) which were related to three glacial terminal events. In our opinion many parts of their glacial units represent local thickening of the Quaternary deposits, e.g. erosional remnants, which cannot be related to terminal events.

We have paid particular attention to Andfjorden. Even with all the available profiles at hand we feel confident on only one terminal moraine complex. This moraine complex is located c. 50 km from the shelf edge (Figs. 4 and 5). We name this complex the Flesen moraine after some skerries to the east of the moraine complex. Another ridge complex, 20–25 km from the shelf edge, is located on the eastern side of the trough. This ridge-system may be end moraines deposited in front of a glacier moving north along Andfjorden, or from a glacier moving westwards from Senja, or it may represent slide deposits. Additional data are needed before the genesis of this ridge system can be safely interpreted.

The thickness of the total postglacial glaciomarine/marine sediments in Andfjorden has been mapped (Fig. 4); a maximum thickness of about 50 m has been registered. This seismic sequence is characterized by parallel internal reflectors and a smooth surface. It should be noted that the sequence

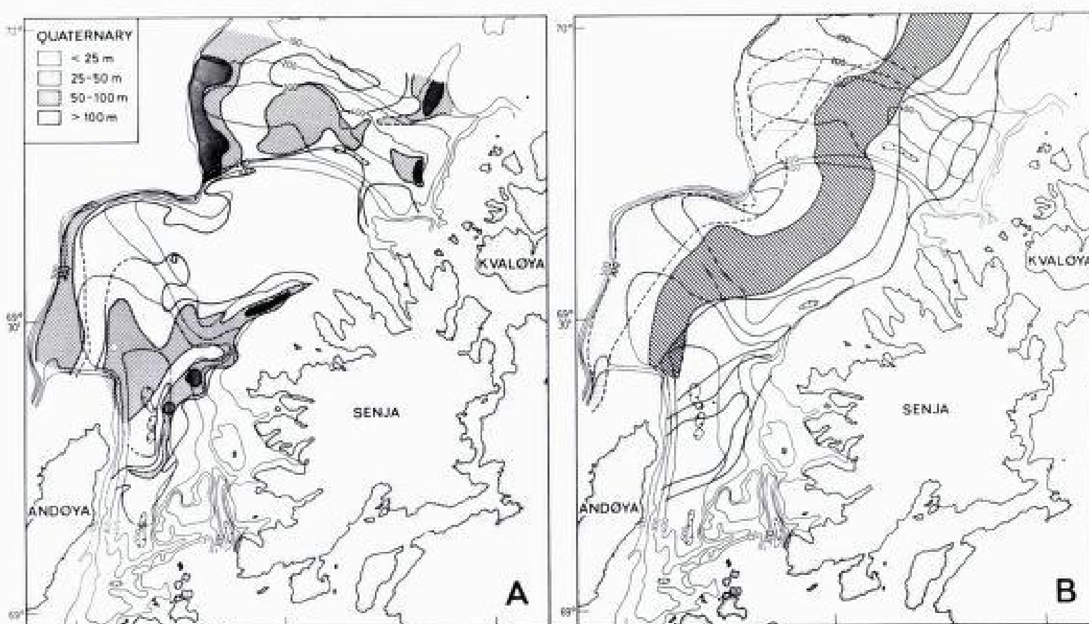

Fig. 3. A: Thickness of Quaternary sediments in milliseconds two-way travel time (equals metres if seismic velocity is 2000 m/sec). B: Glacial units, – After Rokoengen et al. (1979).

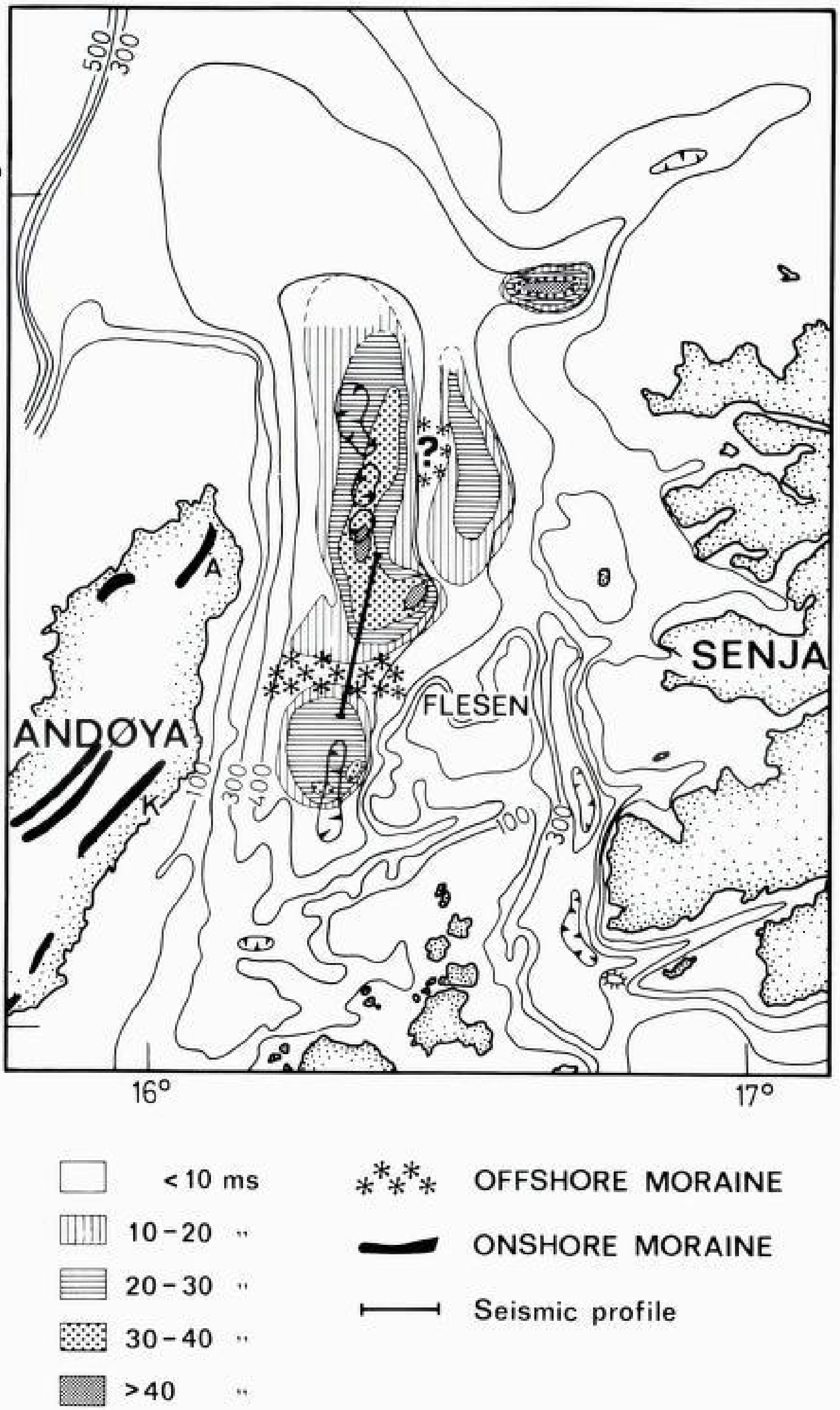

Fig. 4. Thickness of the postglacial glaciomarine/marine sediments in Andfjorden in millisecond two way travel time. Position of the submarine Flesen moraine and the questionable moraine (?) are indicated. The A and K denotes the Aeråsen moraine and the Kirkeraet ridge, respectively. Position of profile shown in Fig. 5 is indicated.

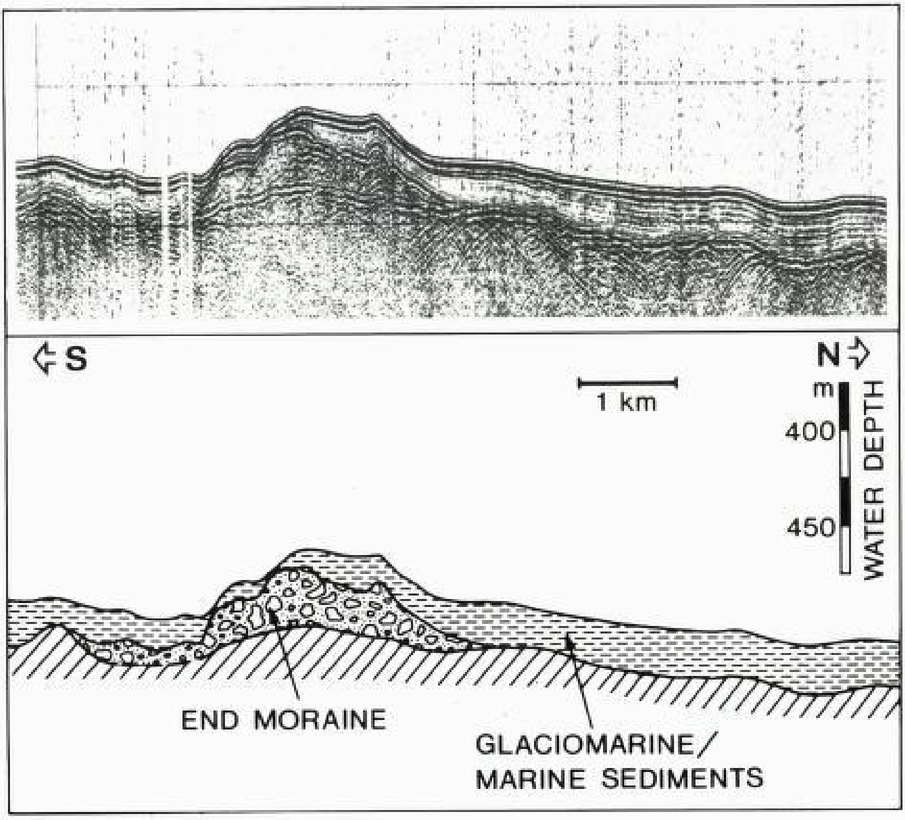

Fig. 5. Seismic (sparker) profile, with interpretation, across the Flesen moraine. Location is shown in Fig. 4.

overlies the Flesen moraine (Fig. 5) but thins on the proximal flank. The low boundary of the sequence is probably diachronous, becoming progressively younger towards the south. In the outer and marginal part of Andfjorden the sequence is lacking. This is partly due to bottom current erosion.

Lithostratigraphy

A generalized composite stratigraphy based on cores from the outer and central part of Andfjorden show nine lithostratigraphic units (Fig. 6). The same stratigraphic units, except units tG and tH, are found in Malangsdjupet. In the following a brief description and interpretation of the lithostratigraphic units is given; a more comprehensive discussion is given in Vorren et al. (1982).

The unit tA subcrop map (Fig. 7), indicates that units lying directly beneath unit tA become progressively older seawards. Thus there exists a hiatus of increasing length seawards (and towards the basin margins) between the late Holocene tA unit and underlying units. This situation has

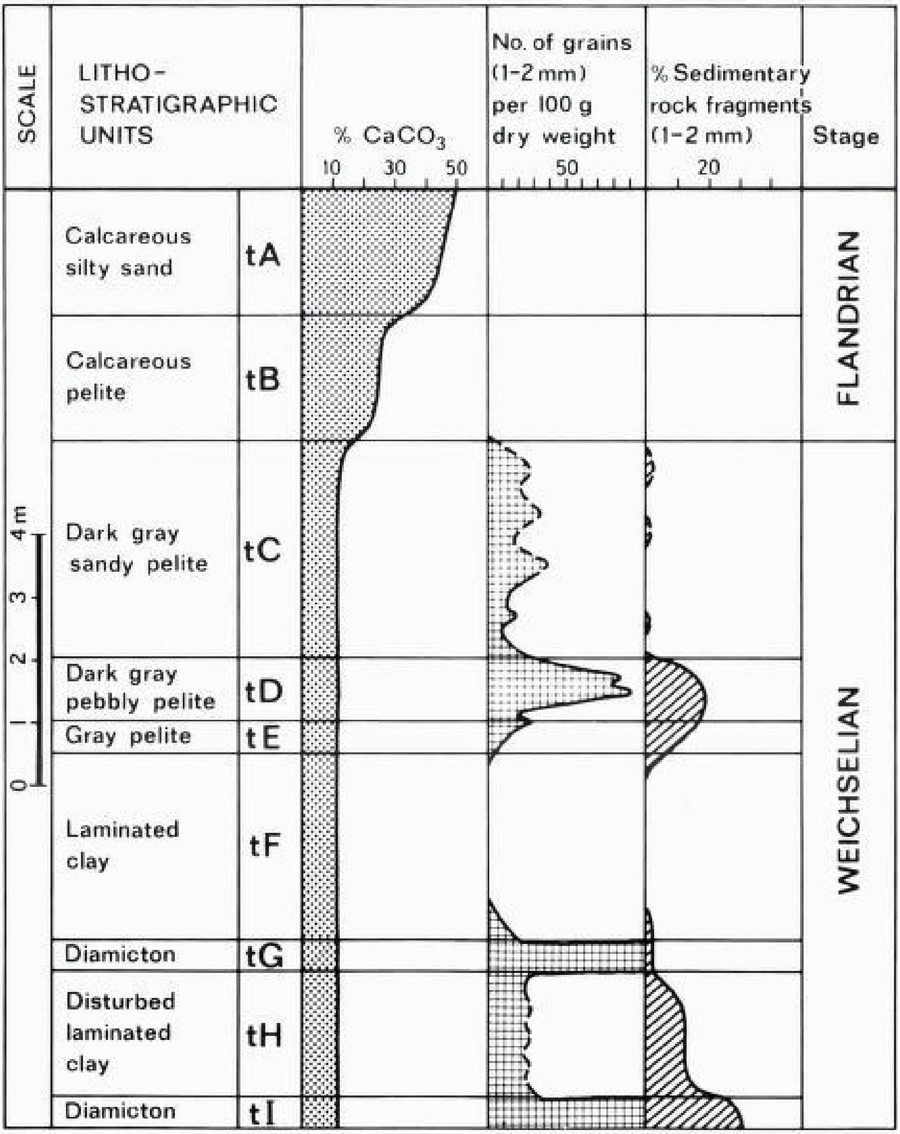

Fig. 6. Generalized composite lithostratigraphy of the upper beds in the outer and central reaches of Andfjorden.

given us the opportunity to construct a rather long stratigraphy based on only 5 m long cores. However, in most of the deeper parts of the throughs the Holocene sequence is so thick that we are not able to recover Weichselian sediments.

The window with unit tC sediments west of Flesen must be due to local thinning of the Holocene sequence over the Flesen moraine ridge (Fig. 5 and 6). The diamictons in the outer parts of Andfjorden and Malangsdjupet are represented by the tI and tG-units. In the inner parts of Andfjorden the diamictons are of a different composition.

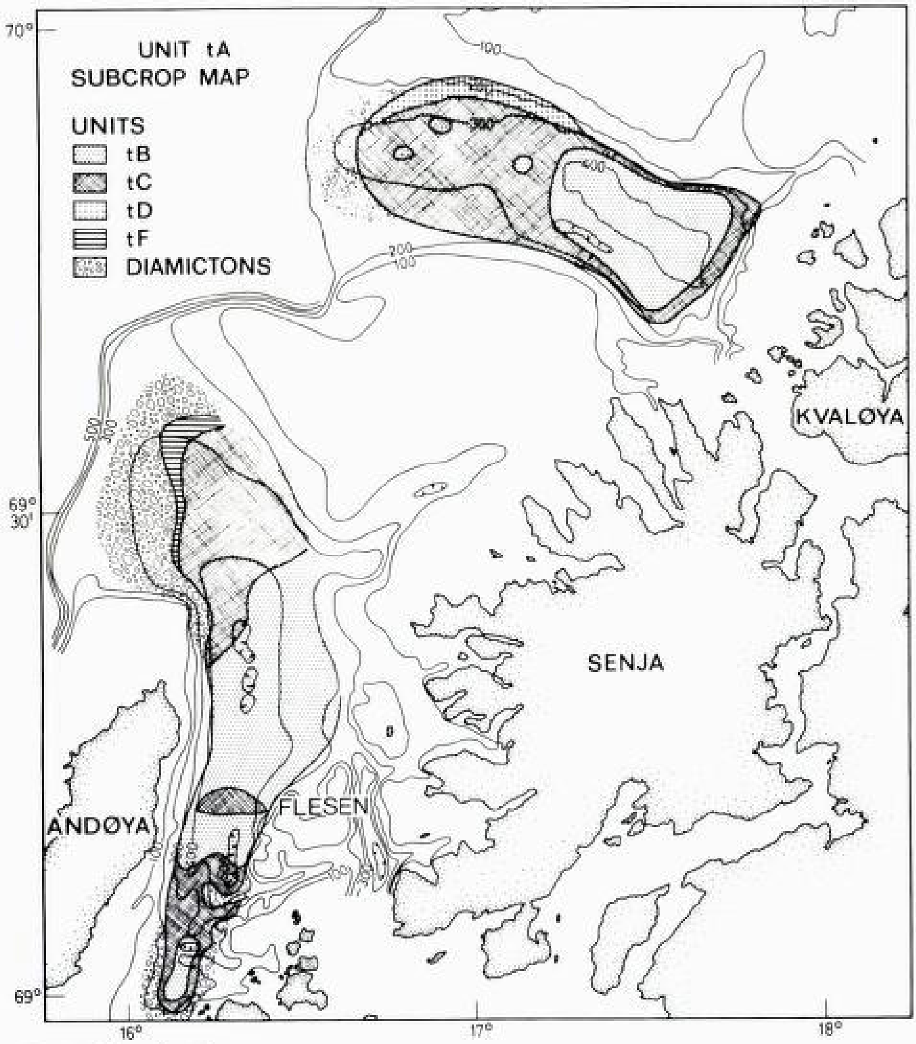

Fig. 7. Unit tA subcrop map.

Unit tI is a very dark grey diamicton. The upper boundary which is sharp, is sometimes represented by a sandy gravelly horizon. The low water content (15–20%) and the shear strength indicate overconsolidation. Macrofossils of boreo-arctic species occur as fragments, many of which are abraded. We are fairly certain that unit tI represents a true basal till.

Unit tH is a disturbed laminated clay found in outer Andfjorden. It is overlain by unit tG which is a dark grey diamicton. Compared with unit tI, unit tG contains less lithoclasts and has lower frequencies of sedimentary rocks. Like unit tI, unit tG only contains transported macrofossils. We believe that the most likely genesis of unit tG is as a basal till, although a proximal glaciomarine deposit should not be totally ruled out.

Unit tF is a (dark) olive-grey laminated clay. The maximum thickness found in the cores is 74 cm in Malangsdjupet and 195 cm in Andfjorden. The natural water content is 28–35% and the shear strength is normally between 0.4 and 1.0 t/m². The clay and silt which predominates in the grain-size composition is probably derived from suspension.

Ice-rafted grains occur only in the lower and upper part of this unit in Andfjorden. No macrofossils are found and the foraminiferal assemblage is characterized by a very low number of specimens per gram sediment. Taxa that at present are restricted to polar shelf areas dominate the lower part of unit tF, whilst there is an increase in the cosmopolitan *C. laevigata* upwards. The undisturbed lamination and the foraminiferal assemblage indicate a reatricted benthic fauna reflecting rather unfavourable ecological conditions. The fine texture indicates a low energy regime. We thus interpret these sediments as having been deposited in an environment which was covered with sea-ice, at least seasonally.

Unit tE is a 20–40 cm thick, light-coloured marine clay. The boundary with unit tF is transitional; it is placed where lamination is no longer visible. Unit tE resembles tF, except for a slightly higher dropstone content, and a massive structure. The massive structure, we believe, is the result of bioturbation. Small pectinids and sponge spicules occur, indicating a more diverse fauna than in tF-time.

Unit tD is a very dark grey pebbly pelite. The boundary with tE can be seen by a colour change. The thickness is 60–70 cm in Malangsdjupet and 70–90 cm in the outer reaches of Andfjorden, increasing shorewards to at least 190 cm. In the outer reaches this unit is massive; in the middle reaches it contains scattered light-coloured laminae; and in the inner reaches it is weakly bedded and laminated. The natural water content is 38–35% and the undrained shear strength about 0.7 t/m². This unit is characterized by a relatively high content of dropstones and by a high frequency of sedimentary rocks among the clasts in the outer reaches, decreasing southwards in Andfjorden.

Yoldiella spp., and the low saline indicator *Elphidium excavatum* (Smith 1970, Ellison & Nichols 1976) dominate the macrofauna and microfauna respectively. We believe that the tD-unit was deposited in a cold, near-glacier environment with a high iceberg influx.

Unit tC is a very dark grey sandy pelite. The lower boundary in some cores is marked by a thin (<0.5 cm) sandy layer/lens, but mostly it is barely visible. The boundary is defined by a minimum in dropstone content which is also the level at which sedimentary rock clasts almost disappear. Scattered lenses of sand and burrows occur frequently in this unit. Turbidites are found in the high relief parts of the inner reaches. Generally this unit contains fewer dropstones than tD and has a higher sand content which increases upwards. The natural water content is 25–35% and the shear strength 0.4–1.5 t/m² *Bathyarca glacialis* is a characteristic pelecypod and *Cassidulina reniforme* and *Nonion labradoricum* dominates the foraminifera assemblage. These species have at present an affinity for polar waters. We interpret the tC-unit as having been deposited in an iceberg environment, but with water masses being more temperate and dynamic than during tD-time.

Unit tB and tA comprise Holocene calcareous mud and sand, rich in shells of boreal molluscs and foraminifera reflecting the present hydrographic conditions. In the outer reaches unit tB is missing. The older part of the Holocene there is represented by a winnowing period with lag development.

Age

Radiocarbon dates have been obtained from units which contain macrofossils in sufficiently large quantities. These dates number 19 from the pre-Holocene units in the 1979 cores (Fig. 8 & Table 1). In order to obtain sufficient dating material some samples had to be taken from several cores. The cores were then very carefully correlated assuring that the material was recovered from corresponding stratigraphic levels. The dates fall in two distinct groups; between 10.000 YBP and 14.000 YBP and older than 36.000 YBP.

Fifteen of the dates are from material from unit tC. The two dates T-3634 and T-4035 are from material close to the lower boundary while T-3637 is close to the upper boundary. These date unit tC to between 10.000 and 13.000 YBP.

Three dates were obtained from unit tD. Dating T-3638 ($37.580 \pm ^{2210}_{1720}$) is from a crushed and abraded, obviously resedimented fragment of *Mya truncata*. The dated material from Malangsdjupet (T-3234) is collected from several cores, some of which had a very thin (or missing) overlying tC-unit. The age is definitely too young, possibly due to burrowing of more recent fauna into the underlying unit tD. The T-3633 dating accords with the dates from unit tC. The high standard deviation is due to the small sample size. An estimate of the sedimentation rate suggests that the lower boundary of unit tD is about 14,000 years old.

Dates from unit tI, T-3511, $36{,}760 \pm ^{1870}_{1510}$ and T-2499, $38{,}850 \pm ^{4070}_{2780}$ (Elvsborg 1979), if they represent finite ages, indicate that the glacier advanced over the outer shelf area after approximately 36,000 YBP.

The ages of the undated units tE, tF, tG and tH can only be estimated. The sedimentation rate seems to have been higher for unit tE and especially for tF than for the overlying units. As a rough estimate we believe that tE and tF were deposited within a period of a thousand years or so.

The interpretation of unit tG as a basal till implies that the lower boundary is a priori an erosional unconformity representing a hiatus of unknown length. The sandy-gravelly layer on top of this unit may represent a period of winnowing and, thus, also this boundary represents a hiatus of unknown length. This interpretation is supported by the presence of an up to 50 m-thick postglacial glaciomarine/marine sequence of mostly pre-tE sediments in the inner part of the trough (Fig. 4). Either the sedimentation rate must have been much higher in the inner parts during tF-time, or another possibility is a period during pre tE-time with erosion/non-deposition in the outer areas and sedimentation in the inner areas.

Table 1. Radiocarbon dates of pre-Holocene fossils from cores recovered from Andfjorden and Malangsdjupet in 1979. The T-number denotes the Trondheim-laboratory number.

No.	Core	Fossils	Weight	% Dated	Unit	YBP
T-3233	2-9	Astarte cf. crenata	2,4	90%	tC	11 820 ± 340
T-3234	2-3, 10	Y. intermedia, N. pernula				
		Y. lenticula, Scaphopoda indet	3.0	90%	tD	12 150 ± 340
		B. glacialis				
T-3390	56-1, 2, 3 & 4	B. glacialis, N. pernula	3.9	90%	tC	11 250 ± 270
		Astarte sp., Thyasira sp.				
		Scaphopoda indet, Y. intermedia				
T-3388	56-1 & 2	B. glacialis	1.8	90%	tC	12 430 ± 620
T-3389	56-1	Astarte cf. crenata	4.2	90%	tC	12 380 ± 270
T-3511	2-3, 4, 5, 7, 8, 9, 10,	Shell fragments	13.–	80%	tI	36 760 $\pm\frac{1870}{1510}$
	3-1, 2, 4-1, T-5-1					
T-3633	15-1, 2, 3, 4 & 5	Y. intermedia, Y. lenticula	1.3	95%	tD	13 630 ± 1250
T-3634	15-1, 2, 3, 4 & 5	Ophiura sp., Yoldia sp.	1.7	95%	tC	12 910 ± 420
	17-1	B. glacialis, N. pernula				
		Lunatia sp.				
T-3635	51-3	B. glacialis	1.8	95%	tC	12 320 ± 350
T-3636	51-4	B. glacialis	1.7	95%	tC	10 940 ± 390
T-3637	25-3	A. cf. crenata, B. glacialis				
		Y. lenticula, N. minuta	1.2	95%	tC	10 240 ± 510
		N. pernula, Ophiura sp.				
T-3638	9-1	M. truncata	6.1	90%	tD	37 580 $\pm\frac{2210}{1720}$
T-4030	33-1	B. glacialis	1.6	95%	tC	11 310 ± 280
T-4029	33-1	B. glacialis	3.2	90%	tC	12 140 ± 310
T-4033	33-1	B. glacialis, N. pernula	1.8	90%	tC	12 200 ± 350
T-4034	33-2	B. glacialis	1.1	95%	tC	11 240 ± 430
T-4035	33-4	B. glacialis	2.0	90%	tC	13 050 ± 350
	33-4					
T-4032	33-5	B. glacialis	2.8	90%	tC	11 920 ± 280
T-4031	31-1	B. glacialis	1.5	95%	tC	11 080 ± 380

Discussion and conclusions

Based on the results of the lithostratigraphic and seismic study we have constructed a tentative time-distance diagram (Fig. 9). Focusing on Andfjorden where the stratigraphy is most complete; four (or possibly five) glacial events can be discerned. They are represented by: two tills (the tI and tG-events), one (two) end moraine; (the Flesen and (?) events) and the glaciomarine unit tD (the D-event). These events are discussed briefly below and tentatively correlated with the terrestrial moraines on Andøya. Several people have contributed to the moraine studies on Andøya (Reusch 1903, Enquist 1918, Holmsen 1924, Undås 1938, 1967, Grønlie 1941, Møller & Sollid 1972, Bergstrøm 1973). It should, however, be noted that there is no general agreement, either on such fundamental questions as the genesis of the accumulations, or on correlation and chronology.

At the time of deposition of the tills tI and tG, the continental ice sheet probably reached the shelf edge. How far the ice front receded during the

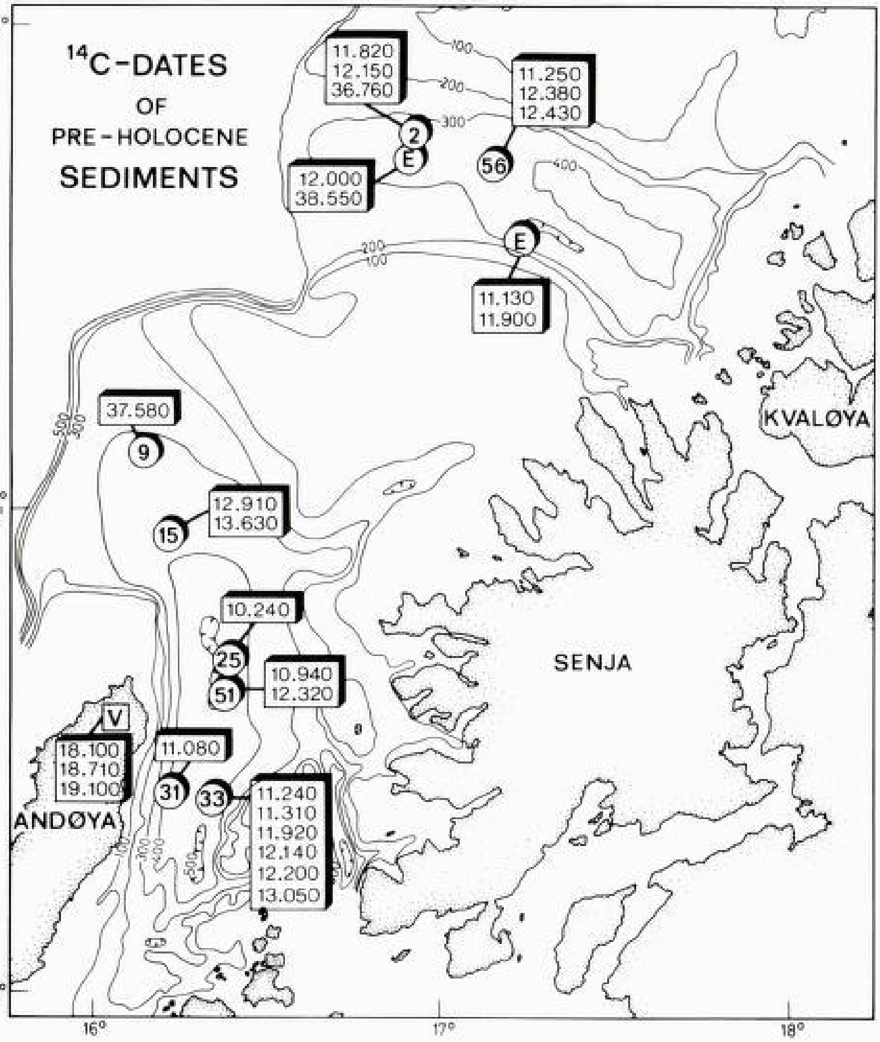

Fig. 8. Radiocarbon dates of Late Weichselian and older continental shelf sediments, cf. Table 1. The E-localities in Malangsdjupet are from Elvsborg (1979). The oldest postglacial dates onshore, V, are from K. D. Vorren (1978).

deposition of the laminated clay, unit tH, is unknown. According to the radiocarbon dates these two events must be older than c. 14,000 YBP and possibly younger than 36,000 YBP (if T-3511 and T-2499 are finite dates). Radiocarbon dates from the northern tip of Andøya (Fig. 8) indicate that this area was deglaciated before 18–19,000 YBP (K. D. Vorren 1978). It seems doubtful that the whole of Andfjorden could be glaciated without the ice sheet covering the northern tip of Andøya at the same time. Thus, the tI- and tG-events may also be older than 18–19,000 YBP. Another, slight possibility is that the tG-units are correlatable with the oldest indisputable continental end moraine on Andøya, namely the Aeråsen moraine (Fig. 4), which probably just predates the 18–19,000 YBP-dates (K. D. Vorren 1978).

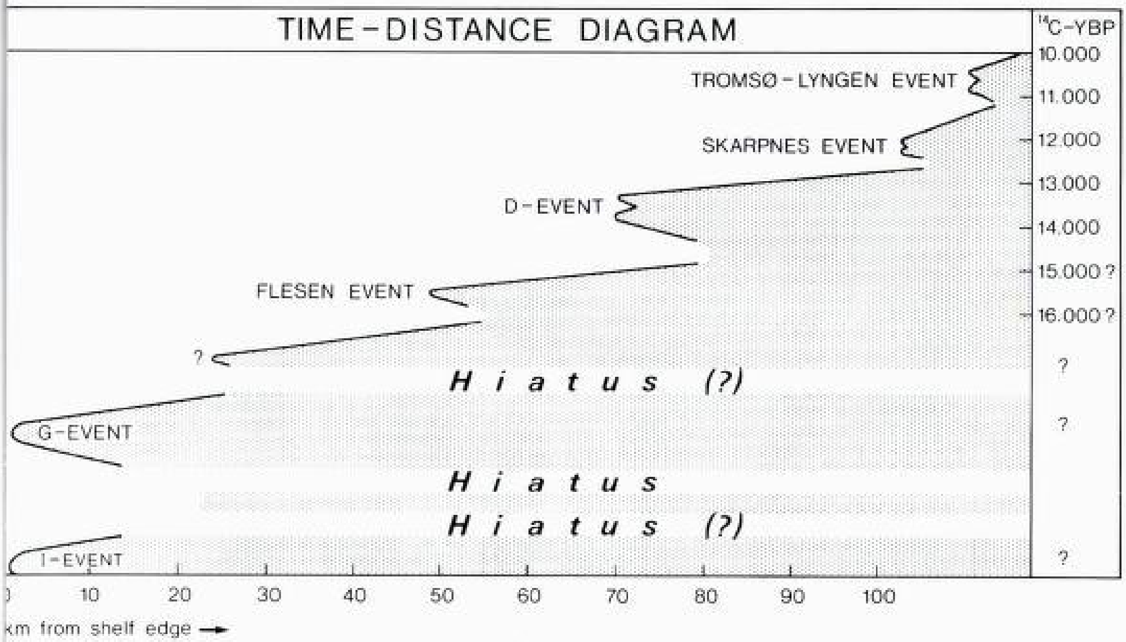

Fig. 9. Time- distance diagram for the margin of the continental ice-sheet in Andfjorden and adjacent shoreward areas.

Both the uncertain end moraine (20–25 km from the shelf edge) and the Flesen end moraine are overlain by unit tD-sediments. Thus the moraines must be older than c. 14,000 YBP. K. D. Vorren (1978) has recorded a cool event at c. 15,000–16,000 YBP. This may possibly be correlated with the Flesen event. The Flesen end moraine may possibly be correlated with the accumulation «Kirkeraet» on Andøya (Fig. 4). This accumulation is believed to be an end moraine by Holmsen (1924), Grønlie (1941) and Møller & Sollid (1972). Reusch (1903) and Bergstrøm (1973), however, interpret it as a shore phenomenon.

During later phases of deposition of unit tF and tE the ice front probably receded landward of the sedimentary/crystalline boundary (Fig. 1). This is indicated by a high content of land derived clay minerals in the upper part of unit tF and in unit tE. The extent of the continental ice sheet during the D-event is indicated by the dropstone composition. Large amounts of sedimentary dropstones in Malangsdjupet indicate that the ice sheet terminated beyond the sedimentary/crystalline boundary. The situation is somewhat more complicated in Andfjorden. There we find relatively high frequencies of sedimentary rock fragments in the outer part and very low frequencies in the inner part. The sedimentary rock fragments in Andfjorden must have been drifted in by icebergs from Malangsdjupet and elsewhere. Topographic conditions indicate that the ice margin may have halted at the submarine ridge system between Senja and Andøya and in the inner shallow part of Andfjorden. Possibly the end moraines on southern Andøya at Åse, Bjørnskinn and Bjørnskinnsmyra (Møller & Sollid, 1972, Bergstrøm 1973) may correlate with the D-event.

Two glacier advances marked by pronounced end moraines in the fjord areas have been dated to 12,500–12,000 YBP (Skarpnes event) and 11,000–10,000 YBP (Tromsø–Lyngen event), respectively (Andersen 1968, Vorren & Elvsborg 1979).

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REFERENCES

Ahlmann, H. W:son 1919: Geomorphological studies in Norway. *Geogr. Annaler 1,* 1–148, 193–252.

Andersen, B. G. 1968: Glacial geology of western Troms, North Norway. *Norges geol. Unders. 256,* 160 pp.

Bergstrøm, E. 1973: Den prerecenta lokalglaciationens utbredningshistoria inom Skanderna. *Stockholm Univ. Naturgeogr. Inst. Forskningsrapport 16.* 214 pp.

Bugge, T. & Rokoengen, K. 1976: Geologisk kartlegging av de øvre lag på kontinentalsokkelen utenfor Troms. *Inst. Kontinentalsokkelundersøkelser 85,* 51 pp.

Dahl, E. 1955: Biogeographic and geological indications of unglaciated areas in Scandinavia during the glacial ages. *Geol. Soc. America Bull. 66,* 1499–1519.

Ellison, R. L. & Nichols, M. M. 1976: Modern and Holocene foraminifera in the Chesapeake Bay region. In: Shafer, C. T. & Pelletier, B. R. (eds.) *First Int. Symp. on benthonic foraminifera of continental margins, 1, Halifax 1975, Part A: Ecology and Biology, Maritime sediments, Spec. Publ. 1.*

Elvsborg, A. 1979: Late Quaternary sedimentation in a glacial trough on the continental shelf off Troms, northern Norway. *Norsk geol. Tidsskr. 59,* 309–325.

Enquist, F. 1918: Die glaziale Entwicklungsgeschichte Nordwest-skandinaviens. *Sveriges Geol. Undersøkelser ser. C. 285.*

Gross, M. G. 1971: Carbon determination. In Carver, R. E. (ed.): *Procedures in Sedimentary Petrology.* Wiley Interscience. N.Y. 653 pp.

Grønlie, O. 1941: On the traces of the ice ages in Nordland, Troms and the southwestern part of Finnmark in northern Norway. *Norsk geol. Tidsskr. 20,* 1–69.

Hansbo, S. 1957: A new approach to the determination of the shear strength of clay by the fall cone text. *Royal Swedish Geotechn. Inst. Proc. 14.*

Holmsen, G. 1924: Kvartærgeologiske iakttagelser på Andøen. *Norges geol. Unders. 122,* 80–85.

Ives, J. D. 1975: Delimitation of surface weathering zones in eastern Baffin Islands, northern Labrador and Arctic Norway: a discussion. *Geol. Soc. America Bull. 86,* 1096–1100.

Lien, R. & Myhre, L. 1977: Investigations of the sea bed cover on the continental shelf offshore Troms and western Finnmark. *Inst. Kontinentalsokkelundersøkelser 92,* 33 pp.

Mosby, H. 1968: Surrounding seas. In Sømme, A. (ed.): *A Geography of Norden.* J. W. Cappelens forlag, Oslo, 343 pp.

Møller, J. J. & Sollid, J. L. 1972: Deglaciation chronology of Lofoten – Vesterålen – Ofoten, North Norway. *Norges geogr. Tidsskr. 26,* 101–133.

Reusch, H. 1903: Nogle optegnelser fra Andøya. *Norges geol. Unders. 36,* 26–37.

Rokoengen, K., Bugge, T. & Løfaldli, M. 1979: Quaternary geology and deglaciation of the continental shelf off Troms, north Norway. *Boreas 8,* 217–227.

Smith, R. K. 1970: Late glacial foraminifera from southeast Alaska and British Columbia and a world-wide high northern latitude shallow-water fauna. *Archives des Sci., 23 (3),* 675–705.

Vorren, K.-D. 1978: Late and Middle Weichselian stratigraphy of Andøya, north Norway. *Boreas 7*, 19–38.

Vorren, T. O. & Elvsborg, A. 1979: Late Weichselian deglaciation and paleoenvironment of the shelf and coastal areas of troms, north Norway – a review. *Boreas 8*, 247–253,

Vorren, T. O., Hald, M., Edvardsen, M. & Lind-Hansen, O.-W. 1982: Glacigenic sediments and sedimentary environments on continental shelves: general principles with a case study from the Norwegian shelf. In Ehlers. J. (ed.): *Glacial Deposits in Northern Europe*. A. A. Balkema, Rotterdam.

Undås, J. 1938: Kvartærstudier i Vestfinnmark og Vesterålen. *Norsk geol. Tidsskr. 18*, 81–218.

Undås, J. 1967: *Om maksimum av siste innlandsis i ytre Vesterålen*. J. W. Eides boktrykkeri A.S., Bergen. 31 pp.

Eemian and Weichselian Deposits at Bø on Karmøy, SW Norway: A Preliminary Report

B. G. ANDERSEN, H. P. SEJRUP & Ø. KIRKHUS

Andersen, B. G., Sejrup, H. P. & Kirkhus, Ø. 1983: Eemian and Weichselian deposits at Bø on Karmøy, SW Norway: a preliminary report. *Norges geol. Unders. 380*, 189–201.

Lithological and biostratigraphical studies, including studies of pollen, foraminifera and molluscs, together with amino acid analysis, radiocarbon datings and Uranium–Thorium datings of the deposits in a 21 m section with marine beds and tills at Bø on Karmøy gave the following main stratigraphical results: *Haugesund Stadial:* 6 m unit of clayey till and 5 m of tectonized marine beds (Late Weichselian and possibly late Middle Weichselian). *Bø Interstadial:* 5 m unit of marine sand and gravel with a cool climate fauna and flora (early Middle Weichselian or late Early Weichselian). *Karmøy Stadial:* 1.5 m sandy, gravelly till with large striated boulders and 1 m tectonized marine beds. *Torvastad Interstadial:* 0.5 m brown organical sand with a cool-climate flora and fauna (Early Weichselian). *Avaldsnes Interglacial:* 3 m yellow marine sand with a warm climate fauna and flora (Eemian).

Bjørn G. Andersen, Institutt for geology, Univ. i Oslo, P.O. Box 1047, Blindern, Oslo 3, Norway.

Hans Petter Sejrup, Geologisk inst. avd. B, Allégt. 41, N-5014 Bergen-Universitetet, Norway.

Øyvind Kirkhus, Statoil, P.O. Box 300, N-4001 Stavanger, Norway.

Introduction

The presented results are based on studies from two excavations made on the floor of the old Bø clay pit at Karmøy (Fig. 1). The first excavation was dug in 1979 during extremely bad weather conditions. Interglacial deposits supposed to be of Eemian age were reached at a depth of 6 m – 10 m, 1 m to 7 m below sea level. In the following year a new excavation was made only 10 m from the former site. However, this time the very bouldery till which overlies the deposits of supposed Early Weichselian and Eemian age could not be penetrated, and only the beds above the till were studied.

Review of previous studies

BØ, KARMØY

Øyen (1905), Ringen (1964) and Andersen et al. (1981) observed the 6 m to 8 m thick bed of clayey till with shell fragments which overlies the marine beds at Bø. Øyen considered the till to be of pre-Weichselian age, but Ringen suggested that it is of Weichselian age and deposited by a glacier which moved in westerly direction. The investigations carried out by Andersen et al. (1981) supported Ringens observations. Below the till they observed *Mya* beds, and they correlated the beds with the Nygaard Interstadial somewhere between 40000 and 50000 years B.P. old. On the basis of amino-acid analysis Miller et al. (in press) arrived at the conclusion that the *Mya* beds are most likely of late Early Weichselian age.

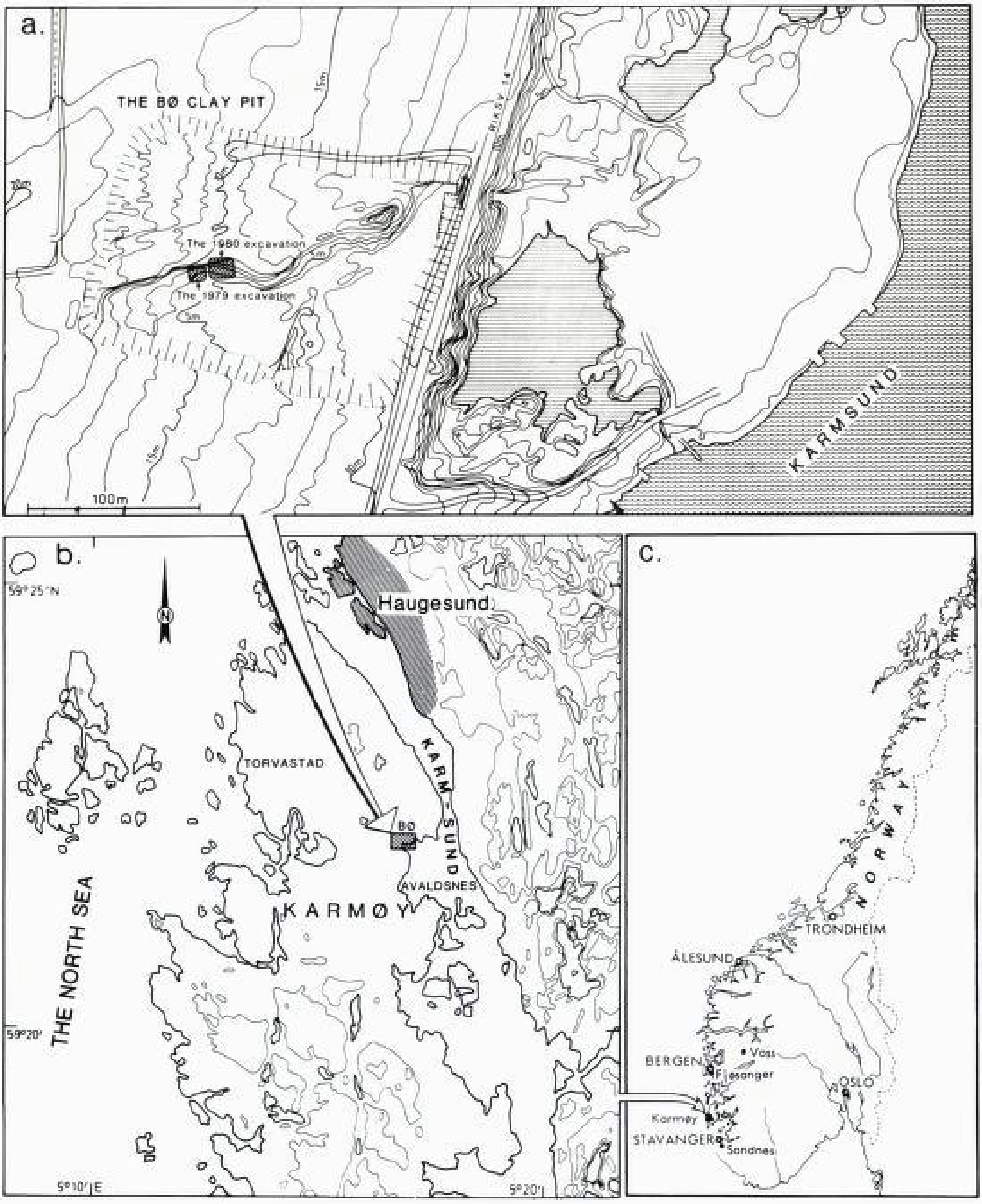

Fig. 1. Key map
a) Location of Bø clay pit and of the two excavations.
b) Location of Bø.

NYGAARD, KARMØY

In a clay pit at Nygaard, which lies only 4 km from Bø, Øyen (1905) recorded *Mya* beds which rest on a very bouldery till and are overlain by glacially tectonized *Portlandia arctica* beds. Andersen et al. (1981) suggested that the *Mya* beds represent an interstadial, the Nygaard Interstadial, and radiocarbon-dated *Mya* shells indicated an age approximately 40 000–45 000 years B.P. for the beds. The bouldery till below the *Mya* beds were supposed to represent an Early Weichselian stadial, a Karmøy Stadial.

Lithostratigraphy

The lithostratigraphy exposed in the two excavations at Bø is shown in Figs. 2 and 3. Altogether 5 formations were distinguished: Haugesund Diamicton, Bø Sand, Karmøy Diamicton, Torvastad Sand and Avaldsnes Sand. Fig. 4 shows the results of grain-size analysis and of $CaCO_3$ determinations of samples from both excavations.

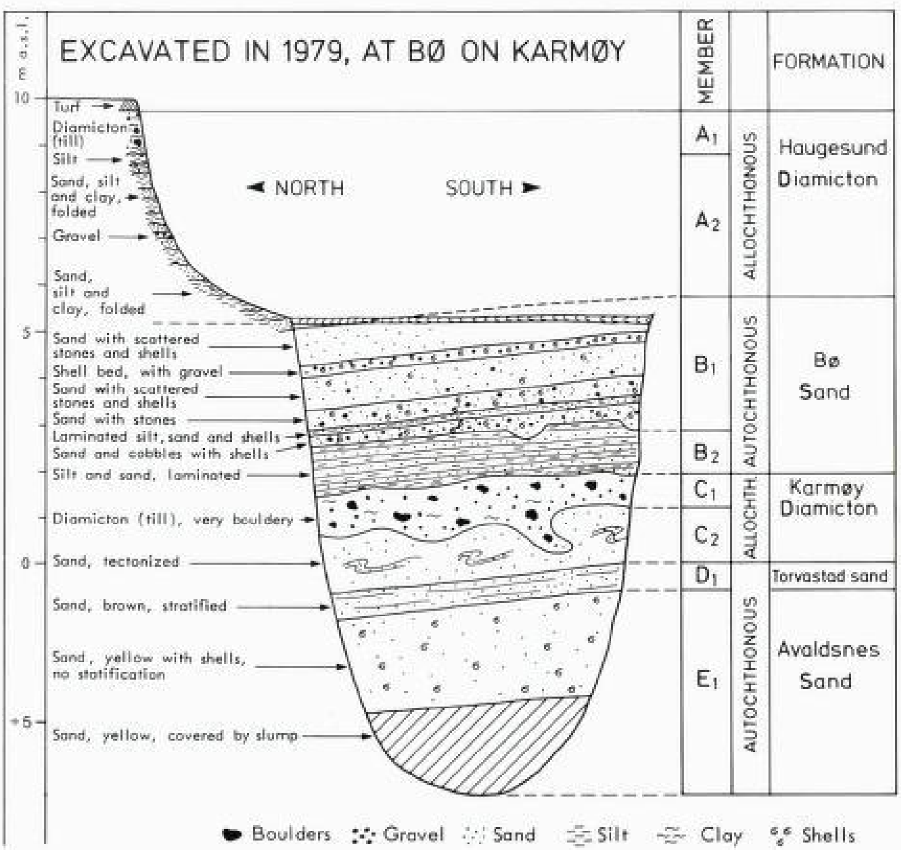

Fig. 2. Section exposed in 1979.

The Avaldsnes Sand is the oldest formation. About 4 m of this sand was exposed from 1 to 6 m below sea leavel, but only the upper 3 m could be sampled. It is a yellow fine sand with no clear stratification. Marine shells, including some bivalves in living position were seen throughout the formation. A high $CaCO_3$ content (20%) and a low loss of ignition (1.2–1.4%) could indicate relatively favourable current conditions. This together with the interpretation of the molluscan fauna (p. 197) suggest that there could have been a connection with the open sea across the island, and the sea level was most likely more tan 15 m–20 m higher than at present.

The Torvastad Sand is a c. 0.4 m thick formation of brown, partly stratified sand with much organic material of plant remains. Marine molluscs were present in the sand, but not as common as in the Avaldsnes Sand. The grain-size distribution was similar to the distribution in the Avaldsnes Sand (Fig. 4). A sheltered brackish-water environment, possibly a lagoon environment, seems most likely for this sand. The amino-acid chronology (Miller et al. in press) suggests a considerable hiatus between Torvastad Sand and the Avaldsnes Sand, but this hiatus was not observed in the excavated profile.

The Karmøy Diamicton was divided into two members C_1 and C_2. Member C_1 is a very bouldery till and C_2 is composed of strongly tectonized and folded sand beds with till lenses. All sand beds within C_2 seem to have been derived from the Torvastad and Avaldsnes Sands. Some of the erratics within C_1 are very large, 1 m to 3 m in diameter, and many of them are nicely striated. Both

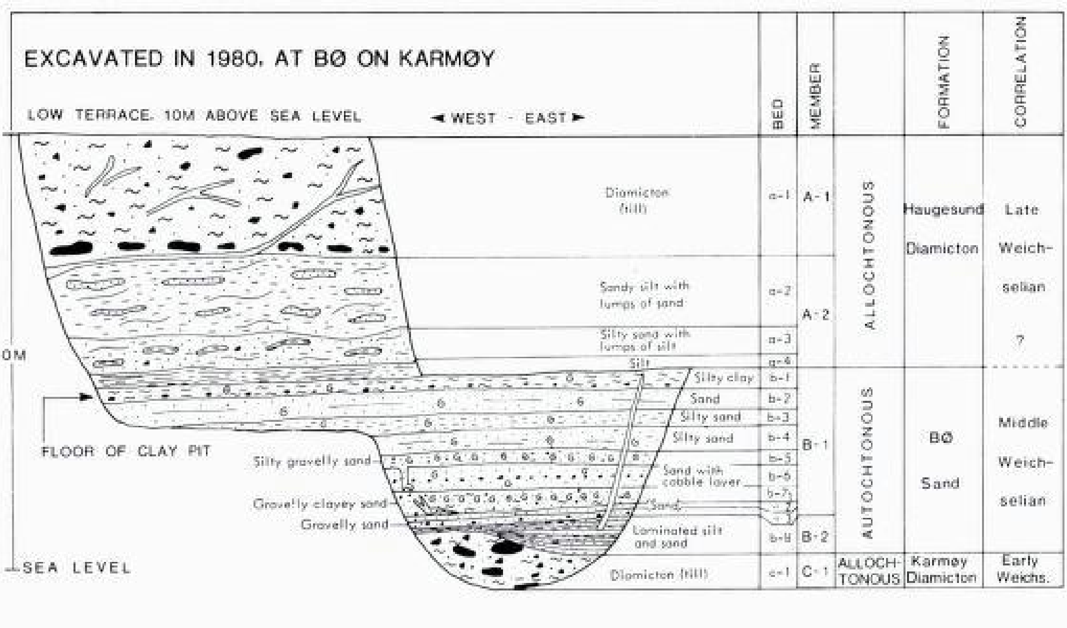

Fig. 3. Section exposed in 1980.

pebble counts, fabric analysis and the striation show that the Karmøy till was deposited by glaciers which moved in southwesterly, westerly and northerly directions at various phases. The Karmøy Diamicton represents a stadial older than 50 000 years B.P. (see the following discussions).

The Bø Sand consists of 8 beds with a combined thickness of about 5 m (See Figs. 2 and 3). There is a hiatus between beds b_8 and b_7, and this hiatus separates member B_1 from B_2.

Member B_2 (bed b_8) is a laminated silt and fine sand which lies comformably on the surface of the Karmøy Diamicton. The thickness of b_8 varies from 1.5 m in the depressions to about 0.2 m on the higher parts of the till surface. A few scattered stones within the member are most likely dropstones. This suggests that the silt was deposited near an ice front. The lack of fosssils and organic carbon in b_8 together with the lamination could indicate that the bed was deposited in fresh or brackish water. Bed b_8 therefore most likely represents a phase during the glacial retreat immediately after the deposition of Karmøy till.

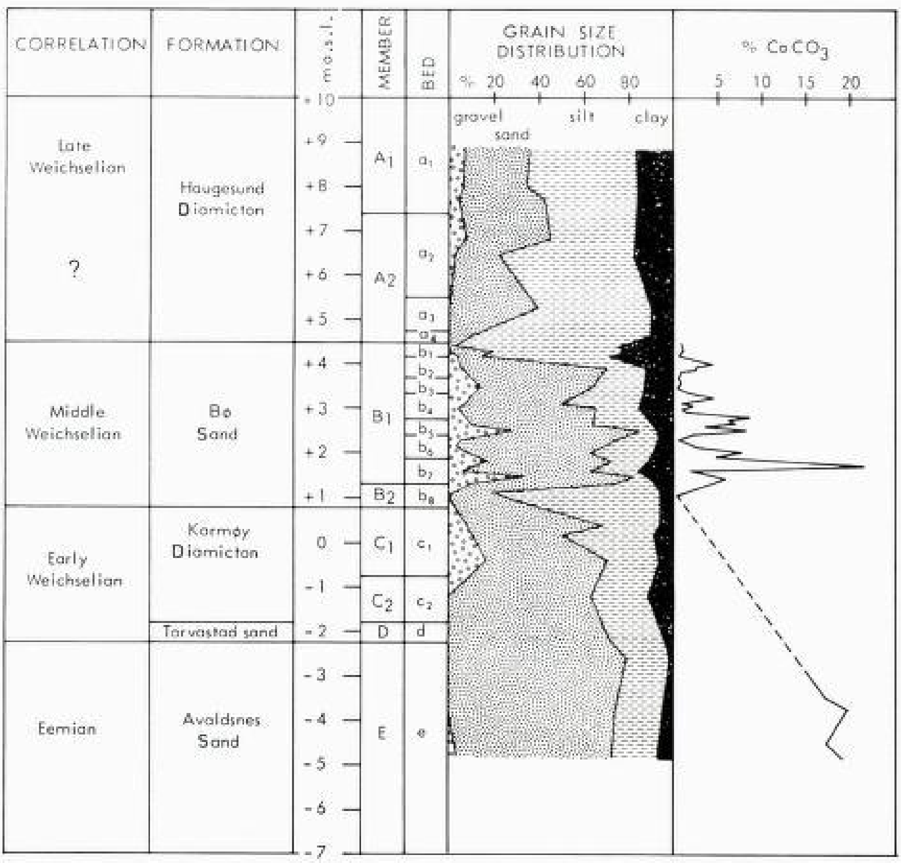

Fig. 4. Grain-size distribution and $CaCO_3$ content.

Member B_1 (beds b_7–b_2)

The upper laminae within bed b_8 are truncated by a bed of gravelly sand at the base of bed b_7. In addition about 0.2 m deep erosion channels were observed (Fig. 2). The erosion could have taken place below the sea level, but seen in connection with bed b_8 the erosion could as well occurred at or above sea level. Bed b_7 is very fossiliferous and it supposedly represents the most favourable part of an interstadial. Therefore, the hiatus below bed b_7 could represent a considerable part of an early phase of this interstadial.

Silty sand is the main constituent in all of the beds in B_1 (see Fig. 4). In addition some cobbles, gravel and clay occur, and stones with rounded or subrounded edges dominate in all beds. Some stones are striated which suggests a glacial origin. Paired bivalves were seen in all beds, and in some beds the shells are concentrated in zones which correspond with the peaks on the graph for the $CaCO_3$ content. Particularly bed b_7 is very fossiliferous.

Bed b_1 is a stratified silty clay with many dropstones and a molluscan fauna with *Portlandia arctica*. This is clearly a glaciomarine near-ice deposit.

The Haugesund Diamicton was divided into two members, A_1 and A_2. *Member A_1* is a clayey till with shell fragments. Only the lower 2 m of the about 6 m thick till unit was exposed in the excavated sections. The till fabric (Ringen, 1964) together with the striation direction on bedrock show that it was deposited by a glacier which moved almost due west. Radiocarbon dates and Uranium–Thorium dates on thick *Mya truncata* shell fragments within the till suggest an age between 35 000 years B.P. and 50 000 years B.P. for the shells, and the till must be younger.

Member A_2 consists of strongly tectonized beds of sand, silt and clay. The beds have obviously been moved in a westerly direction by the overriding glacier.

Pollen stratigraphy

Samples from the Avaldsnes, Torvastad and Bø formations were analyzed. There were numerous pollen grains in the sand from the two first mentioned formations, and very few in the beds from the Bø Formation. In beds b_8, b_6, b_4 and b_2 there were too few grains for statistical treatment.

Most of the grains were corroded, and 5%–15% could not be identified. Calculations are based on a pollen sum (P) for aboreal pollen (AP) and non-arboreal pollen (NAP). Relatively few samples have been analyzed, and the presented diagram shows only the preliminary results.

POLLEN ZONES WITHIN THE AVALDSNES–TORVASTAD SANDS

The two oldest zones are based on only two spectra and they are therefore questionable.

The Q.M. – Corylus assemblage zone

The zone is defined by an 8% NAP value, 12% Q.M. and 10% Corylus values. The Betula value is as low as 18%. If the presented analysis is representative this zone seems to represent the most favourable conditions.

The Betula – Poaceae assemblage zone

A 24% NAP value and a 34% Betula value define the zone. The Q.M. and Pinus values are respectively 2% and 5%. If the analysis is representative this zone represents a less favourable phase than the zones below and above. The high NAP value suggests that some deforestation had taken place.

The Pinus assemblage zone

Pinus values between 30% and 45%, Q.M. values at about 10% and NAP values less than 10% define this zones, which represents a favourable climate with a closed forest.

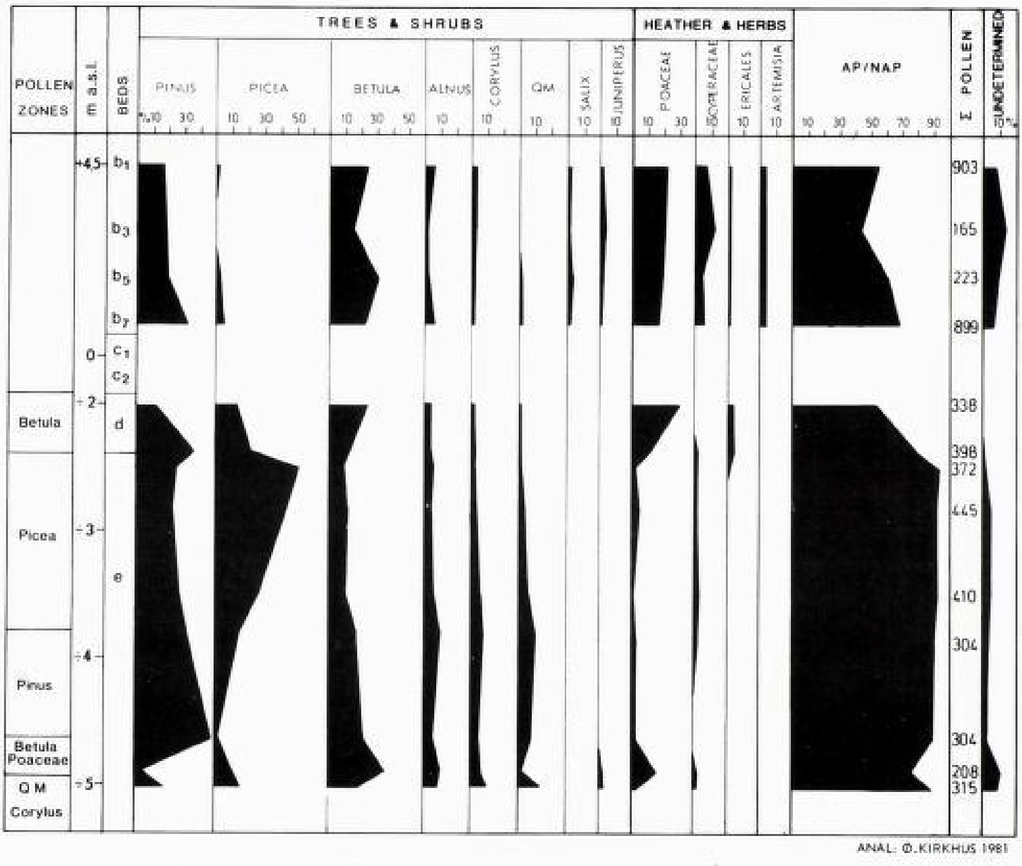

Fig. 5. Pollen diagram. The presented values for beds b_1, b_3, b_5 and b_7 are means of, respectively, 3, 2, 2 and 7 analysed samples.

The Picea assemblage zone

The zone is defined by 2%–10% NAP values, and 20%–52% *Picea* values. In addition the *Pinus* and Q.M. values are relatively high, respectively 20%–35% and 4%–8%. The zone represents a boreal Picea–Pinus forest with some leave trees.

The Betula assemblage zone

A rise in Betula and NAP values from respectively 12% to 25% and 20% to 53%, a drop in *Pinus* and *Picea* values to about 12% and the disappearance of Q.M. characterize this zone, which must represent a cool period, possibly with no trees during the latest phase. The amino-acid chronology (p. 199) suggests an Early Weichselian age for the *Betula* assemblage zone, and it suggests a considerable age difference between this zone and the *Picea* assemblage zone (p. 199).

THE BØ SAND

Many of the pollen grains in the analyzed samples are most likely resedimented, and this in addition to the low number of grains in the samples makes the interpretation problematic. The presence of *Artemisia* (about 4%), *Salix* (about 2%), *Juniperus* (about 3%), *Betula* (20%–30%) and NAP values of (30%–55%) suggest an open, cold climate vegetation. Pollen of *Q.M.*, *Corylus*, *Pinus* and *Picea* must be secondary resedimented pollen. This conclusion is supported by the fact that they were all, except for Q.M., found in bed b_1 which contains a high-artic *Portlandia arctic* molluscan fauna.

Correlation with the pollen flora at Fjøsanger

Mangerud et al. (1981) recorded interglacial deposits between beds of glacial deposits at Fjøsanger near Bergen. The corresponding interglacial was called the Fjøsangerian Interglacial and it was correlated with the Eemian. The pollen diagram from the Fjøsangerian deposits shows a division into 5 zones. The youngest zone is a *Picea* zone with 20%–40% *Picea*, 5%–25% *Pinus*, less than 10% *Q.M.*, generally less than 15% *Corylus* and 5%–10% *Betula*. The *Picea* zone overlies a *Corylus–Quercus–Alnus* (C–Q–A) zone with very high *Corylus* values (max. 30%), Q.M. values (max. 25%), *Alnus* values (max. 15%) and *Pinus* values (max. 40%). The *Picea* values lie below 10% and the *Picea* curve starts in this zone.

The *Picea* zone at Karmøy resembles very much the *Picea* zone at Fjøsanger. In both zones the *Picea* and the *Pinus* values are very high and *Q.M.*, *Alnus Corylus* and *Betula* values are relatively low (5%–15%). The *Pinus* zone and the older zones at Karmøy are most likely younger than the middle of the C–Q–A zone at Fjøsanger since they all contain *Picea* and in general low values of *Corylus*. The highest *Corylus* values (10%) and *Q.M.* values (15%) at Karmøy were found in the *Q.M.* – *Corylus* zone, which could correspond to the upper part of the C–Q–A zone at Fjøsanger. In addition, the very low values of both *Pinus* and *Picea* immediately above the two zones support this correlation. Thus, the pollen assemblages found in the *Avaldsnes Sand* may be correlated with parts of the *Corylus* – *Quercus* – *Alnus* and the *Picea* assemblage zones of the Fjøsangerian.

Marine molluscs

Fig. 6 shows the distribution of seven selected bivalves and one gastropod in a composite diagram, which is based on material collected from both excavations. A more detailed description and discussion of the molluscs and foraminifera faunas will be given elsewhere (Sejrup et al. in prep.). In the diagram the present-day distribution of the presented species is shown by the zoogeographic division used by Feyling-Hanssen (1955). The use of the symbols rare to frequent is based on a subjective judgement of data from samples and observations in field. Marine molluscs were found in all the beds at Bø except from the laminated silt and sand in bed b-8. All the marine fossils found at Bø are of species that are living in the world's ocean today. This suggests that the deposits are not older than the Hoxnian in England, since many of the species in older faunas are now extinct (West 1968).

The Avaldsnes Sand contains a mollusc fauna rich in individuals and species. This together with the presence of the species *Ostrea edulis* (Linné) and the lack of cold water indicators, suggest that the temperature conditions during deposition probably were similar or warmer than today (Dons 1936, Feyling-Hanssen 1955, 1960).

Warm demanding faunas of this kind could not have existed in this area without the presence of warm Atlantic surface water in the Norwegian Sea. According to Kellogg et al. (1978) such conditions have only prevailed two times in the Norwegian Sea during the last 440.000 years, in oxygen-isotope stage 5 e and in the Holocene. On the basis of the marine fauna the Avaldsnes Interglacial is correlated with oxygen-isotope stage 5 e. A similar fauna found by Mangerud et al. (1979) at Fjøsanger near Bergen was correlated with stage 5 e and the Eemian, and they recorded a pollen flora which supported this correlation.

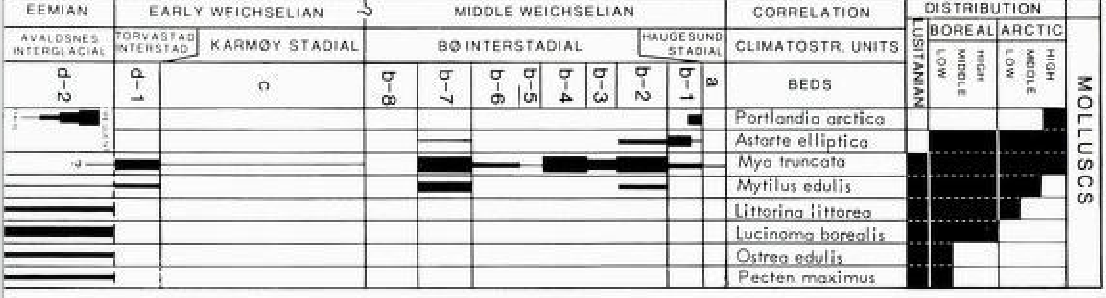

Fig. 6. The distribution of seven selected bivalves and one gastropod.

The presence of *Mytilus edulis* (Linné) and *Littornia littorea* (Linné) could indicate a deposition in the tidal zone or just below, but rather large frequencies of *Lucinoma borealis* (Linné), which is most common at depths between 20 and 50 m (Jensen & Spärck 1934), suggest that the extreme shallow-water species could have been redeposited. No safe statement can thus be made about the depth of deposition.

As mentioned before, the boundary between the Avaldsnes Sand and the superimposed *Torvastad Sand* (d-2) is probably erosional and represents a major hiatus. In the Torvastad Sand all the termophileous molluscs have been replaced by species wich tolerate rather cold water. However, the lithology shows no evidence of glaciomarine conditions, and this together with the presence of *Mytilus edulis* suggest that Torvastad Sand was deposited in a period when at least the costal areas of Western Norway were ice free. The marine conditions could have been similar to those found in the middle to low arctic region today.

Ockelman (1958) explains the northern limit of *Mytilus edulis* in arctic areas as a result of long seasons with sea-ice destroying the shallow-water faunas. In this period, which is called the Torvastad Interstadial, the sea level was most likely about 0–5 m above the present sea-level at Bø.

The Karmøy Diamicton (beds c_1 and c_2) represents a glacial advance, the Karmøy Stadial. Abraided fragments of marine molluscs found in these deposits show that the glacier, prior to the deposition of the till, picked up material from older deposits.

The Bø Sand. No marine macrofossils were found in the laminated sand-silt in bed b–8. A possibility is that this bed represents a very low-salinity depositional environment just in front of the retreating glacier. The beds b_7 to b_2 are rich in marine molluscs, and many molluscs lay in living position. The relative abundance between the different species is changing from bed to bed, but this might be a result of small local changes in the sedimentation environment. No thermophileos species were found, but the fauna suggests that the marine condition in this period might have been slightly more favorable than in the period when the Torvastad Sand was deposited. The presence of *Mytilus edulis* indicates shallow water during deposition of beds b_7 to b_2.

Both the lithology (Fig. 4) and the mollusc fauna of bed b–1 differ from those of bed b_8 to b_2. The fauna is poorer in species, and in the upper part the high-arctic species *Portlandia arctica* dominates. This species can live under the extreme conditions near calving glaciers (Ockelman 1958). Therefore both the sediments and the lithology in b–1 suggest that a glacier was close to to the Bø locality when this bed was deposited.

The Haugesund Diamicton. Only scattered shell fragments were found in the two members of this formation. Fragments of *Mya truncata* are most common. They were most likely picked up by the glacier from the Bø Sand.

Dating and correlations

Radiocarbon dates. Shells of *Mya truncata* from the Bø Sand, beds b_7–b_5–b_3, and b_1, have been radiocarbon dated (Fig. 7). Their radiocarbon age is about the same, about 39 000–42 000 years B.P. The same age was obtained for *Mya* shell fragments from the Haugesund Diamicton (a_1, 3 samples) and from the *Mya* beds at Nygaard (3 samples), Andersen et al. (1981). For a shell from the Avaldsnes Sand an age of about 49 000 years B.P. was obtained, but this is considered to be a minimum age (cf. Fig. 7).

U-Th dates of samples from the different beds will be carried out. Three samples of *Mya truncata* fragments from the clayey Haugesund till (a_1) were U-Th dated (Andersen et al. 1981). Three fractions of each sample were dated, and the ages of the inner fractions range between 36 200 and 40 000 years B.P. The apparent radiocarbon age of the three samples range between 34 000 and 38 000 years B.P.: see Andersen et al. (1981). Since the shell fragments were most likely derived from the *Mya* beds in the Bø Sand, the dates support the conclusion about a Middle Weichselian age for this formation.

AMINO-ACID GEOCHRONOLOGY

The results of the amino-acid analysis on the material from the two excavations at Bø have been published by Miller et al. (in press). The degree of isoleucine epimerization was measured in different species of marine molluscs and benthonic foraminifera. Fig. 7 shows the D-alloisoleucine/L-isoleucine ratios (aJle/Jle) in the total fraction of *Mya truncata* and *Cibicides lobatulus* from the marine beds at Bø. For comparison the ratios in the same species from Fjøsanger are also shown.

STRATIGRAPHY GENERAL	STRATIGRAPHY BØ/FJØSANGER	AMINO ACID GEOCHRON. (aJle/Jle) BØ		AMINO ACID GEOCHRON. (aJle/Jle) FJØSANGER		RADIOCARBON BØ
		Mya tr.	Cib. lob	Mya tr.	Cib. lob.	Years B.P.
late Late Weichselian	late Late Weichselian	0.076×	0.044×	0.076×	0.044×	
Middle Weichselian	Bø Interst.	0.14	0.102			b_1 37.000 $^{+800}_{-700}$ b_2 41.300 $^{+1200}_{-1000}$ b_4 39.600 $^{+1100}_{-1000}$ b_7 40.600 $^{+1200}_{-1000}$
Early Weichselian	Torvastad Interst. Fana Interst. Gulstein Stadial	0.21	0.123	0.27 0.28		
Eemian late / early	Avaldsnes/Fjøsanger Intergl.		0.169	0.32 0.37	0.164 0.205	49.900 $^{+5500}_{-3300}$ ××

Fig. 7. Radiocarbon dates and the D-alloisoleucine/L.-isoleucine ratios in *Mya truncata* and *Cibicides lobatulus*. Ratios for the same species in beds at Fjøsanger are plotted for comparison.

The similar ratios in *Cibicides lobatulus* from the *Picea zone at Karmøy and at the top of the Eemian at Fjøsanger (also from the Picea* zone) strongly support the correlation of these two interglacial sequences.

Based on the assumption that the Avaldsnes and Fjøsangerian Interglacials represents the last interglacial, the Eemian, and that this ended about 120 000 years B.P., age-estimates were made for the two interstadials at Bø. The results indicate that the Torvastad Sand was deposited 78 000 ± 7 000 years B.P. and the Bø Sand 52 000 ± 12 000 years B.P. (Miller et al. in press.).

Summary of main results

Avaldsnes Interglacial: The fauna and flora assemblages observed in the Avaldsnes Sand show that this sand represents a warm interglacial. On the basis of fauna, flora and amino-acid ratios the Avaldsnes Interglacial was correlated with the Fjøsangerian, oxygen-isotope stage 5e and the Eemian. The sea level was most likely more than 15 m–20 m higher than today at Bø.

Torvastad Interglacial: The flora and fauna assemblages in the Torvastad Sand together with the age estimate on the basis of the amino-acid ratio suggest that the sand represents a cool Early Weichselian interstadial. The sea level was about 0–5 m higher than today at Bø.

Karmøy Stadial: The Karmøy diamicton with large striated boulders represents a true till. This till was deposited by an ice sheet which covered southwestern Norway and moved in southwesterly and westerly to northerly directions at Bø. The age estimates suggest that this happened in Early Weichselian time.

Bø Interstadial: Both the fauna, flora and the lithology of the Bø Sand suggest a cool to cold climate. The ice front was located near Bø during the early and late phases of the Bø Interstadial, which is most likely of early Middle Weichselian or late Early Weichselian age. The sea level was not much higher than 5 m above the present.

Haugesund Stadial: The clayey Haugesund diamicton is a clayey till deposited after 40 000–60 000 years B.P., most likely by the Late Weichselian glacier which covered the entire southwest coast of Norway.

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REFERENCES

Andersen, B. G., Nydal, R., Wangen, O. P. & Østmo, S. R. 1981: Weichselian before 15.000 years BP at Jæren – Karmøy in southwestern Norway. *Boreas 10*, 297–314.

Dons, C. 1936: Norges strandfauna XIII. Muslinger 1. *Det Kgl. Norske Vid. Selsk. Forhandl. Bd. 9, nr. 42.*

Feyling-Hanssen, R. W. 1960: *In:* Holtedahl, O. (ed.) Geology of Norway. *Norges geol. Unders. 208*, 384.

Feyling-Hanssen, R. W. 1955: Stratigraphy of the marin Late-Pleistocene of Billefjorden, Vestspitsbergen. *Nor. Polarinst. Skr. 107*, 226 p.

Jensen, A. S. & Spärck, R. 1934: Bløddyr II. Saltvandsmuslinger. *Danm. Fauna, 40*, 208 p.

Kellogg, T. B., Duplessy, J. C. & Shacleton, N. J. 1978: Planktonic foraminiferal and oxygen isotopic stratigraphy of Norwegian deep-sea cores. *Boreas 7*, 61–73.

Mangerud, J., Sønstegaard, E., Sejrup, H.-P. & Haldorsen, S., 1981: A contineous Eemian–Early Weichselian sequence containing pollen and marine fossils at Fjøsanger, western Norway. *Boreas 10*, 137–208.

Ockelman, W. K. 1958: The zoology of East Greenland. Marine Lamellibranchiata. *Med. Grønnland 122*, 256 p.

Miller, G. H., Sejrup, H.-P., Mangerud, J. & Andersen, B. G. (in press): Amino acid ratiois in Quaternary molluscs and foraminifera from western Norway: Correlation. geochronology and paleotemperature estimates. *Boreas.*

Ringen, E. 1964: Om drumliner og Skagerakmorenene på Karmøy. *Nor. geogr. Tidsskr. 19*, 205–228.

West, R. G. 1968: Pleistocene Geology and Biology. 377 pp, Longmans, Green & Co. Ltd.

Øyen, P. A. 1905: Nygaardsprofilet på Karmøen. *Christiania Vidensk.-Selsk. Forhandl. 8*, 1–15.

Diatom Stratigraphy Related to Late Weichselian Sea-level Changes in Sunnmøre, Western Norway

SVEN ERIK LIE, BJØRG STABELL & JAN MANGERUD

Lie, S. E., Stabell, B. & Mangerud, J. 1983: Diatom stratigraphy related to Late Weichselian sea-level changes in Sunnmøre, Western Norway. *Norges geol. Unders. 380,* 203–219.

In sediment cores from four small basins from 44 to 26 m above sea-level, the boundaries between marine, brackish and lacustrine sediments are precisely identified through their distinct diatom stratigraphy. In all basins the diatom successions demonstrate a development from marine to lacustrine environments. The basins, today being lakes or bogs, were isolated from the sea between 12.400 and 9.200 radiocarbon years B.P., at the end of the Late Weichselian. The emergence was fast during the Bølling and Allerød (12.000–11.000 yr BP), and slower during the Younger Dryas (11.000–10.000 yr BP). The slow emergence was probably a main reason for the formation of the morphological distinct Tounger Dryas shore-line.

Sven Erik Lie, Norsk Hydro, Petroleum Division, Boks 641, N-9401 Harstad, Norway

Bjørg Stabell, Universitetet i Oslo, Institutt for geologi, Boks 1047, Blindern, Oslo 3, Norway

Jan Mangerud, Universitetet i Bergen, Geologisk Institutt, avd. B, Allégt. 41, N-5000 Bergen, Norway

Introduction

Raised former beaches are frequent in Sunnmøre in Western Norway (Fig. 1). One such prominent level is the Younger Dryas shore-line, correlated with many cirque moraines (Reite 1967, Mangerud et al. 1979). However, it was earlier not known whether the better morphological development of this line was due to a transgression, a stillstand in relative sea-level changes, or faster erosional depositional processes during the Younger Dryas. The main object of our study was therefore to establish a relative sea-level curve based on stratigraphical investigations for the area, and thereby solve the problem.

During the last few years a Younger Dryas transgression has been demonstrated further south in Western Norway (Anundsen 1978, Anundsen & Fjeldskaar, in press, Krzywinski & Stabell, in press, Thomsen, 1982) where a Younger Dryas re-advance of the ice-sheet is known (Mangerud 1977, Sindre 1980). In Sunnmøre the front of the ice-sheet was situated much further inland (Sollid & Sørbel 1979) and only cirque glaciers were formed along the coast (Mangerud 1980). Sea-level curves from this area would therefore be of major significance for understanding the geological and geophysical processes governing sea-level changes along glaciated coasts. In this paper we will mainly present the results of diatom studies from the Ålesund area. The

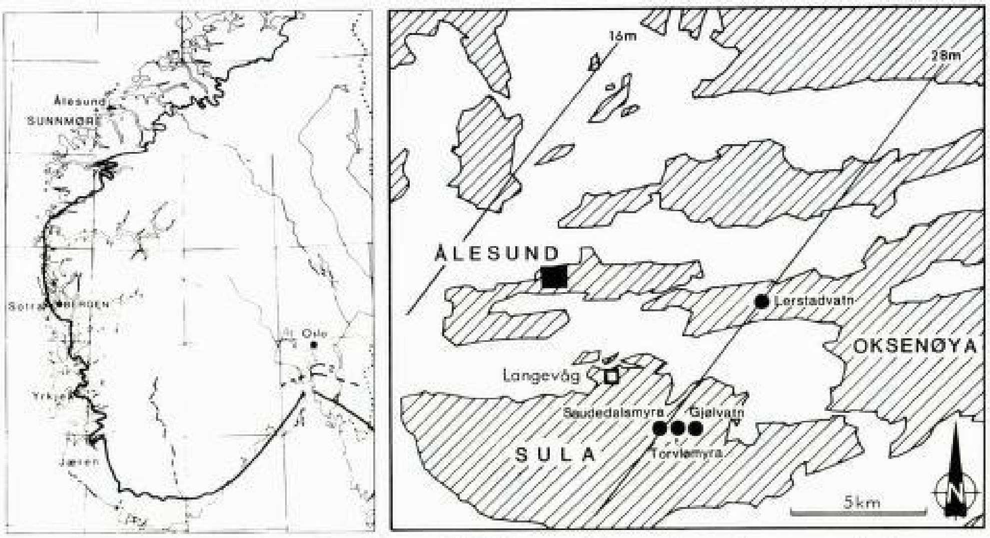

Fig. 1. To left a key map of southern Norway. The Younger Dryas end moraines are marked with heavy lines. To the right a map of the Ålesund area, land areas hatched. The Younger Dryas isobases (16 and 28 m) are indicated according to Reite (1967).

lithostratigraphy, pollen-stratigraphy, radiocarbon datings and a more extensive discussion of the sea-level curves and shore-lines will follow in subsequent papers.

Methods

For establishing relative sea-level curves we used the classical method of coring lakes at different altitudes (e.g. Hafsten 1960, Kjemperud 1981a) to determine the sequence of marine and lacustrine beds, and to date these changes. We used diatoms for interpretations of whether the sedimentary environments were saline, brackish or fresh.

This study includes four basins situated below the marine limit, which by means of terraces is determined to 45 m a.s.l. around Ålesund. Three localities lie less than one kilometre apart on the island Sula, south of the town of Ålesund. The fourth (Lerstadvatn) is situated on the same isobase (Reite 1967, Sollid & Kjenstad 1980), but 5 km further north (Fig. 1).

The chronostratigraphical classification follows Mangerud et al. (1974), where the chronozones were defined in radiocarbon years. For the identification of the boundaries we have obtained approximately 20 radiocarbon dates (Lie & Lømo 1981, Kristiansen pers. comm.) which will be published on a later occasion.

Lithostratigraphy

The sediments in the four investigated basins are so similar that a lithostratigraphical correlation can easily be made, and we will therefore give a common description. The total sequence is subdivided into five informal

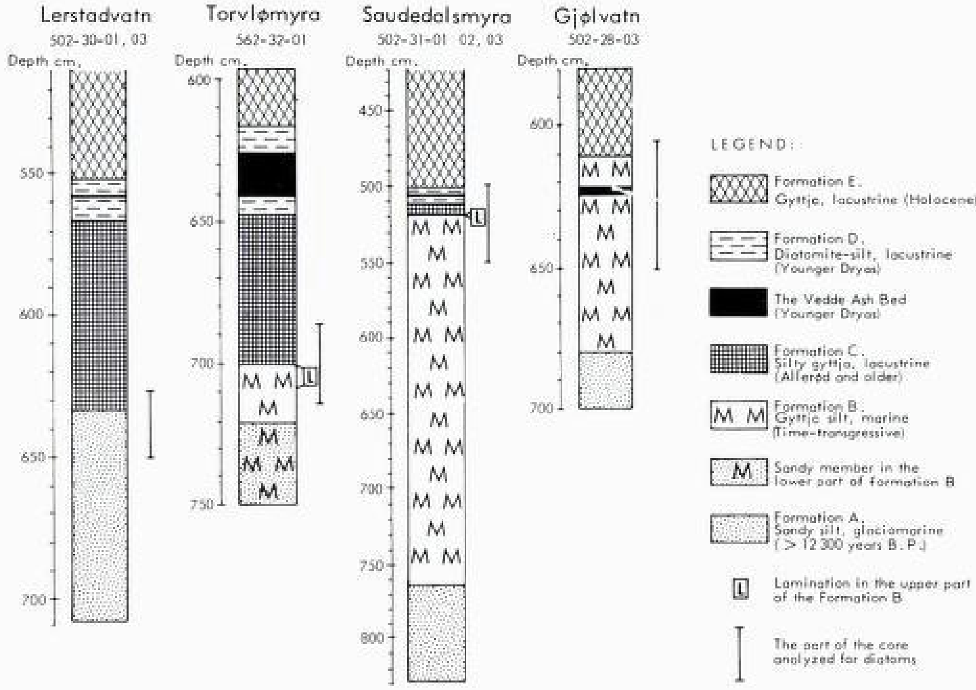

Fig. 2. The lithostratigraphy of the investigated cores. The numbers below the locality names are the core numbers. Note the smaller depth scale used for Saudedalsmyra compared to the three other localities.

formations (Fig. 2), lettered from A to E from the base. Only in Saudedalsmyra are all five formations present.

Formation A. Sandy silt. This formation consists mainly of grey silt, with beds of silty sand and clayey silt. The loss on ignition is <2%, and macrofossils have not been found. The formation is interpreted to be of glaciomarine origin, and older than 12.300 years B.P.

Formation B. Gyttja silt. Brownish grey to grey gyttja silt. The main difference from formation A is the higher organic content and the brownish colour. The formation is generally homogeneous, but at some sites well defined red, red-brown and green laminae occur in the upper (brackish) part. Fragments of marine shells are common, except in the upper part. The sediments are marine and, at the top, brackish.

Formation C. Silty gyttja. The main characteristic is its much higher organic content than the underlying and directly overlying formations. The colour varies between greyish, greenish and brownish, but brownish colours dominate. Beds with abundant plant remains occur. The lower boundary is normally a smooth transition, whereas the upper boundary is sharp. This formation is entirely lacustrine, and its age Allerød and partly older.

Formation D. Diatomite-silt. This formation consists of a pale green to yellowish-grey silt. It has a very high content of diatom frustules, thus the name diatomite-silt. The Vedde Ash Bed with an age of 10.600 ± 60 years

B.P. (Mangerud et al. in press.) is found near the midpoint of this formation in three of the basins. The sediments are lacustrine, and of Younger Dryas Age.

Formation E. Gyttja. This is a brown, fine detritus gyttja. It is lacustrine and of Holocene age.

Diatom analyses

The sediment samples for diatom analyses were prepared after a simplified version of the procedure described by Schrader (1973).

1) Each sample was boiled in 35% H_2O_2 in a small beaker.
2) A part of the solution was poured into a centrifuge glass, distilled water added and centrifuged for 2 min. at a speed of 1200 rev. per min. Water with suspended clay was decanted.
3) Water was again added and the rest of point 2 repeated 4 times.
4) A few drops of the residue were spread on a cover-glass, evaporated and embedded in Hyrax mounting medium.

The analyses were carried out using a Zeiss research microscope with the objectives 100/1.30 and 63/1.4 with phase contrast. The calculation basis for the relative frequencies was approximately 200 valves in each spectrum. *Fragilaria* spp. were counted separately and are not included in this sum. The results are presented in two types of diagrams. The construction of these diagrams is based on Florin (1946). One diagram shows the succession of the most dominant species. In the other diagram the taxa are grouped with regard to their preference for the salt content of the water, according to the halobion system of Kolbe (1927) modifiewd by Hustedt (1957). The polyhalobous taxa prefer salinities higher than 30‰, the mesohalobous between 30 and 0.2‰, and oligohalobous taxa can live in both brackish and fresh water. The oligohalobous halophilous show an optimum in slightly saline water, and the oligohalobous indifferent taxa prefer fresh water. The halophobous group includes exclusively freshwater species which have an upper tolerance boundary at 0.2‰ S. The diagrams include mainly the sediments around the transition zone from marine to lacustrine sedimentary environments (Fig. 2). Samples from other parts of the cores have been examined without quantitative counting of the diatoms, to look for major changes in the salinity of the environment. The marine, brackish, 'halophilous', and freshwater diatom zones are based on the change in the diatom composition (Kjemperud 1981b). More precisely they may be classified as paleo-environmental zones.

The transition from polyhalobous/mesohalobous to oligohalobous assemblages indicates the change in sedimentary environment from marine/ brackish to freshwater, often called the isolation contact, because at that time the lake was isolated from the sea. The isolation contact represents in the sediment sequence the time of emergence of the lake's threshold above high tide (Ingmar 1973) (see discussion in Stabell 1982). In this area the present tidal range is about 2 m. The constructed curve of shore-level displacement (Fig. 12) represents the high tide, and the contemporary mean sea-level can be estimated at about 1 m below.

Investigated basins

Lerstadvatn, 44 m a.s.l.

This site lies at the west end of Oksenøya about 7 km east of Ålesund (UTM coordinates 610 295). The basin is about 0.16 km², and it is subdivided into two sub-basins by a rock sill. The deeper basin is the modern lake, while the shallower basin is located below the bog near the western shore.

The analysed sediment core was taken from the bog surface and represents the sediments between 718 cm and 546 cm below the bog surface (Fig. 2). Eight samples between 650 cm to 627 cm in formations A and C were analysed for diatoms (Figs. 3 and 4). The four lowest spectra define the marine diatom zone. The dominating taxa are the polyhalobous types *Rhabdonema minutum, Pinnularia quadratarea, Cocconeis scutellum, Diploneis smithii*

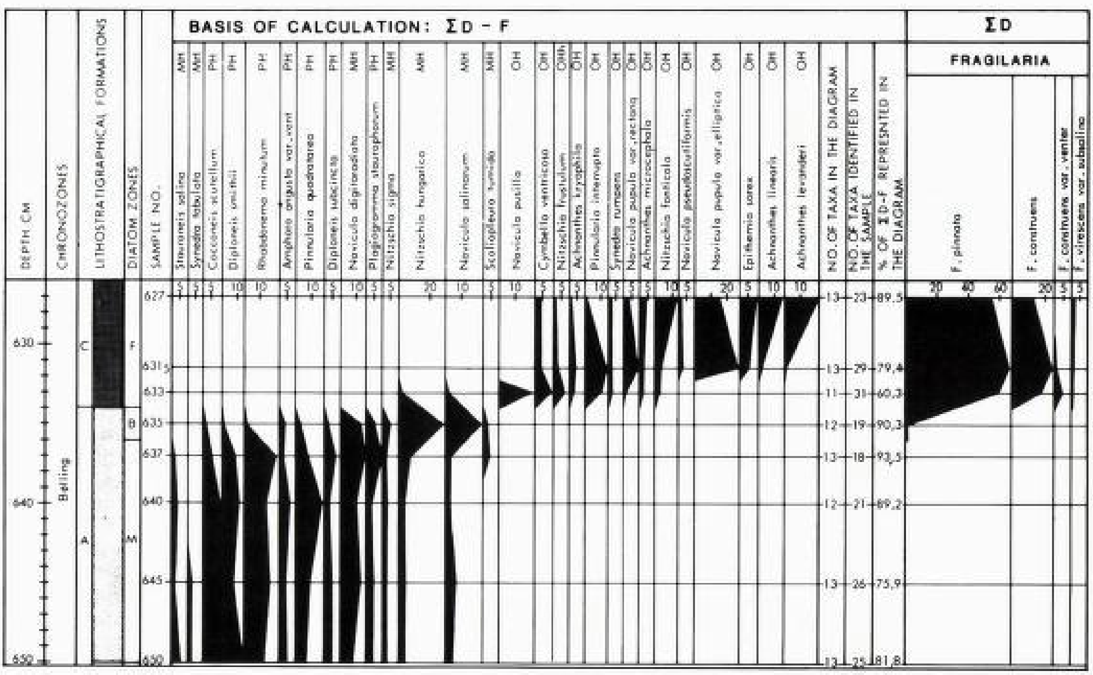

Fig. 3. Lerstadvatn 44 m a.s.l. Succession and distribution of the most dominant diatom taxa. Basis for calculation of the taxa other than *Fragilaria* are D−F, where D is total number of diatoms counted and F is number of *Fragilaria* spp. counted. Diatom zones: M = marine, B = brackish, H = 'halophilous' and F = freshwater. Capital letter after name of taxon denotes salinity group. PH = polyhalobous (marine), MH = mesohalobous (brackish), OHh = oligohalobous halophilous (brackish/fresh), OH = oligohalobous indifferent (mainly freshwater) and HF = halophobous (freshwater). The horizontal lines in the 'loss on ignition' columns indicate the level of ignited sample. Diatom sample no, is identical to depth.

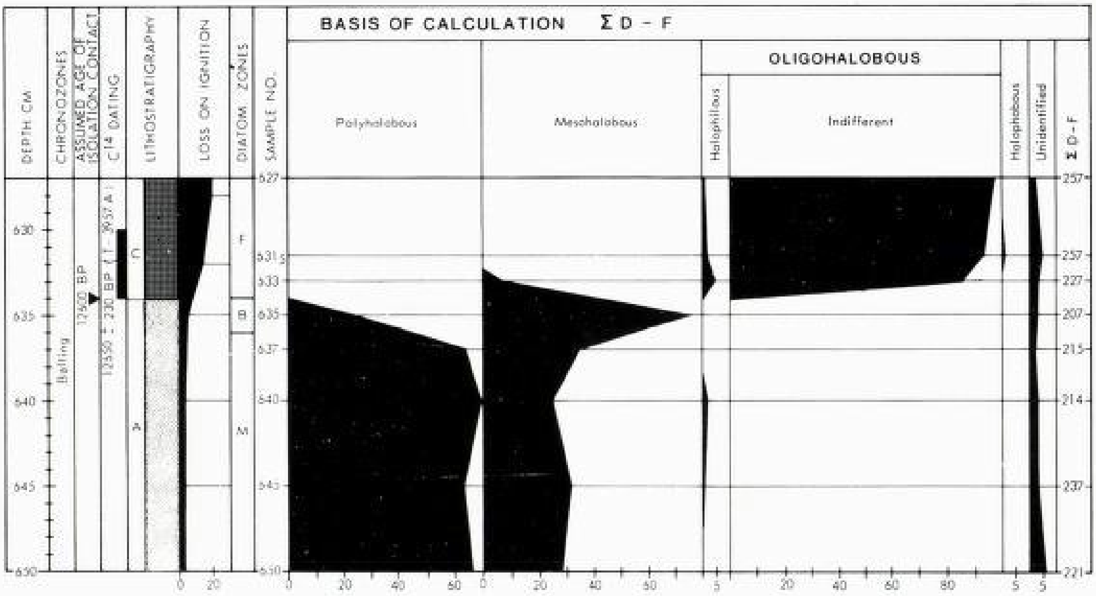

Fig. 4. Diatom diagram from Lerstadvatn with taxa grouped with regard to their preference for the salt content of the water. See text to fig. 3.

and the mesohalobous *Navicula digitoradiata*. The occurrence of *N. digitoradiata* could indicate a periodic lowering of the salinity in the fiord, possibly due to meltwater from the retreating ice-sheet.

At 635 cm there is a peak of the mesohalobous group to 75% occurrence, mainly as a result of the flourishing of *Navicula salinarum* and *Nitzschia hungarica*. This diatom assemblage defines the brackish diatom zone. The spectrum above (633 cm) is almost totally dominated by oligohalobous indifferent taxa. *Navicula pusilla* is described as oligohalobous indifferent (Simonsen 1962) but by Hendey (1964) observed as common in brackish water. It has a very limited stratigraphical distribution just above the isolation contact. The peak of *N. pusilla* and a 6% occurrence of *Nitzschia frustulum* may define a slightly saline environment. However, due to the rich flourishing of typical freshwater species such as *Cymbella ventricosa* and *Pinnularia interrupta* at the same level, we do not define this as a halophilous zone. The isolation contact is defined between 635 and 633 cm and coincides with the lithological transition from formation A to formation C at 634 cm. Sediments just above this boundary (634–630 cm) are C^{14}-dated to 12.650 ± 230 yr BP (T-3957 A). Many shell datings indicate the deglaciation of this area to about 12.400 yr BP (Mangerud et al. 1981 and unpublished) and we therefore assume that an age of 12.400 to 12.300 yr BP is most likely for the isolation contact in Lerstadvatn. This is only slightly outside one standard deviation of the obtained date.

Torvlømyra, 35 m a.s.l.

The bog Torvlømyra is at the north-east end of the island Sula (Fig. 1) (UTM 576 258). The main basin is about 100 m long and only 10 m wide. Continuous Late Weichselian stratigraphy is found only in the north-eastern part of the bog. The analysed sediment cores range from 750 to 560 cm below the bog surface (Fig. 2). The lower half of formation B is a sandy member, suggesting a high-energy sedimentary environment. This has been caused by a tidal current (E–W) when the sea-level was above the thresholds in both ends of the basin. The difference in elevation between the thresholds is only one metre.

The ten analysed diatom samples represent the beds between 715 and 687 cm in the formations B and C (Figs. 5 and 6). The two lowermost spectra show a marine diatom assemblage. *Rhabdonema minutum* is the dominating species, which besides *Diploneis subcinta* and *Cocconeis scutellum* give a high percentage of the polyhalobous group. At level 710 cm the polyhalobous is still dominating, but both the mesohalobous *Navicula digitoradiata* and *Hyalodiscus scoticus* show high frequencies and indicate a transition to brackish conditions. The mesohalobous group reaches a peak of about 55% at 705 and 703 cm. A flourishing of *Achnanthes hauckiana*, *Navicula salinarum* and *N. halophila* is followed by *Synedra tabulata*, *Achnanthes delicatula* and *Mastogloia elliptica*. Species of the 'Clypeus-flora' (Florin 1946), which are typical for brackish/freshwater-transition sediments around the Baltic, are not registered. However, both *N. halophila* and *M. elliptica* can be included in that group (Ingmar 1973). The two last named are common in Late- and post-Weichselian sediments in Sunnmøre, and their frequency peaks are very often in the

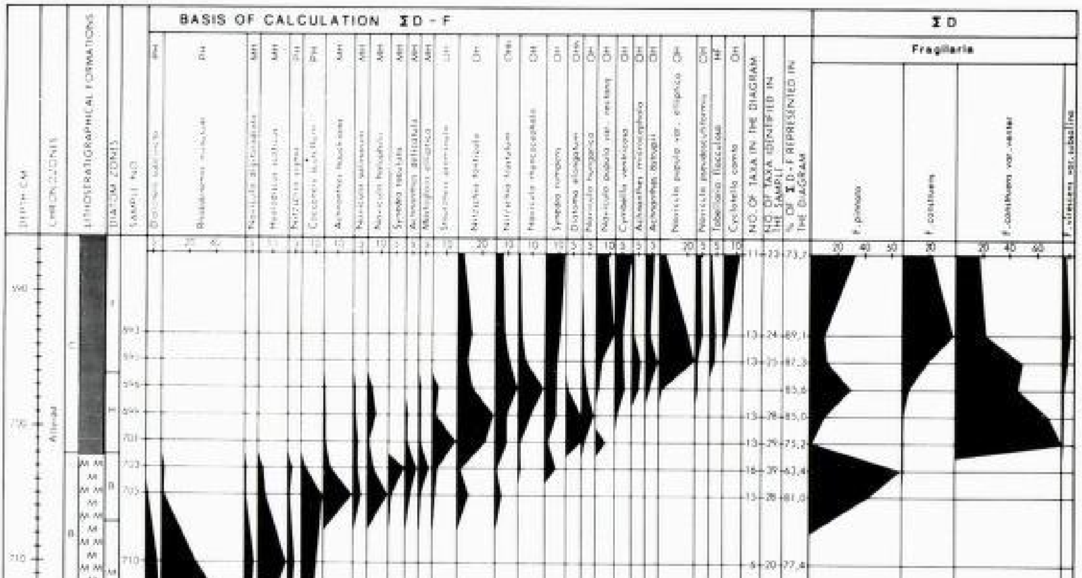

Fig. 5. Torvlømyra 35 m a.s.l. Succession and distribution of the most dominant diatom taxa. See text to fig. 3.

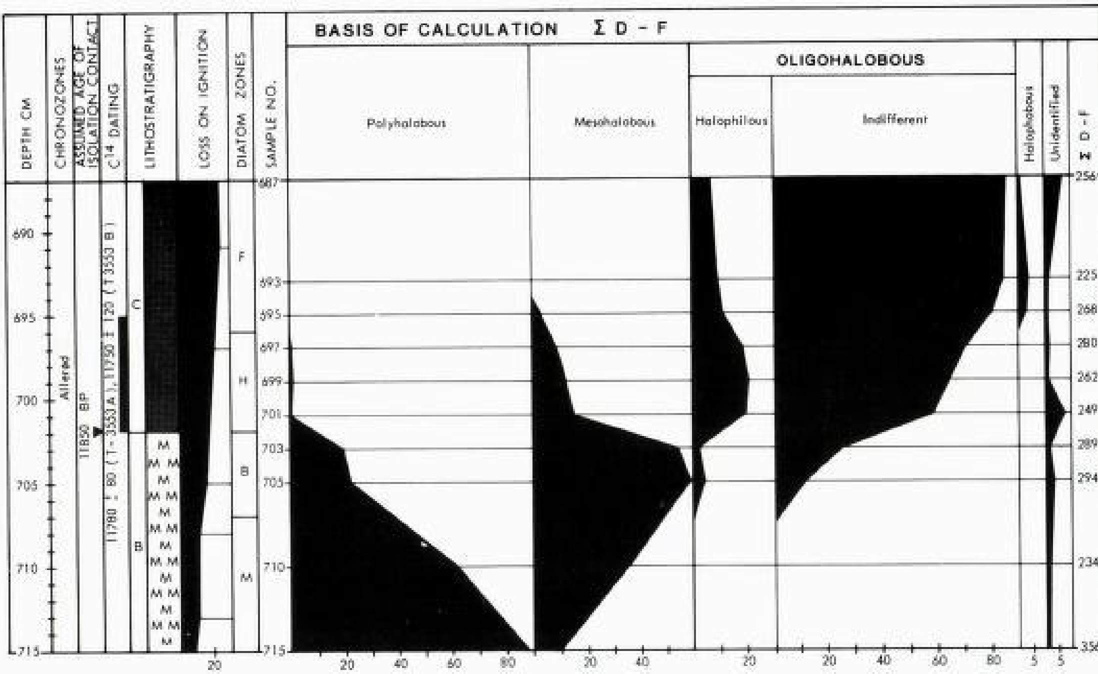

Fig. 6. Diagram from Torvlømyra with taxa grouped with regard to their preference for the salt content of the water. Se text to fig. 3.

sediments just beneath the isolation contact. A marked increase of the oligohalobous group at 701 cm indicates the upper limit of the brackish zone and thus defines the isolation contact.

The mesohalobous group still reaches an occurrence decreasing from 15 to 9 percent in the three lowest levels in the freshwater zone. This, together with a high frequency of the halophilous species *Nitzschia frustulum, Diatoma elongatum* and the occurrence of *Pinnularia globiceps* and *Rhopalodia gibberula* support the definition of a halophilous diatom zone. Only 1 to 2 percent of the polyhalobous group are registered through this zone, which indicate just a slight input of saline water. During the time of isolation from the sea the lake was surrounded by fiords. A sporadic supply of sea water by wind and waves in rough weather may have influenced the salinity in the basin for some time after the threshold rose above the littoral zone. A stratigraphically well defined halophilous diatom zone is favoured by a slow land uplift, which also may explain that this zone is more than 5 cm thick in Torvlømyra. The upper limit of the halophilous zone is defined by the first appearance of the halophobous taxon *Tabellaria flocculosa.* A sample of the silty gyttja between 701 and 695 cm was radiocarbon-dated to 11.780 ± 80 yr BP (T-3553 A, NaOH soluble) and 11.750 ± 120 yr BP, (T-3553 B, insoluble in NaOH). The radiocarbon dates give the ages of approximately the centre of the sample. Based on a probable rate of sedimentation, the age of the isolation contact (702 cm) can be estimated to 11.850 yr BP.

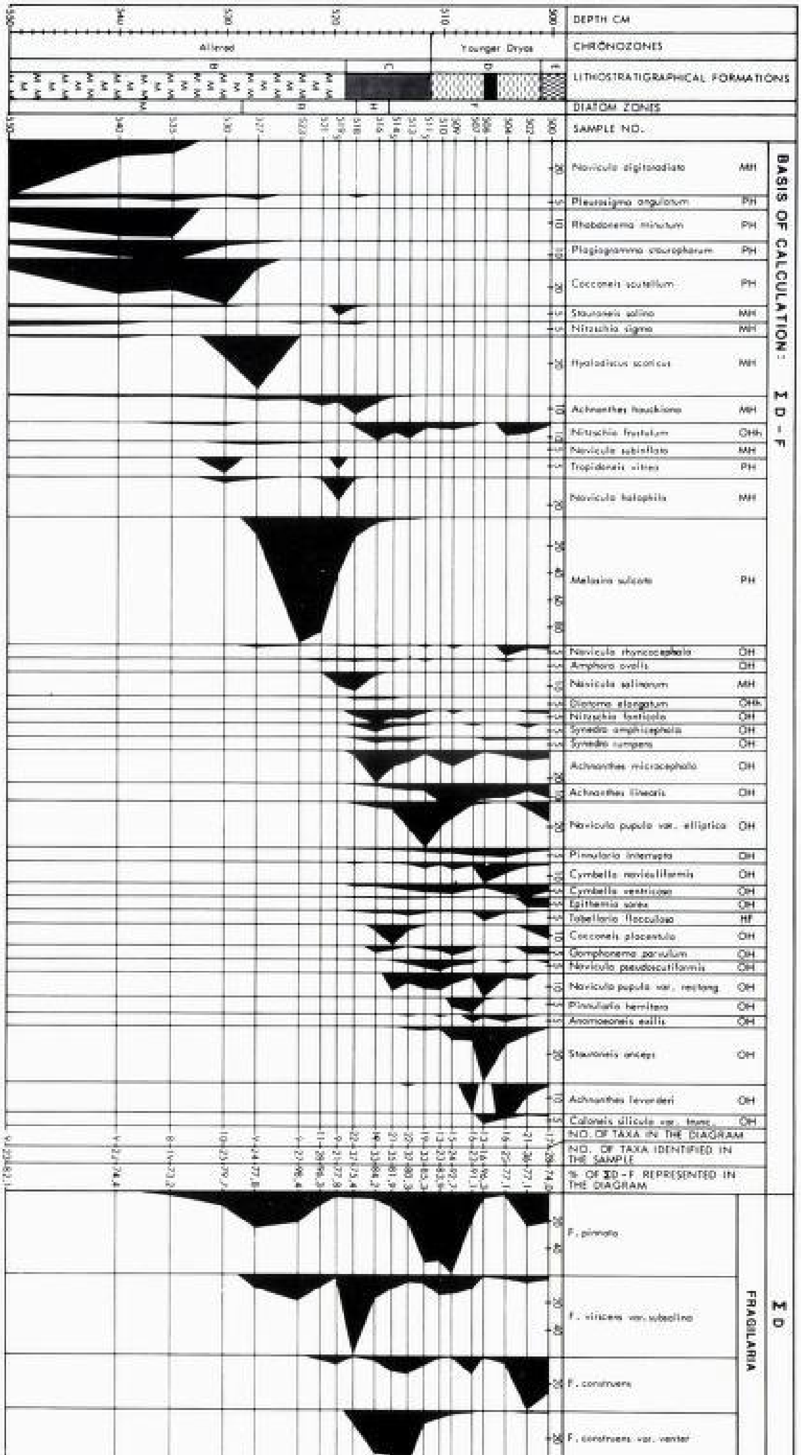

Fig. 7. Saudedalsmyra 30.2 m a.s.l. Succession and distribution of the most dominant diatom taxa. See text to fig. 3.

Saudedalsmyra, 30.2 m a.s.l.
Saudedalsmyra is a 140 m long and 65 m wide bog situated about 50 m downstream of Torvlømyra (UTM 575 253). The cores were taken in the deepest part of the basin near the centre of the bog, and represent the sediment between 828 to 380 cm beneath the bog surface (Fig. 2).

The diatom analyses represent the beds between 550 to 500 cm in formations B, C, D and E (figs. 7 and 8). The marine diatom zone include the four lowermost samples. As in Lerstadvatn the mesohalobous *Navicula digitoradiata* shows high frequencies among the typical marine species such as *Rhabdonema minutum* and *Plagiogramma staurophorum*. The brackish zone is defined by an increase of mesohalobos taxa, both in diversity and total occurrence. But even though the mesohalobous group is dominant in number of taxa, the polyhalobous *Melosira sulcata* (syn. *Paralia sulcata*) is the most common species in this zone. At the levels 523 and 521 cm it almost totally dominates the assemblage (more than 90%). *M. sulcata* is a marine species tolerating salinities as low as 5–3‰ (Simonsen 1962). It is a bottom dwelling form, but sometimes found among the marine plankton, particularly after winter gales (Hendey 1964). A 5% occurrence in a freshwater zone, as recorded by Stabell (1980), may be explained as allochthonous. However, the great dominans of *M. sulcata* in Saudedalsmyra must represent the autochthonous flora.

The large number of *Melosira sulcata* gives a peak of the polyhalobous group just before the isolation from the sea (Fig. 8), and causes low relative frequencies of the mesohalobous taxa (Fig. 7). However, the mesohalobous group shows a succession similar to the brackish zone at other localities, and the zone boundaries are therefore defined disregarding the *Melosira* peak. The isolation contact is in all cases defined to 518 cm depth.

The halophilous zone is only represented by sample 516 cm. At the time of isolation the basin was connected to the sea by a 300 m long and 10 to 30 m wide channel. Later, when it became a lake, it was therefore well protected against the influence of sea-water. This can explain why the halophilous diatom zone is less developed on this site compared with Torvlømyra. A 3 cm thick sample from the silty gyttja above is radiocarbon-dated to 11.130±140 yr BP (T-3958 A), and from that the date of the isolation estimated to have been approximately 11.150 yr BP.

The sea level of the transition Younger Dryas/Preboreal time is registered only 3 m below the threshold of Saudedalsmyra (see Gjølvatn below). Saudedalsmyra is therefore very sensitive for possible transgressions during the Younger Dryas. A transgression of that age is known further south in Western Norway (Anundsen 1978, Anundsen & Fjeldskaar, in press, Krzywinski & Stabell, in press). The beds representing the Late Allerød and Younger Dryas where therefore carefully analysed in order to discover possible transgression sequences within this period. However, no marine influence was detected in the sediments above the isolation contact.

The diatom succession in the freshwater diatom zone shows a noticeable change at the lower boundary of formation D. The number of taxa is redused more than 30%. A decrease in the occurrence of *Navicula pupula var. elliptica*,

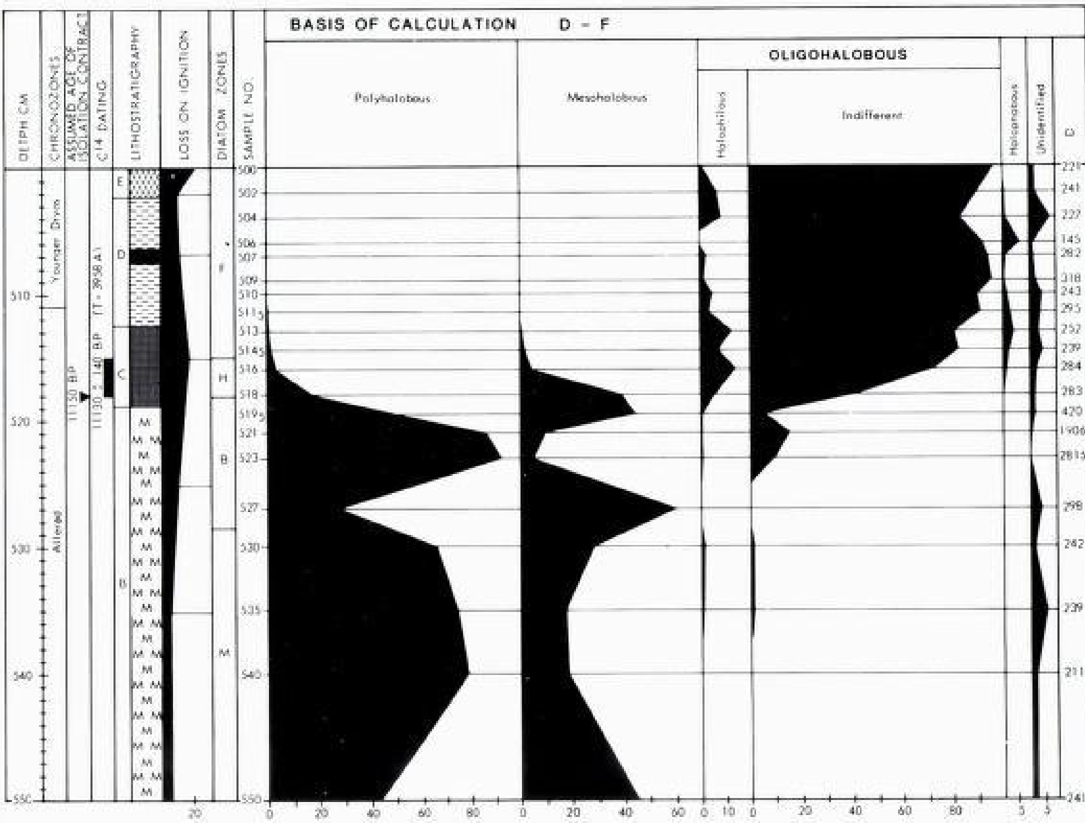

Fig. 8. Diagram from Saudedalsmyra with taxa grouped with regard to their preference for the salt content of the water. See text to fig. 3.

Achnanthes microcephala, Cocconeis placentula and *Epithemia zorex* is registered simultanously with an increase of *Achnanthes linearis, Pinnularia interrupta, P. hemiptera, Anomoeoneis exilis, Stauroneis anceps* and *Achnanthes levanderi.* An almost reverse change occurs near the upper boundary of formation D. Many studies (summarized in Stabell 1982) show the lacustrine diatoms to be unsuitable as climatic indicators. The diatom succession through formation D is probably a secondary response to the climatic deterioration during the Younger Dryas, and reflects a change of the nutrient condition in the paleo-lake. A decrease in production of other water plants, and a change of the vegetation around the basin may explain the event.

The Vedde Ash Bed was deposited as a several cm thick layer at the bottom of the paleo-lake (Mangerud et al. in press.), and certainly influenced the organic production. This is also seen in the sample (506) from the ash layer. Only 16 taxa are registered, compared to 23 taxa in the sample below. *Stauroneis anceps* is the dominant species, and the curves of both *Navicula pupula* var. *rectangularis* and *Cymbella naviculoides* show a peak at this level. *Achnanthes linearis* which dominates the samples just below and above is not observed et all.

Gjølvatn, 23.3 m a.s.l.

The lake Gjølvatn is situated about 700 m east of Saudedalsmyra (UTM 582 252). The total basin area is 0.06 km². The present lake constitutes less than half of the basin. The core is from the west end of the lake and represents the sediments between 700 to 540 cm below the water surface (Fig. 2).

The mesohalobous *Hyalodiscus scoticus* dominates the brackish zone. A rela- between 660 and 606 cm in formations B and E (Figs. 9 and 10). The marine diatom zone is dominated by *Trachyneis aspera* and *Diploneis subcincta.*

The mesohalobous *Hyalodiscus scoticus* dominates the brackish zone. A relative high occurrence of the polyhalobous *Cocconeis scutellum* and even a frequency of nearly 5 percent *Rhabodonema minutum* may indicate open fiord conditions and good circulation of marine coastal water. However, between 5 and 10 percent of the oligohalobous group are recorded throughouth this zone, and a flouorishing of the mesohalobous *Navicula salinarum, N. halophila* and *Nitzschia hungarica* is typical in the brackish zone also at the other localities.

Unique for the marine and brackish flora in Gjølvatn is the representation of a small *Diatomella* sp. (Fig. 11). It is recorded in four samples between 621.5 and 614 cm with frequencies of 1 to 3 percent. Marine taxa of *Diatomella* sp. are described by Voigt (1957), and the description of *Diatomella salina* fits our species well. The form has later been recorded at more sites in Sunnmøre, so far only in Younger Dryas and Perboreal marine deposits.

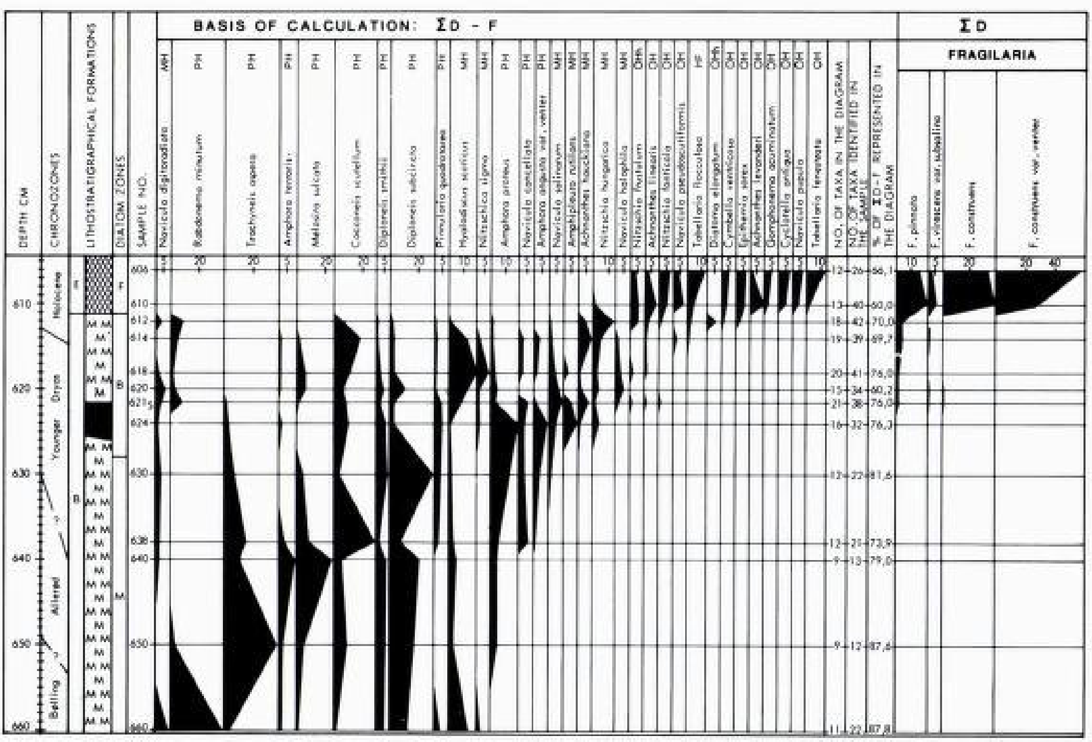

Fig. 9. Gjølvatn 26.3 m a.s.l. Succession and distribution of the most dominant diatom taxa. See text to fig. 3. Sample 624 cm represents the entire ash bed. The transition between diatom zone M and B is therefore placed between the last marine sample and the lower limit of the ash layer.

An abrupt transition to an assemblage almost totally dominated by lacustrine taxa can be correlated with the marked lithological boundary between formation B and formation E. A rapid regression in the relative sea-level may explain the lack of a halophilous zone.

Radiocarbon datings of the gyttja above the isolation contact gave the ages 9760±110 yr BP (T-3552 B, NaOH insoluble) and 10.000±140 yr BP (T-5352 A, NaOH soluble) The isolation is estimated to 9900 yr BP.

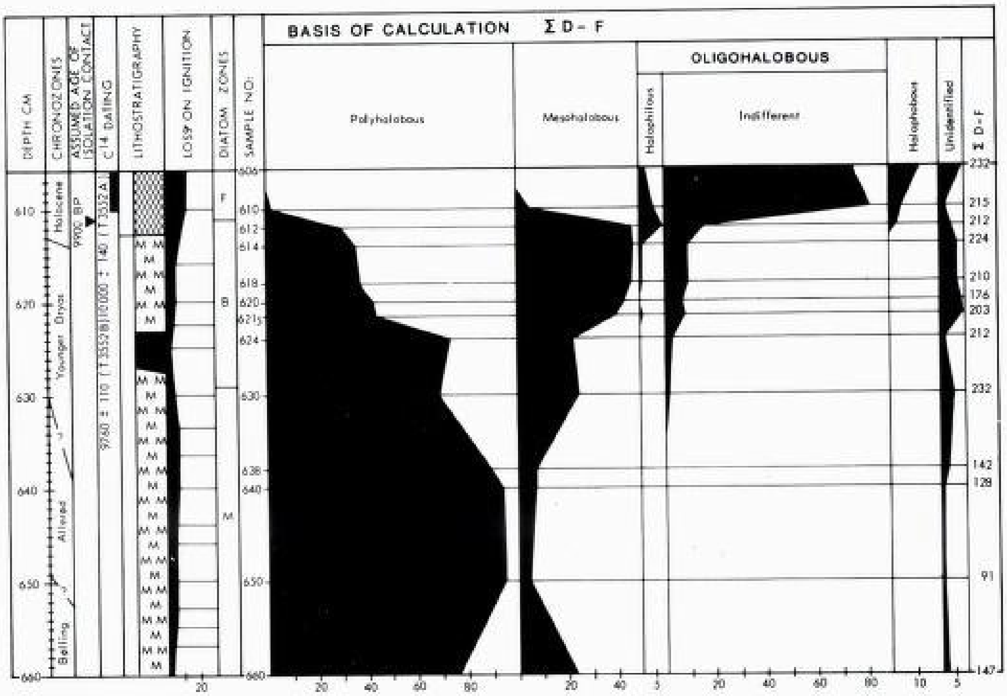

Fig. 10. Diagram from Gjølvatn with taxa grouped with regard to their preference for the salt content of the water. See text to fig. 3.

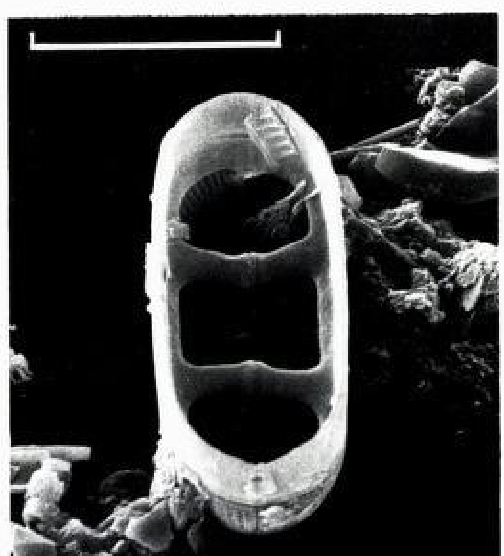

Fig. 11. Diatomella? salina Voigt. Note the narrow septae, distinguishing it from the freshwater species *D. balfouriana* Grev. Line is 10 μm.

The Fragilaria succession

The isolation contact has been defined without regard to *Fragilaria* spp. Even though they occur in great quantities in the marine/lacustrine transition zone and often dominate the lacustrine floras, it is difficult to point out a general succession for the *Fragilaria* species. *F. pinnata* and *F. virescens* var. *subsalina* show high frequencies both in marine/brackish and lacustrine environments in Saudedalsmyra while they occur only in the freshwater zone at Lerstadvatn. For the species *F. construens* var. *venter* a peak is registered just above the isolation in Lerstadvatn, Torvlømyra and Saudedalsmyra and it may be used as an indicator for the first lacustrine sedimentation. Also in Gjølsvatn this taxon increases rapidly above the isolation contact,but here it continues to increase upwards. Both because of their uncertain salt tolerance boundaries and their total dominance at some levels, *Fragilaria* spp. are kept out of the calculation basis for the diagrams. In the dissolved diagrams, the most commonly occurring *Fragilaria* taxa are presented.

Conclusions

The boundaries between marine, brackish and lacustrine sediments are precisely identified by means of the diatom stratigraphy. At three of the localities the boundary between the brackish and freshwater diatom zones coincides with lithological changes. In Saudedalsmyra the lithostratigraphical boundary between formations B and C is slightly below the isolation contact. Before the diatom analyses were carried out, there were in all basins other alternative interpretations for the position of the boundary between marine and lacustrine sediments.

In some of the cores, and in many cores from Sunnmøre not described in this paper, the brackish sediment is an algal gyttja, distinctly laminated with reddish, greenish and blackish colours. This has earlier been described from the area around Bergen (Kaland et al. in press), and appears to be an important criterion for the marine-lacustrine transition in the type of sedimentary environments used for this study. However, the exact position of the isolation contact within a laminated blackish sequence is still to be determined by diatom analysis.

Using the age obtained for the isolation contact to define the time the sea-level (high-tide) was at the level of the threshold of each of the lakes, we obtain a relatively sea-level curve as shown in Fig. 12. The ages are partly based on palynological correlations by Lømo (in Lie & Lømo 1981) and Kristiansen (pers. comm.) which will be published later.

The sea-level curve shows an early, rapid regressive shore displacement following the deglaciation at approximately 12.300–12.400 years B.P. The speed of shore displacement slowed down considerably during the Younger Dryas. This slow relative shore displacement change is probably the main reason for the morphologically well developed Younger Dryas shoreline.

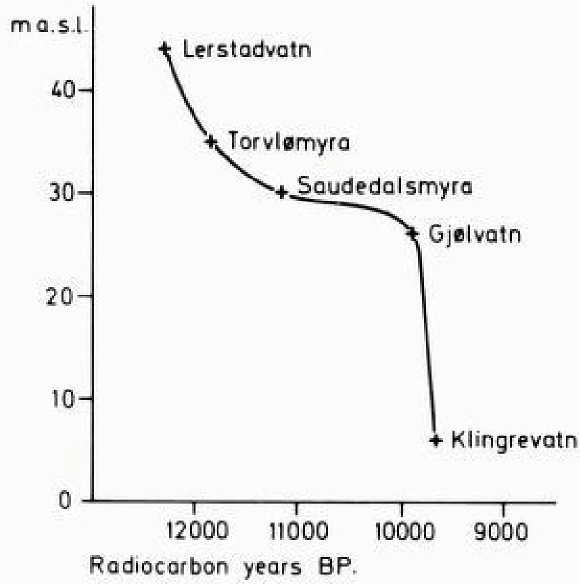

Fig. 12. Shore displacement curve for the Ålesund area. Note that the bend at about 10 000 yr BP has not been precisely dated. Gjølvatn is believed to have been isolated during a period of rapid regression. The course of the curve after the isolation of Gjølvatn is based on unpublished material collected by Kristiansen & Mangerud. The regression minimum (approximately 9000 yr BP) is situated below present sea-level (Hafsten 1979).

However, in many places, e.g. at the front of the numerous glaciers ending in or near the sea, the depositional processes certainly worked faster than during the Allerød.

We have cored a large number of lakes, both further inland (where basins at the critical altitudes are few) and towards the coast. The preliminary results indicate that the 11.000 year shoreline is slightly steeper than the 10.000 year line. Thus the shore displacement during Younger Dryas was faster further inland than around Ålesund. The lines probably crossed west of Ålesund, so at the outermost coast there was a transgression during the Younger Dryas.

The sea-level history is thus very different from Yrkje (Anundsen 1978, Anundsen & Fjeldskaar, in press) and Sotra (Krzywinski & Stabell, in press). This will not be discussed further here, as it will be treated in detail in a forthcoming article on the shore displacement, but a possible cause is the difference in the behaviour of the front of the Scandinavian ice-sheet (Mangerud 1980). In the Yrkje–Sotra area there was a major Younger Dryas re-advance contributing to the transgression (Fjeldskaar & Kanestrøm 1980, Fjeldskaar 1981). Such a transgression is lacking at Ålesund where a re-advance is not known.

Acknowledgements – This paper resulted from a work project (led by Mangerud) where also Inger-Lise Kristiansen and Leif Lømo were actively participating in coring, lithostratigraphical work, pollen-analysis, correlation, etc. The conclusions are therefore partly based on their unpublished work. Eivind Sønstegaard and Hans Petter Sejrup also participated in collection of some of the cores. All radiocarbon dates were performed at the Trondheim laboratory, under the supervision of Reidar Nydal and Steinar Gulliksen.

Urve Miller in consultation with Maj-Britt Florin, Michael Talbot and Jørn Thiede read through the manuscript critically. Michael Talbot corrected the English language.

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REFERENCES

Anundsen, K. 1978: Marine transgression in Younger Dryas in Norway. *Boreas 7*, 49–60.

Anundsen, K. & Fjeldskaar, W. in press: Observed and theoretical Late Weichselian shorelevel changes related to glacier oscillations at Yrkje, southwest Norway.

Fjeldskaar, W. 1981: Late-glacial movements of sea level and crust in Fennoscandia. Dr. sient. thesis 1981, Seismological Observatory, University of Bergen, Norway.

Fjeldskaar, W. & Kanestrøm, R. 1980: Younger Dryas Geoid-deformation caused by deglaciation in Fennoscandia. In: Mørner, N. A. (ed.): Earth rheology, isostasy and eustasy. John Wiley & Sons, 569–574.

Florin, M.-B. 1946: Clypeusfloraen i postglacial fornsjölagerföljder i östra Mellansverige. Geol. Fören. Stockh. Förh. *68*, 429–458.

Hafsten, U. 1960: Pollen-analytic investigation in South Norway. In: Holtedahl, O. (ed.): Geology of Norway. *Norges geol. Unders. 208*, 434–462.

Hafsten, U. 1979: Late and Post-Weichselian shore level changes in South Norway. In: E. Oele, R. T. E. Shüttenhelm & A. J. Wiggers (editors): The Quaternary History of the North Sea. *Acta Univ. Ups. Symp. Univ. Ups. Annum Quingentesium Celebrantis 2*, 45–59.

Hendey, N. I. 1964: An Introductory Account of the Smaller Algae of British Coastal Waters. Part V: Bacillariophyceae (Diatoms). Ministry of Agriculture, Fisheries and Food, Fishery Investigations, Ser. *4*, 317 pp.

Hustvedt, F. 1957: Die Diatomeen des Fluss-Systems der Weser im Gebiet der Hansestadt Bremen. Abh. Naturw. Ver. Bremen *34*, 181–440.

Ingmar, T. 1973: Sjöinsnörningar från aktualgeologiska synspunkter. En översikt. University of Lund, Department of Quaternary Geology, Report 3, 48–90.

Kaland, P. E., Krzywinski, K. & Stabell, B. in press: Radiocarbon Dating of Transitions between Marine and Lacustrine Sediments and their Relation to the development of Lakes. *Boreas* 1983.

Kjemperud, A. 1980a: A shoreline displacement investigation from Frosta in Trondheimsfjorden, Nord-Trøndelag, Norway. *Norsk geol. Tidsskr. 61*, 1–15.

Kjemperud, A. 1980b: Diatom changes in sediments of basins possessing marine/lacustrine transitions in Frosta, Nord-Trøndelag, Norway. *Boreas 10*, 27–38.

Kolbe, R. W. 1927: Zur Ökologie, Morphologie und Systematik der Brackwasser–Diatomeen. Die Kieselalgen des Sperenberger Salzgebietes. Pflanzenforsh. 7, 146 pp.

Krzywinski, K. & Stabell, B. in press: Late Weichselian sea level changes at Sotra, Hordaland, western Norway. *Boreas* 1983.

Lie, S. E. & Lømo, L. 1981: En lito- og biostratigrafisk undersøkelse av marine og limniske sedimenter i Ålesund-området. Cand. real. thesis, Department of Geology, University of Bergen, Norway.

Mangerud, J. 1980: Ice-front variations of different parts of the Scandinavian Ice Sheet, 13 000–10 000 Years BP. In: Lowe, J. J., Gray, J. M. and Robinson, J. E. (eds.): Studies in the Late-Glacial of NW Europe, 23–30. Pergamon Press.

Mangerud, J., Andersen, S. T., Berglund, B. E. & Donner, J. J. 1974: Quaternary stratigraphy of Norden, a proposal for terminology and classification. *Boreas 3*, 109–128.

Mangerud, J., Larsen, E., Longva, O. & Sønstegaard, E. 1979: Glacial history of western Norway 15,000–10,000 BP. *Boreas 8*, 179–187.

Mangerud, J., Gulliksen, S., Larsen, E., Longva, O., Miller, G., Sejrup, H.-P. & Sønstegaard, E. 1981: A Middle Weichselian ice-free period in Western Norway: The Ålesund Interstadial. *Boreas 10*, 447–462.

Mangerud, J., Lie, S. E., Furnes, H., Kristiansen, I.-L. & Lømo, L. in press: A Younger Dryas ash bed in Western Norway and its possible correlations with the tephra in cores from the Norwegian Sea and the North Atlantic. *Quaternary Research.*

Reite, A. J. 1967: Lokalglaciasjon på Sunnmøre. *Norges geol. nders. 247*, 262–287.

Schrader, H. J. 1973: Proposals for a standardized method of cleaning diatom-bearing deep-sea and land-exposed marine sediments. *Nova Hedwigia Beih. 45*, 403–409.

Simonsen, R. 1962: Untersuchungen zur Systematik und Ökologie der Bodendiatoméen der westlichen Ostsee. *Int. Rev. Ges. Hydrobiol. Syst. Beih. 1*, 144 pp.

Sindre, E. 1980: Late Weichselian ice-front oscillations in the Hardanger–Sunnhordland District, West Norway. Continental Shelf Institute. Publication no. 102, March 1980. 16 pp.

Sollid, J. L. & Sørbel, L. 1979: Deglaciation of western Central Norway. *Boreas 8*, 233–239.

Sollid, J. L. & Kjenstad, K. 1980: Hovedflaten (Yngre Dryas' Havnivå) som basis for kvartær kronologi i Midt-Norge. Et metodeforsøk. *Norsk geogr. Tidsskr. 34,* 93–96.

Stabell, B. 1980: Holocene shorelevel displacement in Telemark, southern Norway. *Norsk geol. Tidsskr. 60,* 71–80.

Stabell, B. 1982: Changes in diatom floras in Late Quaternary western and southern Norwegian marine and freshwater sediments: Response to basin isolation from the sea. Dr. scient. thesis 1982, Department of Geology, University of Oslo, Norway.

Thomsen, H. 1982: Late Weichselian shore-level displacement on Nord-Jæren, south-west Norway. *Geol. Fören. Stockh. Förh. 103,* (for 1981), 447–468.

Voigt, M. 1957: Note sur le genre «Diatomella». *Revue algologique 3,* 68–71.

Chemistry and Flow Patterns in some Groundwaters of Southeastern Norway

JENS-OLAF ENGLUND

Englund, J.-O. 1983: Chemistry and flow patterns in some groundwaters of southeastern Norway. *Norges geol. Unders. 380*, 221–234.

Groundwaters in Quaternary deposits and in bedrock of southeastern Norway down to 120 m below surface range in chemical composition from very dilute waters to brackish waters. The total contents of dissolved solids ($Ca^{2+} + Mg^{2+} + Na^{+} + K^{+} + HCO_3^{-} + SO_4^{2-} + Cl^{-}$) are usually lower in recharge areas than in deep circulating groundwaters and in springs.

Chemical differences are explained by two main modes of groundwater flow patterns: (a) flow through shallow aquifers with active flushing, and ending up in springs; (b) long and deep flow, from Quaternary deposits through deep circulating groundwaters in bedrock, and ending up in springs and rivers. The highest pH and concentration values along any one flow path usually occur in the spring waters.

Acidification due to acid precipitation does not seem to have occurred because of the high buffer capacity of the deep groundwaters investigated, deeper than 15–25 m below surface.

Jens-Olaf Englund, Institutt for geologi, Norges Landbrukshøgskole, Box 21, N-1432 ÅS-NLH, Norway

Introduction

The term groundwater is usually reserved to denote the water present beneath the surface of the ground in the zone of saturation. In Norway such water occurs in Quaternary deposits and in fractures in bedrock (e.g. Skjeseth 1953, 1956, Bryn 1961, Englund 1966). The purpose of this paper is to demonstrate that the groundwaters of southeastern Norway, down to 100–120 m below the surface, vary widely in chemical composition; from very dilute waters to brackish waters. This is illustrated by data (Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}, HCO_3^{-}, SO_4^{2-}, Cl^{-} and pH) collected during the years 1975–1981 from the following areas (Fig. 1): Åstadalen, Lillehammer – Brøttum, Stensengbekken watershed at Mjøsa, Ås, Moss and Jeløy. The last three areas are situated below the Late Glacial sea-level, while the other areas are above this level. The mean annual precipitation is highest at Moss, Jeløy, Ås and in Åstadalen with 800–1000 mm, and lowest in the Stensengbekken watershed with 600–700 mm (Aune 1981).

Since seasonal variations are quite common in groundwater chemistry, especially in shallow aquifers, and as only large-scale chemical changes are discussed here, the data used are annual average results. 934 water samples from 113 wells and springs have been analysed (Table 1); these were sampled during winter, spring, summer and autumn. Additional information from 45

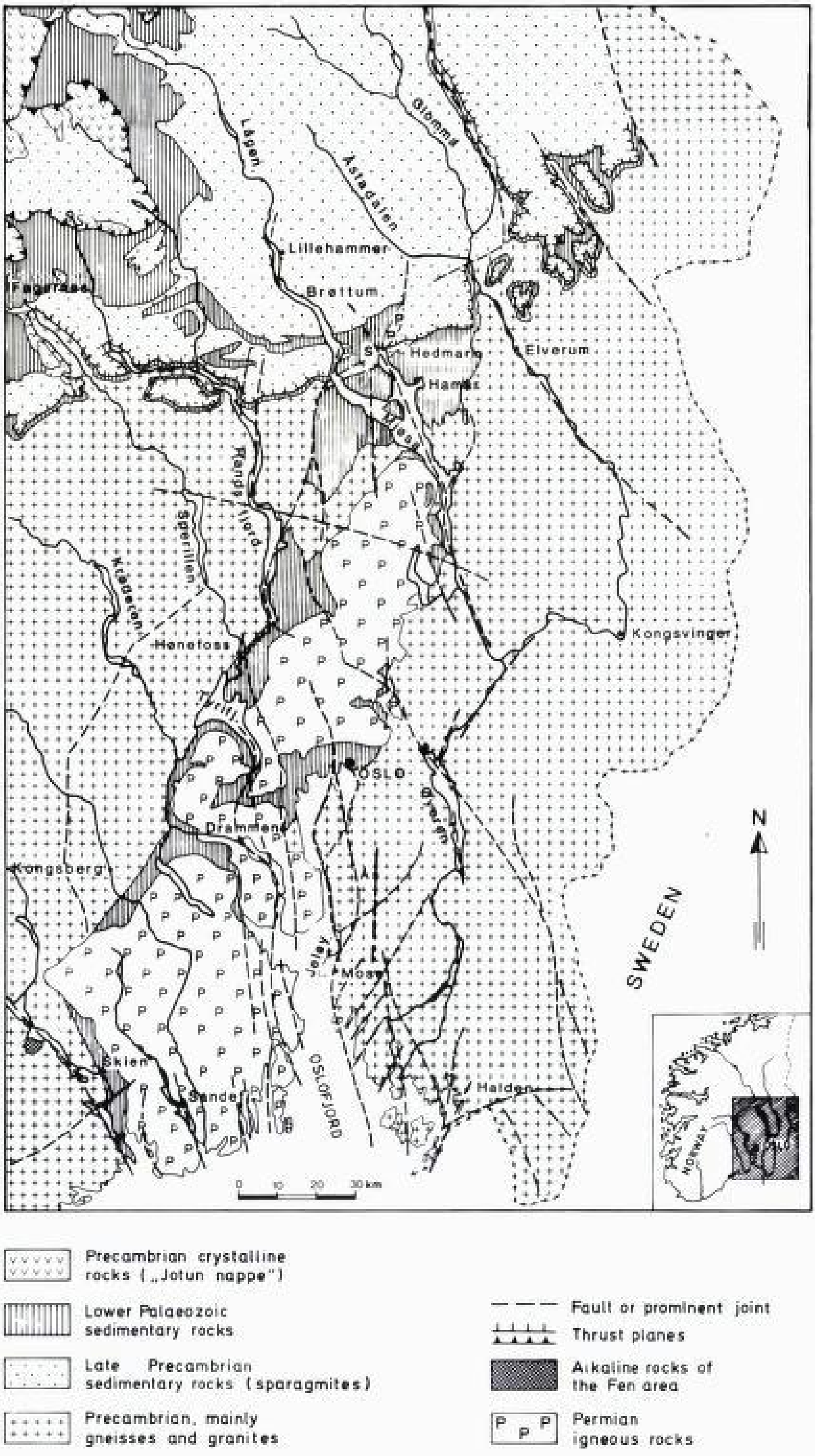

Fig. 1. Location map of areas in which the groundwater investigations were conducted: Jeløy, Moss, Ås, S = Stensengbekken, Brøttum–Lillehammer, Åstadalen. Geology mainly after Holtedahl & Dons (1960) and Ramberg & Larsen (1977).

water samples taken in 20 wells and springs has been used in working out the figures 6 and 7. Alle analyses were carried out by conventional methods on unfiltrated samples. Parts of the analytical data have earlier been published by Englund & Myhrstad (1980) and Englund & Meyer (1980).

General considerations

Nearly all the groundwaters of Scandinavia originate from rain or melting snow that infiltrates through soil or directly into rock fractures. The quality of groundwater at a given point in an aquifer thus depends on the integrated effects of a number of factors, such as: 1) the chemistry of the input water; 2) the pathway of water through the aquifer; 3) subsurface soil and rock types encountered along the pathway; 4) the residence time or flow velocity variations along the pathway; 5) hydrodynamic dispersion; 6) rates and threshold values of physical, chemical and biological processes. This paper is concerned mainly with the effects of the above factores 2, 3 and 4. The chemistry of the input water, however, varies between the investigated areas, due to variations in atmospheric deposition, agricultural activities, domestic sources, and leaching processes in the unsaturated zone.

Geochemical data from groundwaters in Scandinavia have been studied using theoretically calculated stability diagrams, of the types given by e.g. Helgeson et al. (1973); from crystalline Precambrian rocks in Sweden (Jacks 1973) and southeastern Norway (Englund & Myhrstad 1980). These studies indicate that alteration of feldspars, chlorites, micas and also calcite is a widespread process in groundwaters down to 120 m from ground surface. The cations released into the water are mainly Ca^{2+}, Mg^{2+}, Na^{+} and to some extent K^{+}. Another consequence of these processes is a rise in the silica concentration as well as a rise in pH. These trends are also found in Late Precambrian and Cambro-Silurian sedimentary rocks (Englund et al. 1977) although saturation with respect to calcite often occurs in groundwaters from calcareous shales.

Groundwater chemistry in areas below the Late Glacial sea-level is also influenced by remnants of sea-water and salts(?), left over in marine sediments or in rock fractures from the time when areas were covered by the sea (Englund & Meyer 1980). Sodium and chloride are then the dominant ions in the groundwater. Such water is generally found in bedrock below clay deposits where the groundwater moves slowly or is almost stagnant.

Calcium and bicarbonate are, however, the dominant ions in most of the groundwaters studied here (Table 1, Fig. 2). They occur in approximately equal amounts; commonly, however, the amount of bicarbonate is more in harmony with the sum $Ca^{2+} + Mg^{2+}$ (Fig. 3). Magnesium is relatively more important above than below the Late Glacial sea-level.

The HCO_3^- content in groundwaters is derived from soil-zone CO_2 and from dissolution of calcite and dolomite. CO_2 in the soil comes from the atmosphere, via plant respiration and decay of organic material.

Table 1. Average chemical composition of some groundwaters from southeastern Norway. Observation period 1975–1981

	Area	Geology/ Land use	Occurrence of groundwater/ Depth below surface	pH	µeq/l							Number of wells/ springs	Number of analyses
					Ca	Mg	Na	K	HCO_3	SO_4	Cl		
Above the Late Glacial sea-level	15) Åstadalen	Glaciofluvial, till	Springs	5.9	178.5	48.5	57.5	8.5	222.5	25.5	30.0	3	35
	14) "	Dominantly sandst., some congl., shale	"	6.0	149.5	28.5	53.5	5.5	195.0	32.5	18.5	25	49
	13) Lillehammer	Sandst., congl.	In wells, 25–100 m b.s.	7.5	1605	238	191	18	1690	344	85	4	53
	12) "	Sandst., dark shale	In wells, 25–70 m b.s.	7.5	2190	590	491	36	1980	1085	176	9	96
	11) Stensengbekken, at lake Mjøsa	Till/Forested land	Springs/wells 0–2 m b.s.	6.6	1005	213	157	14	933	313	82	4	25
	10) "	Till/Arable land	"	6.6	1480	590	278	85	988	517	423	6	48
	9) "	Sandst., sandy limest., grey shale	In wells 25–60 m b.s.	7.2	3155	1287	757	120	4060	446	890	5	32
	8) "	Limestone	In wells 25–55 m b.s.	7.6	3265	1131	1252	77	4550	794	321	2	11
	7) "	Dark shale, limest.	In wells 25–45 m b.s.	7.1	7605	2795	1087	238	7590	2210	1256	2	12
Below the Late Glacial sea-level	6) Moss	Ice-edge deposits/ Forested land	In wells 0–5 m b.s.	6.4	300	172	543	115	380	417	372	2	4
	5) Ås	Ice-edge deposits/ Arable land	"	7.4	2910	795	987	105	3360	821	727	4	9
	4) Ås, Moss	Gneiss, amphibolite	In wells 30–100 m b.s.	7.5	1400	467	1695	120	1786	562	732	24	51
	3) Jeløy	Volcanic rocks, sandst.	"	7.4	1850	492	1217	64	2459	438	563	13	32
	2) Moss, Jeløy	Volcanic rocks, gneiss, amphibolite	Springs	7.1	2715	639	1383	61	2972	463	656	5	436
	1) Ås, Moss	Gneiss, amphibolite	In wells 25–90 m b.s.	7.2	3645	934	9491	261	66	1421	14456	5	41
	Ocean water (Rankama & Sahama 1950)			~ 8.0	20800	106147	467478	9974	2377	56271	544930		

Abbreviations: Sandst. = sandstone, congl. = conglomerate, limest. = limestone

Where the aquifers contain pyrite, oxygenated recharge water attacks the pyrite to produce SO_4^{2-}. Sulphate-bearing minerals such as gypsum and anhydrite occur in places. They dissolve readily when in contact with water.

Major-ion Evolution Sequences

As groundwater moves along its flow paths in the saturated zone, the contents of total dissolved solids and most of the major ions normally increase. Lower concentrations are usually found in groundwaters from recharge areas rather than in deep circulating groundwaters and in springs.

From the present investigation three main chemical evolution trends can be demonstrated, dependent upon the aquifer type in which the groundwaters occur.

ÅSTADALEN AND LILLEHAMMER–BRØTTUM AREA

The groundwaters are located in Late Precambrian sandstones, conglomerates and dark shales (Brøttum Formation), and in overlying Quaternary deposits such as tills and glaciofluvial sediments. The areas are little disturbed by man.

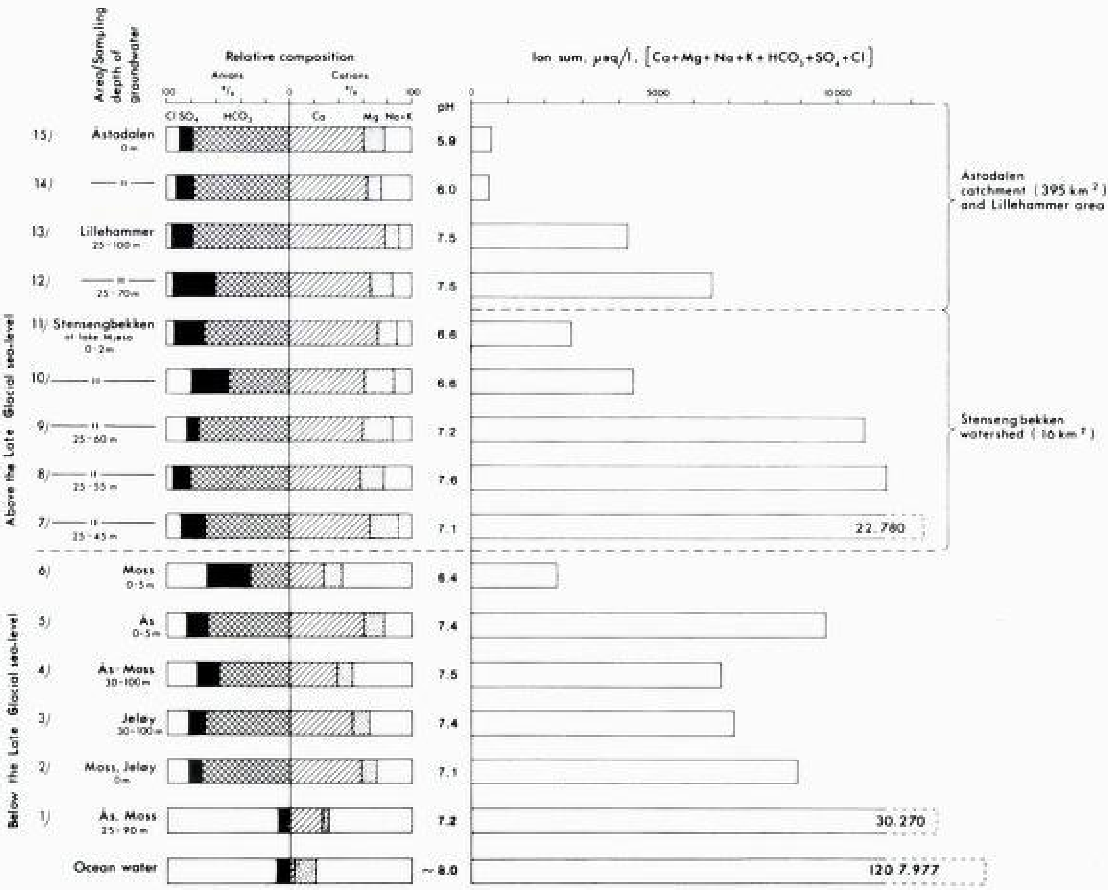

Fig. 2. Average groundwater compositions from southeastern Norway, as presented in Table 1, compared with ocean water after Rankama & Sahama (1950).
1) Precambrian gneiss, amphibolite (fresh-to-brackish water). 2) Springs from Precambrian and Permian rocks. 3) Permian volcanic rocks. Downtonian sandstone. 4) Precambrian gneiss, amphibolite. 5) Quaternary deposits, mainly ice-marginal deposits. Arable land. 6) Quaternary deposits, mainly ice-marginal deposits. Forested land. 7) Cambro-Ordovician dark shale, limestone. 8) Ordovician-Silurian limestone. 9)Ordovician sandstone, sandy limestone, grey shale. 10) Springs/wells from Quaternary deposits, mainly till. Arable land. 11) Springs/wells from Quaternary deposits, mainly till. Forested land. 12) Late Precambrian sandstone, dark shale (Brøttum Formation). 13) Late Precambrian sandstone, conglomerate (Brøttum Formation). 14) Springs from dominantly sandstones with some conglomerates and shales (Brøttum Formation). 15) Springs from Quaternary deposits, mainly glaciofluvial and till.

From a hydrogeological viewpoint the investigated springs of Åstadalen (Table 1, Fig. 2) are draining shallow zones with active flushing, 0–30 m below surface; while in the Lillehammer–Brøttum area the groundwater is older and with a more sluggish flow, down to 50–120 m below surface (Fig. 6).

The groundwaters studied belong to the bicarbonate facies (Fig. 4). There is, however, a clear chemical evolution from water in Quaternary deposits (loc.no. 15), through water in sandstones (loc.no. 14, 13), to the deep water in sandstones alternating with dark shale beds (loc.no. 12); an increase in the

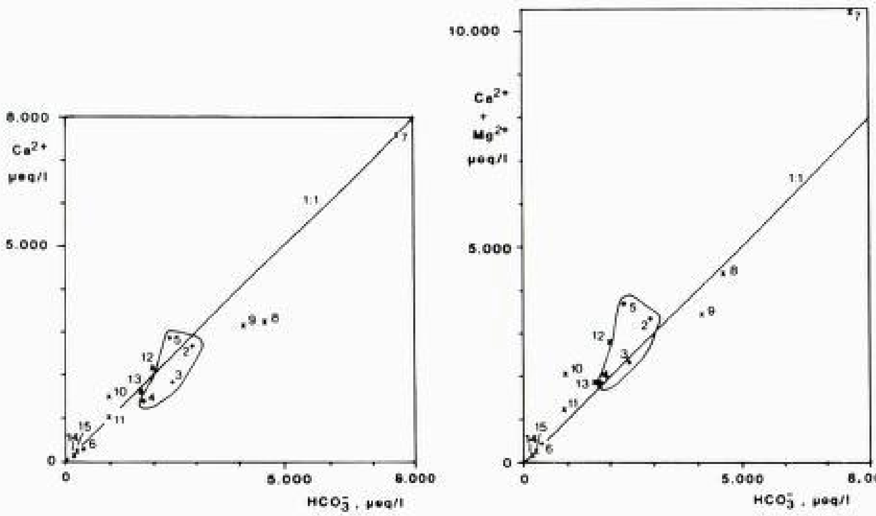

Fig. 3. Relationship between some important ions in groundwaters from southeastern Norway. • data from below, and × data from above the Late Glacial sea-level. Location numbers: see Table 1 and Fig. 2.

total ion content, relative increase of SO_4^{2-} compared to HCO_3^-, and an increase of pH from 5.9 to 7.5, and further to pH 7.7 in springs (Fig. 7). The total ion increase, coupled with the rise in pH, reflect an increased time of contact between water and rock towards depth, and chemical reactions whereby H^+ in the water is being consumed.

The occurrence of sulphate in the groundwater reflects mainly the existence of pyrite in the bedrock (Englund 1972, 1973). Sulphate-bearing weathering products on exposures of these rock units are known from many places, e.g. Fåvang–Ringebu in Gudbrandsdalen (Broch 1931, Caillere & Prost 1968).

STENSENGBEKKEN WATERSHED

The groundwaters occur in Cambro-Silurian shales, sandstones and limestones, and in overlying tills. The watershed represents an ecosystem which has been manipulated by man. Therefore, some of the groundwater, like the river Stensengbekken (Bjerve et al. 1981), is more or less polluted from agriculture, reflected in high concentrations of e.g. potassium and nitrate.

The groundwaters belong to the bicarbonate facies (Fig. 4), showing a chemical evolution with a relative increase in HCO_3^- concentration compared to SO_4^{2-} and Cl^- when going from Quaternary deposits (loc.no. 10–11) to the deep circulating waters in bedrock (loc.no. 7–9). At the same time the total ion content and pH are increasing (Table 1, figs. 2, 6 and 7). The increase of pH is largely coupled with a corresponding increase in the HCO_3^- concentration.

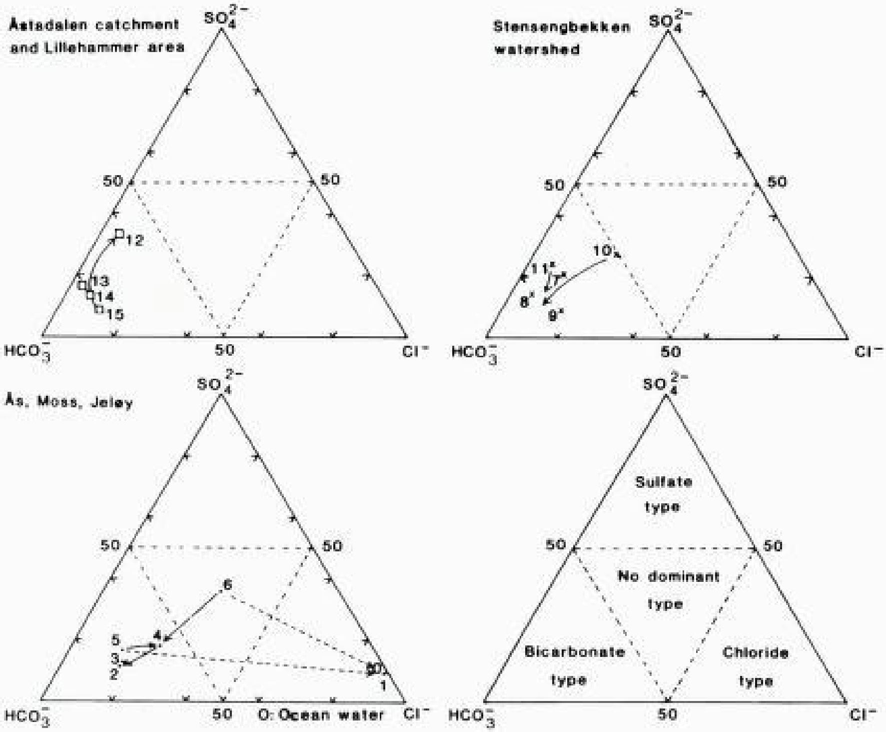

Fig. 4. Classification of the investigated groundwaters from southeastern Norway, based on the percentages of major anions. Water types are designated according to their position on the diagram segments (after Morgan & Winner 1962, and Back 1966). Arrows indicate evolution of water types along flow path. Location numbers: see Table 1 and Fig. 2.

The hydrochemistry of grundwaters in Cambro-Silurian sedimentary rocks is always strongly influenced by the rock types in which they occur (Englund & Myhrstad 1980). Especially important is the solution of calcite and dolomite, resulting in an increase of the HCO_3^- contents towards depth, as well as an increase in the contents of Ca^{2+} and Mg^{2+}.

Sulphate reflects largely the existence of pyrite in bedrock. The increased concentration in chloride towards depth, especially within dark shales alternating with carbonate beds (Table 1), indicates the existence of geological chloride sources (Englund & Myhrstad 1980, p. 43). However, chloride in the shallow groundwaters of the Quaternary deposits is mainly due to agricultural activities, e.g. fertilizers and animals.

ÅS–MOSS

The groundwaters are located in Precambrian gneisses which are commonly alternating with amphibolites, and in overlying Quaternary sediments, largely ice-marginal deposits. Generally, groundwaters in crystalline rocks from various parts of the world have low major-ion concentrations (Matthess 1973, p. 227, Freeze & Cherry 1979, p. 275). Within the present areas, however, waters are found ranging from the bicarbonate facies to the chloride facies (Fig. 4); this is mainly due to the following factors:

1. Permian mineralization along fractures (e.g. carbonate minerals).
2. Remnants of Late Glacial/Holocene sea-water and sea salts(?) in rock fractures and/or in marine sediments.
3. Pollution from agriculture.

Two different chemical evolution trends have been found (Fig. 4): (a) a relative increase in HCO_3^- concentration compared to SO_4^{2-} and Cl^- during flow (loc.no. 6, 4 and 2); (b) transformation of HCO_3^--rich water into brackish water with a high content of chloride (loc.no. 1). It is believed that the first path occur within rocks carrying carbonate minerals, while the second path is mainly due to the supply of old sea-water/sea salts (factor 2 above). Both evolution trends are coupled with a corresponding increase in the content of total ions (Table 1, Figs. 2 and 6).

JELØY

Within the volcanic rocks of this island one chemical evolution trend has been found (Fig. 4): a relative increase in HCO_3^- concentration compared to SO_4^{2-} and Cl^- when going from deep groundwaters in wells and to springs (loc.no. 3 and 2). At the same time the total ion content is increasing (Table 1 and Fig. 2).

SOME CONCLUSIONS

The total contents of dissolved solids ($Ca^{2+} + Mg^{2+} + Na^+ + K^+ + HCO_3^- + SO_4^{2-} + Cl^-$) are usually lower in recharge areas than in deep circulating groundwaters and in springs (Table 1, Figs. 2 and 6). There is simultaneously a change in the relative anion composition, following three main trends (Fig. 4): (a) an increase of SO_4^{2-} compared with HCO_3^- and Cl^- (in pyrite carrying sandstones/shales); (b) an increase of HCO_3^- compared with SO_4^{2-} and Cl^- (in carbonate-bearing rocks); (c) an increase of Cl^- compared with HCO_3^- and SO_4^{2-} (mainly due to the supply of Late Glacial/Holocene sea-water and sea salts(?)).

Many groundwaters of the world show chemical evolution during flow, as described by Chebotarev (1955); from shallow zones of active flushing (HCO_3^--rich water) through intermediate zones ($HCO_3^- + SO_4^{2-}$-rich water) into zones where water flow is very sluggish and the water is old (Cl^--rich water). Most of the water investigated in this work belongs to the upper zone of Chebotarev. The different anion-evolution sequences are determined either by the sediment/rock types in which the groundwaters occur, or by the supply of old sea-water/sea salts (point *c* above).

Quality and Flow Patterns

A schematic picture of the groundwater systems investigated is presented in Fig. 5. The logical location of water samples along flow lines, representing

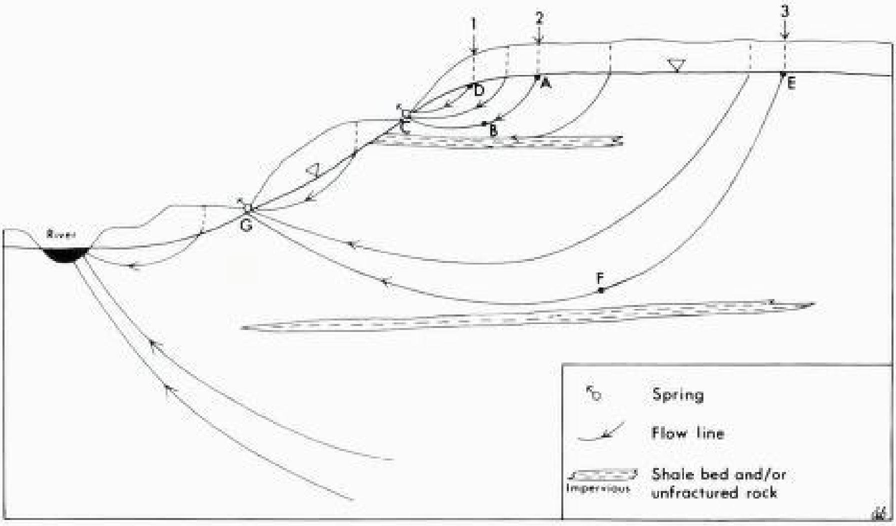

Fig. 5. Hypothetical flow lines in groundwater flow systems within fractured rocks. For further explanation, see text.

entrance (points A, D, E), average (points B, F) and final values (points C, G), are after Engelen (1981). From the present investigation the following statements can be made:

ÅSTADALEN

The flow through the aquifers is following shallow flow lines, mainly of the types 1 and 2 in Figs. 5–7. There are small chemical changes along the flow lines, from entrance values points A and D to the final or outlet values of the springs, point C. This is due to a rather short time of contact between water and rock, as well as to slow dissolution of the minerals.

During winter and dry seasons the flow to the springs is dominantly following flow lines of type 2, while heavy rain and snowmelt causes much flow along the very shallow flow line of type 1. Generally, the base flow of type 2 shows higher contents of geologically derived solids (e.g. Ca^{2+}) than the very shallow flowage along flow line 1.

LILLEHAMMER–BRØTTUM AREA

The groundwaters investigated appear to have followed long and deep flow lines, dominantly of type 3 in Figs. 5–7. At the considered depths (50–120 m below surface) the flow is slower than in Åstadalen (20–30 m below surface), resulting in higher ion concentrations in the groundwater than in Åstadalen (Table 1, Figs. 2 and 6).

STENSENGBEKKEN WATERSHED

Assuming the flow pattern to be dominantly of the types 2 and 3 in Figs. 5–7, Observed chemical changes then correspond to the changes from A to C or from E to G; an increase in ion concentration within Quaternary deposits, as well as from Quaternary deposits to bedrock, and during flow within the bedrock aquifers (Table 1, Figs. 2 and 6).

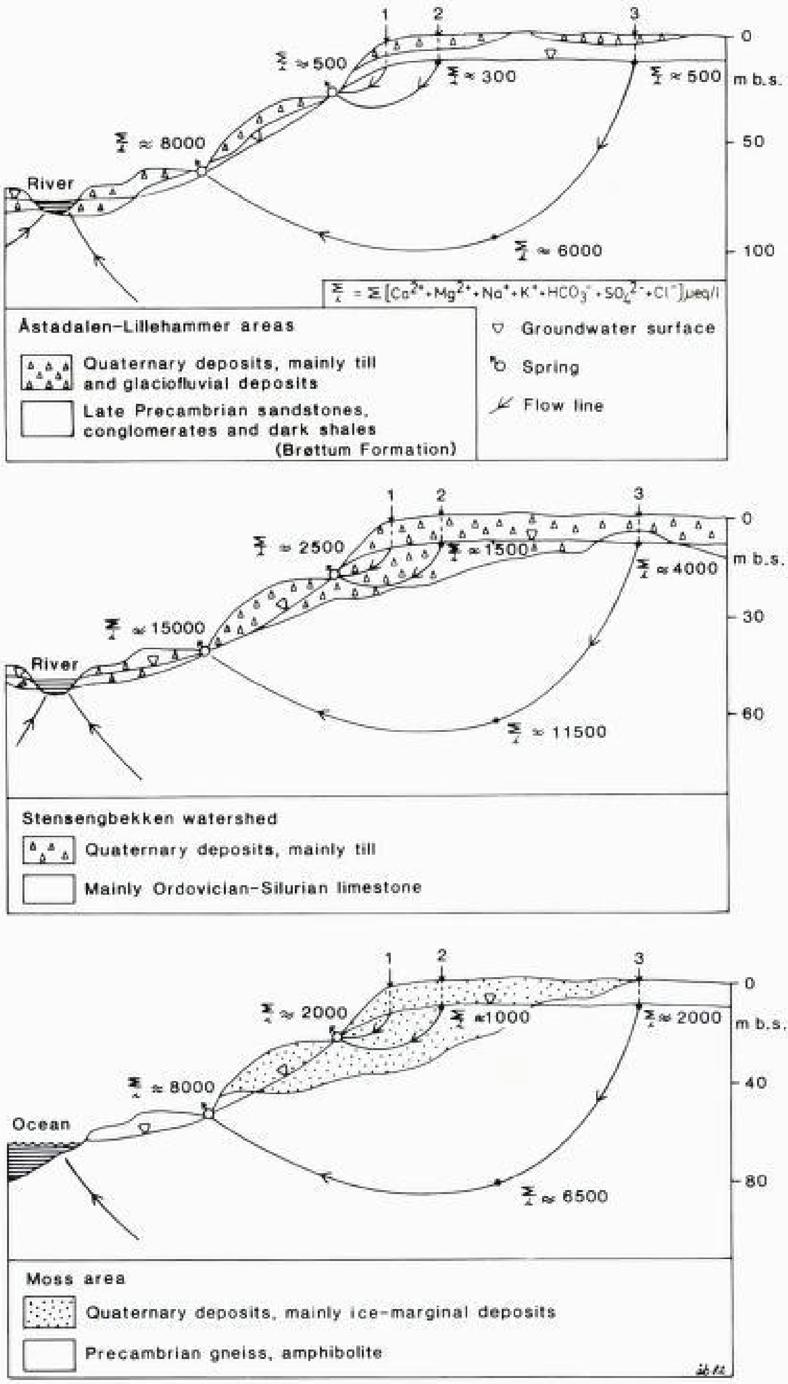

Fig. 6. Some generalized pictures of groundwater flow systems and the resulting changes in ion concentrations. Observation period 1975–1980.

ÅS - MOSS

Two main modes of flow are considered, following flow lines 2 and 3 in Figs. 5–7. Much of the water following flow line 3 has seeped through Quaternary deposits.

SUMMARY

Observed chemical differences are explained by two main modes of groundwater flow patterns: (a) flow through shallow aquifers with active flushing, and ending up in springs (Åstadalen, Stensengbekken watershed, Moss area); (b) long and deep flow from Quaternary deposits or exposed bedrock through deep circulating groundwaters in bedrock, and ending up in springs and rivers (Lillehammer–Brøttum area, Stensengbekken watershed, Ås–Moss areas. The highest concentration value along any one flow path is usually found in the spring waters.

The Acidity of Groundwater

The pH of investigated water types generally increases from about 4.40–4.45 in the precipitation to 5.5–6.5 near or just below, the groundwater surface. Deeper groundwater is usually buffered within the pH range of 7.0–8.5. The magnitude of these changes is dependent on (Fig. 7): (1) the flow pattern considered, and (2) the soil and rock types encountered along the flow path.

The decrease in H^+-concentration in percolating water and during groundwater flow is due to H^+ exchange with the cation exchange reservoir in the soil, and due to chemical weathering reactions with primary and secondary minerals. The extent of these reactions is highly dependent on soil and rock types.

Most investigated groundwaters have high buffer capacities against strong acids, as shown by the high concentrations of HCO_3^- (Table 1). Thus, acid precipitation seems not to have measurable effects on the acidity of the deep groundwaters investigated here; i.e. those deeper than 15–25 m below surface. The only effect observed which could be due to acid precipitation is in the springs of Åstadalen during the time of snow-melting; a decrease of pH from about 6.0 to 5.5 (Englund, unpublished).

Acid groundwaters are, however, known from Scandinavia. Groundwaters with pH$<$6.0, down to 4.0, have been reported in lime-poor sandy soils and in granites in southern Sweden, and have been interpreted as a result of acid precipitation (Eriksson 1981, Hultberg & Johansson 1981). Also in regions from southern Norway where lakes are acidified, shallow groundwater in superficial deposits appears to be acidified, but less so than in neighbouring lakes (Henriksen & Kirkhusmo 1982).

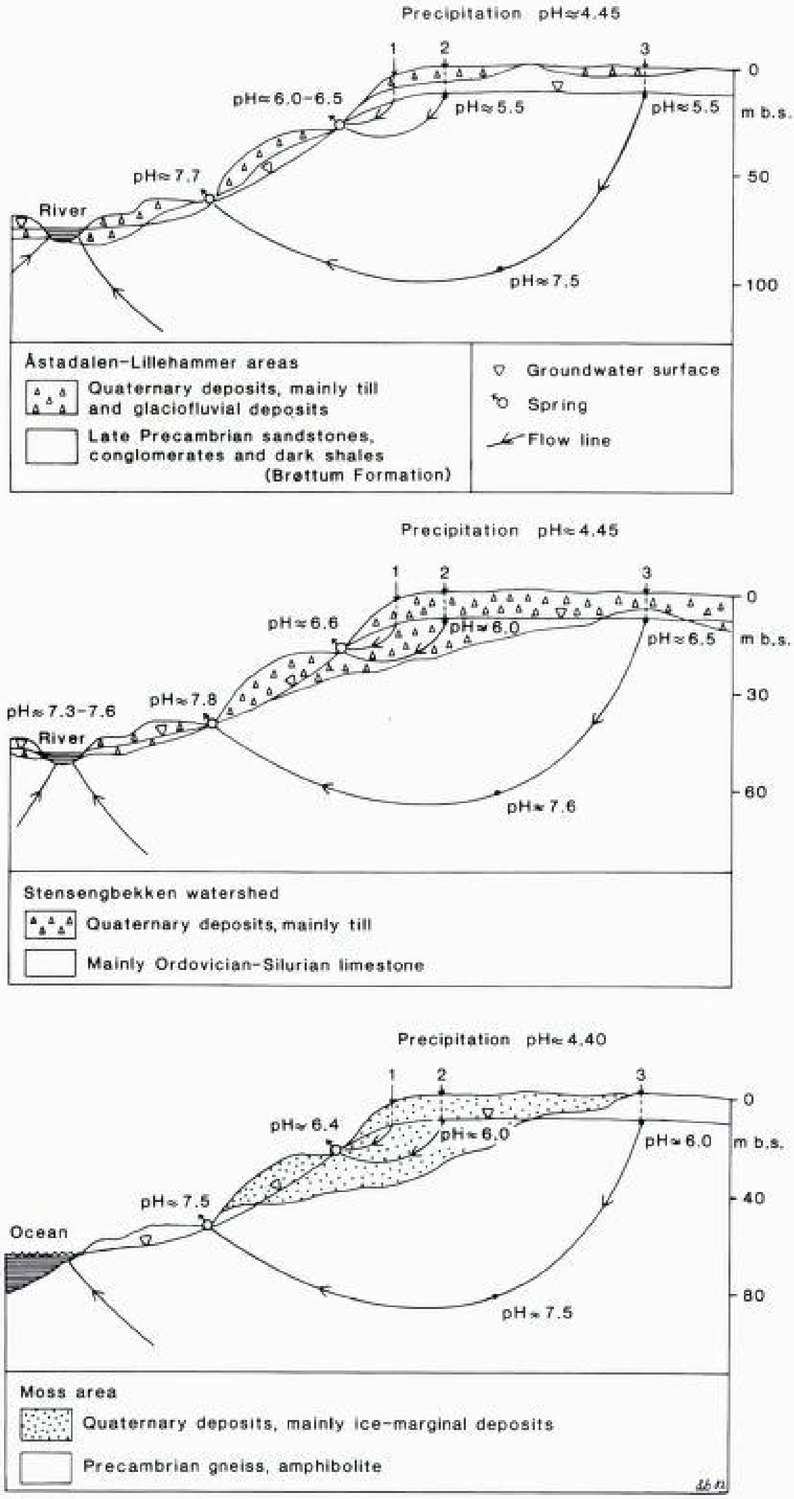

Fig. 7. Some generalized pictures of groundwater flow systems and the resulting changes in pH. Observation period 1975–1980.

Concluding remarks

1. Groundwaters in Quaternary deposits such as glaciofluvial, ice-marginal and till units, and in fractures in gneisses, amphibolites, volcanic rocks, sandstones, shales and limestones have been investigated during the years 1975–1981 from the following areas in southeastern Norway: Åstadalen, Lillehammer–Brøttum, Stensengbekken watershed, Ås, Moss and Jeløy.

2. Since seasonal variations are quite common in groundwater chemistry, especially in shallow aquifers, and since only large-scale chemical changes are discussed here, the data used are annual average results. 979 water samples from 133 wells and springs have been analysed, sampled during winter, spring, summer and autumn.

3. The chemical compositions down to 120 m below surface range from very dilute waters to brackish waters. Calcium and bicarbonate are generally the dominant ions. In areas below the Late Glacial sea-level (Ås, Moss and Jeløy), however, sodium and chloride ions are in places predominant.

4. The total content of dissolved solids (Ca^{2+} + Mg^{2+} + Na^{+} + K^{+} + HCO_3^- + SO_4^{2-} + Cl^-) are usually lower in recharge areas, than in deep circulating groundwaters and in springs. There is, simultanously, a change in the relative anion composition following three main trends: (a) an increase of SO_4^{2-} compared with HCO_3^- and Cl^- (in pyrite-bearing sandstones/shales); (b) an increase of HCO_3^- compared with SO_4^{2-} and Cl^- (in carbonate-bearing rocks); (c) an increase of Cl^- compared with HCO_3^- and SO_4^{2-} (mainly due to the supply of Late Glacial/Holocene sea-water and sea salts(?)). These different anion-evolution sequences are determined either by the sediment/rock types in which the groundwaters occur or by the supply of old sea-water/sea salts.

5. Observed chemical differences are explained by two main modes of groundwater flow patterns: (a) flow through shallow aquifers with active flushing, and ending up in springs (Åstadalen, Stensengbekken watershed, Moss area); (b) long and deep flow, from Quaternary deposits through deep circulating groundwaters in bedrock, and ending up in springs and rivers (Lillehammer–Brøttum area, Stensengbekken watershed, Ås–Moss areas). The highest concentration value along any one flow path is usually found in the spring waters.

6. Generally, the pH of water increases from about 4.40–4.45 in the precipation to about 5.5–6.5 near, or just below, the groundwater surface. Deeper groundwater is usually buffered within the pH range of 7.0–8.5.

7. Acid precipitation does not seem to have measurable effects on the acidity of deep groundwaters; deeper than 15–25 m below surface. The only effect observed which could be due to acid precipitation is found in springs from shallow aquifers in Åstadalen. During melting of snow the pH in the spring waters is decreasing from about 6.0 to about 5.5.

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REFERENCES

Aune, B. 1981: Map. Mean annual observed precipitation in Norway. *The Norwegian Meteorological Institute.*

Back, W. 1966: Hydrochemical facies and ground-water flow patterns in northern part of Atlantic Coastal Plain. *U.S. Geol. Surv. Prof. Paper. 498-A.* 42 pp.

Bjerve, L., Englund, J. O. & Jørgensen, P. 1981: Hydrology and water chemistry in a small watershed. *In:* van Duijvenbooden, W., Glasbergen, P. & van Lelyveld, H. (eds.) Quality of Groundwater. Studies in Environmental Science 17. Elsevier. pp. 863–866.

Broch, O. A. 1931: Pickeringite from Fåvang. *Norsk Geol. Tidsskr. 12,* pp. 117–122.

Bryn, K. Ø. 1961: Grunnvann øst for Oslo-feltet. *Norges geol. Unders. 213,* pp. 5–19.

Caillere, S. & Prost, A. 1968: Occurrence of halotrichite at Sør-Fron (Oppland). *Norsk Geol. Tidsskr. 48, no. 4,* pp. 253–258.

Chebotarev, I. I. 1955: Metamorphism of natural waters in the crust of wathering. *Geochim. Cosmochim. Acta, 8,* pp. 22–48, 137–170, 198–212.

Engelen, G. B. 1981: A systems approach to ground water quality. *In:* van Duijvenbooden, W., Glasbergen, P. & van Lelyveld, H. (eds.) Quality of Grundwater. Studies in Environmental Science 17. Elsevier. pp. 1–25.

Englund, J. O. 1966: Grunnvann i spragmittgruppens bergarter i Syd-Norge. *Norges geol. unders. 242,* pp. 5–18.

Englund, J. O., Jørgensen, P., Roaldset, E. & Aagaard, P. 1977: Composition of water and sediments in Lake Mjøsa, South Norway, in relation to weathering processes. Proceedings of an international symposium held at Amsterdam, The Netherlands, September 6–10, 1976. «Interaction between sediments and fresh water». pp. 125–132. Dr. W. Junk B. V. Publishers, The Hague.

Englund, J. O. & Myhrstad, J. A. 1980: Groundwater chemistry of some selected areas in Southeastern Norway. *Nordic Hydrology, 11,* pp. 33–54.

Englund, J. O. & Meyer, K. F. 1980: Groundwater discharge through springs with well-defined outlets. A case study from Jeløya–Moss, S. Norway. *Nordic Hydrology, 11,* pp. 145–158.

Eriksson, E. 1981: Aluminium in Groundwater. Possible Solution Equilibria. *Nordic Hydrology, 12,* No. 1, pp. 43–50.

Freeze, R. A. & Cherry, J. A. 1979: Groundwater. Prentice-Hall, Inc., Englewood Cliffs, N. J. 604 pp.

Helgeson, H. C., Brown, T. H. & Leeper, R. H. 1973: Handbook of Theoretical Activity Diagrams. Freeman, Cooper and Company. 253 pp.

Henriksen, A. & Kirkhusmo, L. A. 1982: Acidification of Groundwater in Norway. *Nordic Hydrology, 13,* pp. 183–192.

Holtedahl, O. & Dons, J. A. 1960: *In:* Holtedahl, O. (ed.) Geology of Norway. *Norges geol. Unders. 208.*

Hultberg, H. & Johansson, S. 1981: Acid Groundwater. *Nordic Hydrology 12, No. 1,* pp. 51–64.

Jacks, G. 1973: Chemistry of some groundwaters in igneous rocks. *Nordic Hydrology 4,* pp. 207–236.

Matthess, G. 1973: Die Beschaffenheit des Grundwassers. Gebrüder Borntraeger. Berlin–Stuttgart. 324 pp.

Morgan, C. O. & Winner, Jr., M. D. 1962: Hydrochemical facies in the 400 foot and 600 foot sands of the Baton Rouge Area, Louisiana. *U.S. Geol. Surv. Prof. Paper 450-B,* pp. B. 120–121.

Ramberg, I. & Larsen, B. T. 1977: Tectonic map., Oslo graben. *Norges geol. Unders. 337.*

Rankama, K. & Sahama, Th. G. 1950: Geochemistry. University Press, Chicago. 912 pp.

Skjeseth, S. 1953: Vannboringer utført i traktene omkring Mjøsa og Randsfjorden 1950–52. *Norges geol. Unders. 184,* pp. 12–22.

Skjeseth, S. 1956: geologi og vannboring. Kambro-Silurbergartenes hydrologi i Mjøstraktene. *Norges geol. Unders. 195.* pp. 15–36.

A Study of the Earth's Crust in the Island Area of Lofoten–Vesterålen, Northern Norway

MARKVARD A. SELLEVOLL

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The present study is based on a reinterpretation of seismic refraction data as well as gravity data collected by the Seismological Observatory, University of Bergen in the island area of the Lofoten–Vesterålen, Northern Norway. The study shows a good agreement between the seismic and gravity modelling of the Moho depths along the profile line between the two shotpoints Hamnøy (Lofoten) and Stø (Vesterålen). A maximum Moho-depth of 26 km is observed about 85 km from Stø, and it would seem that the rise of the Moho towards Stø is moderate. On the other hand there is a distinct shallowing of the Moho beneath the Lofoten area.

M. A. Sellevoll, Seismological Observatory, University of Bergen, Allégt. 41, N-5014 Bergen–Univ., Norway

Introduction

The crustal structure of the Lofoten–Vesterålen region of Northern Norway was the subject of a special study during the seismic program conducted in the summer 1965 by the U.S. Geological Survey (Branch of Crustal Studies) and the Seismological Observatory, University of Bergen. The Seismological Observatory has since continued to carry out seismic and gravimetric measurements in this island area as well as on the adjacent continental margin. The gravimetric field measurement on land was carried out in 1968 and the deep seismic sounding was further carried out in 1969, 1970 and 1972. The data from the investigations on the Lofoten–Vesterålen islands have been the subject of several M.Sc. theses (Kjenes 1970, Svela 1971, Hansen 1972, Enoksen 1973). Two papers based on the collected data have been published (Thanvarachorn 1975, Sellevoll & Thanvarachorn 1977). The main intention of the present paper is to present the results of a reinterpretation of the available seismic and gravimetric data.

Data acquisition and analysis

The two shotpoints ('Stø' and 'Hamnøy') together with the recording sites are plotted in Fig. 1. The charges were detonated on the sea floor at a dept of 47 m (Stø) and 45 m (Hamnøy). Tables 1a and 1b give information about shotpoint location, shot numbers, date, explosion time and charge size. Seismic signals were detected by a three component geophon system (HS-1/4,5 Hz) and recorded on analogue magnetic tape, together with radio time signals

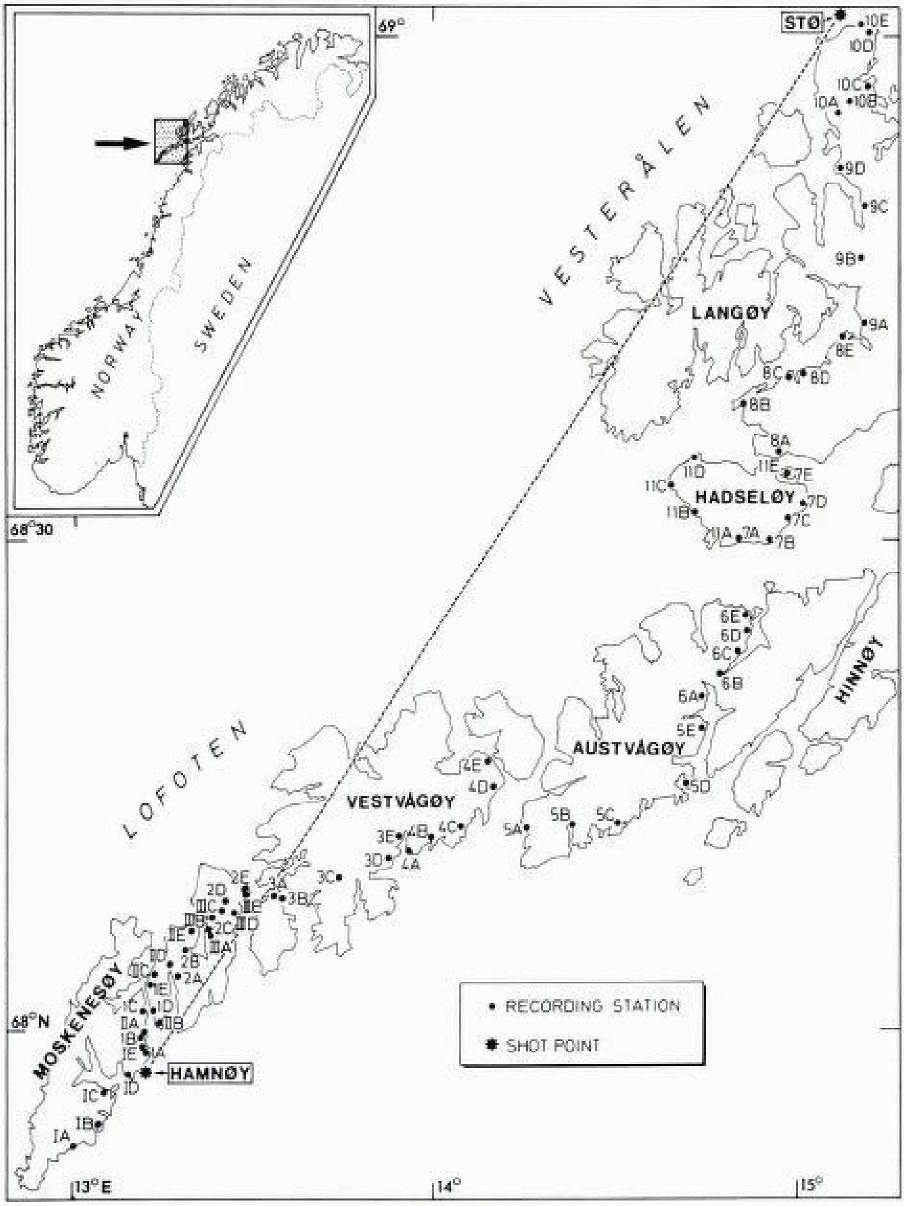

Fig. 1. Location of seismic stations and shotpoints.

(Tryti & Sellevoll 1977). Five recording units (Mars 66, marked A, B, C, D and E) were used during the field measurements and the data obtained were manually digitized. A frequency-analysis (Enoksen 1973) showed that a good signal to noise ratio could be obtained by band-pass filtering between 6 and 16 Hz. After applying this filter the Z-components were plotted in two seismic sections using a reduction velocity of 6 km/sec. The two seismic sections – one for the Hamnøy shotpoint and the other for the Stø shotpoint – are presented in Fig. 2.

Table 1. Shots and shotpoints.

(A) SHOTPOINT: HAMNØY (67°57'42" N, 13°10'12" E) Explosion depth: 45 m)				(B) SHOTPOINT: STØ (69°00'19,2" N, 15°03'23" E) (Explosion depth: 47 m)			
Shot No.	Date	Expl. Time (hms)	Charge (kg)	Shot No.	Date	Expl. Time (hms)	Charge (kg)
1/69	15–8–69	12–28–34.45	(25)	1/70	13–8–70	18–29–32.33	125
2/69	15–8–69	18–29–35.04	(25)	2/70	13–8–70	12–29–53.18	125
3/69	16–8–69		(50)	3/70	12–8–70	12–19–31.72	75
4/69	16–8–69	18–29–34.49	(50)	4/70	12–8–70	18–29–31.34	75
5/69	18–8–69	12–29–34.66	(50)	5a/70	11–8–70	12–29–30.76	75
6/69	18–8–69	18–29–34.34	(75)	5b/70	14–8–70	18–29–32.47	75
8/69	19–8–69	12–29–34.34	(100)	8/70	10–8–70	12–29–32.46	50
9/69	19–8–69	18–29–34.46	(100)	9/70	08–8–70	12–31–55.10	50
10/69	20–8–69	07–59–34.41	(100)	10/70	07–8–70	18–31–57.46	25
				11/70	10–8–70	18–29–31.61	50
10/70	07–8–70	18–29–34.02	100				
9/70	08–8–70	12–29–33.00	100	I/72	31–8–72	18–29–31.92	175
5/70	11–8–70	12–31–31.51	50	II/72	31–8–72	12–29–31.25	150
2/70	13–8–70	12–31–28.95	25	III/72	30–8–72	18–29–31.80	125

Wave pattern, phase correlation and seismic modelling

The first seismic model of the crustal structure of the Lofoten–Vesterålen region was based on seismic records from several large shots fired at a single shotpoint north of the Lofoten–Vesterålen islands (20 km west of Tromsø). A two-layer crust with P-velocities of 6.10 and 6.67 km/s was deduced. The apparent P_n velocity was found to be 8.26 km/s in the distance range 160–290 km from the shotpoint. The elongated gravity high, which is especially well developed in the Lofoten area, was assumed to be mainly a result of crustal thinning (Sellevoll 1967).

Since 1965 new seismic and gravity measurements have been carried out on land in the Lofoten–Vesterålen area. These measurements have given additional information concerning the crustal structure of the Lofoten–Vesterålen island region which in general supported the main feature of the preliminary model based on the 1965 measurements. In the following, a reinterpretation is presented of the data obtained during the 1969–72 seismic investigations in the Lofoten–Vesterålen area. The kinematic seismic modelling has been carried out by application of a ray-tracing program developed by Pajchel (1980).

The P-wave velocity associated with the uppermost part of the crust (P_{g1} in Fig. 2) has been found to be 6.05 km/s. Beyond 25 km from shot-point 'Stø' the P_{g1}-phase seems to disappear and a new strong first onset is clearly observed at a distance of 50–60 km. This supposedly new phase has been designated as P_{g2}. A completely analogous travel-time feature has not been observed on the seismic section from the countershot at shotpoint 'Hamnøy'.

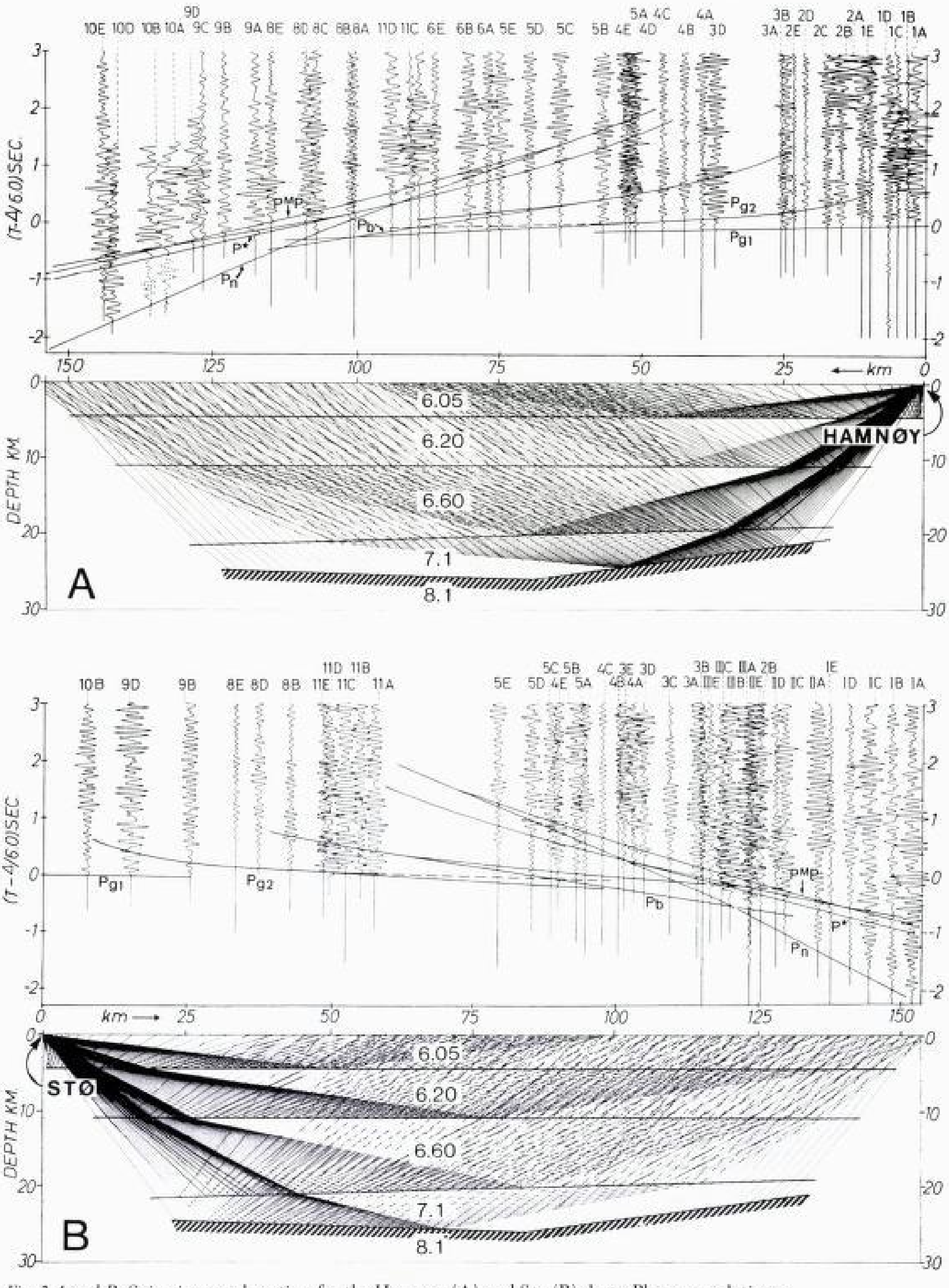

Fig. 2 A and B. Seismic record section for the Hamnøy (A) and Stø (B) shots. Phase correlations and identifications together with ray tracing, layer P-velocities (km/s) as well as the crustal structure are shown on the figure. The Z-components at the stations 10A and B have been destroyed and are partly replaced by the recordings on the NS-components (dashed).

The P_{g1}-phase at this shotpoint, however, is observed beyond 25 km, but certain similarities exist for the P_{g2}-phase from the two shotpoints. The P_{g2}-phase may be explained in several ways, but a preferred interpretation is that the P_{g2}-phase originates from below a thin surface (4–5 km) layer where the P velocity increases from about 6.05 to 6.2 km/s. The data are, however, considered insuffivient for reaching a firm conclusion regarding the structure at the depth interval 4–11 km.

A P-phase (marked P_b), assumed to originate from an intra-crustal discontinuity, is weak but well recorded as the first onset in the seismogram section (Fig. 2 B) between 100 and 120 km from shotpoint 'Stø'. The correlation of the P_b onset for the countershots (Fig. 2 A) is uncertain, but the result obtained indicates an almost horizontal discontinuity at a depth of 11 km and with a phase velocity of 6.6 km/s. This conclusion does not agree with previous interpretations involving an apparent southward shallowing of this discontinuity along the profile in the Lofoten region.

A strong and distinct P-phase, termed P^x, is observed in the seismogram section, Fig. 2 A. A corresponding phase is seen on the seismogram section from the countershot, but this is not so well recorded (Fig. 2 B). The apparent velocities and intercept-times obtained suggest that this phase originates from a layer located at a depth of about 20–22 km in the Vesterålen region. The depth of this layer decreases towards the Lofoten region.

A distinct P_n-phase is observed from shotpoint 'Stø' by the stations located in the Lofoten area (Fig. 2 B). The high apparent velocity of this phase (8.8 km/s) indicates a Moho rising towards the south in the Lofoten area. The correlation of the corresponding P_n-phase (with an apparent velocity of 7.3 km/s) observed from the countershot at 'Hamnøy' can be called into question; but the resulting real P_n-velocity of 8.05 km/s is in agreement with what should be expected. Combining the P_n apparent velocities observed from both shotpoints show that the Moho dips at 5° to the northeast beneath the Lofoten region along the line of profile.

Concerning the Vesterålen region (shotpoint 'Stø') and the correlation of the P_n-phase from the two shotpoints we are unable to draw any definite conclusions about the apparent velocities in both directions from the two shot-points. The observations suggest, however, a weak shallowing of the Moho in a northward direction along the profile (Fig 2).

Discussion and conclusion

The distribution of the shotpoints and recording sites used during these seismic reflection–refraction experiments in the Lofoten–Vesterålen area does not meet the requirement that shot and recording sites should lie on a 'straight line' (Fig. 1). Most of the seismic stations lie on the east side of the straight line between the two shotpoints. The station location 'offsets' are in some cases as much as 30 km. The physical conditions along the real ray paths from the shotpoints to the recording stations may be different from the physical conditions along the shortest line between the two shotpoints (Fig. 1). Such possible variations in physical conditions may change both the kinematic and dynamic wave pattern. The seismic modelling of the crustal structure carried out in the present study is consequently circumscribed by this uncertainty. A normalisation of the seismic recordings shown on Fig. 2 has for various reasons not been possible, and this fact reduces to some extent the reliability concerning the phase correlation as well as the possibilities for detailed seismic amplitude modelling of the crustal structure.

In general the geological-geophysical patterns along the Lofoten–Vesterålen island belt as well as on the adjacent continental shelf (margin) suggest that the underlying continental crust is to some degree segmented into 'blocks' which are observed as basins and crystalline basement highs on the adjacent Lofoten–Vesterålen shelf area. The Lofoten–Vesterålen islands appear to have undergone a rather extensive uplift as compared with the surrounding sedimentary sequences. This differential vertical movement, which has developed faults, is especially well observed as rather strong reflection–diffraction waves on the multichannel seismic sections obtained in the near coastal areas along the Lofoten–Vesterålen islands (Nysæther et al. 1969).

Gravity modelling has been carried out in order to investigate whether the crustal model obtained on the basis of the present reinterpretation of the seismic data is in agreement with the gravity anomaly observed along the Lofoten–Vesterålen islands. The modelling program applied (Hjelle 1979) is a prism-oriented program where the gravity effect from each prism is calculated and added in order to obtain the total gravity value at each point on the surface used in the gravity calculation. The program works in two modes: 1) a direct calculation of the gravity effect from a body where its shape and density distribution are known; 2) an inverse calculation of the shape based on a known gravity anomaly field and density distribution. The last mentioned type of calculation has been applied for the modelling presented here.

The gravity anomaly map which constitutes the basis for the calculation is shown in Fig. 3. The seismic stations and shotpoints are distributed in such a pattern that the gravity edge effect is rather limited (Figs. 1 and 3). The area within the frame in Fig. 3 has been divided into 42 km x 20 km 'units' and the Bouguer anomaly values are estimated for each of the 840 'units'. The depth to each of the layers is assumed to be the same for the whole area as shown on the seismic crustal model presented in Fig. 2. The desities which have been applied for the gravity modelling are assumed constant for each layer, and the densities used are based on the seismic velocities (P-velocities) by application of the following formula (Talwani et al. 1959):

$$\varrho = 1.7 + 0.2\ V_p$$

The seismic crustal model presented in Fig. 2 has no lateral density variation above 18 km and the observed gravity values were reduced by 67 m.gal in order to correct for the masses above this depth (18 km). The gravity effect

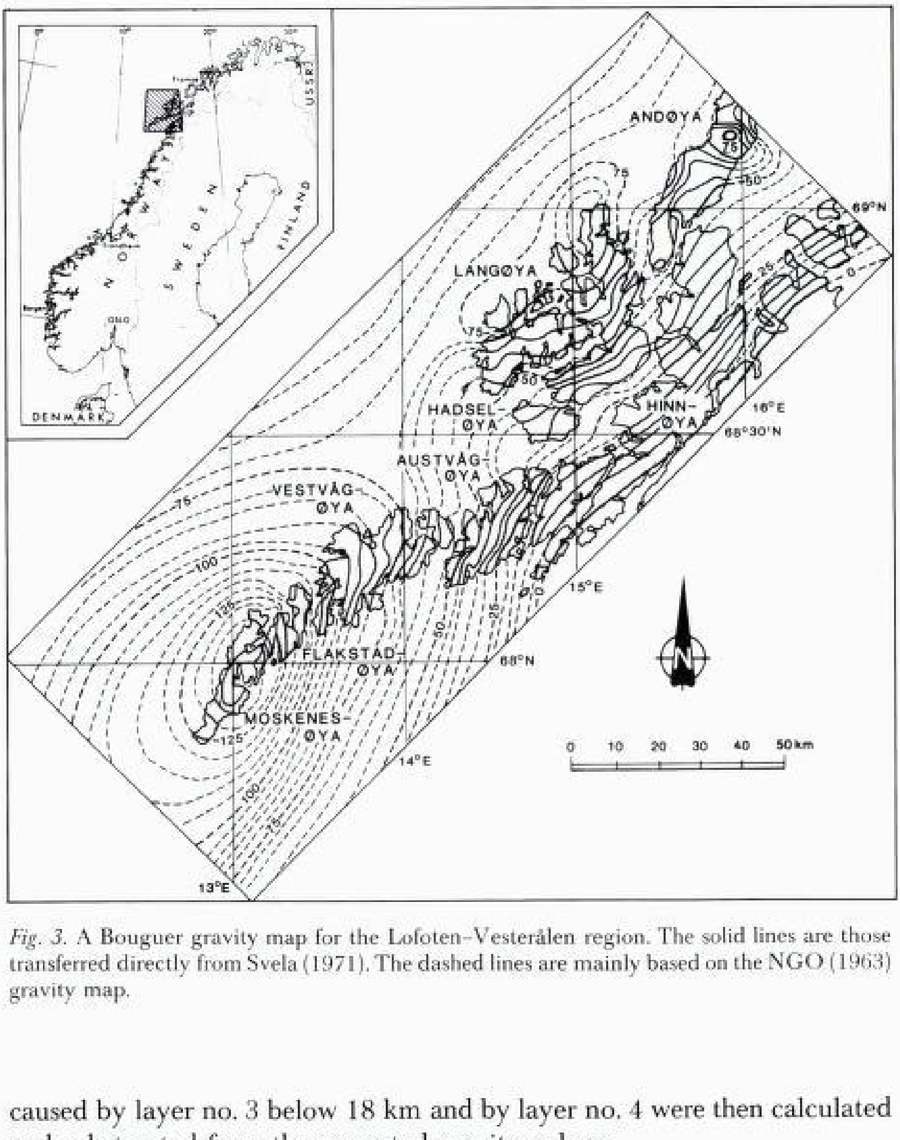

Fig. 3. A Bouguer gravity map for the Lofoten-Vesterålen region. The solid lines are those transferred directly from Svela (1971). The dashed lines are mainly based on the NGO (1963) gravity map.

caused by layer no. 3 below 18 km and by layer no. 4 were then calculated and substracted from the corrected gravity values.

The regional gravity effect was calculated and finally removed, and the gravity residual effect then obtained was utilized for the calculation of the depth of the Moho. The depths of the Moho from the seismic model were used to initiate the gravity modelling. Adjustments to these depths were made in the area to satisfy the residual gravity effect. The computed depths along the central part of the 'profile line' (location of seismic stations and shot-points) are plotted in Fig. 4 (filled circles). The figure shows that the seismic and gravity modelling of the Moho-depths are in good agreement with each other.

Magnetic measurements carried out by the Geological Survey of Norway in the Lofoten-Vesterålen and adjacent regions also show many features of interest. The Lofoten area belongs to a distinctive magnetic high. This can be

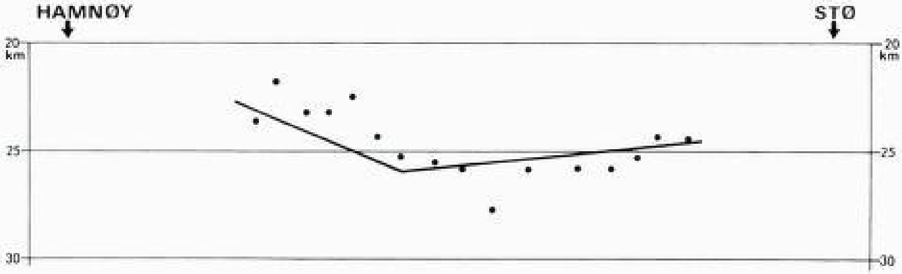

Fig. 4. A correlation of the 'seismic and gravimetric modelling' of the depth to Moho. Solid line – seismic. Dots – gravimetric.

explained by the magnetic properties of the rock, occurring at the surface, if these are considered to extend to a depth of around 20 km (Åm 1975).

Finally attention is drawn to the geological history of the Lofoten–Vesterålen region. The Lofoten–Vesterålen terrain exposes an unusually deep section through the Archaen and Proterozoic crust, the geological history of which (Griffin et al. 1978) has involved repeated vertical movements of considerable magnitude (at least 30 km). There have been two major periods in which new material has been added to the crust, one at ca. 2700 m.y, the other a protracted period from ca. 2100 to 1700 m.y. Important metamorphic events occurred at ca. 2700–2600 m.y, ca. 1830 m.y and ca. 1100–900 m.y Minor heating, intrusion of pegmatites and thrust-faulting occurred during the time of the Caledonian orogeny. Present evidence suggests that Lofoten–Vesterålen formed part of the Baltic plate during the Caledonian orogeny, but escaped deformation because it lay beneath the overriding nappe complex.

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REFERENCES

Enoksen, T., 1973: *Refraksjonsseismiske undersøkelser i Lofoten–Vesterålen området.* Unpubl. Cand. real thesis, Seismological Observatory, University of Bergen.

Griffin, W. L., Taylor, P. N. & Hakkinen, J. W., 1978: Archean and Proterozoic crustal evolution in Lofoten–Vesterålen, North Norway. *Jour. Geol. Soc. 135,* 629–647.

Hansen, R., 1972: *En geofysisk undersøkelse av jura-kritt feltene på Andøya i Vesterålen.* Unpubl. Cand. real thesis, Seismological Observatory, University of Bergen.

Hjelle, K., 1979: *En marin-geofysisk undersøkelse av det kontinentale randområdet utenfor Lofoten–Vesterålen.* Unpubl. Cand. real. thesis, Seismological Observatory, University of Bergen.

Kjenes, H., 1970: *Seismiske undersøkelser av jordskorpen i Lofoten–Vesterålen området.* Unpubl. Cand. real. thesis, Seismological Observatory, University of Bergen.

Nysæther, E., Eldholm,O. & Sundvor, E., 1969: Seismiske undersøkelser av den norske kontinentalsokkel, Sklinnabanken–Andøya. Tkn. Rapp. nr. 3, *Jordskjelvstasjonen,* Universitetet i Bergen.

Pajchel, J., 1980: Kinematic modelling of reflected-refracted and diffracted waves in inhomogenous media with curve-linear boundaries. *Internal report, Seismological Observatory*, University of Bergen.

Sellevoll, M. A. & Thanvarachorn, P., 1977: A seismic reconnaissance study of the earth's crust in the Lofoten-Vesterålen area, Northern Norway. The Norwegian Geotraverse Project, K. S. Heier (ed.), 101–113.

Sellevoll, M. A., 1967: Seismic measurements in Norway 1965, Lofoten– Vesterålen region. (Sørnes & Sellevoll: Final Scientific report, ARPA No. 612-1, available at Seismological Observatory, University of Bergen.)

Svela, P., 1971: *Gravimetriske undersøkelser av Lofoten-Vesterålen området.* Unpubl. Cand. real thesis, Seismological Observatory, University of Bergen.

Talwani, M., Sutten, G. M. & Wovzel, J. L., 1959: A crustal section across the Puerto Rico trench. *Jour. Geophys. Res.*, 1545–55.

Thanvarachorn, P., 1975: A report of preliminary seismic and gravity study of the Lofoten–Vesterålen region, Norway. Seismological Observatory, University of Bergen.

Tryti, J. & Sellevoll, M., 1977: Seismic crustal study of the Oslo Rift. *Pageoph. 115/4*, 1061–85.

Åm, K., 1975: Aeromagnetic basement complex mapping north of latitude 62°N, Norway. *Norges geol. Unders. 316*, - .