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Timing of late- to post-tectonic Sveconorwegian granitic magmatism in South Norway

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Dating of late- to post-tectonic Sveconorwegian granitic intrusions from South Norway by the SIMS U-Pb method on zircons and by internal Pb-Pb isochrons on rock-forming minerals indicates a major event of granitic magmatism all across southern Norway in the period 950 to 920 Ma. This magmatic event included emplacement of mantle-derived magma into the source region of granitic magmas in the lower crust east of the Mandal-Ustaoset shear zone, and formation of hybrid magmas containing crustal and mantle-derived components. West of the Mandal-Ustaoset shear zone, granitic magmatism started earlier, at c. 1030 Ma. A distinct group of granites, characterized by low Sr concentration and a high Rb/Sr ratio, is restricted to central Telemark, and shows evidence of involvement of a component related to the ca. 1500 Ma metarhyolite of the Telemark supracrustal sequence. Whereas one of these granites clearly belongs to the 920-950 Ma age group, two of the intrusions dated in this study (Otternes and Gunnarstul) are significantly older and may be genetically related to an earlier event of anorogenic magmatism in the region at c. 1120 to 1150 Ma.

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Introduction

Granitic intrusions in the continental crust are important indicators of tectonic regimes in the past, as well as potential sources of information on the composition and history of deep crustal protoliths. Recent studies on Sveconorwegian granites from South Norway have shed light on the timing of anorogenic and shortening stages of the evolution of this orogenic belt (Bingen & van Breemen 1998, Bingen et al. 1993), and on the nature and distribution of geochemical components in the deep crust at the end of the Sveconorwegian orogeny (Andersen et al. 1994, 2001a, Andersen 1997, Andersen & Knudsen 2000). Whereas most granitic rocks in southern Norway contain clear evidence of tectonic deformation (granitic gneiss, augen gneiss), a distinct group of granites is unfoliated or has only a weak and non-persistent foliation. These granites have anorogenic geochemical signatures, and are traditionally known as 'post-tectonic' or 'postkinematic' Sveconorwegian granites. Their Sr, Nd, Pb, and Hf isotope systematics suggest that they formed by mixing of a mantle-derived magma with regionally defined crustal components (Andersen 1997, Andersen et al. 2001a, 2002b).

In the present study, SIMS U-Pb zircon ages and internal Pb-Pb isochron ages are presented for a selection of inferred late- to post-tectonic Sveconorwegian granites from South Norway. The aim of the study has been to determine the timing of anorogenic granitic magmatism in the different parts of South Norway, and to search for evidence of inherited zircons in these granites for information about the source region of granitic magmas in the deep crust.

Geologic setting

The Precambrian areas of South Norway form part of the Southwest Scandinavian Domain of the Baltic Shield (Gaál & Gorbatschev 1987). The area of interest in this paper is restricted to the east by the Phanerozoic Oslo Rift, and to the west and northwest by the Caledonian nappes. Whereas the oldest dated rocks in South Norway west of the Oslo Rift are c. 1.60-1.66 Ga orthogneisses (Jacobsen & Heier 1975, Andersen et al. 2001b, Ragnhildstveit et al. 1994), indirect evidence from age distributions of detrital zircons in metasedimentary rocks (Knudsen et al. 1997b, Birkeland et al. 1997, Haas et al. 1999, Bingen et al. 2001) and radioisotope data from metasedimentary (Andersen et al. 1995, Knudsen et al. 1997a) and granitic rocks (Andersen et al. 1994, 2001a, Andersen 1997) point to the existence of protoliths of 1.7-1.9 Ga age, i.e. to rocks whose ages are similar to intrusions of the Transscandinavian Igneous Belt (TIB) and the Svecofennian domain of south and central Sweden (e.g. Åhäll & Larson 2000). In parts of South Norway (e.g. the Bamble and Telemark sectors, see below), the tectonometamorphic habit is largely a result of processes during the Sveconorwegian (i.e. Grenvillian) orogeny (1.2-0-9 Ga).

South Norway is a mosaic of crustal domains separated by Sveconorwegian shear zones (Fig. 1). These domains have traditionally been referred to as sectors, a non-genetic term which is retained here, in the absence of a generally accepted tectonic terrane model for the area (see discussion by Haas et al. 1999 and Andersen et al. 2001b). The medium- to high-grade metamorphic *Bamble sector* (Fig 1) consists of metasedimentary rocks and amphibolite intruded by

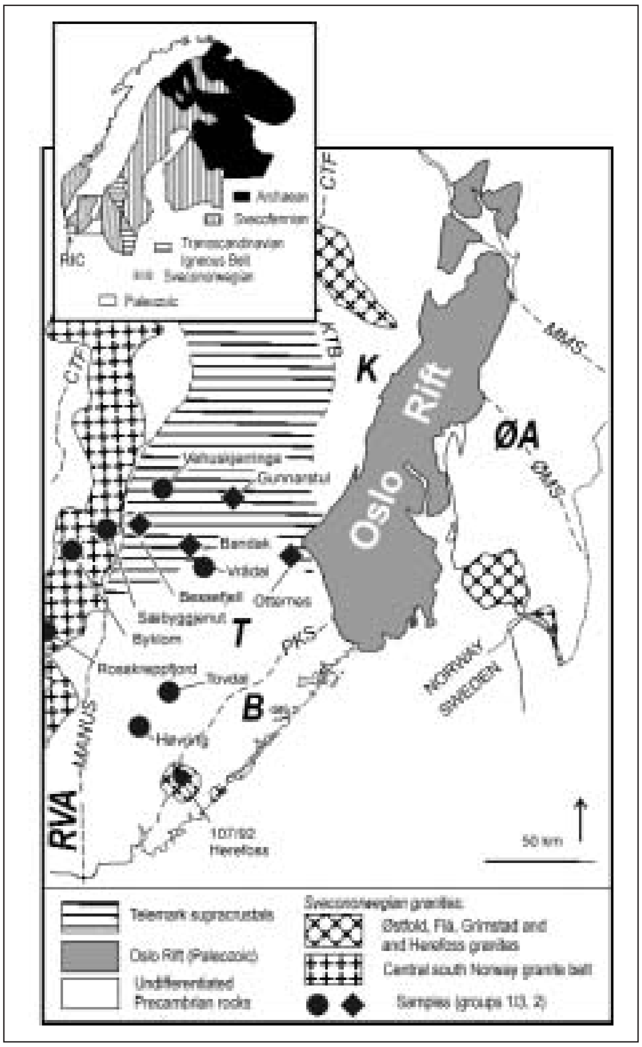

Fig. 1. Simplified geologic map of South Norway, showing main structural division of the Precambrian crust and position of intrusions sampled for dating. Sectors (large font): *B:* Bamble sector, *T:* Telemark sector, *K:* Kongsberg sector, *ØA:* Østfold-Akershus sector, *RVA:* Rogaland-Vest Agder sector. Shear zones (broken lines): PKS: Kristiandsand-Porsgrunn, MANUS: Mandal-Ustaoset, KTB: 'Kongsberg-Telemark boundary', MMS: Mjøsa-Magnor, ØMS: Ørje mylonite zone, CTF: Caledonian thrust front. Inset: Major crustal domains of the Baltic Shield. RIC: Rogaland Igneous Complex.

Sveconorwegian (1.2-0.9 Ga) gabbro, granite, and charnockite (Starmer 1985, Padget 1990, de Haas 1992, 1997). Granodioritic to tonalitic gneiss lenses having possible tectonic contacts with metasupracrustal rocks, constitute a minor but characteristic component of the Bamble sector (Andersen et al. 2001b). The *Kongsberg sector* (including the Begna sector of Bingen et al. 2001) is generally assumed to be related to the Bamble Sector (e.g. Starmer 1985 and references therein), but granodioritic to tonalitic gneiss is more abundant, especially in the southeastern part (Jacobsen & Heier 1978, Dons & Jorde 1978). The northeastern part of the Kongsberg sector is made up of poorly known gneisses, including rocks of supracrustal as well as intrusive origin, probably older than 1.6 Ga (Nordgulen 1999). The Bamble and Kongsberg sectors are separated from the *Telemark sector* by a system of Precambrian ductile shear zones (PKS-KTB in Fig. 1), whose southern part (the Porsgrunn-Kristiansand shear zone, PKS) shows evidence of Sveconorwegian thrusting, (Andresen & Bergundhaugen 2001, Bergundhaugen 2002). The Telemark sector is characterized by granitic gneiss of ambiguous origin in the south (Kleppe 1980, Falkum 1998) and a well-preserved, low-grade supracrustal sequence (Telemark supracrustal rocks) to the north (Dons 1960, Sigmond et al. 1997). The oldest part of the supracrustal sequence is the c. 1.5 Ga Rjukan Group rhyolite (Dahlgren et al. 1990) that was deposited in one or more extensional basins possibly related to a Mesoproterozoic continental rift (Sigmond et al. 1997). The Mandal-Ustaoset shear zone (*MANUS* line) separates the Telemark sector from the *Rogaland-Vest Agder sector*, which consists mainly of granodioritic to granitic gneiss and foliated to massive granite, with minor amounts of metasupracrustal rocks (Sigmond 1975, 1998, Bingen et al. 1993, de Haas et al. 1999), and which contains voluminous c. 930 Ma anorogenic anorthositic and mafic intrusions (the Rogaland Igneous Complex (RIC in inset to Fig. 1), Schärer et al. 1996).

So far, four geochemically and chronologically defined associations of Sveconorwegian granitic intrusions have been recognized in South Norway. Bingen & van Breemen (1998) identified three regionally important groups of deformed granitic intrusions, the *Gjerstad, Feda*, and *Fennefoss* augen gneiss suites. The Gjerstad suite has an anorogenic geochemical character and was emplaced in the time interval 1.19 to 1.12 Ga. Gjerstad suite intrusions are present in the Bamble, Telemark, and Rogaland – Vest Agder sectors (Hagelia 1989, Kullerud & Machado 1991, Andersen et al. 1994, Zhou et al. 1995, Simonsen 1997). The c. 1.05 Ga Feda suite is restricted to the Rogaland – Vest Agder sector and consists of high-K calc-alkaline granitoids, a composition which suggests that active subduction took place off SW Norway at this time. The c. 1035 Ma Fennefoss suite in the Telemark sector has a transitional geochemical signature between orogenic and anorogenic (Bingen & van Breemen 1998).

In addition to the deformed granitoids characterized by Bingen & van Breemen (1998), massive or weakly foliated Sveconorwegian granites exist in all crustal sectors of South Norway, comprising a fourth 'suite' of intrusions (Killeen & Heier 1973, Andersen 1997, Andersen et al. 2001a). They include two large batholiths, the Østfold (or Iddefjord) granite (which is a northern extension of the Båhus Batholith of SW Sweden) east of the Oslo Rift and the Flå granite in the Kongsberg sector. Granitic intrusions also make up a large N-S trending belt west of the Mandal–Ustaoset shear zone (the 'Central south Norway granite belt' in Fig. 1), as well as numerous smaller intrusive bodies in the Telemark and Bamble sectors. Some late- to post-kinematic granites have been dated by U-Pb or Pb-Pb systematics (Båhus: 922±5 Ma, Eliasson & Schöberg 1991; Flå: 928±3 Ma, Nordgulen et al.

1997; Herefoss: 926±8 Ma, Andersen 1997; Grimstad: 989±8 Ma, Kullerud & Machado 1991; Lyngdal: 946±5 Ma; Hidra charnockite: 933±3 Ma, Farsund: 946±7 Ma, Pasteels et al. 1979), or by whole-rock Rb-Sr isochrons (e.g. Pedersen & Falkum 1975, Killeen & Heier 1975, Kleppe 1980, Pedersen & Konnerup-Madsen 2000), but several remain undated.

Material studied

The samples dated in the present study were selected from the collection of Andersen et al. (2001a); field and petrographic data on the individual samples analysed are summarized in Appendix 1. Most samples appear unfoliated on outcrop and hand-specimen scale, notable exceptions being the samples from Herefoss and Otternes. Parts of the Herefoss granite have a well-developed foliation parallel to the intrusive contacts, which has been shown to be related to flow during emplacement (Elders 1963). On the other hand, the foliation in the Otternes sample may have formed in response to tectonic deformation (see discussion below).

Based on the radioisotopic and trace element signatures of late Sveconorwegian granites from South Norway, Andersen et al. (2001a) defined three different petrogenetic groups of late- to post-tectonic granites.

Group 1: Granites belonging to this group crop out throughout South Norway, including north and central Telemark. These rocks are characterized by strontium contents above 150 ppm Sr, $^{87}Rb/^{86}Sr < 5$, $^{87}Sr/^{86}Sr_{0.93Ga} < 0.710$ and $\varepsilon_{Nd}<0$. A crustal component involved in their petrogenesis has resided in a moderately LILE-enriched environment in the continental crust since 1.7-1.9 Ga. Group 1 includes the Østfold (Båhus), Flå and Herefoss granites, the intrusions of the central-south Norway Granite Belt, and several intrusions in the Telemark sector. Samples from three intrusions from the Rogaland-Vest Agder sector (Rosskreppfjord, Byklom, Sæbyggjenut) and a sample from the Herefoss granite (Andersen 1997) have been included in the present U-Pb SIMS study. In addition, two Group 1 granites from the Telemark sector (Vehuskjerringa, Vrådal) have been dated by internal Pb-Pb isochrons.

Group 2: Granites belonging to group 2 are restricted to the central Telemark sector. They are characterized by less than 150 ppm Sr, $^{87}Rb/^{86}Sr > 5$, $^{87}Sr/^{86}Sr_{0.93Ga} > 0.710$ and $\varepsilon_{Nd}<0$. The Group 2 'Low Sr concentration' granites are, in general, spatially associated with low-Sr metarhyolite of the Telemark supracrustal sequence (e.g. Menuge & Brewer 1996); a c. 1.5 Ga crustal component with anomalously low Sr concentration has contributed to the petrogenesis of these granites (Andersen et al. 2001a). In the present study, four Group 2 intrusions (Bessefjell, Bandak, Gunnarstul, and Otternes granites) have been analysed.

Group 3: Among the intrusions studied by Andersen et al. (2001a), only one granite has a juvenile geochemical signature, suggesting insignificant contributions of old crustal material. This intrusion (the Tovdal granite) was accordingly placed in a group of its own. It is characterized by $^{87}Sr/^{86}Sr_{0.93Ga} < 0.705$ and $\varepsilon_{Nd}>0$.

Analytical methods

The present study is based on 5-8 kg samples of homogenous granite. The samples were crushed to a grain-size of less than 250 μm using a jaw crusher and a percussion mill. *Zircons* were separated from the <250 μm fraction by a combination of Wilfley-table washing, heavy liquid separation (1,1,2,2- tetrabromoethane and diiodomethane) and magnetic separation. The final, non-magnetic zircon fraction was then purified by hand picking under a binocular microscope, and selected grains were mounted on doubly adhesive tape, cast in epoxy and polished for the ion microprobe study. U-Pb dating of zircons was performed in the NORDSIM laboratory located at the Swedish Museum of Natural History in Stockholm during 1998-2000, using a CAMECA IMS1270 ion microprobe; analytical conditions and data reduction procedures are described by Whitehouse et al. (1997, 1999). Corrections for common lead were made using measured $^{206}Pb/^{204}Pb$ ratios and present-day average crustal lead of Stacey & Kramers (1975). This correction works well for moderate common-lead levels, but becomes problematic when common-^{206}Pb exceeds ca. 2 percent of total ^{206}Pb. U-Pb data are given in Table 1.

Additional separates of rock-forming minerals for the Pb-Pb isochron study were prepared by a combination of heavy liquid and magnetic separation, followed by hand picking. Lead was separated and analysed by methods described by Andersen (1997). Whole-rock and K-feldspar lead isotope data are taken from Andersen et al. (2001a), data are given in Table 2. All geochronological calculations have been made using Isoplot/Ex version 2.32 (Ludwig 2000). The preferred age estimate for multi-grain populations of concordant and equivalent zircons is the 'concordia age' of the population, which is a weighted average incorporating information from the $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ ratios in all grains (Ludwig 2000). Most of the samples have more complex zircon populations that lost lead late in the history of the sample. Such lead loss may have occurred during the Caledonian orogeny, or during the Late Paleozoic rifting event which affected the southwestern parts of the Baltic Shield or in recent time. Regression lines for zircon populations which did not give meaningful, unconstrained lower intercepts, were accordingly calculated with a forced intercept of 250 ± 250 Ma.

Results

Zircons

Zircons in the study samples have moderately elongated, prismatic habits. BSE images reveal more or less strongly developed oscillatory 'magmatic' zoning (Fig. 2b,d,e), as well as more complex internal structures, suggesting the presence of different types of xenocrystic cores (Fig. 2a,c,e). Some grains also have thin, BSE-bright overgrowths (Fig. 2c)

Table 1. SIMS U-Pb data on zircons.

Spot 1	$^{207}Pb/^{206}Pb$	±σ %	$^{207}Pb/^{235}U$	±σ %	$^{206}Pb/^{238}U$	±σ %	ρ Error correlation	$^{208}Pb/^{232}Th$	±σ %	Disc. %	U ppm	Th ppm	Pb ppm	Th/U	$^{206}Pb/^{204}Pb$ meas.	f_{206}%	$^{207}Pb/^{206}Pb$ Ma	±σ Ma	$^{207}Pb/^{235}U$ Ma	±σ Ma	$^{206}Pb/^{238}U$ Ma	±σ Ma	$^{208}Pb/^{232}Th$ Ma	±σ Ma
Tovdal																								
n442-01a	0.07034	0.8	1.5175	1.6	0.15646	1.4	0.87	n.d.		0	182	23	32	0.13	35160	0.1	938	17	937	10	937	12		
n442-02a	0.07226	1.5	1.5304	2.0	0.15361	1.2	0.66	n.d.		-6	249	34	43	0.13	10500	0.2	993	31	943	12	921	10		
n442-03a	0.07099	0.9	1.4726	1.7	0.15045	1.4	0.87	n.d.		-3	127	49	23	0.38	33370	0.1	957	18	919	10	903	12		
n442-04a	0.07082	0.7	1.5686	1.6	0.16065	1.4	0.90	n.d.		0	139	39	26	0.28	106910	0.0	952	14	958	10	960	13		
n442-05a	0.07070	1.0	1.4936	1.8	0.15322	1.5	0.84	n.d.		0	79	18	14	0.23	>106	0.0	949	20	928	11	919	13		
n442-06a	0.06957	1.9	1.4987	2.8	0.15624	2.0	0.76	n.d.		0	43	33	9	0.78	15700	0.1	916	40	930	17	936	17		
n442-07a	0.06998	1.3	1.5066	2.9	0.15615	2.6	0.92	n.d.		0	54	68	12	1.28	15320	0.1	928	26	933	18	935	22		
n442-08a	0.06956	1.0	1.5293	1.8	0.15945	1.5	0.82	n.d.		1	84	20	15	0.24	>106	0.0	915	21	942	11	954	13		
Herefoss																								
n440-02a	0.07004	0.8	1.6631	1.4	0.17222	1.1	0.87	n.d.		9	124	82	27	0.67	9160	0.2	929	16	995	9	1024	10		
n440-03a	0.06858	1.1	1.5054	2.5	0.15921	2.2	0.93	n.d.		3	53	67	12	1.26	9120	0.2	886	23	933	15	952	20		
n440-04a	0.07026	0.5	1.5980	1.1	0.16496	1.0	0.91	n.d.		3	258	25	47	0.10	79550	0.0	936	10	969	7	984	9		
n440-04b	0.07046	0.5	1.5694	1.1	0.16155	0.9	0.87	n.d.		1	211	126	43	0.60	52300	0.0	942	11	958	7	965	8		
n440-05a	0.06922	1.1	1.5560	2.3	0.16304	2.0	0.90	n.d.		4	76	62	16	0.81	22330	0.1	905	23	953	14	974	18		
n440-06a	0.06953	1.0	1.5340	1.6	0.16002	1.3	0.87	n.d.		2	83	59	17	0.71	20710	0.1	914	20	944	10	957	11		
n440-09a	0.07079	1.0	1.4766	1.9	0.15127	1.7	0.89	n.d.		-2	57	21	10	0.37	19150	0.1	951	21	921	12	908	14		
n440-10a	0.07072	1.6	1.5969	2.7	0.16377	2.2	0.87	n.d.		0	26	36	6	1.36	9590	0.2	949	33	969	17	978	20		
n440-11a	0.07072	1.8	1.5518	3.0	0.15914	2.4	0.80	n.d.		0	28	35	6	1.27	>1e6	0.0	949	37	951	19	952	21		
n440-12a	0.07062	1.3	1.5350	2.3	0.15765	1.9	0.83	n.d.		0	72	82	16	1.14	68120	0.0	946	27	945	14	944	17		
Høvring																								
n716-01a	0.06835	1.8	1.3374	2.6	0.14192	1.9	0.73	0.04346	4.1	-3	135	139	27	1.03	1970	1.0	879	37	862	15	855	15	860	35
n716-02a	0.09160	0.5	2.5700	2.0	0.20350	1.9	0.96	0.06453	4.0	-20	319	105	78	0.33	3028	0.6	1459	10	1292	15	1194	21	1264	48
n716-04a	0.07011	2.3	1.3287	3.0	0.13744	1.9	0.63	0.04329	4.0	-12	116	117	22	1.01	923	2.0	932	47	858	17	830	15	857	34
n716-05a	0.07085	1.2	1.3622	2.3	0.13946	1.9	0.85	0.04787	3.9	-12	461	422	89	0.91	921	2.0	953	25	873	13	842	15	945	36
n716-06a	0.07099	0.6	1.4575	2.0	0.14892	1.9	0.95	0.04561	4.0	-7	259	84	45	0.33	13965	0.1	957	13	913	12	895	16	901	35
Sæbyggjenut																								
n447-01a	0.07029	1.3	1.4867	2.1	0.15340	1.7	0.93	n.d.		0	231	217	49	0.94	3460	0.5	937	27	925	13	920	14		
n447-03a	0.07208	2.6	1.5363	4.4	0.15458	3.5	0.82	n.d.		0	30	31	6	1.05	>106	0.0	988	52	945	27	927	30		
n-705-01a	0.07293	1.5	1.5240	3.4	0.15156	3.0	0.89	0.04672	7.1	-11	39	36	8	0.94	4205	—	1012	31	940	21	910	25	923	64
n705-02a	0.06715	2.3	1.3094	3.8	0.14141	3.0	0.79	0.04262	7.0	1	57	50	11	0.88	2948	0.6	843	48	850	22	853	24	844	58
n705-03a	0.06950	1.5	1.3488	3.4	0.14076	3.1	0.90	0.04501	7.0	-8	117	110	23	0.94	7042	0.3	913	30	867	20	849	24	890	61
n705-05a	0.07204	2.3	1.3657	3.8	0.13749	3.0	0.79	0.04183	7.1	-17	134	108	24	0.81	1386	1.4	987	46	874	22	830	23	828	58
n705-06a	0.06838	2.0	1.2305	3.6	0.13051	3.0	0.83	0.04121	7.0	-11	80	89	15	1.10	3663	0.5	880	42	815	21	791	22	816	56
n705-07a	0.07108	1.5	1.3814	3.4	0.14095	3.0	0.89	0.04425	6.9	-12	176	180	34	1.02	9107	0.2	960	31	881	20	850	24	875	59
Rosskreppfjord																								
n443-01a	0.07321	1.1	1.7126	2.3	0.16965	2.0	0.97	n.d.		0	265	24	50	0.09	1880	1.0	1020	22	1013	14	1010	19		
n443-02a	0.06514	0.9	0.8951	1.2	0.09966	0.8	0.81	n.d.		-21	2191	935	263	0.43	3460	0.5	779	19	649	6	612	5		
n443-03a	0.07394	0.7	1.7394	1.4	0.17062	1.2	0.87	n.d.		0	151	89	32	0.59	>106	0.0	1040	14	1023	9	1016	11		
n443-03b	0.07349	0.8	1.7102	1.5	0.16877	1.3	0.88	n.d.		0	154	109	33	0.71	29260	0.1	1028	16	1012	10	1005	12		
Byklom																								
n441-01a	0.07259	1.4	1.6806	3.6	0.16791	3.3	0.92	n.d.		0	40	79	11	1.98	>1e6	0.0	1003	28	1001	23	1001	31		
n441-02a	0.07330	1.4	1.6070	2.1	0.15900	1.5	0.76	n.d.		-5	96	177	25	1.83	33320	0.1	1022	28	973	13	951	14		
n441-04a	0.07168	0.7	1.7103	1.3	0.17306	1.2	0.88	n.d.		3	119	75	26	0.63	88420	0.0	977	13	1012	9	1029	11		
n441-06a	0.07106	0.8	1.6332	1.0	0.16669	0.7	0.94	n.d.		2	444	925	127	2.08	5420	0.3	959	16	983	6	994	6		
n441-08a	0.07257	2.4	1.5920	3.2	0.15911	2.0	0.86	n.d.		-2	61	108	16	1.78	2200	0.8	1002	49	967	20	952	18		
n441-09a	0.07119	1.1	1.6146	1.9	0.16450	1.6	0.91	n.d.		0	152	269	40	1.77	11230	0.2	963	22	976	12	982	15		
n441-10a	0.07199	1.5	1.5254	1.9	0.15369	1.2	0.92	n.d.		-5	318	541	80	1.70	1150	1.6	986	30	941	12	922	10		
n441-11a	0.07211	1.9	1.6298	3.0	0.16392	2.2	0.85	n.d.		0	41	51	10	1.26	7400	0.3	989	40	982	19	979	20		
n441-12a	0.07274	1.2	1.5970	1.7	0.15922	1.3	0.73	n.d.		-3	141	150	31	1.07	55680	0.0	1007	25	969	11	952	11		
n441-13a	0.07310	3.8	1.6260	4.2	0.16132	1.7	0.73	n.d.		-2	409	860	113	2.10	3960	0.5	1017	77	980	26	964	15		
Bessefjell																								
n448-01a	0.06937	0.9	1.5944	1.6	0.16669	1.4	0.88	n.d.		7	94	137	23	1.47	34900	0.1	910	18	968	10	994	13		
n448-02a	0.07127	1.3	1.6477	2.6	0.16767	2.3	0.90	n.d.		0	72	72	17	1.01	17230	0.1	965	26	989	17	999	21		
n448-02b	0.07096	0.7	1.5454	1.7	0.15795	1.5	0.93	n.d.		0	247	191	50	0.77	42300	0.0	956	14	949	10	945	13		
n448-04a	0.07048	2.3	1.4520	3.4	0.14942	2.5	0.92	n.d.		0	47	49	10	1.04	3210	0.6	942	48	911	20	898	21		
n448-08a	0.06771	2.2	0.9970	3.1	0.10679	2.2	0.90	n.d.		-22	270	115	36	0.43	5970	0.3	860	45	702	16	654	14		
n448-10a	0.07060	1.2	1.4787	1.9	0.15191	1.5	0.78	n.d.		-1	183	156	37	0.85	>106	0.0	946	24	922	12	912	13		
n448-12a	0.06398	1.0	0.6295	1.5	0.07136	1.1	0.90	n.d.		-40	2101	811	181	0.39	4100	0.5	741	21	496	6	444	5		
Bandak																								
n708-01a	0.08800	1.0	2.3404	3.2	0.19290	3.0	0.95	0.05160	7.2	-19	1359	625	317	0.460	689	2.72	1382	20	1225	23	1137	31	1017	71
n708-02a	0.08593	0.4	2.0681	3.0	0.17455	3.0	0.99	0.05156	6.8	-24	2029	679	420	0.335	4490	0.42	1337	8	1138	21	1037	29	1016	68
n708-03a	0.08470	0.5	1.7917	3.1	0.15343	3.0	0.99	0.04923	6.8	-32	1835	349	323	0.190	5350	0.35	1309	9	1042	20	920	26	971	65
n708-04a	0.06346	0.7	0.7600	3.1	0.08686	3.1	0.97	0.01086	7.6	-27	5312	1122	503	0.211	1140	1.64	724	16	574	14	537	16	218	16
n708-05a	0.06356	0.9	0.7684	3.2	0.08768	3.0	0.96	0.01661	8.9	-27	5723	790	547	0.138	673	2.78	727	19	579	14	542	16	333	30
n708-06a	0.06103	0.5	0.6840	3.1	0.08129	3.0	0.99	0.02325	7.4	-22	11600	875	1021	0.075	1623	1.15	640	10	529	13	504	15	464	34
n708-07a	0.09060	1.1	2.1305	3.3	0.17054	3.1	0.94	0.02742	7.4	-32	1720	1161	352	0.675	935	2	1438	20	1159	23	1015	29	547	40
n708-08a	0.09331	0.9	3.1292	3.2	0.24321	3.0	0.95	0.06905	7.1	-7	724	363	218	0.501	1545	1.21	1494	18	1440	25	1403	38	1350	92
n708-09a	0.07404	0.6	0.9796	3.1	0.09596	3.0	0.98	0.02715	6.8	-45	2666	1385	312	0.520	1819	1.03	1043	12	693	15	591	17	541	37
n708-10a	0.08105	1.1	1.0710	3.4	0.09584	3.2	0.95	0.01861	7.6	-54	1981	1649	236	0.833	808	2.31	1223	21	739	18	590	18	373	28
Gunnarstul																								
n449-01a	0.07735	0.3	2.0557	0.8	0.19276	0.8	0.97	n.d.		0	1318	850	325	0.64	2330	0.8	1130	7	1134	6	1136	8		
n449-02a	0.07703	0.4	2.0725	1.1	0.19514	1.0	0.96	n.d.		1	754	766	199	1.02	11420	0.2	1122	7	1140	7	1149	10		
n449-03a	0.07828	1.0	2.0054	1.2	0.18581	0.7	0.82	n.d.		-4	970	675	235	0.70	1090	1.7	1154	20	1117	8	1099	7		
n449-04a	0.07928	0.7	2.0605	1.3	0.18850	1.1	0.90	n.d.		-4	812	501	192	0.62	2090	0.9	1179	14	1136	9	1113	11		
n449-05a	0.07765	1.4	1.9603	1.9	0.18310	1.3	0.96	n.d.		-3	272	94	59	0.35	1670	1.1	1138	28	1102	13	1084	13		
n449-07a	0.07857	1.2	2.0132	2.0	0.18584	1.7	0.84	n.d.		-3	156	63	34	0.40	30600	0.1	1161	23	1120	14	1099	17		
n449-08a	0.07730	2.2	1.9913	3.1	0.18683	2.2	0.98	n.d.		0	152	55	34	0.36	3840	0.5	1129	43	1113	21	1104	22		
Otternes																								
n450-02a	0.07245	0.5	1.2879	0.8	0.12892	0.7	0.98	n.d.		-22	2738	1200	427	0.44	2080	0.9	999	10	840	5	782	5		
n450-03a	0.07565	1.4	1.4580	1.8	0.13979	1.1	0.97	n.d.		-22	1437	904	246	0.63	1170	1.6	1086	28	913	11	843	9		
n450-04a	0.07758	0.4	1.7422	1.0	0.16288	0.9	0.95	n.d.		-14	2653	1112	520	0.42	4400	0.4	1136	8	1024	6	973	8		
n450-06a	0.08005	0.9	2.1098	2.3	0.19115	2.2	0.95	n.d.		-2	377	151	87	0.40	8180	0.2	1198	18	1152	16	1128	22		

1: Analysed spots are identified by Nordsim laboratory log numbers

n.d.: Not determined

Table 2. Lead isotope data for minerals and whole-rocks.

	$^{206}Pb/^{204}Pb$	2σ	$^{207}Pb/^{204}Pb$	2σ	$^{208}Pb/^{204}Pb$	2σ
Byklom 083096-3						
Biotite	17.819	0.020	15.546	0.021	50.803	0.081
Magnetite	17.515	0.014	15.475	0.018	40.348	0.064
Apatite	24.830	0.020	16.046	0.019	46.044	0.073
Titanite	46.959	0.037	17.614	0.021	54.026	0.086
K-feldspar	16.772	0.013	15.429	0.018	36.415	0.058
Whole-rock[1]	17.308	0.014	15.502	0.019	39.874	0.062
Rosskreppfjord 080296-4						
Titanite	493.1	0.4	50.17	0.07	210.36	0.37
Hornblende	21.629	0.020	15.808	0.022	37.636	0.067
Magnetite	21.251	0.019	15.845	0.022	41.255	0.073
Whole-rock[1]	19.880	0.016	15.694	0.019	38.966	0.061
Vrådal 081696-1						
Biotite	19.635	0.018	15.619	0.021	40.450	0.071
Magnetite	20.994	0.019	15.719	0.021	40.223	0.071
Apatite	19.527	0.018	15.619	0.021	40.313	0.071
K-feldspar	16.717	0.015	15.383	0.015	36.173	0.015
Monazite	127.2	0.5	23.194	0.086	1312	5
Whole-rock[1]	17.478	0.014	15.506	0.019	37.674	0.059
Vehuskjerringa 072406-1						
Biotite	20.620	0.016	15.682	0.019	40.429	0.063
Pyrite	19.209	0.015	15.609	0.019	39.210	0.061
Apatite	29.814	0.023	16.331	0.020	51.827	0.081
Titanite	211.8	0.2	29.002	0.035	236.2	0.4
K-feldspar	17.047	0.014	15.409	0.020	36.451	0.061
Whole-rock[1]	18.734	0.015	15.565	0.019	38.169	0.060

[1]: Data from Andersen et al. (2001a)

and embayments (Fig. 2b). The zircons from the Tovdal granite (Fig. 2d) have more elongated, slender prismatic habits, with less complex zoning patterns than zircons from any of the other granites, and they lack cores. The most frequent type of core has near-euhedral shapes, well-developed internal oscillatory zoning (Fig. 2a, c), and is discordantly overgrown by the enclosing grains. These cores lack abraded surfaces and lack evidence of strong metamictization. Cores of this type are typical in most of the Group 1 granites; the examples shown in Fig. 2 come from Herefoss (Fig. 2a) and Rosskreppfjord (Fig. 2c). Other cores are heterogeneous, fractured, and more or less metamict (Fig. 2e). The example in Fig. 2e comes from the Sæbyggjenut granite (Group 1 intrusion from the Rogaland-Vest Agder sector), but similar cores are abundant in the Group 2 granites as well.

U-Pb data

Group 1 and 3 intrusions. Five concordant zircons and one weakly inversely discordant zircon from the Tovdal granite give a concordia age of 940 ± 10 Ma (Fig. 3a). Two normally discordant (3 and 6 %) grains give higher ages, but still within uncertainty of the concordia age. A regression line calculated for all zircons together gives an upper intercept age of 947 +21/-17 Ma, assuming a forced lower intercept at 250 ± 250 Ma. We believe the concordia age (940 ± 10 Ma) is the best estimate of the emplacement age of the Tovdal granite.

Of 14 grains analysed from the Herefoss granite, four were discarded because of high contents of common lead. The remaining 11 grains define a regression line with an upper intercept of 920 +16/-27 Ma (forced lower intercept, Fig. 3b). This age is indistinguishable from the internal Pb-Pb lead isochron age of 926 ± 8 Ma reported by Andersen (1997), which is regarded as the best estimate of the emplacement age. Apparently xenocrystic cores (Fig. 2a) cannot be distinguished from the bulk of the zircon fraction in terms of U-Pb systematics.

Four discordant zircons from the Høvring granite yield an age of 971 +63/-34 Ma, assuming lead loss at 250 ± 250 Ma (Fig. 2c). This age is indistinguishable from a Rb-Sr whole-rock isochron age of 945 ± 53 Ma reported by Pedersen (1981) and Pedersen & Konnerup-Madsen (2000). A single, strongly discordant but distinctly older grain lies on a discordia line to an upper intercept at 2082 +340/-200 Ma (assuming lower intercept at 971 Ma) and must have been inherited from an Early Proterozoic source.

Among the intrusions from the Rogaland-Vest Agder sector, the Sæbyggjenut granite gives an upper intercept age of 959 +50/-32 Ma, based on seven concordant to normally discordant zircons and a forced intercept at 250 ± 250 Ma (Fig. 3d). Two additional analyses were discarded due to high common lead contents. Three of the four zircons analysed from the Rosskreppfjord intrusion are concordant at 1020 Ma, but with a high MSWD. The fourth zircon is strongly discordant, a regression line gives an upper intercept of 1036+23/-22 Ma with a Paleozoic unconstrained lower intercept (Fig. 3e); this is regarded as the best estimate of the emplacement age. The Byklom intrusion gives an upper intercept age indistinguishable from Sæbyggjenut, which may be given as 979 +9/-12 Ma, with a negative intercept, or 970 +14/-18 Ma with a forced intercept (Fig. 3f). The former is accepted as the best estimate of its emplacement age.

Group 2 granites. The Bessefjell granite was dated by a Rb-Sr whole-rock isochron to 904 ± 16 Ma (Killeen & Heier 1975, recalculated using $\lambda^{87}Rb=1.42 \times 10^{-11}\ a^{-1}$). Seven zircons, ranging from concordant to 40 % discordant define a regression line with intercepts at 209 Ma and 940 ± 19 Ma (Fig. 4a), which is regarded as the emplacement age.

The Bandak granite (Fig. 4b) has a zircon population dominated by zircons whose data plot along a poorly defined lead-loss line from an upper intercept at c. 1500 Ma to a Paleozoic lower intercept. Some grains fall to the left of this line, indicating additional lead loss, or growth of new zircon in Sveconorwegian time. Two of these plot near a possible lead-loss line from a late Sveconorwegian upper intercept, suggesting that crystallization of new zircon in Sveconorwegian time did actually take place. The Bandak granite was dated by a Rb-Sr whole-rock isochron age of 1002 ± 76 Ma (MSWD=0.57, $^{87}Sr/^{86}Sr_i$ = 0.765 ± 0.021) by

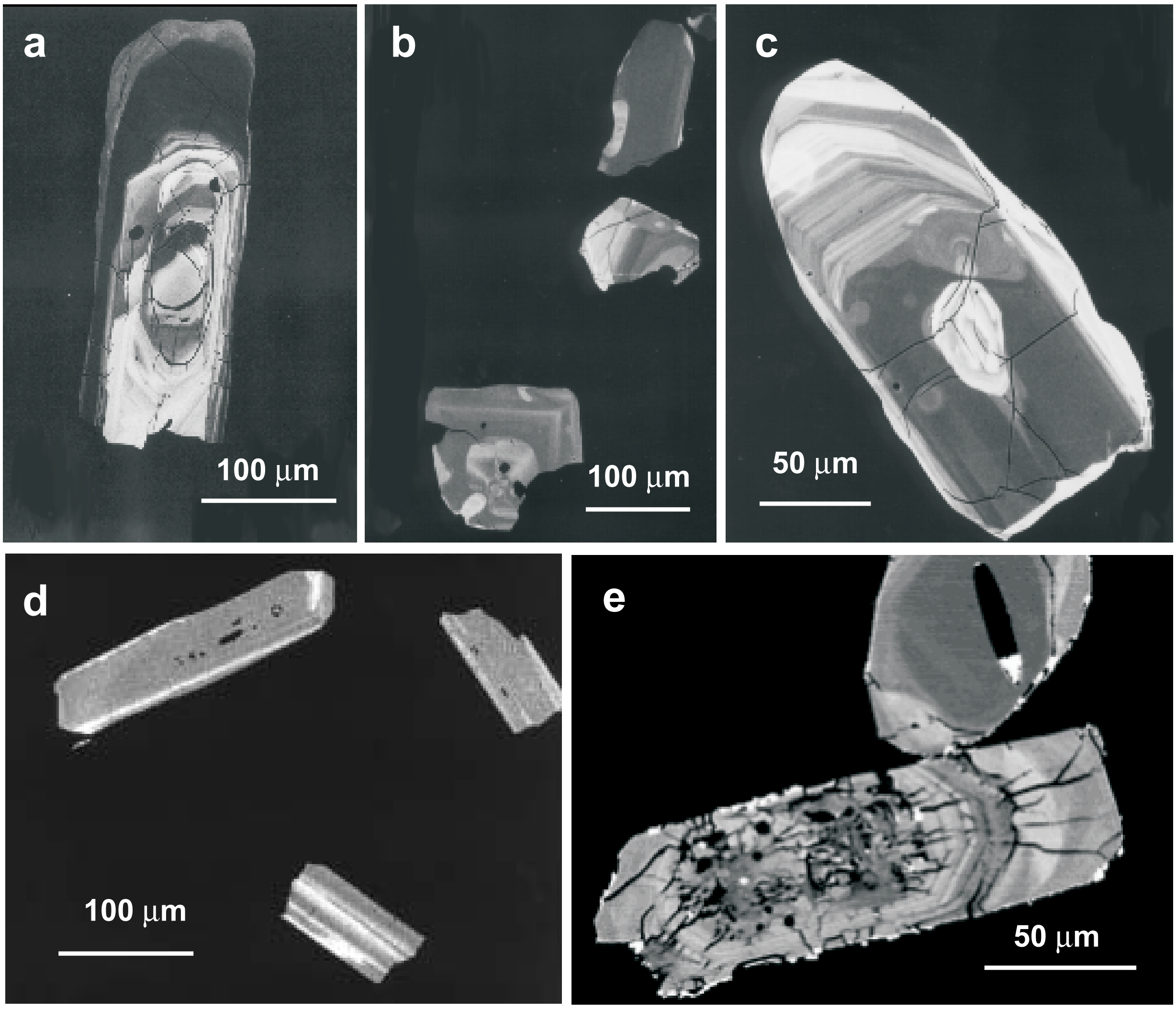

Fig. 2. Electron backscatter photomicrographs of zircons from late- and post-tectonic Sveconorwegian granites in South Norway. *(a)* Zircon from the Herefoss granite, containing an apparent xenocrystic core. Both core and enclosing grain have oscillatory magmatic zoning. Ages of zircon cores in the Herefoss granite are indistinguishable from enclosing zircon. *(b)* Fragments of zircons from Byklom granite, characterized by zircons with simple magmatic zoning and absence of cored grains. Lowermost grain has a small, BSE-bright embayment (bottom left), probably due to late- or post-magmatic hydrothermal alteration. *(c)* Zircon from Rosskreppfjord granite, with magmatic zoning and a thin BSE-bright overgrowth. BSE-bright central area is part of the magmatic zoning and not a xenocrystic core. BSE-bright rim is too thin for dating and probably formed by late- or post-magmatic hydrothermal alteration. *(d)* One terminated zircon and two fragments from Tovdal granite. Zircons from this sample differ from all others studied by a more slender habit, simple magmatic zoning with few zones, and absence of xenocrystic cores. *(e)* Two zircons from Sæbyggjenut granite, illustrating two morphological types in this granite: One type (upper grain) is internally homogeneous or lacks magmatic zoning. This particular grain contains a low-atomic number inclusion (BSE dark). The other type (lower grain) contains a heterogeneous, fractured and partly metamict core. White spots at grain surfaces and along fractures are parts of the gold coating which could not be removed by polishing after the SIMS session.

Kleppe (1980). Its Sveconorwegian Sr, Nd and Pb isotope composition resembles the other Sveconorwegian Group 2 granites and is distinct from that of the Mid Proterozoic Tinn granite (Andersen et al. 2001a). A population of undated single zircons give hafnium isotope model ages similar to those observed in other Group 2 granites (Andersen et al. 2002b). The Bandak granite is thus interpreted as a Sveconorwegian granite which has suffered heavy contamination by Mid Proterozoic crustal material, at least locally. The zircons analysed in the present study do not give new information on its emplacement age.

Seven zircons from the Gunnarstul granite, ranging from concordant to 4 % discordant, give upper intercept ages of 1133 +6/-6 Ma with no defined lower intercept, or 1134 ± 21 Ma with a forced intercept at 250 ± 250 Ma (Fig. 4c). An early Sveconorwegian age is also obtained from the Otternes granite, where six variably discordant zircons define a poorly defined regression line with a model 2 (Ludwig 2000) age of 1233 ± 90 Ma (Fig. 4d).

$^{208}Pb/^{232}Th$ ratios were only reported in analyses done in 2000. Data for the Høvring, Sæbyggjenut, and Bandak granites are presented in Table 1. $^{208}Pb/^{232}Th$ ages are, in general,

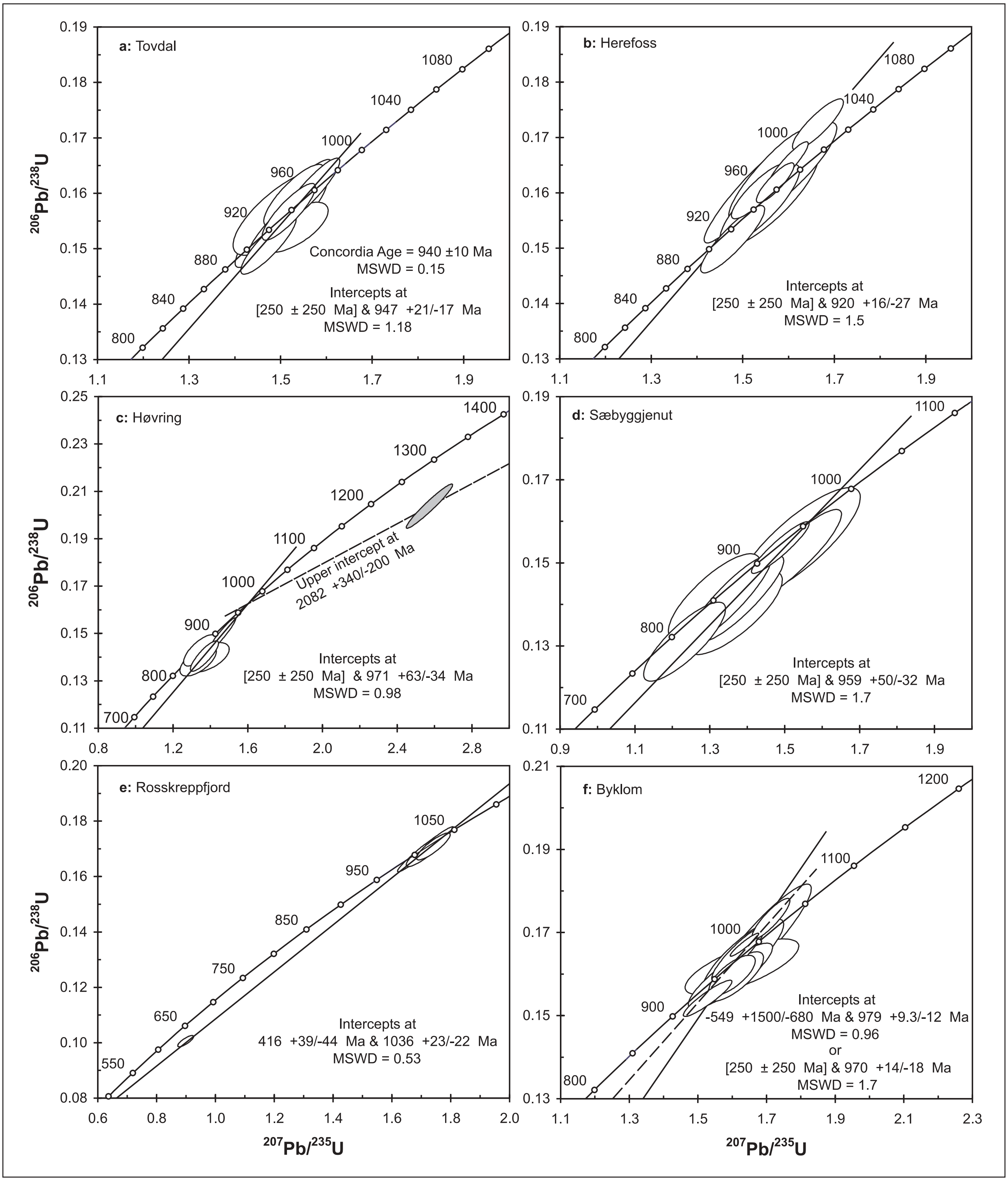

Fig. 3. Concordia diagrams for Group 1 and Group 3 granites. Single SIMS spot analyses are plotted with 2 σ error ellipses.

indistinguishable from the corresponding $^{206}Pb/^{238}U$ ages. Some of the grains from the Bandak granite have significantly lower $^{208}Pb/^{232}Th$ ages, however, which is a characteristic feature in zircons that have lost a significant amount of their lead in a late event (e.g. Andersen 2002).

Lead isochrons

Lead isotope data for rock-forming minerals and their corresponding whole-rocks are given in Table 2, and isochron diagrams are shown in Fig. 5. None of the study samples has a perfect fit to mineral isochrons; ages have therefore been cal-

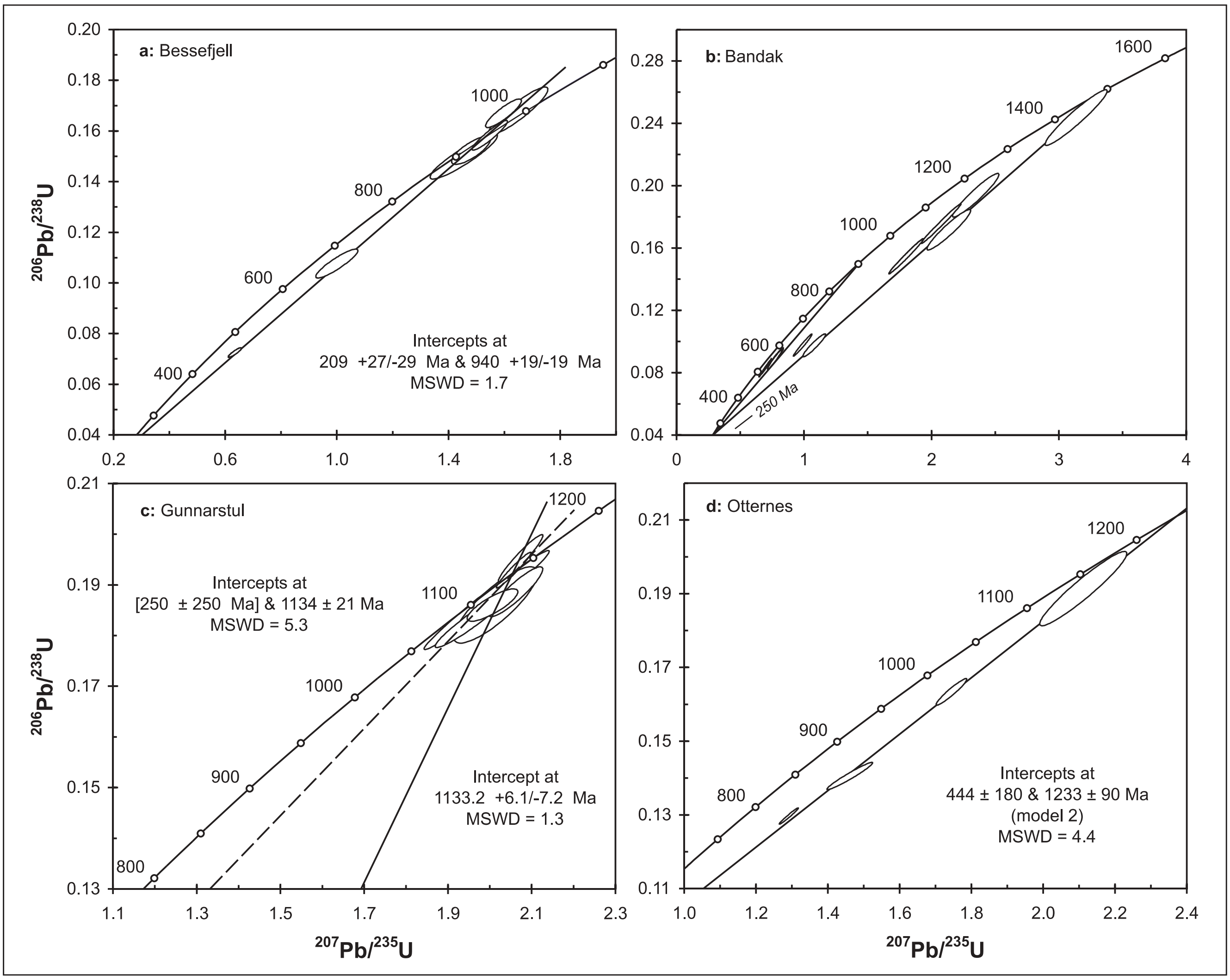

Fig. 4. Concordia diagrams for Group 2 granites. Single SIMS spot analyses are plotted with 2 σ error ellipses.

culated using Model 2 of Isoplot (Ludwig 2000), which assigns equal weight and uncorrelated errors to all points. Three of the four isochrons in Fig. 5 (i.e. Fig. 5b,c,d) are strongly controlled by a single, highly radiogenic mineral, which is titanite in the case of the Byklom, Rosskreppfjord, and Vehuskjerringa granites, and monazite in the Vrådal granite. The *Byklom* granite (Fig. 5a) has less total spread in lead isotope composition than the other three samples, and both apatite and titanite have moderately high radiogenic compositions. The age calculated from the line (983±68 Ma) has a large uncertainty, reflecting the moderate spread, but it agrees with the 980±18 Ma SIMS zircon U-Pb age (Fig. 3f). The *Rosskreppfjord* granite has a larger internal spread in radiogenic isotope composition; the age of 1009 ± 10 Ma is heavily controlled by titanite at $^{206}Pb/^{204}Pb$=493 (Fig. 5b), and agrees within uncertainty with the SIMS U-Pb age (Fig. 3e). The magnetite falls slightly off the isochron line. It should be noted that a K-feldspar separate was not analysed from this granite.

The agreement between U-Pb zircon and Pb-Pb mineral isochron ages for the Byklom and Rosskreppfjord granites is good, suggesting that an internal Pb-Pb isochron based on rock-forming minerals may in fact give a good representation of the emplacement age of a granite, in cases where no metamorphic recrystallization has taken place. Where a granite has been metamorphosed, a Pb-Pb mineral isochron probably reflects the metamorphic recrystallization age, however, rather than the emplacement age (e.g. Andersen et al. 2002a).

Zircon U-Pb ages are not available from the Vrådal and Vehuskjerringa granites, which are Group 1 intrusions from central Telemark. The *Vrådal* intrusion (Sylvester 1964, 1998) gives a Pb-Pb isochron age of 939 ± 20 Ma (Fig. 5c), based on seven points, but which is largely controlled by radiogenic monazite. Kleppe (1980) reported a Rb-Sr whole-rock isochron age of 895±38 for the Vrådal intrusion, whereas Sylvester (1964) published mineral Rb-Sr ages at 877±27 and 908±20 Ma (recalculated to $\lambda^{87}Rb$ =1.42 x 10^{-11} a^{-1}). The present Pb-Pb isochron age of 939 ± 20 Ma is assumed to be the best estimate of the intrusion's age. The *Vehuskjerringa*

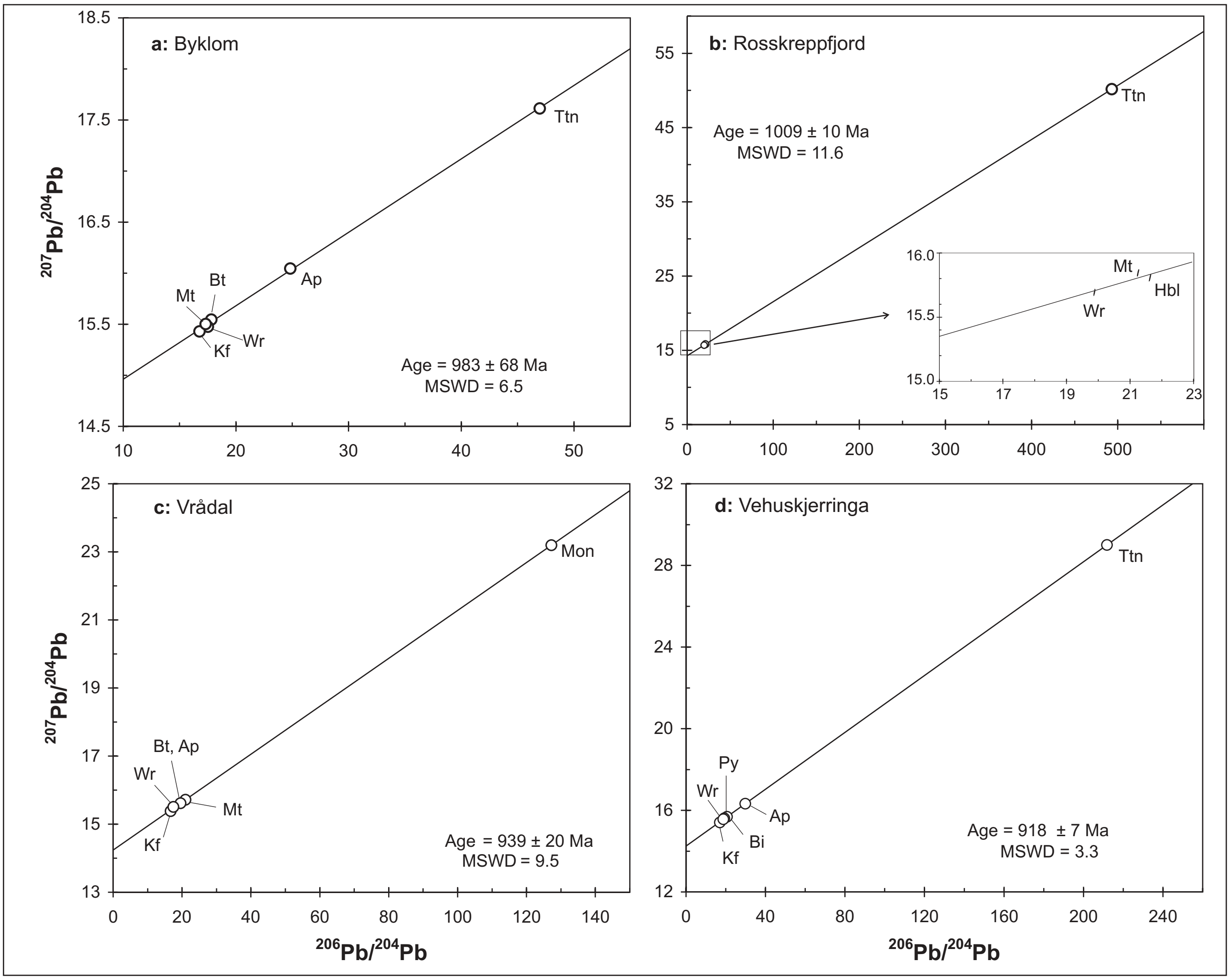

Fig. 5. Internal Pb-Pb isochrons. The circles exaggerate the actual size of the error ellipses. Isochrons are calculated using regression model 2 of Isoplot (Ludwig 2000).

granite yields a better-fitted reference line (MSWD=3.3, Fig. 5d) and an age of 918 ± 7 Ma, which is assumed to represent its emplacement age.

Discussion

Timing of Sveconorwegian granitic magmatism

The present age data (Table 3) are summarized in Fig. 6, which also includes published U-Pb and Pb-Pb ages or ranges of ages for Sveconorwegian magmatic rocks from adjacent parts of South Norway. The compilation of published data in Fig. 6 indicates that 'post-tectonic' granites from across South Norway concentrate in the period 920-950 Ma, which also includes the emplacement of anorthosite, charnockite, and granite in Rogaland (Pasteels et al. 1979) and granites in Caledonian basement windows and the Western Gneiss Region (Corfu 1980). Of the 11 granites dated in the present study, five were emplaced within this age interval (Tovdal, Herefoss, Vrådal, Vehuskjerringa, Bessefjell), and two other intrusions (Høvring and Sæbyggjenut) have age uncertainty ranges extending into this period, although their nominal ages are older. The emplacement of the juvenile Tovdal granite at 940 Ma indicates that mafic magmas were ponded in the lower crust at this time. Although the Tovdal granite is the only known example in the region of a late Sveconorwegian granite with a juvenile geochemical signature, the presence of Group 1 granites containing a well-defined mantle-derived component in all sectors of South Norway suggests that the mafic 'underplating' event was of regional importance.

By contrast, the granites from the Rogaland-Vest Agder sector dated in this study, and possibly the Høvring granite from the southwestern part of the Telemark sector, yield emplacement ages between the main group of post-tectonic granites and the age of the synorogenic granites of the Feda suite. A granitic intrusion from Jølster in the Western Gneiss Region dated by Skår (1998) also falls within this interval. The Grimstad granite in the Bamble sector has been dated at 989 ± 9 Ma by U-Pb on zircons (Kullerud & Machado 1991), but insufficient data have been published for the significance of this date to be evaluated. Onset of Group 1

Table 3. Summary of geochronological results.

		Ages (Ma)							Mineral			
		Concordia		Intercept ages					isochron age		MSWD	Comment
Sample	Sector		2σ	Upper	+2σ	-2σ	Lower[1]	2σ		2σ		
Group 1 and 3 granites												
Tovdal	T	940	10	947							0.15	
					21	17	[250]	[250]			1.18	
Herefoss	B			920	16	27	[250]	[250]			1.5	
Høvring	T			971	63	34	[250]	[250]			0.98	
				2082	340	200	[971]	[60]				Single, inherited grain
Vrådal	T								939	20	9.5	Model 2 age
Vehuskjerringa	T								918	7	3.3	Model 2 age
Sæbyggjenut	RVA			959	50	32	[250]	[250]			1.7	
Rosskreppfjord	RVA			1036	23	22	416	+39/-44			0.53	
		1020	13								10.2	
									1009	10		Model 2 age
Byklom	RVA			979	9	12					0.96	Negative lower intercept
				970	14	18	[250]	[250]			1.7	
									983	68	6.5	Model 2 age
Group 2 granites												
Bessefjell	T			940	19	19					1.7	
Gunnarstul	T			1133	6	7					1.3	Lower intercept undefined
				1134	21	21	[250]	[250]			5.3	
Otternes	T			1233	90	90	444	180			4.4	Model 2 age

[1]: Lower intercepts ages given in [square brackets] are forced lower intercepts.

granitic magmatism west of the Mandal - Ustaoset shear zone as early as c. 1036 Ma indicates that this region has seen a transition from a shortening to an anorogenic (extensional ?) tectonic regime within c. 15 Ma after the Feda magmatism at c. 1050 Ma.

Sources of inherited zircons

Except for one Early Proterozoic zircon in the Høvring granite and the dominantly inherited population in the Bandak granite, apparently xenocrystic cores do not give ages significantly different from zircons without visible cores (Herefoss, Sæbyggjenut Figs. 2,3). This observation indicates that most of the cores observed formed during early crystallization of the magma itself, or that they lost all radiogenic lead while residing in the granitic magma. The source of the single, old zircon in the Høvring granite cannot be positively identified, but it should be noted that the age of the TIB granitoids would fall within its very wide uncertainty range. The source of inherited zircons in the Bandak granite must have been rocks of c. 1500 Ma age, i.e. of the same age as the rhyolite of the Rjukan Group of the Telemark supracrustal sequence (Dahlgren et al. 1990; Sigmond 1998). The Bandak granite intruded metarhyolite which has not yet been dated, and it is locally rich in xenoliths of metarhyolite (Appendix 1). The presence of abundant, c.1500 Ma zircons in the granite suggests that the country rocks are equivalents of the c. 1500 Ma Rjukan Group metarhyolite, rather than the c.1150 Ma rhyolite of the Bandak Group (Dahlgren et al. 1990, Laajoki et al. 2000).

Age and significance of the Group 2 granites

The Group 2 granites can be divided into an older and a younger group. The 940 Ma Bessefjell granite is coeval with the main group of post-tectonic Group 1 granites. The Gunnarstul and Otternes granites are much older; Gunnarstul is coeval with the anorogenic Gjerstad suite augen gneisses (Bingen & van Breemen 1998) and the imprecise age of the Otternes granite overlaps both these rocks and the younger, c. 1150 Ma volcanism in Telemark (Dahlgren et al. 1990, Laajoki et al. 2000). The Otternes granite has a distinct foliation, which may have been formed in the same tectonic event that formed the gneissic foliation in the c.1150 Ma Gjerstad suite augen gneiss intrusions in the Bamble and Telemark sectors. The Gunnarstul granite lacks this foliation, which may suggest that it was emplaced after the cessation of this tectonic event, or that Sveconorwegian deformation was not pervasive in the eastern part of the Telemark sector.

The wide range of ages for the Group 2 granites indicates that the granitic magmas with a distinct Group 2 geochemical signature did not form in a separate event. Sr, Nd and Pb isotope data suggest collectively that Group 2 granites plot on a mixing line between a juvenile, Sveconorwegian component and a Mid Proterozoic crustal component with highly elevated Rb-Sr ratio (Andersen et al. 2001a). The trace element and isotopic properties of this component resemble both the c. 1500 Ma Rjukan group rhyolite (Menuge & Breweer 1996, Brewer & Menuge 1998) and the 1476 Ma Tinn granite (Andersen et al. 2001a, 2002a), and the presence of Group 2 granites in parts of the Telemark

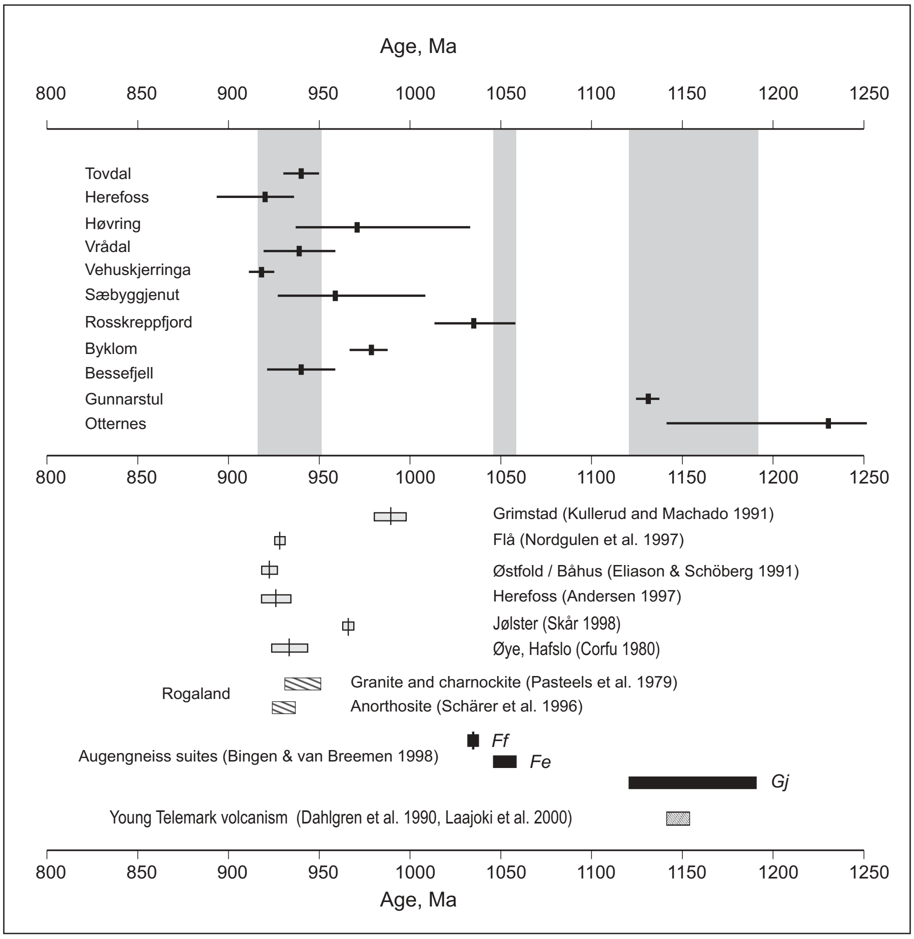

Fig. 6. Summary of geochronologic data from present study, compared to published data. Upper part of figure contains data from present study, given as recommended age ± 2σ (Table 3). Shaded fields represent the time of emplacement of main suite of post-tectonic granites (see text), and of orogenic Feda and anorogenic Gjerstad suite augen gneisses (Bingen & van Breemen 1998). Lower part of the figure summarizes published age data of mid- to late Sveconorwegian igneous rocks from South Norway (including relevant data from basement windows in the Caledonides and from the Western Gneiss Region). The data are given as either 'boxplots' with mean age and ± 2σ confidence interval, where such are available, or as bars indicating total duration of magmatic event. Abbreviations: *Ff:* Fennefoss augen gneiss (Telemark sector), *Fe:* Feda suite augen gneiss, and *Gj:* Gjerstad suite augen gneiss.

sector suggests the existence of deep crustal protoliths of similar chemistry at the time of granitic magmatism.

Tectonic setting

The main event of granite emplacement in the Telemark sector is contemporaneous both with emplacement of large granitic batholiths in the Kongsberg sector (Flå granite) and the area east of the Oslo rift (Østfold / Båhus granite) and with emplacement of granite/charnockite, mafic rocks and anorthosite in Rogaland. Deep crustal melting and emplacement of granitic intrusions throughout south Norway in this period appears to be associated with mafic underplating. On the other hand, the parent magma of the Rogaland anorthosite is assumed to have formed by partial melting of mafic protoliths in the deep crust, without the participation of juvenile, mantle-derived magma (Schärer et al. 1996, Schiellerup et al. 2000).

A common interpretation in orogenic belts is to link post-orogenic plutons to orogenic collapse and decompressional melting (England & Houseman 1988). Such a hypothesis has also been suggested for the generation of post-Sveconorwegian granites in South Norway (Eliasson 1992), but independent evidence for orogenic collapse was not presented. Such data now appear to exist. One line of evidence is provided by the shear zone separating the Bamble and Telemark sectors (Fig. 1). Structural observations across this shear zone demonstrate that it initiated as a top-toward-NW shortening shear around 1090 Ma (Andresen & Bergundhaugen 2001), followed by top-toward-SE extensional shearing and crustal thinning. Another line of evidence in support of collapse of the Sveconorwegian orogen exists in southern Sweden. There, exhumation of eclogite, dated at c. 970 - 956 Ma, has been linked to late-/post-orogenic collapse by Möller (1998, 1999). Thus, temporal and spatial links between extension and post-orogenic granites exist. The juvenile component in some of the granites also suggests that crustal thinning, rather than crustal thickening, is a likely mechanism for magma-generation in the deep crust.

Conclusions

New SIMS U-Pb data on zircons from Sveconorwegian granites from South Norway confirm the existence of an event of granitic magmatism across the region at 920 to 950 Ma. The emplacement of juvenile, mantle-derived material into the crust in this period is demonstrated by the 940 Ma Tovdal granite and by the presence of distinct, mantle-derived components in the other granites. In the western part of South Norway (west of the Mandal-Ustaoset shear zone), this magmatism started earlier, possibly as early as c. 1030 Ma, and was of longer duration. Granites formed in this event include both intrusions with normal trace element distributions (Group 1 of Andersen et al. 2001a) and intrusions with distinct, negative Sr concentration anomalies and high initial $^{87}Sr/^{86}Sr$ ratios (Group 2 granites of Andersen et al. 2001a). The 'low Sr' granites thus reflect the presence in the deep crust of a protolith of suitable composition (e.g. rocks related to the c. 1500 Ma metarhyolite of the Rjukan Group) and not the existence of a separate event of granitic magmatism. Group 2 granites also formed much earlier (before 1120 Ma), however, simultaneous with emplacement of anorogenic granite of the Gjerstad augen gneiss suite. This conclusion suggests that granites with a Group 2 geochemical signature could form in the region whenever mafic, mantle-derived magmas came into contact with suitable protoliths in the lower crust. The emplacement of mafic magmas into the crust and the resulting anorogenic granitic magmatism is most likely related to extensional tectonic events, one of which may have been of regional extent in South Norway (the 'Gjerstad augen gneiss event'), the other an orogen-wide event of late-/post-orogenic collapse.

The present study gives a somewhat two-sided impression of the importance of inherited zircons in these granites. Whereas one of the granites studied is totally dominated by inherited zircons, to the extent that its emplacement age cannot be determined from the SIMS data obtained, apparent xenocrystic cores in other granites have ages indistinguishable from the bulk of the zircon populations. This may suggest that the cores formed at an early stage of the crystallization history of the magma itself, or that their U-Pb system was totally reset in the magma. Spatially resolved Hf isotope data by laser ablation ICPMS are needed to distinguish between these two mechanisms.

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Appendix 1

Petrography and field relationships of samples analysed

Sample numbers are taken from Andersen (1997) and Andersen et al. (2001a). Localities are given by UTM coordinates (map datum: WGS84).

Bandak 072196-3 (32VML658849): Fine-grained and weakly foliated, two-feldspar biotite leucogranite, with traces of secondary chlorite and epidote. Zircon and monazite are accessory minerals. The pluton intruded metavolcanic supracrustal rocks, which also occur as xenolithic screens within the intrusion.

Bessefjell 072496-2 (32VML412952): Massive, medium-grained, two-feldspar biotite granite, intruding quartzite and metabasalt of the Telemark Supracrustal sequence. Dated to 904±16 Ma by Rb-Sr (Killeen & Heier 1975, recalculated with λ^{87}Rb=1.42 x10^{-11} a^{-1}).

Byklom 083096-3 (32VML083833): Medium-grained, massive, gray, two-feldspar granite. Titanite and magnetite are minor minerals; pyrite, zircon and allanite are accessory phases.

Gunnarstul 083193-1 (32VMM879073): Fine-grained, massive, two-feldspar biotite granite, with secondary muscovite. The plagioclase is extensively sericitized, and biotite is chloritized. The Gunnarstul granite intruded mafic metavolcanic rocks of the Telemark supracrustal sequence.

Herefoss 107/92 (32VVMK594804): Coarse-grained, foliated, biotite-hornblende granite (Andersen 1997).

Høvring 082996-2 (32VMK396014) Two-feldspar biotite granite with minor hornblende and accessory titanite, allanite, apatite, pyrite, and zircon. The Høvring (or Høvringsvatn) complex is a composite intrusion, consisting of granite, and younger, monzonitic intrusive facies. Granites in the complex have been dated at 945 ± 53 by a whole-rock Rb-Sr isochron (Pedersen 1981).

Otternes 072696-2 (32VNL149794): Foliated, two-feldspar leucogranite with minor titanite, biotite, magnetite, and allanite. Intrudes granitic gneiss and quartzite.

Rosskreppfjord 080296-4 (32VLL939463): Coarse-grained, two-feldspar, biotite-hornblende granite with minor titanite and magnetite, and accessory zircon.

Sæbyggjenut 072496-3 (32VML254926): Coarse-grained 'tricolor' biotite - hornblende granite with pink perthite, pale green plagioclase, and clear quartz. Apatite, titanite, magnetite, and pyrite are accessory minerals.

Tovdal 072396-3 (32VML544170): Weakly foliated, biotite-hornblende granite with minor titanite, apatite, magnetite, and fluorite, and accessory allanite and zircon. The Tovdal granite intruded banded gneiss, probably of supracrustal origin.

Vehuskjerringa 072496-1 (32VMM546120): Medium grained biotite granite with magnetite, titanite, and apatite, intruding quartzite, quartz schists, and metabasalt of the Bandak Group.

Vrådal 081696 (32VML740747): Two-feldspar biotite granite intruding granitic gneiss (Sylvester 1964, 1998).

Age and petrogenesis of the Tinn granite, Telemark, South Norway, and its geochemical relationship to metarhyolite of the Rjukan Group

TOM ANDERSEN, ARTHUR G. SYLVESTER & ARILD ANDRESEN

Andersen, T., Andresen, A. & Sylvester, A.G. 2002: Age and petrogenesis of the Tinn granite, Telemark, South Norway, and its geochemical relationship to metarhyolite of Rjukan Group. *Norges geologiske undersøkelse Bulletin 440*, 19-26.

The Tinn granite is a Mid Proterozoic, foliated pluton, situated in the central part of the Telemark Sector, South Norway. It is spatially associated with metarhyolite belonging to the Tuddal Formation of the Rjukan Group of the Telemark supracrustal sequence. SIMS U-Pb dating indicates an age of 1476 ± 13 Ma, which is slightly younger than the 1500-1514 Ma eruption interval for the Tuddal Formation rhyolite. Rare xenocrystic zircon cores give an age of 1506 ± 10 Ma, which is indistinguishable from the age of the Tuddal Formation. The absence of older inherited zircons and evidence from whole-rock Nd isotopes suggest that no source component older than the Rjukan Group is needed in the source region of the Tinn granite magma. The preferred petrogenetic model for the Tinn granite is a partial melting-mixing process at moderate depth in the crust, within the Rjukan Group volcanic pile, A mafic magma acted as a source of heat and contributed to the bulk chemistry of the granitic magma. Resetting of the lead isotope system of the granite at mineral scale took place in Sveconorwegian time, at 1031 ± 32 Ma.

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Introduction

Granitic intrusions make up a substantial component of the continental crust in the southwestern part of the Baltic Shield. Among the Precambrian granitic rocks of South Norway, a group of granitic orthogneisses in the Telemark sector stands out as especially poorly understood. These rocks are spatially associated with the Mid to Late Proterozoic Telemark Supracrustal sequence (Sigmond et al. 1997), and include the voluminous south Telemark Gneisses situated south of the main outcrop area of the supracrustal sequence (Ploquin 1972, Dons & Jorde 1978, Kleppe 1980) and the Tinn granite in the north (Sigmond 1998).

This study presents new SIMS U-Pb data for the emplacement age of the Tinn granite, and clarifies its relationship to the Telemark metarhyolite and to other possible source rocks.

Geologic setting

The *Telemark Supracrustal sequence* is a well-preserved sequence of Mid Proterozoic metavolcanic and metasedimentary rocks, situated in the central part of the Telemark Sector (Dons & Jorde 1978, Sigmond 1998), surrounded by strongly deformed, higher-grade ortho- and paragneisses (Fig. 1). The supracrustal sequence consists of three lithostratigraphic groups separated by angular unconformities: the *Rjukan, Seljord* and *Bandak* Groups, and a fourth group, the Heddal Group, conformably overlying the Seljord Group in the eastern part of the outcrop area (Dons 1960, Sigmond 1998). The *Rjukan Group* is entirely metavolcanic, with a thick sequence of metarhyolite (the *Tuddal Formation*) overlain by a thinner metabasaltic formation (the *Vemork Formation*). The rhyolitic rocks were deposited in extensional basins, possibly as part of a continental rift system (Sigmond et al. 1997), on a migmatitic basement of unknown age. From the north end of lake Tinnsjø to the Caledonian nappe front, the Rjukan Group is cut by younger mafic to granitic intrusions of the *Uvdal plutonic belt* (Sigmond et al. 1997, Sigmond 1998). The Seljord Group consists of quartzite and conglomerate and the Heddal Group of quartz arenite with subordinate metavolcanic rocks, whereas the Bandak Group is a mixed, volcanic-sedimentary sequence comprising several formations.

Sigmond (1998) reported conventional U-Pb zircon ages for the Tuddal Formation rhyolite of 1512 +9/–8 Ma and 1499 ± 39 Ma, and 1509 +19/–3 for an intrusion of the Uvdal plutonic belt: the best estimate of the duration of the Rjukan Group volcanism was given as 1500-1514 Ma. A volcanic formation of the Bandak Group has been dated at c. 1150 Ma (Dahlgren et al. 1990). From detrital zircon systematics, Haas et al. (1999) inferred a maximum depositional age of the Seljord Group at 1450 Ma, but new U-Pb age data from rhyolite underlying the type profile of the Seljord Group suggest that some of the sedimentary rocks previously assigned to the Seljord Group may be younger than 1155 Ma (Laajoki et al. 2000).

The Tinn granite makes up the southernmost part of the Uvdal plutonic belt (Fig. 1). It is a fine-grained, pale pink, leucocratic two-feldspar granite, with minor dark brown biotite

Fig. 1. Simplified geologic map of the Tinn granite and surrounding rocks, compiled from Dons & Jorde (1978) and Sigmond (1998). Overview map: Simplified map of South Norway showing regional subdivision used in this study: A: Østfold-Akershus sector; K: Kongsberg sector; T: Telemark sector; B: Bamble sector; R: Rogaland-Vest Agder sector. Major shear zones: MMS: Mjøsa-Magnor shear zone, ØMZ: Ørje mylonite zone; PKS: Porsgrunn-Kristiansand shear zone; KTB: Kongsberg-Telemark boundary; MANUS: Mandal-Ustaoset line; CTF (broken line): Caledonian thrust front.

and magnetite, and accessory zircon, titanite and apatite. Field observations do not define unambiguous age relationships between the granite and metarhyolite of the Tuddal Formation: The contact between granite and metarhyolite is gradational, and enclaves or dikes of one in the other are nowhere observed. The grain size of the metarhyolite increases toward the contact with the granite, however, which suggests a local thermal imprint related to the emplacement of the granite (Sigmond 1998). Both units are foliated parallel to the contact. This may be a result of deformation during granite emplacement, as has been suggested for other granites in southern Norway (e.g. Elders 1963, Sylvester 1998), or a result of later tectonic deformation locally controlled by the more competent granite.

The Tinn granite is a moderately silica-rich granite (SiO_2= 68.3-72.4 wt%, Table 1). Its atomic (Na+K)/Al ratio is well below 1.0. The normative mineralogy is highly leucocratic, with a differentiation index (normative *qz+ab+qz+ne*) well above 80. The samples plot well within the 'granite *sensu stricto*' fields in pq and *Ab-Or-An* classification diagrams (e.g. Rollinson 1993). Compared to the average of the Tuddal Formation metarhyolite, the Tinn granite is low in SiO_2 and high in CaO and Na_2O (Table 1). Furthermore, the Tinn granite is metaluminous ((Ca+Na+K)/Al of 1.05-1.08, normative *co*=0), in contrast to the peraluminous composition of the average metarhyolite (normative *co*=1.9).

Analytical methods

The present study is based on two 5-8 kg samples of the Tinn granite (083196-2 and 071996-2), for which whole-rock Sr, Nd, and Pb isotope data were published by Andersen et al. (2001). The samples were crushed to a grain size of less than 250 µm using a jaw crusher and a percussion mill. *Zircons* were separated from the <250 µm fraction by a combination of Wilfley-table washing, heavy liquid separation (1,1,2,2- tetrabromoethane and diiodomethane) and magnetic separation. The final, non-magnetic zircon fraction was then purified by hand picking under a binocular microscope, and selected grains were mounted on doubly adhesive tape, cast in epoxy and polished for the ion microprobe study. Electron backscatter imaging (BSE) in a scanning electron microscope (Department of Geology, Oslo) and an electron microprobe (Macquarie University, Sydney) was used both as a preliminary survey before analysis, and to document individual grains after analysis. The U-Pb zircon dating was performed in the NORDSIM laboratory located at the Swedish Museum of Natural History in Stockholm, using a CAMECA IMS1270 ion microprobe; analytical conditions and data reduction procedures are described by Whitehouse et al. (1997, 1999). U-Pb data are listed with 1 σ errors in Table 1, whereas the derived ages are given with 95% confidence errors. Additional separates of rock-forming minerals for the Pb-Pb isochron study were made by a combination of heavy liquid and magnetic separation, followed by hand picking. Lead was separated and analysed by methods described by Andersen (1997). Whole-rock and K-feldspar lead isotope data are taken from Andersen et al. (2001).

All geochronologic calculations have been made using Isoplot/Ex version 2.32 (Ludwig 2000).

Table 1. Geochemical data on the Tinn granite.

Sample	083196-2	071996-2	Tuddal Fm average		083196-2	071996-2	Tuddal Fm average
	UTM-reference						
UTM E	4862	4908					
UTM N	66509	66523					
	Wt% Oxides				*CIPW-Norm*		
SiO_2	68.30	72.41	75.07	qz	23.67	29.01	36.9
TiO_2	0.66	0.21	0.27	co	0	0	1.9
$Al2O_3$	14.06	13.10	12.52	or	30.73	30.73	30.6
$Fe2O_3$	1.16	0.66	0.73	ab	27.76	29.47	24.2
FeO	2.10	1.19	1.47	an	8.27	4.75	0.9
MnO	0.05	0.04	0.03	di	1.27	1.31	
MgO	0.54	0.13	0.40	hs	2.92	1.15	2.7
CaO	2.27	1.31	0.25	ilm	1.25	0.40	0.5
Na_2O	3.28	3.48	2.86	mt	1.57	0.89	1.1
K_2O	5.20	5.20	5.19	ap	0.55	0.10	0.1
P_2O_5	0.23	0.04	0.03				
LOI	1.58	0.83		an%	0.23	0.14	0.04
Sum	99.43	98.61					
	Trace elements, parts per million						
Rb	288	310	184				
Ba	364	319	491				
Pb	17	17	9				
Sr	32	41	33				
Eu	0.46	0.65					
	Radiogenic isotopes						
$^{147}Sm/^{144}Nd$	0.1229	0.1049					
$^{143}Nd/^{144}Nd$	0.512083	0.511829					
$\pm 2\sigma$	16	10					
t_{DM}	1.60	1.69					
$^{87}Rb/^{86}Sr$	28.0653	22.9046					
$^{87}Sr/^{86}Sr$	1.282289	1.172639					
$\pm 2\sigma$	14	11					

Major element analysis by XRF (Department of Geology, University of Oslo), trace elements by ICPMS (Actlabs, Canada). Fe2O3/FeO estimated according to Rollinson (1993). Whole-rock radiogenic isotope data from Andersen et al. (2001), and data on Tuddal Fm. metarhyolites from Brewer & Menuge (1998). Nd model ages are calculated using the depleted mantle reservoir of De Paolo (1981).

Morphology and internal structure of zircons

Zircons in the Tinn granite are moderately elongated prisms. BSE images reveal a well-developed oscillatory magmatic zoning, in most grains overgrown by a thin and discontinuous, BSE-bright outer zone (Fig. 2a). These overgrowths were too thin to be analysed, but were most likely formed during metamorphic recrystallization of the granite. *Xenocrystic cores* predating the main zircon-forming event, with boundaries clearly discordant to the magmatic zoning, are rare. Among more than 200 grains mounted for analysis, only four single crystals contained cores which were visible in BSE images, two of which are shown in Fig. 2b and c.

Geochronology

SIMS U-Pb data and the age of emplacement

Twenty-nine spots on 25 selected zircon grains were analysed by secondary ion mass spectrometry (Table 2). Individual spot analyses are identified by NORDSIM laboratory log numbers. Core-rim pairs (denoted by *a* and *b* in Table 2) were analysed in grains where inherited cores were

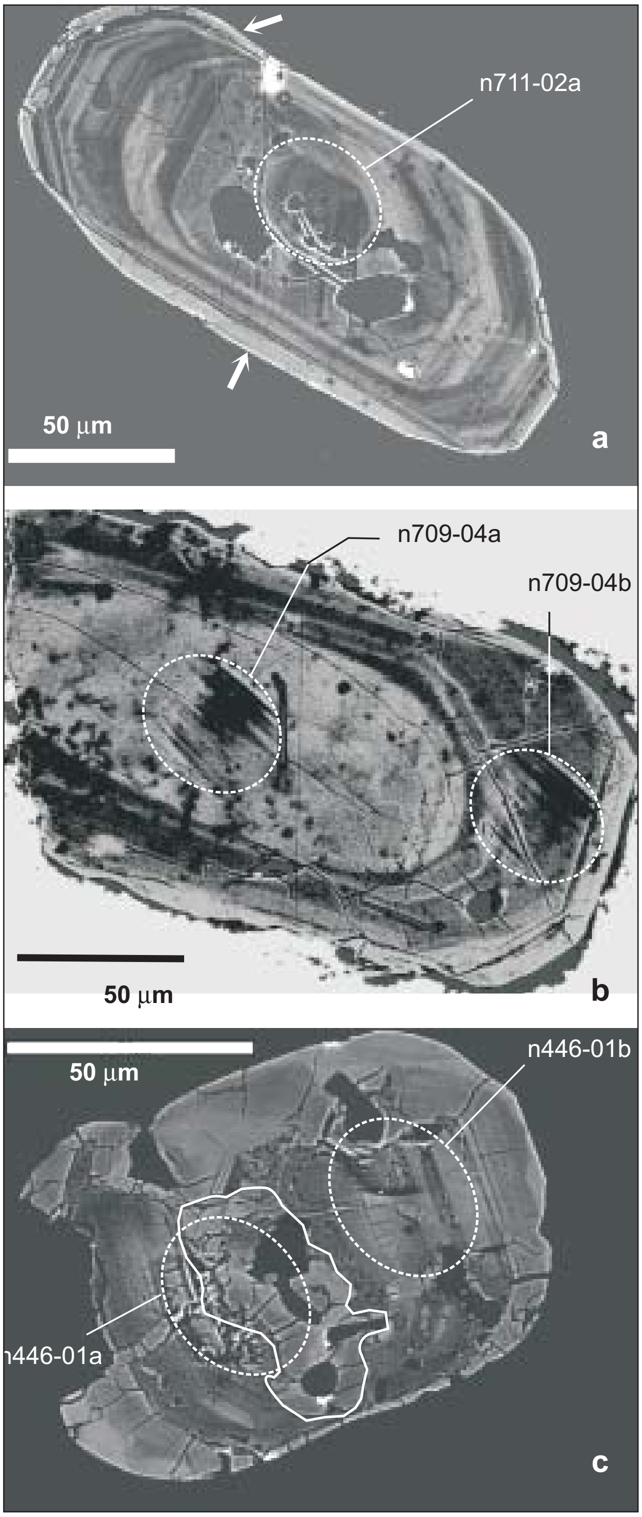

Fig. 2. Backscatter electron images of zircons from the Tinn granite. Location of SIMS analytical spots indicated (stippled ellipses) with numbers referring to Table 1. Images made after analysis, using a Cameca SX-50 electron microprobe at GEMOC National Key Centre, Macquarie University, Australia. a: Most common zircon in Tinn granite consists of central domain with oscillatory, magmatic zoning and a thin, discontinuous, BSE-bright overgrowth (arrows). Sample 071996-2. b: Zircon crystal with well-rounded, xenocrystic core in oscillatory zoned host. Note weak, oscillatory zoning in core (right part) cut by interface between core and host. Sample 083196-2 c: Zircon with a corroded xenocrystic core (strengthened by outline). BSE-bright overgrowth is more strongly developed in this crystal than in that in b, but it has been partly broken off (left part) during crushing. Sample 083196-2.

Table 2. SIMS U-Pb data for the Tinn granite.

Sample Spot	^{207}Pb/^{206}Pb	±σ %	^{207}Pb/^{235}U	±σ %	^{206}Pb/^{238}U	±σ %	ρ Error correlation	^{208}Pb/^{232}Th	±σ %	Disc. %	U ppm	Th ppm	Pb ppm	Th/U	^{06}Pb/^{204}Pb	f_{206}%	^{207}Pb/^{206}Pb Ma	±σ Ma	^{207}Pb/^{235}U Ma	±σ Ma	^{206}Pb/^{238}U Ma	±σ Ma	^{208}Pb/^{232}Th Ma	±σ Ma
083196-2																								
n446-01a	***0.09367***	***0.4***	***3.5086***	***1.2***	***0.27168***	***1.1***	***0.95***			***1***	***259***	***148***	***89***	***0.57***	***186880***	***0.0***	***1501***	***7***	***1529***	***9***	***1549***	***15***		
n446-01b	+ 0.09297	0.3	3.6029	0.8	0.28106	0.8	0.92			7	277	149	98	0.54	56240	0.0	1487	6	1550	7	1597	11		
n446-02a	*0.06268*	*0.5*	*0.6768*	*0.7*	*0.07832*	*0.5*	*0.93*			*-31*	*3889*	*2760*	*375*	*0.71*	*1590*	*1.2*	*697*	*10*	*525*	*3*	*486*	*2*		
n446-02b	*0.08708*	*2.7*	*1.9590*	*2.8*	*0.16315*	*1.0*	*1.00*			*-29*	*903*	*4068*	*191*	*4.51*	*1300*	*1.4*	*1362*	*51*	*1102*	*19*	*974*	*9*		
n446-03a	*0.07440*	*0.9*	*1.0245*	*1.0*	*0.09986*	*0.5*	*0.98*			*-43*	*2430*	*2640*	*309*	*1.09*	*1090*	*1.7*	*1052*	*18*	*716*	*5*	*614*	*3*		
n446-03b	*0.07351*	*0.8*	*0.8177*	*1.0*	*0.08067*	*0.5*	*1.00*			*-53*	*3444*	*4278*	*365*	*1.24*	*980*	*1.9*	*1028*	*17*	*607*	*4*	*500*	*3*		
083196-2																								
n709-01a	*0.07750*	*0.4*	*1.3477*	*2.6*	*0.12612*	*2.6*	*0.99*	*0.0286*	*10*	*-34*	*3229*	*878*	*464*	*0.27*	*4805*	*0.4*	*1134*	*8*	*867*	*15*	*766*	*18*	*570*	*57*
n709-02a	+ 0.09078	0.3	2.4729	2.6	0.19756	2.5	0.99	0.02877	7.6	-21	813	632	193	0.78	12228	0.2	1442	6	1264	19	1162	27	573	43
n709-03a	+ 0.09296	0.4	2.3622	2.6	0.18430	2.5	0.99	0.01448	7.6	-29	545	891	123	1.64	13077	0.1	1487	8	1231	19	1090	26	291	22
n709-04a_core	***0.09423***	***0.4***	***3.2895***	***2.6***	***0.25318***	***2.6***	***0.99***	***0.08109***	***7.7***	***-4***	***1434***	***682***	***453***	***0.48***	***66534***	***0.0***	***1513***	***8***	***1479***	***21***	***1455***	***34***	***1576***	***117***
n709-04b_ rim	+ 0.09093	0.3	2.5881	2.6	0.20642	2.5	0.99	0.02728	7.5	-18	770	726	193	0.94	8842	0.2	1445	6	1297	19	1210	28	544	40
n709-05a	+ 0.09298	1.2	1.8845	2.8	0.14700	2.6	0.91	0.00930	8.1	-43	1205	3332	225	2.76	516	3.6	1488	22	1076	19	884	21	187	15
n709-06a	+ 0.09334	0.3	2.7464	2.6	0.21339	2.5	0.99	0.03531	7.5	-18	682	588	180	0.86	14507	0.1	1495	7	1341	19	1247	29	701	52
n709-07a	+ 0.09385	0.4	2.5787	2.6	0.19928	2.6	0.99	0.01793	7.7	-24	557	805	136	1.45	25674	0.1	1505	7	1295	19	1171	27	359	27
n709-08a	+ 0.09303	0.2	3.3108	2.5	0.25811	2.5	1.00	0.07615	7.5	-1	931	507	301	0.54	93371	0.0	1489	4	1484	20	1480	34	1483	108
n709-09a	+ 0.09215	0.3	2.8457	2.6	0.22398	2.6	0.99	0.02870	7.5	-13	875	1111	246	1.27	35298	0.1	1470	5	1368	19	1303	30	572	42
n709-10a	+ 0.09240	0.3	2.9505	2.6	0.23160	2.6	0.99	0.03714	8.0	-10	956	936	277	0.98	32852	0.1	1476	5	1395	20	1343	31	737	58
n709-11a	+ 0.09088	0.4	2.7126	2.6	0.21649	2.6	0.99	0.02492	7.5	-14	448	400	116	0.89	13596	0.1	1444	8	1332	20	1263	30	498	37
n709-12a	+ 0.09419	0.6	2.6101	2.6	0.20098	2.5	0.98	0.03421	7.7	-24	353	243	86	0.69	31182	0.1	1512	10	1303	19	1181	27	680	51
071796-2																								
n711-01a	+ 0.09183	0.6	2.6542	2.7	0.20963	2.6	0.97	0.04561	8.2	-18	974	724	256	0.74	13340	0.1	1464	12	1316	20	1227	30	901	72
n711-02a	*0.08781*	*1.4*	*2.6250*	*3.0*	*0.21681*	*2.6*	*0.88*	*0.03204*	*7.8*	*-9*	*821*	*1478*	*241*	*1.80*	*585*	*3.2*	*1378*	*27*	*1308*	*22*	*1265*	*30*	*638*	*49*
n711-03a	*0.08476*	*0.5*	*1.8583*	*2.7*	*0.15901*	*2.7*	*0.98*	*0.00845*	*7.7*	*-29*	*696*	*986*	*128*	*1.42*	*3619*	*0.5*	*1310*	*9*	*1066*	*18*	*951*	*24*	*170*	*13*
n711-04a	+ 0.09345	0.2	3.0782	2.6	0.23889	2.5	1.00	0.06449	7.5	-9	1025	568	304	0.55	31646	0.1	1497	4	1427	20	1381	32	1263	92
n711-06a	+ 0.09097	0.4	2.6031	2.6	0.20754	2.6	0.99	0.03755	7.6	-17	750	581	192	0.77	4170	0.5	1446	8	1302	19	1216	28	745	56
n711-07a	+ 0.09268	0.5	2.1991	2.6	0.17208	2.6	0.98	0.01344	7.7	-33	835	1611	179	1.93	11330	0.2	1481	10	1181	18	1024	24	270	21
n711-08a	+ 0.09184	0.3	2.7849	2.6	0.21993	2.6	0.99	0.03227	7.6	-14	774	905	215	1.17	5107	0.4	1464	7	1351	19	1282	30	642	48
n711-09a	+ 0.09294	0.5	2.8222	2.6	0.22023	2.5	0.98	0.03356	7.7	-15	310	321	85	1.03	4466	0.4	1487	10	1361	20	1283	30	667	51
n711-10a	*0.08541*	*0.3*	*1.9206*	*2.6*	*0.16310*	*2.5*	*0.99*	*0.02305*	*7.6*	*-29*	*1760*	*1981*	*357*	*1.13*	*14102*	*0.1*	*1325*	*6*	*1088*	*17*	*974*	*23*	*461*	*35*
n711-12a	+ 0.09101	0.5	2.6960	2.6	0.21484	2.5	0.98	0.03322	7.5	-15	1467	2338	421	1.59	1212	1.5	1447	9	1327	19	1255	29	661	49

Points n446-01 to n446-03 analysed in February 1999, the other poins in january 2000. Spots with a number ending in a have been analysed in the centre of a grain, b in the rim.
Analysts: A. Andresen and T. Andersen.
+: indicates analyses included in the final estimate of the emplacement age. **Bold** typeface refers to xenocrystic zircon cores

clearly observed or suggested by BSE images; all other analyses are from the central part of oscillatory zoned crystals without visible xenocrystic cores. One of the analysed grains (core-rim pair n446-01a/b) is reversely discordant; the other grains range from near-concordant (1 % discordant) to strongly discordant (>30% discordant). When all points are plotted in a concordia diagram, a majority of points cluster along a lead-loss line from a Mid Proterozoic upper intercept to a lower intercept which is poorly defined, but within analytical uncertainty of 0 Ma (Fig. 3a). Several points fall significantly to the left of this line, however, suggesting that some zircons have also been affected by a lead-loss event related to later metamorphism (Fig. 3b,c). The metamorphic overprint may be of Sveconorwegian (e.g. Andersen & Munz 1995) or, possibly, Caledonian age.

When the grains least affected by Sveconorwegian (or Caledonian) lead-loss are regressed for each sample separately, identical ages of 1476 ± 20 (071996-2, 7 single analyses) and 1476 ± 13 Ma (083196-2, 12 points) are obtained (Fig. 3b,d), assuming recent lead-loss. The less-than-perfect fit of these regressions could be caused by the presence of undetected, slightly older cores in the volumes sampled by the ion-beam, or by the effects of incipient Sveconorwegian lead loss. However, further 'improvement' of these ages by exclusion of more points is not justified. The oscillatory zircon must have grown during crystallization of the Tinn granite magma, and *1476 ± 13 Ma* is regarded as the best estimate of its emplacement age.

Of the four cores, two (n446-02a, n446-03a) come from grains that have been thoroughly reset as indicated by Sveconorwegian ^{207}Pb/^{206}Pb ages (Table 2). The two remaining cores (n446-01a and n709-04a) plot marginally to the right of the main population of zircons (Fig. 3c). Regression through a forced lower intercept at 0 Ma yields an age of 1506 ± 10 Ma (Fig. 3c), which is slightly, but still probably significantly, older than the age of the main population of zircons in sample 083196-2.

Pb-Pb isotope data and timing of metamorphism

K-feldspar, biotite, apatite, magnetite and titanite separated from sample 071996-2 give a large range of lead isotopic compositions (Table 3), from near-initial lead (K-feldspar) to radiogenic lead with ^{206}Pb/^{204}Pb above 90 (titanite). A regression of all data (minerals and both whole-rocks) yields a Pb-Pb scatterchron with an age of 1031 ± 32 Ma (Isoplot Model 2; Ludwig, 2000) and an MSWD of 9.4 (Fig. 4). Removal of the whole-rock 083196-2 from the regression increases the uncertainty to ± 40 Ma and raises the MSWD slightly, but does not affect the age. The 1031 ± 32 Ma correlation line indicates partial homogenization of the lead isotopes in hand specimen as well as intrusion scale in Sveconorwegian

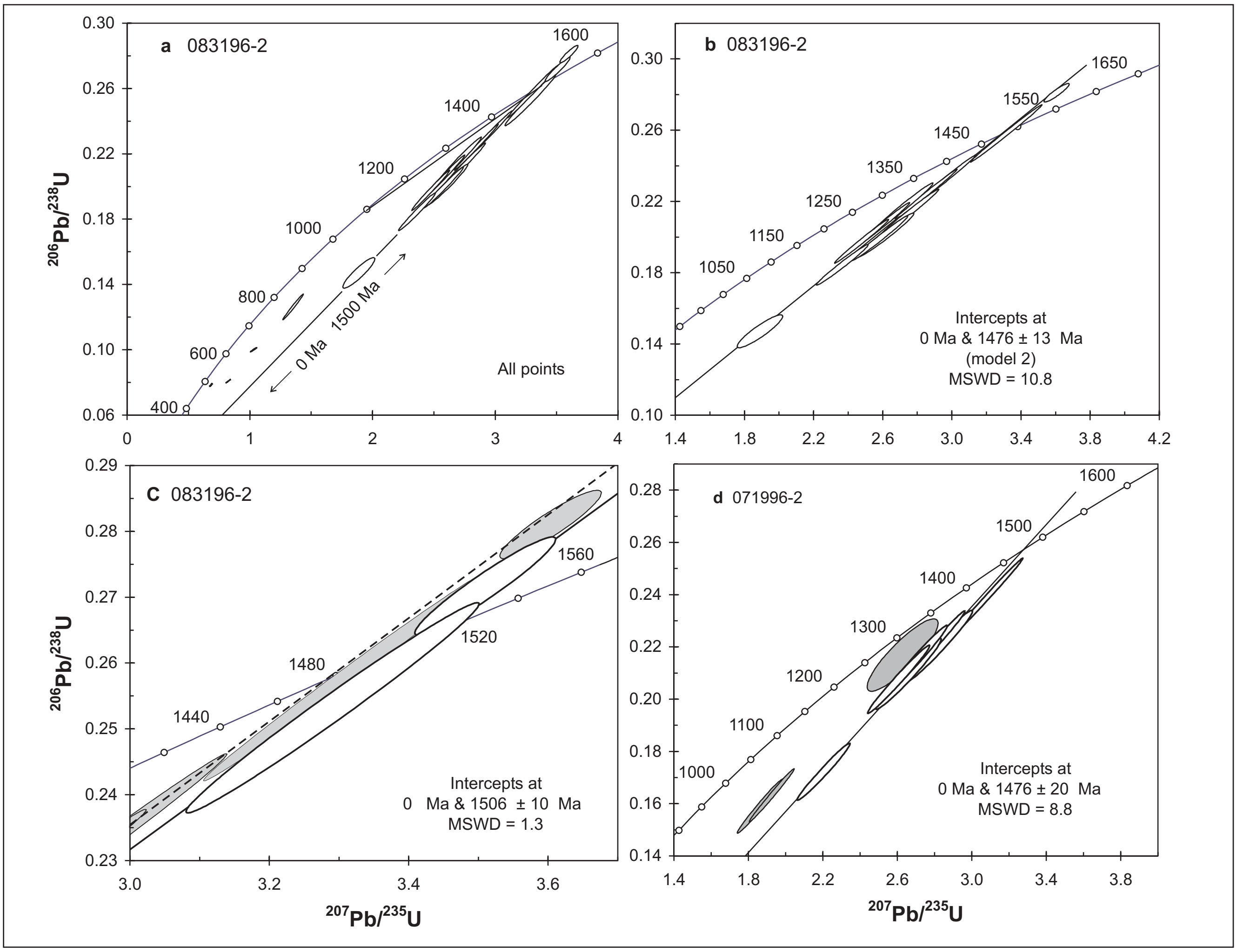

Fig. 3. Concordia diagrams of SIMS U-Pb data for the Tinn granite (Table 1). Data-points are shown with 1s error ellipses. a: All points, with a reference recent lead-loss line drawn to 1500 Ma. Note widely discordant grains, plotting between the reference line and concordia. These grains lost radiogenic lead both in Sveconorwegian and in recent times and are omitted from further consideration. b: Data from sample 083196-2, showing grains without visible cores. Regression line through a forced lower intercept at zero is shown. c: Cores in sample 083196-2 (white) compared to grains of main population in same sample (gray). Regression lines for the main population (dotted, see b) and for cores (forced through zero) are shown. d: Data from sample 071996-2, showing best-fit recent lead-loss line. Points indicated in gray suffered partial lead-loss in the Sveconorwegian orogeny and have been omitted from the regression.

time, and suggests that the partial lead-loss observed in some zircons was indeed due to a Sveconorwegian metamorphic overprint. This age corresponds within overlapping uncertainties with a regional lead isotope resetting event detected in metasedimentary rocks and in other felsic intrusions in South Norway (Heaman & Smalley, 1994; Andersen & Munz 1995, Simonsen 1997).

Table 3. Lead isotope data on whole-rocks and rock-forming minerals of the Tinn granite.

		$^{206}Pb/^{204}Pb$		$^{207}Pb/^{204}Pb$		$^{207}Pb/^{204}Pb$	
071996-2	Whole-rock	22.843	0.018	15.962	0.019	42.043	0.066
	K-feldspar	18.476	0.017	15.640	0.021	37.645	0.067
	Magnetite	66.569	0.060	19.110	0.026	71.859	0.127
	Biotite	59.604	0.054	18.692	0.026	74.026	0.131
	Titanite	96.349	0.087	21.389	0.029	72.433	0.128
	Apatite	20.681	0.018	15.788	0.021	39.253	0.068
083196-2	Whole-rock	22.274	0.017	15.915	0.019	40.713	0.064

Whole-rock and K-feldspar data from Andersen et al. (2001)

Discussion

The age of eruption of the Tuddal formation rhyolite is still not well determined, but the assumption of a c. 14 Ma period of volcanic activity by Sigmond (1998) is reasonable from what is known from modern and recent geologic analogs. In the southwestern United States, numerous rhyolitic volcanic centers formed in response to Cenozoic crustal extension (e.g., Lipman 1992). One of the largest and best studied silicic volcanic centers in the world in the Timber Mountain – Oasis Valley caldera complex in southwestern Nevada. It is about 100 km long and 50 km wide and has existed for 16 Ma. Silicic volcanism predominated between 16 and 6 Ma, the most activity and voluminous magma production was between 12 and 10 Ma, single calderas lasted 1-2 Ma, some

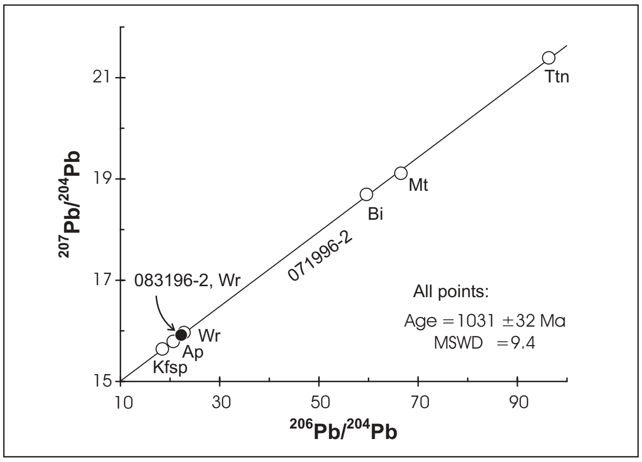

Fig. 4. Lead isotope data for whole rocks and minerals from the Tinn granite; data from Table 2. The age represents an event of Sveconorwegian lead isotope homogenization, whose timing is comparable to other events in South Norway (e.g. Andersen & Munz 1995). Abbreviations: Ttn: titanite; Mt: magnetite; Bi: biotite; Wr: whole rock; Ap: apatite; Kfsp: potassium feldspar (microcline).

were as short as 100,000 years, and the time lapse between individual eruptive events may have been only several tens of years (Byers et al. 1989). The volcanic center is typified by at least 35 separable important eruptive events. Basaltic volcanism began there at about 9 Ma and continues to present (Perry et al. 1998).

The most precise U-Pb age reported by Sigmond (1998) for a Tuddal Formation rhyolite (1512 +9/–8 Ma) and the 1509 +19/–3 age for a crosscutting intrusion combine to suggest that the rhyolitic magmatism had terminated before 1500 Ma. The two zircon cores dated here (1506 ± 10 Ma) are coeval with the rhyolite and may have been inherited from such a source. An emplacement age of 1476 ± 13 Ma thus makes the Tinn granite slightly younger than the Tuddal Formation rhyolite. Metamorphism of the granite post-dated its emplacement by c. 450 Ma, causing only minor lead-loss from zircons. The present geochronologic data thus agree with the interpretation that the Tinn granite intruded the Tuddal Formation, and that the foliation-concordant nature of the contact between the two units is due to deformation during emplacement or to later Sveconorwegian(?) deformation.

At 1476 Ma, the Nd isotopic composition of both of the samples dated in this study falls within the wide range of variation of the Tuddal Formation metarhyolite (Menuge & Brewer 1996, Brewer & Menuge 1998; data for the Tinn granite from Andersen et al. 2001, see Table 1); sample 083196-2 also overlaps with the much more restricted range of variation of the Vemork Group metabasalt (Fig. 5a). Depleted mantle model ages (De Paolo 1981 model) of 1.60 and 1.69 Ga are within the range of the Rjukan Group (Brewer & Menuge 1998).

The Tinn granite has a very radiogenic present-day Sr isotope composition, with $^{87}Sr/^{86}Sr$ well above 1.0. The reason for this is a very high Rb/Sr ratio, which is in turn due to anomalously low Sr contents combined with a normal upper-crustal Rb concentration (Fig 1, see also Andersen et al. 2001). In these features, the Tinn granite resembles a group of post-tectonic Sveconorwegian granites from the Telemark sector ('low-Sr concentration granites' of Andersen et al. 2001), and a range of metasedimentary rocks and gneisses of uncertain origin from the area west of the Oslo Rift (Andersen & Knudsen 2000). The present-day Sr isotope composition of the Tinn granite falls within the upper part of the range of the Tuddal Formation (Fig. 5b). When recalculated to 1476 Ma, the Tinn granite still overlaps with the range of the rhyolite, but at unrealistically low, time-cor-

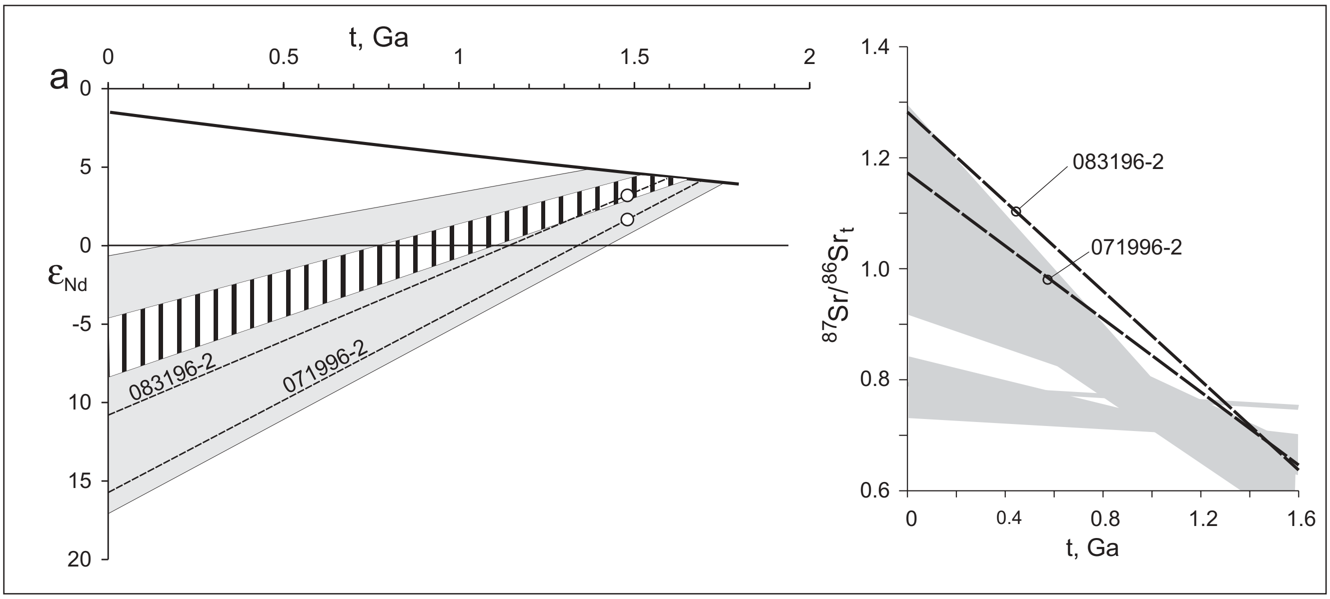

Fig. 5. Sr and Nd isotope evolution diagrams for the Tinn granite compared to the Rjukan Group. a: Nd isotopes. Gray shading represents total range of Tuddal Formation; striped field is corresponding range of Vemork Formation (data from Brewer & Menuge 1998). b: Sr isotopes. Gray shading represents total range of Tuddal Formation. Data from Kleppe (1980) and Verschure et al. (1990). See discussion in text.

rected $^{87}Sr/^{86}Sr$ (< 0.70), indicating that the Sr isotope system of the granite was partly reset in Sveconorwegian time. The high Rb/Sr ratios in some of the Tuddal Formation rhyolite have been attributed to Sveconorwegian Rb-metasomatism (Verschure et al. 1990), but Andersen et al. (2001) argued that the similar Rb/Sr ratios observed in Sveconorwegian low-Sr granites were inherited from the source, because none of these rocks is anomalously enriched in Rb, yet they have consistently low Sr concentrations. The same argument can be used for the Tinn granite.

The presence of 1506 ± 10 Ma inherited zircon cores in the Tinn granite, and its pronounced similarity to the Tuddal Formation in Nd isotope systematics, suggest that material related to the rhyolite of the Tuddal Formation has, indeed, been involved in the petrogenesis of the Tinn granite, as a source for anatectic melt, as a significant contaminant, or as a parent magma. Older, regionally distributed, possible protoliths in South Norway include pre-1.65 Ga TIB equivalents and other, still unidentified rocks with a crustal prehistory back to 1.7-1.9 Ga, which have acted as source terranes for the Seljord Group and other Mid Proterozoic clastic sedimentary rocks (Knudsen et al. 1997, de Haas et al. 1999, Bingen et al. 2001). The Nd isotope systematics of the Tinn granite, and the lack of pre-1.51 Ga inherited zircons indicate that such rocks were not significant as source rocks for the Tinn granite, nor were they important as contaminants.

Petrogenesis of the Tinn granite

The normative mineralogy of the Tuddal Formation rhyolite is strongly dominated by *qz, ab*, and *or*; and normative *an* is consistently very low, with an average of 0.97 % (Table 1, data from Brewer & Menuge 1998). The mean differentiation index is as high as 92 ± 7 (2 σ), which allows differentiation and partial melting of a rhyolitic precursor to be adequately reproduced by the *ab-or-qz* system, for which abundant experimental data are available (e.g. Johannes & Holtz 1996 and references therein). Both fractional crystallization of a rhyolitic magma and partial melting of Tuddal Formation rhyolite would produce melts at the thermal minimum of the quartz-feldspar cotectic boundary in the Ab-Or-Qz system at low to moderate pressures, and at the albite-K-feldspar-quartz eutectic at higher pressures. At low to moderate pressures, minimum melts would be higher in normative *qz* and lower in *an* and *or* than the Tinn granite; at higher pressures (> 5 kbar), minimum melts would be less silicic, but would have significantly higher ab/or ratios than observed (Johannes & Holtz 1996). The observed range of normative *an* in the Tinn granite could be caused by accumulation of alkali feldspar and plagioclase in a Tuddal-like magma, but it is highly improbable that a liquid with less than 1 % normative *an* could accumulate enough calcic plagioclase to increase the *an* content by a factor of 4 to 8. The Tinn granite also has low concentrations of feldspar-compatible trace elements (Ba, Sr, Pb, Eu; Table 1), which does not agree with the presence of accumulated feldspar.

The compositional data thus suggest that the Tinn granite is neither a differentiate of a Tuddal parent magma, a cumulate formed from such a magma, nor a simple anatectic melt of a Tuddal Formation protolith. The granite could represent a retarded batch of an undifferentiated parent magma related to the Tuddal rhyolite, but the time interval between the end of rhyolitic volcanism and emplacement of the granite may be too long for a single silicic magmatic system to have remained active.

The observed compositions of the Tinn granite can be adequately explained by mixing between a minimum melt and a low *qz*- high *an* melt, i.e. between an anatectic melt from a rhyolite-like protolith and a mafic magma. The thickness of the Tuddal Formation has been estimated to be minimum 7 km (Sigmond 1998). Rocks cogenetic with the rhyolite, with anomalously low Sr concentration, must also be present at greater depth in the Telemark area (Andersen et al. 2001, Andersen & Knudsen 2000). 10-20 Ma after eruption of the Tuddal Formation rhyolite, the volcanic pile and related intrusions at deeper levels in the crust would probably be hot enough to partially melt when heated up by injection of mafic magma. There is abundant evidence of mafic to intermediate magmatic activity in Telemark after eruption of the Tuddal Formation rhyolite (mafic members of the Uvdal plutonic belt, Vemork Formation basalts, e.g. Sigmond 1998), providing a source for the necessary extra thermal energy and a mafic component. An open-system process, involving mafic magma and material derived from a crustal protolith related to the Tuddal Formation, is therefore the preferred petrogenetic model for the Tinn granite. The Tinn granite must have formed during a period of crustal extension that permitted mafic magma to ascend to a high crustal level and, thereby, cause partial melting of the upper crust as well as to mix and mingle with the derived silicic melts.

Conclusions

Zircons from the Tinn granite were dated at 1476 ± 13 Ma, which suggests that emplacement of the granite post-dates the felsic volcanism that gave rise to the Tuddal Formation by at least 11 Ma, accepting the 1500-1514 Ma age estimate of the Tuddal Formation by Sigmond (1998). Radiogenic isotope data and rare inherited zircon cores indicate that no material significantly older than the Rjukan Group was involved in its petrogenesis. Whole-rock major element data from the Tinn granite suggest that the granitic magma formed by an open system process in which partial melts from a protolith with age and Nd isotope systematics indistinguishable from the Tuddal Formation were mixed with mafic material. Partial melting may have been induced by emplacement of mafic magma into the middle to upper crust of the Telemark sector after termination of the rhyolitic volcanism, but while the crust still remained hot.

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Devonian ages from $^{40}Ar/^{39}Ar$ dating of plagioclase in dolerite dykes, eastern Varanger Peninsula, North Norway

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Guise, P. G. & Roberts, D. 2002: Devonian ages from $^{40}Ar/^{39}Ar$ dating of plagioclase in dolerite dykes, eastern Varanger Peninsula, North Norway. *Norges geologiske undersøkelse 440*, 27-37.

Gas-release spectra derived from the analysis of plagioclase from three geographically distinct but geochemically comparable dolerite dykes from the eastern part of Varanger Peninsula, northern Norway, show similar features and favour an interpretation that the dykes were intruded in Late Devonian time at around 370 Ma. These particular dykes had previously yielded fairly similar, K-Ar whole-rock ages. As one of the dykes had earlier been traced into the Trollfjorden-Komagelva Fault Zone with the aid of a proton-magnetometer, this would indicate that all significant, displacive movement along this major fault zone had ceased by latest Devonian time. The dyke ages reported here fit into a known pattern of Mid Devonian to Early Carboniferous rifting and sporadic mafic magmatism reported from adjacent parts of Kola Peninsula and neighbouring areas along the northeastern margin of the Fennoscandian Shield.

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Introduction

In the Caledonides of Scandinavia, many of the lithostratigraphical successions in diverse thrust sheets are intruded by mafic dykes. Occurring either in isolation or, in a few places, in swarms, dykes of this type generally relate to important phases of crustal extension and/or rift magmatism or, less commonly, to late-stage emplacements along joints or faults in compressional regimes. Subtle variations in chemical composition also allow for fair assessments to be made of the likely palaeotectonic settings of these hypabyssal rocks. In the few cases where isotopic ages are available, these serve as a bonus in helping us to define the local geological history with greater precision.

On Varanger Peninsula in Finnmark, in the extreme northeast of Norway (Fig. 1), the only sign of igneous activity is provided by dolerite dykes that cut Riphean to Vendian lithostratigraphic successions exposed on either side of the major, WNW-ESE trending, Trollfjorden-Komagelva Fault Zone (TKFZ). The geology of this peninsula is now well known through the systematic mapping and stratigraphic and sedimentological studies of Siedlecka & Siedlecki (1967) and Siedlecki (1980), summarised in Siedlecka & Roberts (1992). However, isotopic dating studies on the dykes of the peninsula have been few.

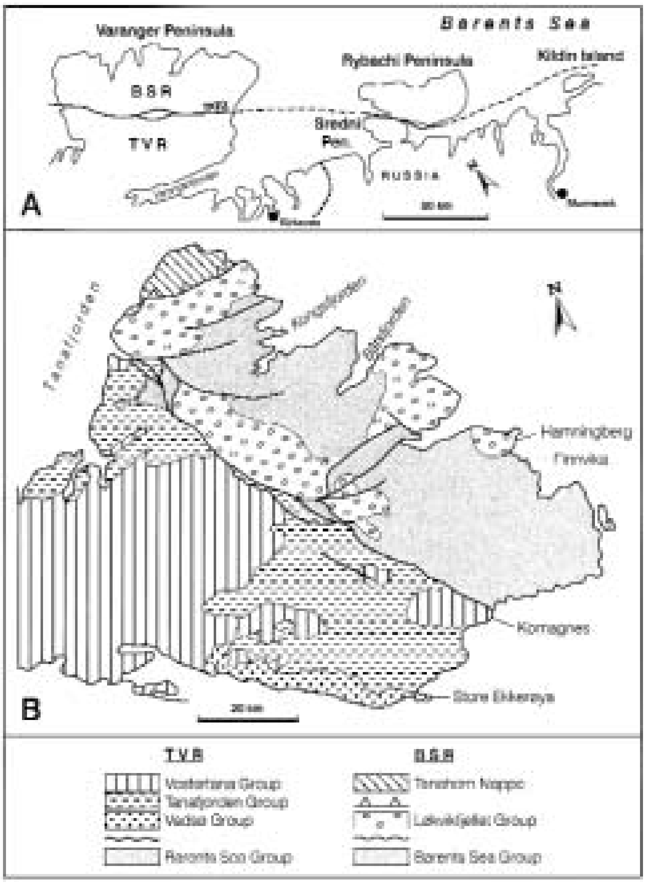

Fig. 1. (a) Outline map showing the location of the Rybachi and Sredni Peninsulas in relation to Varanger Peninsula. BSR – Barents Sea Region; TVR – Tanafjorden-Varangerfjorden Region; TKFZ – Trollfjorden-Komagelva Fault Zone. (b) Much simplified geological map of Varanger Peninsula, showing the locations of the studied dykes at Finnvika, Komagnes and Store Ekkerøya. Separate simplified legends are shown for the TVR and BSR. Major faults are indicated by continuous thick lines.

A K-Ar whole-rock investigation of dolerite/metadolerite dykes from different areas on the peninsula by Beckinsale et al. (1975) is the only detailed work published so far. Other investigations restricted to particular dykes are those of Roberts et al. (1995) and Roberts & Walker (1997). There have also been studies dealing with palaeomagnetic dating of certain dykes (Knutsen 1995, Torsvik et al. 1995).

In this contribution, we present the results of a $^{40}Ar/^{39}Ar$ investigation of plagioclase separated from two prominent dolerite dykes from the southeastern part of Varanger Peninsula, south of the TKFZ, one of which has earlier yielded disparate K-Ar and palaeomagnetic ages – Late Devonian and Vendian, respectively. The $^{40}Ar/^{39}Ar$ analytical data, in our view, help to resolve these differences. We also present plagioclase analytical data from another dyke, sampled from north of the TKFZ. All three dykes have given similar plateau ages.

Geological setting

Our knowledge of the Neoproterozoic-Early Palaeozoic geological evolution of Varanger Peninsula derives from diverse, detailed investigations of lithostratigraphy, tectonic structure, low-grade metamorphism, micropalaeontology and remote sensing applications. Specific accounts or reviews include those of Siedlecka & Siedlecki (1967, 1971), Banks et al. (1971), Roberts (1972), Siedlecka (1975), Johnson et al. (1978), Pickering (1981), Taylor & Pickering (1981), Vidal (1981), Edwards (1984), Bevins et al. (1986), Rice et al. (1989), Karpuz et al. (1993), Rice (1994), Rice & Reiz (1994) and Laajoki (2002). In addition, the effects of Timanian deformation in the easternmost parts of Varanger Peninsula has been discussed by Roberts (1995, 1996).

The Trollfjorden-Komagelva Fault Zone provides a natural, structural divide, separating the peninsula into a northeastern *Barents Sea Region (BSR)* and a southwestern *Tanafjorden-Varangerfjorden Region (TVR)*. The established, formal lithostratigraphies for the two regions are presented in Fig. 2. Although details do not concern us here, the TVR has been termed a *pericratonic* sedimentation domain and the BSR a *basinal* domain (Siedlecka & Roberts 1995), with reference to the developing, northeastern passive margin of Baltica in Riphean-Vendian time (Olovyanishnikov et al. 2000, Roberts & Siedlecka 2002). Importantly, the critical stratigraphic relationship between the two domains, involving an unconformity where the Ekkerøya Formation lies directly upon a steeper dipping Båtsfjord Formation, has been described by Rice (1994) from one small area in western Varanger Peninsula. This unconformity has also been documented by Roberts & Karpuz (1995).

Dolerite dykes are particularly common in certain parts of the Barents Sea Region, especially in northwestern and central areas. On the contrary, dykes are extremely rare in the TVR. Based on their K-Ar results, Beckinsale et al. (1975) distinguished two principal groups of dyke ages: (A) c. 360 Ma, and (B) c. 650 Ma (both ages recalculated after Dalrymple 1979); and a third group (C) of strongly cleaved dykes with questionable 'ages' of 945 to >1900 Ma. Dykes of age-groups B and C are restricted to the rocks of the BSR, whereas the far less common group A dykes occur on either side of the Trollfjorden-Komagelva Fault Zone. Later, unpublished work by Beckinsale (pers. comm. 1977) tended to favour an age closer to 550-560 Ma for the dykes of Group B. Palaeomagnetic studies by Knutsen (1995) also supported a Vendian age. An attempt to provide better age constraints for the dykes of Group A by application of the $^{40}Ar/^{39}Ar$ method to pyroxenes (Roberts et al. 1995) was not success-

1

Age	Lithostratigraphic units and their thicknesses		
		Formation	Member
UPPER RIPHEAN	TANAFJORDEN GROUP 1448-1665m	Grasdalen 280 m	Upper Lower
		Hanglecærro 200 m	
		Vagge 80 m	
		Gamasfjellet 280-300 m	
		Dakkovaire 273-350 m	Ferruginous sandstone 130 m 'k' member 62 m 'j' member 46 m 'i' member 35 m Quartzitic sandstone 60-80 m
		Stangenes 205-255 m	
		Grønneset 130-200 m	
	VADSØ GROUP, 590 - 960 m	Ekkerøya 15-190 m	★ V8
		Golneselva 50-135 m	
		Paddeby 25-120 m	
		Andersby 25-40 m	
		Fugleberget 125 m	
		Klubbnasen 50 m	
		Veinesbotn 300 m	

2

Age	Lithostratigraphic units and their thicknesses		
		Formation	Member
CAMBRIAN - ORDOVICIAN	DIGERMULEN GROUP 1510 - 1555 m	Berlogaissa 300 m	
		Kistedalen 710-735 m	Grey quartzite 200 m Black shale 200 m Black quartzite 10-35 m Sandstone and shale 200 m Quartzite and shale 100m
		Duolbasgaissa 500-520 m	Massive bedded quartzite 300 m Thin-bedded quartzite 200-220 m
	VESTERTANA GROUP 1317 - 1665 m	Breivika 600 m	
VENDIAN		Stappogiedde 505-545 m ★ V1	Manndrapselva 190 m Innerelva 275 m Lillevatnet 40-80 m
		Mortensnes 10-60 m	
		Nyborg 200-400 m	
		Smalfjord 2-50 m	

3

Age	Lithostratigraphic units and their thicknesses		
		Formation	Member
RIPHEAN - VENDIAN	LØKVIKFJELLET GROUP 5710 - 5810 m	Skidnefjellet >800 m	
		Stordalselva 1200 m	
		Skjærgårdsneset 210 m	
		Styret 1500-1600 m	
		Sandfjorden 2000 m	
	BARENTS SEA GROUP 8900 - 10 000 m	Tyvjofjellet 1500 m	
		Båtsfjord 1400-1600 m	Skovika 1100-1300 m Annijokka 300 m
		Båsnæringen 2500-3500 m	Hestman 600-1300 m Godkeila 490-1450 m Seglodden 100-350 m Næringselva 500-1200 m
		Kongsfjord >3500 m ★ V4	Nålneset 2000 m Risfjorden 1000-1500 m

Fig. 2. Lithostratigraphic successions, Varanger Peninsula, showing the locations of the investigated samples of dolerite dykes, V1, V4 and V8. Columns 1 and 2 are from the Tanafjorden-Varangerfjorden Region, southwest of the TKFZ (Fig. 1), and column 3 from the Barents Sea Region, northeast of the fault zone.

ful, though the dataset did show a slight bias towards a possible Devono-Carboniferous age.

The Group C dykes, generally termed metadolerites, are particularly common in the Kongsfjord district (Fig. 1). They carry a penetrative cleavage which is also axial planar to abundant ENE-WSW-trending folds in the country rocks (Roberts 1972, Rice & Reiz 1994), and many of the dykes are boudinaged. One of these Kongsfjord dykes has been dated by the Sm-Nd method to around 550 Ma (data attributed to B.Sundvoll in Andersen & Sundvoll, 1995).

Group B dykes are most profuse in the Båtsfjord area (Fig.1). They are very weakly cleaved, a cleavage which is again parallel to the axial surfaces of open to tight folds in the host metasedimentary rocks. The dykes, cleavage and fold axes all trend approximately ENE-WSW. The Group A dolerites are comparatively fresh and either unmetamorphosed or very slightly altered. They generally trend between N-S and NE-SW.

Geochemically, the Group C metadolerites carry signatures quite close to those of abyssal tholeiites, though somewhat transitional to a continental margin regime (Roberts 1975). The Group A dolerites, on the other hand, have chemical features more akin to those of continental tholeiites developed in a plate-marginal rather than continental interior situation. Some unpublished geochemical data do exist for dykes of Group B (D.Roberts, in prep.), indicating that they are of transitional, oceanic/continental tholeiite character.

In addition to the above, it should be mentioned that just one dolerite dyke sampled from the eastern side of Båtsfjorden has provided indications of a considerably younger age. Based on preliminary Sm-Nd and Rb-Sr analytical data, this particular dyke may possibly have been emplaced during the Jurassic period (B.Sundvoll, written communication 1991).

A few dolerite dykes also occur on the nearby Rybachi and Sredni Peninsulas in NW Russia (Fig. 1a), cutting lithostratigraphical successions that are comparable to those on Varanger Peninsula (Polkanov 1935, Sinitsin 1967, Bekker et al. 1976, Lyubtsov et al. 2000); and similar dykes are also present on the adjacent mainland of the Kola Peninsula cutting Archaean and Palaeoproterozoic gneisses (Fieandt 1912, Hausen 1932, Fedotov & Amelin 1998). Some of these Kola dykes have yielded Vendian ages, whereas others are latest Devonian (Juve et al. 1995, Roberts & Onstott 1995, V. Negrutsa, pers. comm. 1991) to Early Carboniferous (Fedotov & Amelin 1998). In one case, on Rybachi, there is conflicting evidence from $^{40}Ar/^{39}Ar$ and palaeomagnetic dating of one particular swarm of mafic dykes, where either Vendian or Devonian ages have been suggested (cf. Torsvik et al. 1995, Roberts & Onstott 1995).

The investigated dolerite dykes

The dykes investigated in this argon-dating study occur in the TVR near Komagnes (sample V1) and on the island of Store Ekkerøya (V8), and in the BSR close to the small bay Finnvika (V4) (Fig. 1). Their locations in the lithostratigraphical successions are shown in Fig. 2. Although Beckinsale et

Fig. 3. (a) The dolerite dyke just west of Komagnes, cutting thin-bedded shales and mudstones of the Innerelva Member of the Stappogiedde Formation; photo looking almost due north. (b) Foreshore exposure of the dolerite dyke just northwest of Finnvika, cutting low-grade, turbiditic sedimentary rocks of the Kongsfjord Formation, Barents Sea Group.

al. (1975) did not give any precise sampling localities, it seems reasonably certain that all three dykes analysed for the present study belong to their Group A classification, i.e., with K-Ar whole-rock ages of around 360 Ma. With reference to the sample numbers V1, V4 and V8, these appear to correspond to samples R12, R51 and M10 of Beckinsale et al. (1975; cf. their fig.1 and table 1).

Field relationships

Komagnes dyke

This c. 2.5 m-thick dyke is easy to detect in the old, raised cliff Giviaida, north of the main road, c. 1 km west of the promontory Komagnes (Fig. 3), on 1:50,000 map-sheet 2435 II Ekkerøy (4-NOR edition, grid-reference 038 910). The dyke trends c. N-S and dips at 75° east, and cuts through flat-lying, thin-bedded, blue-green shales and reddish-grey mudstones of the Innerelva Member of the Late Vendian, Stappogiedde Formation, the highest part of the Vestertana Group (Siedlecka & Roberts 1992) (Fig. 2). Another, similar, 40 cm-thick dolerite dyke is present in the same raised cliff c. 150 m west of the main dyke, but this is not considered further here.

Closer inspection reveals that the Komagnes dyke in the cliff-face really consists of two parallel dykes with a thin screen of hornfelsed sediment in between. Towards the foreshore, and in the intertidal zone, the dyke (or composite dyke) is nearer to vertical and even west-dipping, and splits into several thinner dykes with offshoots which locally curve into a bedding-parallel orientation. Adjacent to the dyke, or dykes, a near-vertical, widely spaced fracture cleavage is present in the hornfelsed shales. The Innerelva Member otherwise shows little effects of tectonic deformation. There is a good compactional fabric in the shales, however, and this burial diagenetic event has been dated (Rb-Sr on illite subfractions) to c. 560 Ma (Gorokhov et al. 2001).

In thin-sections of the dyke, a uniform mineralogy is dominated by plagioclase and clinopyroxene with ophitic to subophitic texture. The plagioclase is andesine to labradorite and locally shows oscillatory zoning, and the clinopyroxene shows the optical properties of pigeonite (Roberts 1975). Accessory minerals are (titano)magnetite, apatite, rare calcite and traces of interstitial or pyroxene-marginal chlorite. The feldspar shows only incipient stages of sericitisation.

Finnvika dyke

This 9-9.5 m-thick dyke is located c.1 km northwest of Finnvika on the northern coast of Varanger Peninsula (Fig. 1); on 1:50,000 map-sheet 2436 II Syltefjord (4-NOR edition, grid-reference 108 235). The dyke trends between NE-SW and ENE-WSW and here dips at c. 75° southeast, cutting through thick-bedded, immature sandstones and intercalated, cleaved shaly units of the low-grade, turbiditic Kongsfjord Formation, the lowest stratigraphic unit of the Barents Sea Group (Figs. 2 & 3).

The dyke contacts are subparallel to bedding, but in the foreshore exposures the dyke locally transects bedding at a low angle in a left-stepping sense before resuming a bedding-parallel attitude. Cleavage in the pelites varies between vertical and east-southeast dip orientations. The dyke shows no evidence of deformation or metamorphism, other than small offsets along WNW-ESE faults, and clearly postdates the pervasive cleavage and some related, small-scale upright folds. In this same general area and same formation, Taylor & Pickering (1981) reported a Rb-Sr whole-rock isochron age of 520 ± 47 Ma on cleaved mudstones, interpreted to date the folding and associated axial-surface cleavage.

Petrographically, the Finnvika dyke is similar to the one from Komagnes, with plagioclase and clinopyroxene showing subophitic texture; and the feldspar looks to be quite fresh.

Store Ekkerøya dyke

A near-vertical, N-S-trending dolerite dyke reaching up to 16 m in thickness cuts through gently NE-dipping strata of the Ekkerøya Formation, the highest unit of the Vadsø Group, in the southwestern part of the island of Store Ekkerøya (Figs. 1 & 2). The sampling locality is on 1:50,000 map-sheet 2435 II Ekkerøy (4-NOR edition) at grid reference 890 774. The dyke is particularly prominent in the cliffs at Flågan, a nature reserve and bird colony. The Ekkerøya Formation here consists of medium-bedded sandstone and subordinate conglomerate with intercalations of siltstone and mudstone (Siedlecka & Roberts 1992), and is devoid of cleavage or folds.

In thin-section, the central part of the Store Ekkerøya dyke is coarser grained than the other two dolerite dykes and varies in texture from ophitic to locally glomeroporphyritic, with the plagioclase laths occurring in scattered clusters up to 3.5-4 mm across. Both these feldspar clusters and the normal, individual laths show variable degrees of sericitisation. Nearer the dyke margins, grain size is comparable to that in the Komagnes and Finnvika dyke samples, and sericitisation is more advanced. The mineral paragenesis is otherwise the same as for the Komagnes and Finnvika dykes.

$^{40}Ar/^{39}Ar$ dating

Analytical procedure

The plagioclase separates were prepared at the Geological Survey of Norway (NGU), Trondheim, and the samples irradiated at Risø Nuclear Reactor, Roskilde, Denmark. Full procedural details are presented in an Appendix. The fast neutron dose was monitored by Leeds biotite standard Tinto, 409.2 Ma (Rex et al. 1986) and hornblende Hb3gr, 1072 Ma (Turner et al. 1971). Flux variation over the package length was of the order of 3%. The interference correction factors used were; (40/39)K = 0.048, (36/39)Ca = 0.38 and (37/39)Ca = 1492.

Table 1. $^{40}Ar/^{39}Ar$ analytical data.

Temp °C	$^{39}Ar_K$	$^{37}Ar_{Ca}$	$^{38}Ar_{Cl}$	Ca/K	$^{*40}Ar/^{39}Ar_K$	%Atm ^{40}Ar	Age (Ma)	Error (Ma)	$\%^{39}Ar_K$
	{ Vol. x 10^{-9} cm^3 }								
660	0.20	3.1	0.008	32	52.47	58.6	442.7	16.6	4.5
755	0.53	11.7	0.009	44	43.87	40.7	377.1	3.3	12.2
830	0.68	16.6	0.008	49	43.82	23.4	376.7	4.8	15.6
885	0.55	12.5	0.008	45	43.72	14.4	376.0	3.7	12.7
930	0.39	7.7	0.006	39	44.00	35.9	378.2	5.2	9.0
980	0.32	4.3	0.005	27	44.33	39.5	380.7	10.4	7.2
1010	0.24	2.4	0.006	20	44.77	22.2	384.1	9.8	5.5
1060	0.32	4.0	0.006	25	44.86	17.5	384.8	6.1	7.3
1165	0.67	10.3	0.015	31	49.64	15.7	421.4	5.2	15.3
1300	0.46	8.8	0.012	38	52.45	46.1	442.5	5.3	10.5
V1 Plagioclase, run 2461 weight = 0.05322g, J value = 0.00530 ± 0.5 %									
Total gas age 395 ± 3Ma (weight %K = 0.22 , $^{*40}Ar$ = 37.7 x 10^{-7} cm3g^{-1})									
630	1.5	1.8	0.04	2.3	42.61	37.3	367.4	2.8	5.8
760	4.7	15.0	0.07	6.3	42.81	9.6	368.9	0.4	17.9
830	2.9	8.6	0.04	6.0	43.22	9.3	372.1	1.0	10.9
865	1.7	3.5	0.03	4.1	42.94	14.0	369.9	2.0	6.4
905	1.5	2.1	0.03	2.7	42.58	12.2	367.1	1.4	5.8
955	1.8	1.5	0.03	1.6	42.79	17.3	368.8	1.7	7.0
1000	2.0	1.6	0.04	1.6	42.87	16.5	369.4	2.3	7.5
1050	3.5	3.3	0.06	1.9	42.43	8.9	365.9	0.8	13.2
1160	5.9	8.7	0.12	3.0	42.72	7.0	368.2	0.4	22.2
1300	0.9	5.5	0.03	12.2	46.17	32.7	394.9	2.8	3.4
V4 Plagioclase, run 2462 weight = 0.06140g, J value = 0.00530 ± 0.5 %									
Total gas age 369 ± 2Ma (weight %K = 1.2 , $^{*40}Ar$ = 185 x 10^{-7} cm3g^{-1})									
555	1.4	0.5	0.14	0.8	35.22	80.0	308.7	4.2	2.1
680	3.6	3.7	0.17	2.0	42.66	29.2	367.8	0.8	5.5
750	5.4	10.3	0.10	3.8	44.78	9.0	384.2	0.7	8.2
825	5.7	7.1	0.13	2.5	43.98	14.2	378.0	0.6	8.7
870	5.7	3.2	0.11	1.1	43.69	9.2	375.7	0.5	8.7
920	8.7	3.4	0.17	0.8	43.44	8.0	373.8	0.3	13.3
975	9.8	2.5	0.22	0.5	43.54	10.1	374.6	0.2	15.0
1025	10.5	3.3	0.28	0.6	43.60	12.7	375.1	0.2	16.0
1120	10.0	7.3	0.66	1.5	51.96	20.9	438.8	0.4	15.3
1300	4.7	9.1	0.43	3.9	59.08	22.0	491.4	0.6	7.2
V8 Plagioclase, run 2460 weight = 0.08236g, J value = 0.00530 ± 0.5 %									
Total gas age 393 ± 2Ma (weight %K = 2.1 , $^{*40}Ar$ = 366 x 10^{-7} cm3g^{-1})									

Errors are 1σ. $^{*40}Ar$ I volume of Radiogenic ^{40}Ar, gas volumes corrected to STP.

Isotopic analyses were performed with a modified MS10 mass spectrometer; measured atmospheric $^{40}Ar/^{36}Ar$ was 287.8 ± 0.2 and sensitivity 1.1 x 10^{-7} cm^{3}V^{-1}. The J-value uncertainty is included in the errors for the total gas ages but the individual step ages have analytical errors only. All errors are quoted at the 1-sigma level. The analytical data are presented in Table 1. Age spectra and age correlation plots were produced using 'Isoplot/Ex' (Ludwig 2000).

Results and interpretation

V1 plagioclase: The age spectrum shows some disturbance with older ages at both low- and high-temperature, gas release steps. A plateau of 377.6 ± 1.8 Ma is defined by 70% of the gas released (Fig. 4). The isotope correlation plot gives a good linear trend for the plateau steps and yields the same age. The intercept $^{40}Ar/^{36}Ar$ of 296 is close to the accepted atmospheric argon value of 295.5. The last two steps of the experiment lie on the excess argon side of the linear array, evidence for the contribution of older ages of these steps.

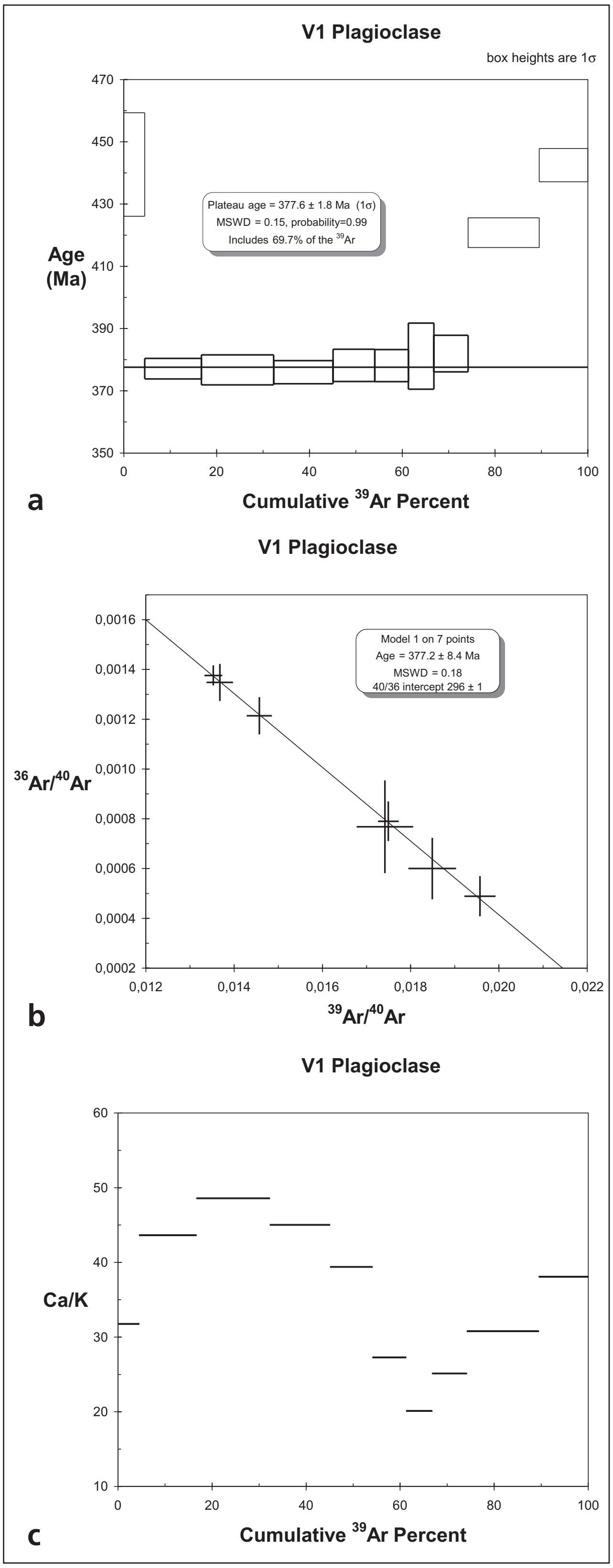

Fig. 4. Plagioclase sample V1: (a) $^{40}Ar/^{39}Ar$ age spectrum; (b) isotope correlation plot; (c) Ca/K plot.

Variations in the Ca/K ratio show no correlation with the age spectrum. This indicates that all the slight compositional variations present in the plagioclases are therefore giving the same age. It should be noted that the K content of this particular plagioclase is much lower than that of the other two samples, resulting in the larger errors on ages for individual steps and higher Ca/K values. This may suggest the hidden presence of subtle differences in the mineralogy of this particular separate, compared with that of V4 and V8.

We interpret the 378 ± 2 Ma plateau age (Fig. 4) as closely corresponding to the actual crystallisation age of this dyke.

V4 plagioclase: The age spectrum of V4 shows a well-defined plateau at 369 Ma consisting of more than 95% of the gas released (Fig. 5). This age is further confirmed by the isotope correlation plot which has a good linear trend with the $^{40}Ar/^{36}Ar$ intercept at 297, again close to the atmospheric argon value. In this sample, there is no indication of excess argon being present. This spectrum conforms to that expected of a sample that underwent rapid cooling and has since remained thermally undisturbed. The variation in the Ca/K ratios of the steps is not reflected in the age spectrum.

The 369 ± 0.23 Ma plateau looks to provide as good an age as one is likely to get from this technique, and this is supported by the inverse isochron plot. Accordingly, we interpret this to represent the crystallisation age of the Finvika dyke.

V8 plagioclase: This sample shows a disturbed spectrum with evidence of argon loss in the low-temperature steps and the presence of excess argon in the high-temperature steps (Fig. 6). The disturbance is such that there is no plateau as defined by the criteria of Isoplot/Ex (Ludwig 2000). However, a fit was forced through the steps shown in Fig. 6 to give an age of 375 ± 1 Ma (95% confidence) with just over 50% 0f the gas released. The isotope correlation plot again gives a linear trend, but confirms the presence of excess argon in the two highest temperature steps. A line fitted through the first eight steps of the experiment gives an age of 377 Ma and an $^{40}Ar/^{36}Ar$ intercept of 278 with large uncertainty. This value is lower than the accepted atmospheric argon value, which would be expected when there has been loss of radiogenic ^{40}Ar. Variation in the Ca/K ratio is approximately mirrored in the age spectrum, indicating that slightly differing compositional variations could also be contributing to the disturbance of the age spectrum.

We interpret the weighted mean age of c. 375 Ma as a likely approximation to the intrusive age of this dyke.

Of the three samples, the age of V4 at 369 Ma can be accorded the highest confidence. The other two samples, V1 and V8, both show varying degrees of disturbance in their spectra. It is noteworthy that the V4 sample was collected from the very centre of the 9 m-thick Finnvika dyke. Sample V1, on the other hand, was taken c. 80 cm in from the margin of the 2.5 m-thick Komagnes dyke; and V8 approximately 1 metre in from the margin of the 16 m-thick Store Ekkerøya dyke. It is therefore possible that the disturbances recorded

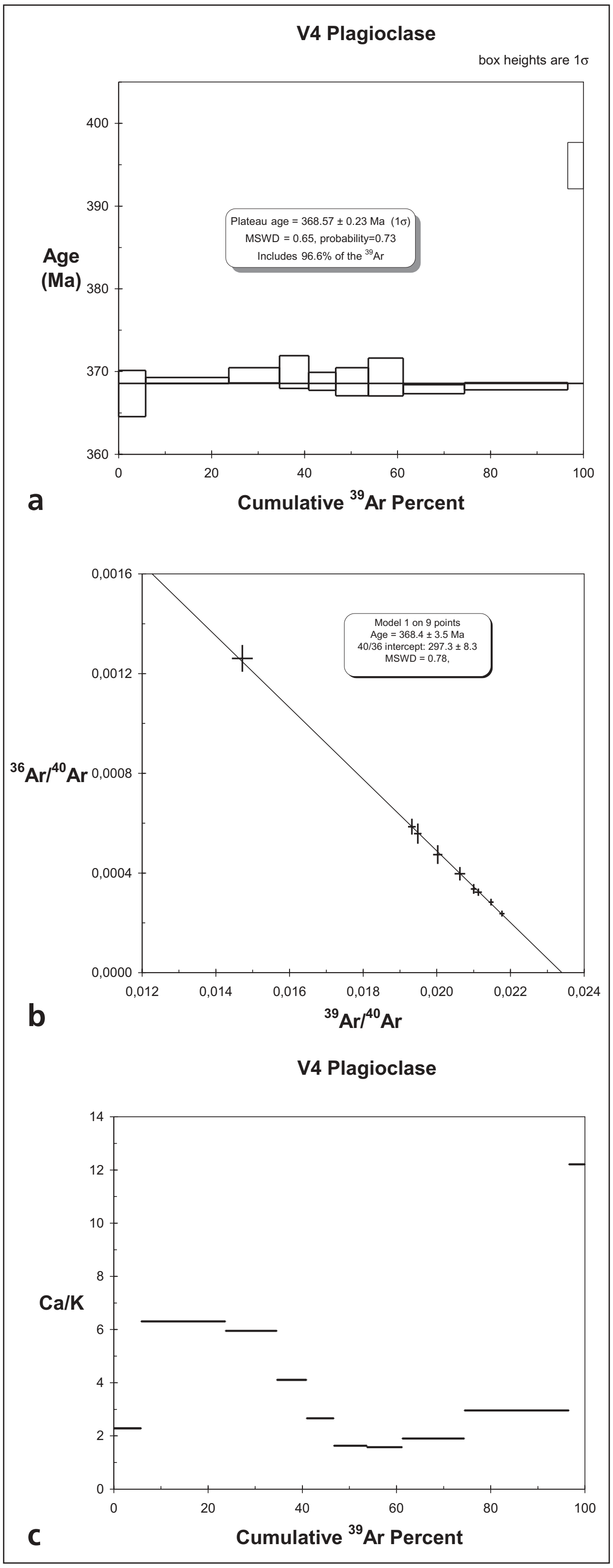

Fig. 5. Plagioclase sample V4: (a) $^{40}Ar/^{39}Ar$ age spectrum; (b) isotope correlation plot; Ca/K plot.

in V1 and V8 may relate to acquisition of radiogenic argon from the country rocks during magma ascent; or from fluids

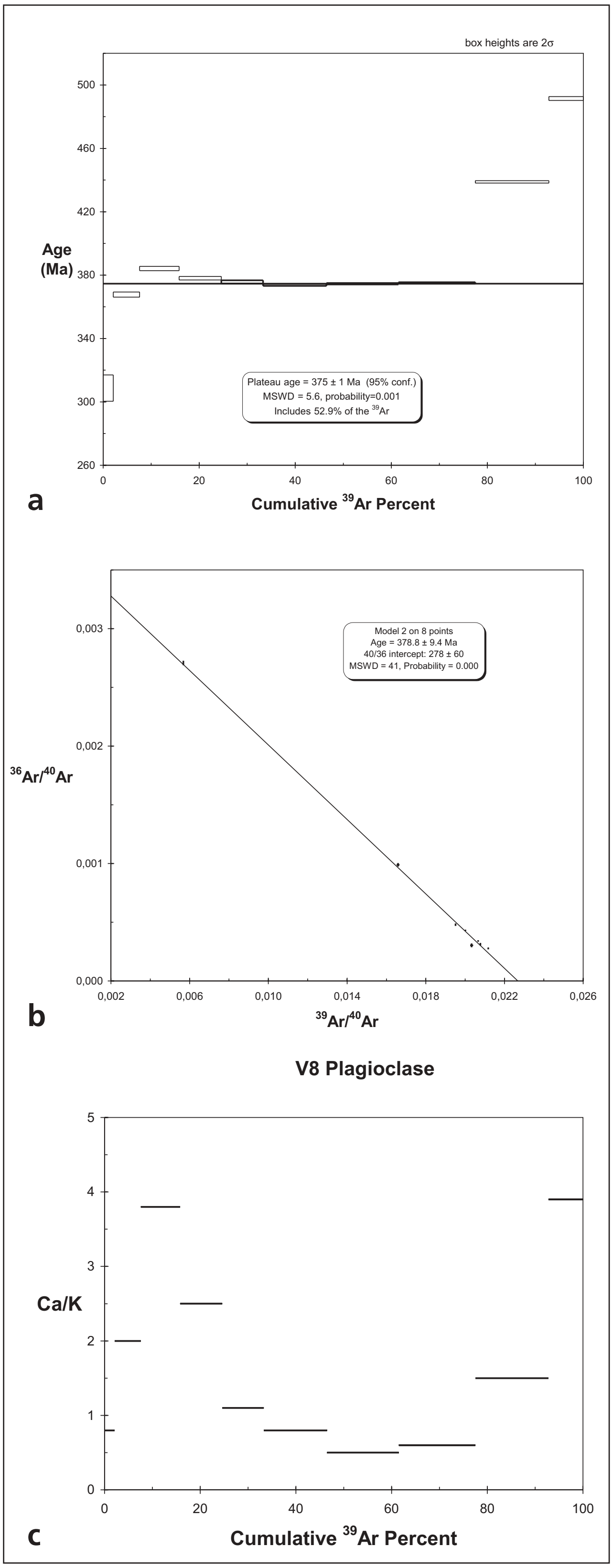

Fig. 6. Plagioclase sample V8: (a) $^{40}Ar/^{39}Ar$ age spectrum; (b) isotope correlation plot; (c) Ca/K plot.

circulating along or close to the dyke margins. Whatever the case, the three, fairly similar ages do tend to confirm the occurrence of a significant event – and mostly likely one of dyke intrusion – at around 370 Ma.

Discussion

Based on all published dating, both isotopic and palaeomagnetic, of mafic dykes cutting Neoproterozoic rocks from the Varanger-North Kola segment of the northern margin of the Fennoscandian Shield, there appear to be two principal ages of dolerite dyke intrusion, namely Vendian and Devonian. The dykes investigated by us belong to Group A of Beckinsale et al. (1975), i.e., those which provided Late Devonian, K-Ar whole-rock ages. One of these dolerite dykes, the V1 dyke from Komagnes, has also been analysed palaeomagnetically, and in this case provided a Vendian age with no indication of any Devonian magnetic resetting (Torsvik et al. 1995). The palaeomagnetic data from the Komagnes dyke are almost identical to those derived from a dolerite dyke from the Sredni Peninsula that, in this case, has provided a Vendian age by both K-Ar and $^{40}Ar/^{39}Ar$ analytical methods.

Notwithstanding this evident conflict of results between the palaeomagnetic and isotopic dating of the Komagnes dyke, the plagioclase age spectra reported here from these three, widely separated dykes are mutually comparable and, together, favour an interpretation that the dykes were emplaced and crystallised at around 370 Ma. This interpretation thus supports the earlier K-Ar dating study of Beckinsale et al. (1975), where these very same dykes yielded concordant maximum ages of around 363 ± 10 Ma (recalculated following Dalrymple 1979). On current Phanerozoic time-scales, the Devonian-Carboniferous boundary is placed at either 362 (Tucker et al. 1998) or 355 Ma (Remane et al. 2000), and the Varanger dykes analysed in this investigation would thus fall in the Famennian stage of Late Devonian time.

In a wider perspective of the Fennoscandian Shield and East European Craton, the eastern and northern margins of Baltica were characterised by a major episode of rifting in Mid Devonian to Early Carboniferous time (Ziegler 1988, Johansen et al. 1993, Nikishin et al. 1996). Rift basin formation along a mainly NW-SE trend occurred beneath the Pechora Basin and eastern Barents Sea (Fig. 7), mimicking the structural trend in the pre-Palaeozoic basement. In the Kola Peninsula region and western parts of the Barents Sea, a more NE-SW rifting trend is evident (Nikishin et al. 1996, Gudlaugsson et al. 1998, Wilson et al. 1999), and a general doming of the Kola-White Sea area occurred in Late Devonian time (Fig. 7). In the Timan-Pechora rift system, Late Devonian basaltic volcanism was widespread, and the Kontozero graben on Kola Peninsula is well known for its alkaline and kimberlitic magmatism during the period 380 to 360 Ma (Kramm et al. 1993). In western parts of Kola, Pb-Zn vein mineralisations of Late Devonian age are commonly associated with N-S to NE-SW trending dolerite dykes (Juve et al. 1995).

An allied topic is the fact that U-Pb zircon data from

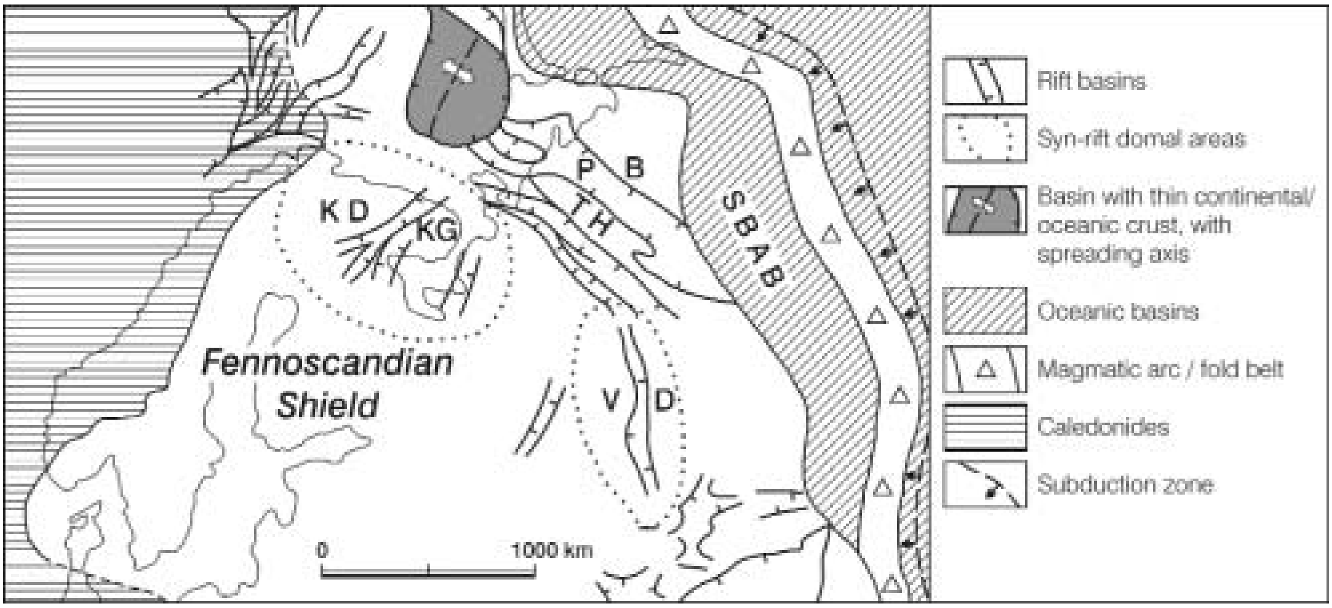

Fig.7 Simplified sketch-map of Baltica showing the main Late Devonian-Early Carboniferous rifts basins and other features. Modified from Nikishin et al. (1996). KD – Kola Dome; KG – Kontozero graben; PB – Pechora Basin; SBAB – Sakmarian back-arc basin; TH – Timan High; VD – Vyatka Dome.

Palaeoproterozoic and Archaean rocks from this northern Fennoscandian domain commonly show Devonian to Early Carboniferous lower intercepts on concordia (e.g., Levchenko et al. 1995, Larson & Tullborg 1998). A similar, Devonian, lower intercept date has been reported from a dolerite dyke with an interpreted Vendian age (U-Pb, zircon) from near Hamningberg in the Barents Sea Region (Roberts & Walker 1997). This recurrent feature has been interpreted by Larson & Tullborg (1998) to relate to the thermal effects of a Devonian foreland basin, with a sediment cover ⩾3 km thick (now removed), arising from rapid erosion of the Caledonian mountain chain. Although support for this notion of a Devonian sedimentary blanket comes from ongoing fission-track studies (G. Murrell, pers. comm. 2002), Nikishin et al. (1996) have stated that the Fennoscandian Shield, except for its rift basins, "may have remained emergent throughout the Devonian". Whether or not a thick, Devonian, foreland basin sedimentary cover existed in this shield area, there is now sufficient evidence from this northern Fennoscandian region that there are mafic dykes of both Devonian and Vendian age, and that the latter may or may not show Devonian overprints.

Returning to Varanger Peninsula, and the Komagnes dyke in particular, Herrevold (1993) and Karpuz et al. (1993) have reported that the dyke could be traced inland with the help of a handborne proton-magnetometer beneath a thin Quaternary cover into the trasé of the Trollfjorden-Komagelva Fault Zone, without change of strike, until its magnetic signature eventually faded and disappeared. No strike-slip offset of the magnetic anomaly could be detected. Accepting that the dyke is almost certainly of Devonian age, then this would indicate that all major strike-slip movements along the fault zone had ceased by Late Devonian time. A similar suggestion was also made by Beckinsale et al. (1975), at that time based on the general, though sporadic occurrence of their Group A dykes on either side of the TKFZ.

Conclusions

Age spectra derived from the analysis of plagioclase separates from three, geochemically similar dolerite dykes from the eastern part of the Varanger Peninsula, northern Norway, show comparable features and lead to an interpretation that the dykes were intruded in Late Devonian time, at c. 370 Ma. These particular dykes had previously yielded quite similar, K-Ar whole-rock ages. An inferred Vendian age for one of the dykes, at Komagnes, based on palaeomagnetic data, should now be dismissed. This particular dyke has been traced beneath thin superficial deposits, by magnetometer, directly into the Trollfjorden-Komagelva Fault Zone, without any visible strike-slip offset of the magnetic anomaly, thus indicating that all major displacive movement along this fault zone had ceased by latest Devonian time.

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APPENDIX

Plagioclases were separated at NGU, Trondheim, using standard procedures. Samples for $^{40}Ar/^{39}Ar$ analysis were individually weighed, wrapped in high-purity aluminium foil and loaded into a Spectrosil phial. Irradiation for 10 hours was carried out at the Riso facility, Roskilde, Denmark. A fast neutron dose of approximately 9 x 10^{17} neutron/cm^2 was given and monitored by Leeds biotite standard Tinto, 409.2 Ma (Rex et al. 1986) and hornblende HB3gr (Turner et al. 1971). The Tinto standard has been cross calibrated against HB3gr, LP-6 (Engels et al. 1971), Fy12a (Roddick 1983) and MMHb-1 (Alexander et al. 1978), ages used for each of these standards as given in Roddick (1983). Flux variation over the package length was of the order of 3%.

Argon was extracted from each sample in a double-vacuum, resistance-heated furnace, developed in Leeds after the ideas of Prof. G. Turner (personal communication) and Staudacher et al. (1978). Samples were loaded into the arms of a glass storage tree above the furnace and the entire system baked overnight at 125°C under vacuum. Following further degassing of the getters and furnace (to 1350°C), a sample was dropped into the crucible and step heating commenced. The temperature of the furnace was monitored with a Minolta/Land™ Cyclops 52 infra-red optical pyrometer and is estimated to be accurate to ± 25°C with reproducibility of ± 5°C. The furnace was allowed to cool for 10 minutes after each 30-minute heating step and the evolved gas purified over two successive getters (mixtures of Ti-Zr metal shavings and Ti sponge), heated to 800°C and then allowed to cool. The gas was then transferred to a small volume inlet section by absorption on charcoal at liquid nitrogen temperature prior to admission to the mass spectrometer. Argon isotope analyses were performed using a modified MS 10 mass spectrometer with 4.2kGauss magnet and voltage peak jumping under computer control. Ion beams were detected by a VG pre-amplifier with 4 x 10^{10} ohm resistor, digitised with a KeithleyTM 2000 voltmeter and stored on computer disc for subsequent processing.

Measured mass spectrometer peak intensities were corrected for the following: Amplifier response and non-linearity; linear extrapolation to gas inlet time; spectrometer mass discrimination; and radioactive decay of ^{39}Ar and ^{37}Ar. Interfering isotopes from neutron reactions on K and Ca corrections used were: (36/39)Ca 0.38, (37/39)Ca 1492 and (40/39)K 0.048. Atmospheric argon extraction blanks dominated mainly by the contribution from the Al sample packet ranged from 5 x 10^{-9} cm^3, ^{40}Ar STP up to 660°C when the Al melts through 3 x 10^{-10} at 900°C and 2 x 10^{-9} at 1350°C.

The mass spectrometer discrimination and sensitivity were monitored by analysing atmospheric argon from a pipette system. The measured atmospheric $^{40}Ar/^{36}Ar$ (typically 287.8 ± 0.2 for these analyses) and the sensitivity (typically 1.1 x 10^{-7} Vcm^{-3} STP) change with filament life. The $^{40}Ar/^{39}Ar$ ratio, age, and errors for each gas fraction were calculated using formulae similar to those given by Dalrymple & Lanphere (1971). Errors in these ratios were evaluated by numerical differentiation of the equation used to determine the isotope ratios and quadratically propagating the errors in the measured ratios. J-value uncertainty is included in the errors quoted on the total gas ages but the individual step ages have analytical errors only. All errors are quoted at the 1-sigma level unless otherwise stated; and ages calculated using the constants recommended by Steiger & Jäger (1977). Age spectra and isotope correlation plots were produced using 'Isoplot/Ex' (Ludwig 2000).

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Mid and Late Weichselian, ice-sheet fluctuations northwest of the Svartisen glacier, Nordland, northern Norway

LARS OLSEN

Olsen, L. 2002: Mid and Late Weichselian, ice-sheet fluctuations northwest of the Svartisen glacier, Nordland, northern Norway. *Norges geologiske undersøkelse Bulletin 440,* 39-52.

During the last decade, general Quaternary geological mapping in Nordland county, northern Norway, has provided field-data, supported by radiocarbon dating, which have been used as a basis for reconstructing the palaeoclimatic and glacial history for the interval 40-10 ka BP. Much of the lateglacial (13-10 ka BP) history was known from previous investigations, but the new results necessitate significant modifications in some areas. These include both ice extension, mainly during the Younger Dryas Chron, and glacial conditions, as well as timing of events. Radiocarbon dating of shells from sediments underlying and overlying till from the previously described ice phase, the Vassdal event, also referred to as Substage A which is represented by the Vega Moraine farther south, indicates that this ice advance represents the Older Dryas ice advance northwest of Svartisen, central Nordland at c. 12.2 ka BP. Earlier knowledge of the glacial variations and palaeoclimate during the late glacial maximum (30-13 ka BP), and the Middle Weichselian intervals before that, was very modest for the central and southern parts of Nordland county. The new data suggest that there were significant glacial and palaeoclimatic variations during most of the interval 40-13 ka BP. Occurrences of shells of the marine mollusc Arctica islandica, e.g., on the island of Åmøya, indicate that temperate Atlantic water reached the coast of Norway at least as far north as 66°47' N latitude at c. 32-38 ka BP. Similar occurrences of shells with an age of c. 41-42 ka BP are reported from sites farther north, e.g. at 67°20' N, 68°48' N and 69°42' N. The associated temperate sea-surface conditions occurred contemporaneously with fern growing in the inland area at Hattfjelldal, southern Nordland. Major ice advances occurred three times separated by significant ice-retreat phases in the area northwest of the Svartisen glacier between 30 and 13 ka BP.

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Introduction

Studies of the Quaternary geology along the coast of central Nordland, northern Norway, some 20–30 years ago revealed a detailed deglaciation history supported by radiocarbon dating of numerous shell samples from this area (Andersen 1975, Andersen et al. 1981, 1982, Rasmussen 1981, 1984). The Geological Survey of Norway carried out Quaternary geological mapping in the fjord and inland areas of Nordland in the same period (e.g., Alstadsæter 1981, Alstadsæter & Hollund 1981a, b, Follestad 1981, 1989, 1990, Sveian & Vallevik 1983, Sveian 1984a, b), and has continued working in these areas also during the last decade (e.g., Follestad 1992, 1993, Bergstrøm 1995, Bargel & Olsen 1996, Olsen et al. 1996a, b, 1997, 2000a, b, c, Olsen 2000, Olsen & Bergstrøm 2000a, b). Some geological reconnaissance studies have also been performed in the Saltfjellet – Svartisen region in recent years (Gjelle et al. 1995). All this effort has provided new data that require some extensions and modifications of the last deglaciation history described by Andersen and co-workers during the period 1975–1984, and reviewed by Andersen et

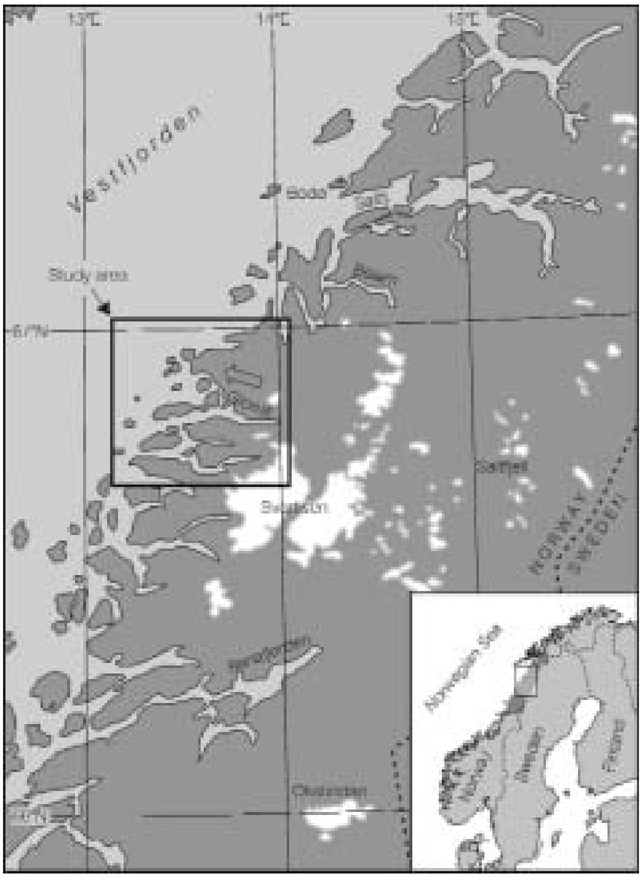

Fig. 1: Location map of the study area. The main ice-flow direction is indicated (broad arrow). Nordland county is indicated by shading (dark grey) on the inset map.

Fig. 2: Location of stratigraphic sites and dates. Inferred position of ice margin at c. 12,200 ^{14}C-yr BP, as well as during the main Younger Dryas (YD) ice advance are indicated. Cirque glaciers during the YD interval, with location both distally and proximally to the main YD ice margin position are also included. The precise age of the early lateglacial moraine (thick line) at Kunna is not known. For names of sites and comments, see Table 1.

al. (1995), but more important, the field data and radiocarbon dating during the last decade have provided a considerably improved understanding of the last glacial intervals prior to the last deglaciation in this area.

The aim of this article is to present some of the Quaternary stratigraphic information and radiocarbon dates acquired during the last decade, mainly from the area northwest of the Svartisen glacier (Fig.1). In addition, palaeoclimatic interpretations and a model of the ice-sheet fluctuations based on the new data are presented, including some comparisons with the newly published model of rapid shifts in glacial extensions during the last glaciation in Norway (Olsen 1997, Olsen et al. 2001a, b, c).

Geology and climate

The bedrock in the coastal area west and northwest of the Svartisen glacier consists of two almost equally extensive (surficially outcropping) groups of rocks, both of which form part of the Caledonian mountain chain. These are the mica schists, mica gneisses and marbles of the Rødingsfjellet Nappe Complex; and subjacent, parautochthonous to allochthonous granites and granodiorites of Precambrian (Mesoproterozoic) age.

The landscape of this region is a typical fjord landscape with fjords of variable dimensions and numerous islands and islets immediately offshore. The fjords are oriented mainly east-west, which made them the preferred routes for rapid ice flow from the east to the shelf during the numerous Pleistocene ice-growth periods. The present climate in this area is typical for the temperate North Atlantic zone, with humid and fairly warm summers, relatively wet winters, and winter temperatures often at or just below 0ºC. Annual mean precipitation ranges from c. 1500 mm in the west to 2500 mm in the inner fjord areas and 3000 – 4000 mm in the mountain areas in the east. Mean temperature for the coldest month (January) is +4 to 0°C at the outer coast, 0 to -4°C in the fjord areas, and -4 to -8°C in the east where the Svartisen glacier is located. Mean temperature for the warmest month (July) is 12–16°C in the fjord areas and 8–12°C in the mountain areas. These data represent the 1931–1960 normal period (e.g., Dannevig 1985), but is not much different today.

Methods

Standard stratigraphical methods employed at the Geological Survey of Norway were used at some few localities during the course of this work (Figs. 2–3; Table 1), but in most cases only reconnaissance studies were carried out. More work is therefore required before attempts are made

Table 1: Stratigraphical sites (1-21) referring to Fig. 2.

No.	Site name	Comments
1	Storvika**	Diamicton (till); late Younger Dryas (YD)
2	Skogreina* (Bolden)	Older Dryas (OD) moraine + MW deposits
3	Stigen*	OD till
4	Åsmoen* (4A, 4B)	OD till + LW/MW deposits
5	Djupvika**	YD till; late YD
6	Neverdalen*	Diamicton/glaciomarine sediment; OD
7	Meløya	LGM 2 till
8	Furumo**, Åmøya	Diamicton; late YD (after Rasmussen 1981)
9	Bogneset** (I, II)	Diamicton (YD) + LW/MW deposits
10	Stamnes*	Diamicton (OD) + Bølling glaciomarine sed.
11	Vargvika*	Diamicton (OD) + Bølling glaciomarine sed.
12	Gammalmunnåga	LW/MW till + older sediments
13	Ytresjøen	LW till + MW till and sediments
14	Vassdal ferry quay	LW/MW till + MW sediments
15	Vassdal	LW/MW till + older sediments
16	Sandvika*	OD till + older till and sediments
17	Aspåsen	LW/MW till
18	Oldra	LW/MW tills and glaciomarine sediments
19	Kjelddal (I-II)	LW/MW tills and glaciomarine sediments
20	Nattmålsåga*	OD/Allerød deglaciation sediments
21	Fonndalen*	YD moraine + Allerød deglaciation sediments

* Stratigraphy associated with the OD glacial oscillation.
** Stratigraphy associated with the ice shelf extension during the YD interval.

to reconstruct maps of detailed ice-flow patterns for each ice-advance phase. However, the preliminary studies involved both ^{14}C and AMS-^{14}C dating of marine mollusc shells, amino acid measurements of a few shell samples, and, with variable success, attempts to find foraminifera and dinoflagellates in some sediment samples. The results of these analyses have been used for preliminary interpretations of palaeoclimatic and glacial variations.

Conventional radiocarbon dating of shell samples was carried out at the Radiological Dating Laboratory in Trondheim, Norway (T numbers; Table 2), and AMS dating was performed at the T. Svedborg Laboratory, Uppsala University, Sweden (TUa numbers) and at the R.J. Van de Graaff Laboratory, Utrecht University, The Netherlands (UtC numbers). Some of the AMS dates were obtained from the dating of bulk organic sediment fractions with > 90–95 % plant remains. The samples for AMS-dating in Uppsala were prepared, and graphite targets produced, at the Radiological Dating Laboratory in Trondheim. All ages cited in this article, if not otherwise indicated, are in radiocarbon years before the present (BP).

Table 2: Radiocarbon dates (conventional and AMS) of shells and sediments (organic fraction with > 90-95 % plant remains).

Locality	Field no.	Lab. no.	Weight of C	Mollusc shell	d^{13}C	^{14}C-yrs.	+/- 1sd
Meløya	13.-98	UtC-8310	1.98 mg	One shell fragment	2.7	38 200	700
Skavika, Ågskaret	M6-92	T-10798		Div. species	*	11 865	60
Stamnes, Åmøya	M5-92	T-10541		Div. species	*	12 420	105
Bogneset I, Åmøya	M4-92	T-10540		Div. species	*	32 100	2600
Bogneset I, Åmøya	2.-6/6-93	TUa-947		Div.	*	40 025	965
Bogneset I, Åmøya	II-6/7-94	TUa-1239		Arctica islandica, a.o.	*	35 940	1455
Bogneset I, Åmøya	III-6/7-94	TUa-1240		Div.	*	28 355	430
Bogneset I, Åmøya	IV-6/7-94	TUa-1241		Arctica islandica, a.o.	*	38 090	1675
Bogneset II, Åmøya	1.-17/6-94	T-11784		Mya	*	11 165	105
Storvika, Gildeskål	2.-14/9-95	UtC-4727		One shell fragment	-6.1	11 110	80
Skogreina	1.-5/6-93	TUa-743		Div. species	*	38 545	835
Skogreina	3.-5/6-93	TUa-946		Div. species	*	37 730	735
Skogreina	2.-5/6-93	TUa-1092		Div. species	*	38 060	710
Stigen	19.-98	UtC-8314	1.98 mg	One shell fragment	1.1	12 200	60
Åsmoen, Ørnes	M7-92	TUa-567		Hiatella arctica	*	28 355	235
Åsmoen, Ørnes	05.06.93	TUa-744		Macoma, a.o.	*	12 520	85
Mosvollelva, Ørnes	08.07.94	TUa-1094		Fragm. of one species	*	29 075	370
Djupvika, Ørnes	M8-92	T-10543		Macoma calcarea	*	10 430	185
Vargvika	M3-92	T-10797		Div. species	*	12 450	195
Gammalmunnåga	M2-92	T-10539		Hiatella, Mya, Macoma	*	> 44.800	
Ytresjøen	22.-98	UtC-8315	2.18 mg	Fragm. of one species	0.4	28 720	240
Ytresjøen	23.-98	UtC-8316	2.19 mg	One shell fragment	1.9	35 500	600
Vassdal ferry quay	1.-6/6-93	TUa-944		Div. species	*	35 280	575
Vassdal	M1-92	T-10796		Div. species	*	30 610	3950
Holmåga	11.-98	UtC-8308	2.16 mg	One shell fragment	1.8	9 059	39
Sandvika	12.-98	UtC-8309	1.95 mg	One shell fragment	0.3	12 600	60
Neverdalsvatnet	1.-7/7-94	T-11785		Chlamys islandica	*	12 520	205
Nattmålsåga	1.-14/9-95	T-12567		Div. species	*	11 975	155
Fonndalen	21.-96	UtC-5465		Fragm. of one species	0.66	11 990	60
Aspåsen	1.-27/7-95	TUa-1386		Div. species	*	36 455	530
Oldra	2.-30/9-93	TUa-745		Div. species	*	32 510	395
Oldra	1.-20/7-95	TUa-1385		Mya truncata	*	33 040	315
Oldra II	3.-20/7-95	TUa-1387		One shell fragment	*	33 975	515
Kjelddal I	15.-98	UtC-8311	2.36 mg	One shell fragment	1.4	35 800	600
Kjelddal II	17.-98	UtC-8312	2.06 mg	One shell fragment	-1	33 700	400

Locality	Field no.	Lab.no.	Fraction	Weight of C	Material	d^{13}C	^{14}C-yrs.	+/- 1sd
Meløya	14.-98	UtC-8456	INS	0.53 mg	>90-95% plant remains	-25.4	17 700	80
Kjelddal I	16.-98	UtC-8457	INS	0.33 mg	>90-95% plant remains	-24.8	18 880	100
Kjelddal II	18.-98	UtC-8313	INS	1.57 mg	>90-95% plant remains	-24.5	24 858	161

d^{13}C: ratio $^{13}C/^{12}C$ in per mil with respect to PDB-reference.
* : Estimated d^{13}C-value.

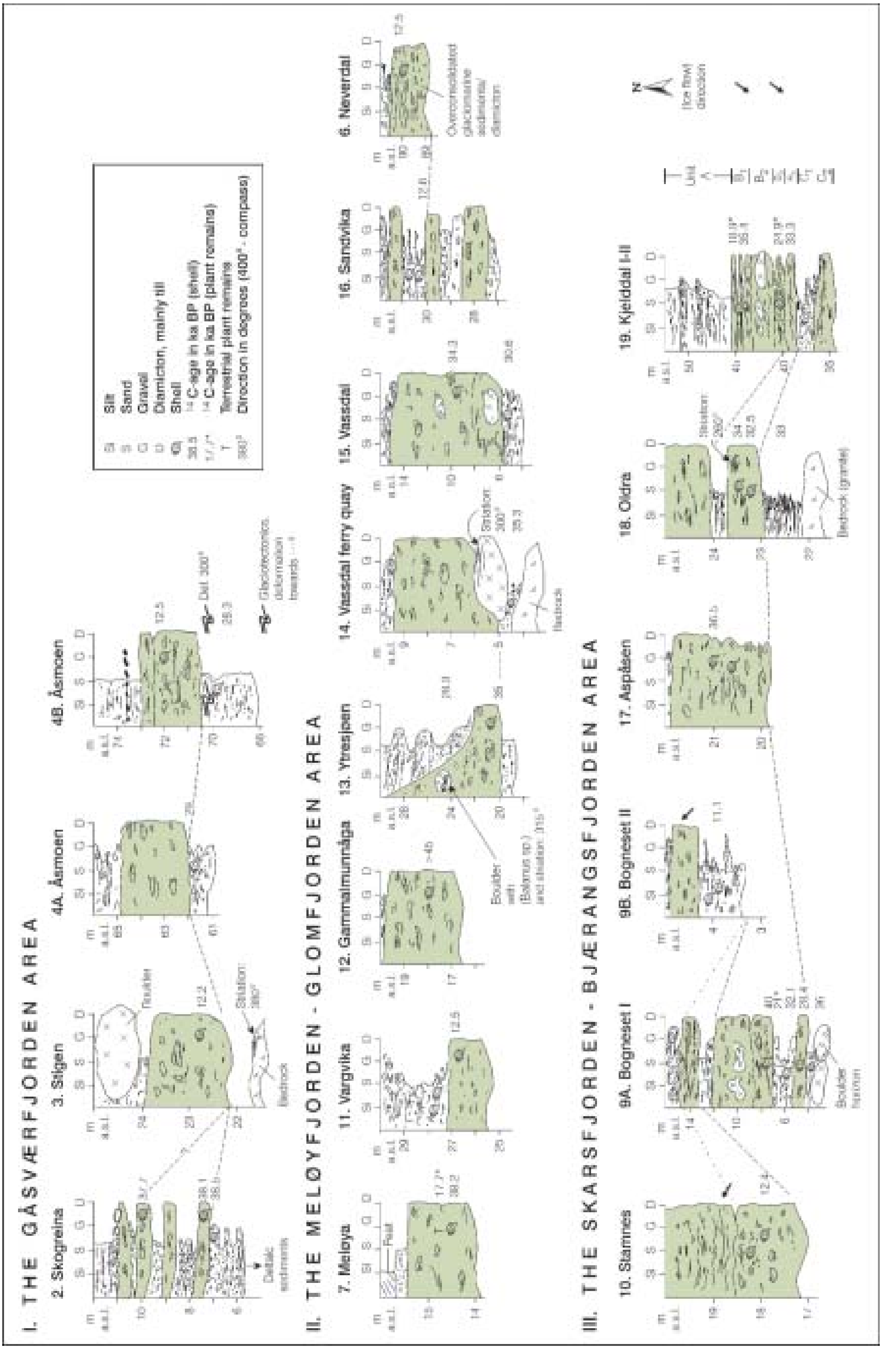

Fig. 3: Generalised stratigraphical logs from various sites in the described areas.

Amino acid racemization (AAR) measurements of shell samples were carried out at the Department of Geology, University of Bergen. Eleven samples were measured, the objective being to distinguish shells dating to between 35,000 and 45,000 yr old and those of Early Weichselian or older age, which is beyond the range of the radiocarbon method. This led to the identification and exclusion of 4 pre-Middle Weichselian samples (Table 3). For details of the analytical and preparation procedures see, e.g., Miller & Brigham-Grette (1989).

Foraminifera were separated following standard methods described by, e.g., Steinsund & Hald (1994). Heavy liquid with specific weight 1.8 g/cm^3 was used during separation. Results of the analyses are briefly mentioned in the text and, for one site (Kjelddal), included in Table 5.

Standard palynological procedure for the preparation and analysis of dinoflagellates (Barss & Williams 1973) was followed. Hydrofluoric acid (50% concentration) and hydrochloric acid (10% conc. HCl) were used to dissolve the minerals. The material was sieved through a 0.01 mm mesh. To calculate the productivity of the dinoflagellate cysts, spores of *Lycopodium clavatum* were added. One tablet contains 13,911 spores, and one tablet per sample was added. Ultrasonic treatment for maximum 60 seconds was used to remove the organic particles which stick to the palynomorphs, and to concentrate the palynomorphs. The organic material was finally cast in Elvacite to produce the preparates.

At least 300 dinocysts are generally counted per sample in order to obtain proper statistics with regard to environmental interpretations. This is possible where the productivity of dinoflagellate cysts is high, but in this work the sediments used have only low and very low productivity. Therefore, during these preliminary studies the counting was stopped at lower levels, but this was high enough to obtain the main trends recorded.

Results and interpretations

At the Arctic Circle in the coastal part of Nordland, some radiocarbon dates of bulk-organic sediment samples and marine mollusc shells in tills and sub-till sediments suggest that the last major ice advances beyond the coastline occurred after the period 18–28 ka BP (Table 2, Fig. 2). Stratigraphic successions and various data from some key areas which are presented briefly below, constitute the main basis for the interpretations given in this article. The areas are described from north to south and the localities, in most cases, from west to east (Fig. 1).

The Gåsværfjorden area

Radiocarbon dating of shells and the stratigraphy from four sites in this area (sites 2–5, Fig. 2), including amino acid measurements of a shell fragment from site 2 (Tables 2–3), reveal a complex palaeoclimatic and glacial history with redeposition of marine sediments from the last interglacial (the Eemian), the Middle Weichselian interstadial-complex (30–39 ka BP), the lateglacial Bølling Chron and the Younger Dryas (YD) Chron.

At *Skogreina* (site 2, also named Bolden, Rasmussen 1981; Fig. 2), an ice-marginal glacial-glaciofluvial complex terrace (the Skogreina Moraine) may encompass most of the above-mentioned history. A glaciofluvial deltaic terrace containing resedimented Eemian – Middle Weichselian marine shells is overlain by 1–4 metres of till with intercalated sand beds (Fig. 3), and overlain by a wave-washed ice-marginal moraine on top in the proximal parts. The surface of this site is located 12–15 m a.s.l., and is covered by shore deposits (mainly boulders, cobbles and pebbles) produced from heavily wave-washed and redeposited glacial-glaciofluvial material. AMS-^{14}C dating and amino acid measurements of fragments of marine shells (Table 2, TUa-743, 946 & 1092, Table 3) found in gravelly sand in the till-covered distal part of the complex indicate Eemian – Mid Weichselian ages of the redeposited sediments. It is supposed that the lower part of the complex represents an ice-marginal formation from a Mid Weichselian glaciation, possibly from the ice-advance that occurred at c. 40 ka BP (Olsen 1997, Olsen et al. 2001a, b), or a younger ice-margin oscillation c. 28–29 ka BP (see below). The moraine-covered proximal part is thought to represent the ice-marginal formation of the lateglacial Substage A (the Older Dryas (OD) cold phase) some 12–12.5 ka BP (Rasmussen 1981, Andersen et al. 1981, Gjelle et al. 1995).

At *Stigen* (site 3, Figs. 2–3), in a proximal position to the

Table 3: Amino acid ratios – site name and inferred age of shells are included.

No.	Lab. refr.	Location	Field no.	m a.s.l.	Species	HYD	Average/ std. dev.	FREE	Average/ std. dev.	Probable age based on the AAR ratios
1	BAL 3392	Skogreina	1-5/6-93	10	*Mya?*	0.158	0.156	0.36	0.361	Early Weichselian -
						0.154	0.003	0.361	0.001	Eemian
2	BAL 3393A	Vassdal	1-1/10-92	5	*Cardium*	0.022	0.022	ND		Postglacial age
					edule	0.021	0.001	ND		
	BAL 3393B	Vassdal	1-1/10-92	5	*Cardium*	0.021	0.018	ND		Postglacial age
					edule	0.015	0.004	ND		
3	BAL 3394	Gml.m.åga	2-1/10-92	20	*Mya*	0.113	0.114	0.382	0.395	Early Weichselian -
						0.115	0.001	0.407	0.018	Eemian
4	BAL 3395	Bogneset	4-25/8-96	8	*Arc. isl.*	0.105	0.1	0.276	0.273	Mid Weichselian
						0.095	0.007	0.27	0.004	

Skogreina ice-marginal deposit and distally to the YD – moraines, there is a coarse-grained till containing shell fragments of lateglacial age. An AMS-^{14}C date of one of these shell fragments yielded an age of 12,200 +/- 60 ^{14}C-yr BP (Table 2, UtC-8314), which indicates that a lateglacial ice advance of pre-YD age (the OD – ice advance) reached beyond this site after 12.2 ka BP.

At *Åsmoen* (sites 4A–4B, Fig. 2), c. 1 km from Ørnes and some few kilometres distally to the Younger Dryas ice marginal zone, till-covered ice-marginal, outwash-sediments indicate, in part, a similar history to that recorded at Skogreina. AMS-^{14}C dating of shell fragments from the subtill sediments indicates that these ice-marginal sediments have a maximum age of c. 28–29 ka BP (Fig. 3, Table 2, TUa-567 & 1094), indicating an ice-marginal position in the vicinity of Åsmoen approximately at this time or later. A radiocarbon dating of a shell fragment from the overlying till gave an age of c. 12,520 +/- 85 ^{14}C-years BP (Table 2, TUa-744), which indicates that the Older Dryas ice advance reached beyond the Åsmoen site after c. 12.5 ka BP.

Radiocarbon-dated shell in till at Djupvika (site 5, Fig. 2, and Table 2, T-10540) indicates that an ice advance in the last part of the Younger Dryas Chron reached at least slightly beyond this site, which is located c. 2–3 km distally to the main ice-marginal zone during the YD interval.

The Meløyfjorden – Glomfjorden area

Radiocarbon dating and amino acid measurements of shells and stratigraphy from eight sites in this area (sites 6, 7 & 11–16, Figs. 2–3) reveal a similar palaeoclimatic and glacial history as in the Gåsværfjorden area (see above).

At *Meløya* (site 7, Figs. 2–3), distally to the Substage A (Older Dryas) moraines, a fine-grained, compact, bluish-grey lodgement till contains redeposited shell fragments and other marine as well as terrestrial organic material (Olsen et al. 2001c). The redeposited material represents a mixed assemblage of sediments of supposed Mid to Late Weichelian age, as indicated by dates giving different ages for different materials (shells vs. terrestrial material; Fig. 3). The organic fraction (> 90–95 % plant remains) of a bulk sediment sample yielded an age of 17,700 +/- 80 ^{14}C-yr BP (Table 2, UtC-8456), which indicates that the last major ice advance flowed across this island (Meløya) after c. 18 ka BP.

Conventional ^{14}C and AMS-^{14}C dating of shells (e.g., *Arctica islandica*) in till and sub-till sediments at *Ytresjøen, Vassdal ferry quay and Vassdal* (sites 13–15, Figs. 2–3) indicate that temperate Atlantic water reached the coast at least as far north as these latitudes c. 28–36 ka ago. This age interval is also a maximum age for the ice advance which is represented by the overlying till at these sites, and the ice advance in question may be the same as that recorded at Åsmoen, and possibly also at Skogreina, at c. 28–29 ka BP.

At *Gammalmunnåga* (site 12, Figs. 2–3), ^{14}C-dating and amino acid measurements of shells in till indicate redeposition of Eemian marine sediments (Table 2, T-10539, and Table 3). ^{14}C-dating of shells from marine sediments overlain by till at *Vargvika* and *Sandvika*, and also overrun by the ice at *Neverdalen* (sites 6, 11 & 16, Figs. 2–3), indicate a last glacial readvance after 12,600 +/- 60, as well as after 12,520 +/- 205 and 12,450 +/- 195 (Table 2, UtC-8309, T-11785 & T-10797) in the outer Glomfjorden area. The ice margin during this readvance reached to Risneset (the Risneset Moraine in Åmnessund) on the northern coast of Åmøya in the west.

The ice-marginal zone during the main (early) Younger Dryas phase crossed the innermost part of Glomfjorden, implying that most of the Glomfjorden area has been ice-free since the Allerød Chron (Fig. 2; Rasmussen 1981, Andersen et al. 1981).

The Skarsfjorden – Bjærangsfjorden area

Radiocarbon dating, amino acid measurements and stratigraphy from this area (sites 9A, 9B, 10 & 17–19, Figs. 2–3) indicate a similar palaeoclimatic and glacial development as in the other two fjord areas described above. However, no *in situ* or redeposited Eemian sediments have so far been recorded here. Both amino acid measurements (Table 3) and radiocarbon dating indicate Mid Weichselian ages for the oldest sub-till sediments found in this area.

Stratigraphical logs from *Stamnes* and *Bogneset* I and II on Åmøya island (sites 9A, 9B & 10, Figs. 2–3; and Fig. 4), indicate a complex stratigraphy with alternating marine or glaciofluvial – glaciomarine –

Chronozone	Lithostratigraphy (locality)	Marine shells	^{14}C-dates, ( )= other sites	Palaeomilieu, - climate and time (BP= before present)
	boulders & stones			Shore deposit, postglacial
Younger Dryas (YD)	diamicton, glaciomarine sediments (Bogneset II)			Ice advance during YD, or c. 11,100 ^{14}C-yr BP, did not reach as far as Åmøya; possible ice shelf development, glaciomarine sedimentation and glacial reworking.
Allerød interstadial	silt-sand (Bogneset II)	*Macoma calcarea, Mya truncata, Hiatella arctica*	11,165 +/- 105 (11,990 +/- 60)	Ice margin retreated inland; glaciomarine - marine, temperate
Older Dryas (OD)	till (Stamnes)			Ice cover; ice advance between 12,200 and 12,000 ^{14}C-yr BP, the ice margin crossed over Åmøya island.
interstadial	silt-sand (Stamnes)	*Macoma calcarea, Mya truncata, Arctica islandica, Hiatella arctica*	12,420 +/- 105 (12,200 +/- 60) (12,600 +/- 60)	Ice margin retreated in the fjord areas; glaciomarine - marine environment, temperate conditions.

Fig. 4: Stratigraphic data from Stamnes and Bogneset II, Åmøya, Nordland, c. 66°47' N. Comments on depositional environment and palaeoclimate are based on lithology and marine shells.

m a.s.l.	Lithostratigraphy		Marine shells (only fragments)	Aminoacid analysis HYD	FREE	^{14}C-dates, conv. and AMS*	Palaeomilieu, - climate and time (BP= before present)
16	1	stones					Shore deposit, postglacial
15 14	2	till					Ice cover; ice advance c. 12,200 ^{14}C-yr BP (between 12,200 and 12,000 ^{14}C-yr BP), 'Older Dryas'.
13	3	S G-S	Shells found in correlative unit at adjacent sites: Shells as in unit 8				Bølling interstadial; 12,200-13,000 ^{14}C-yr BP.
12 11 10 9	4	till					Ice cover; ice advance after 17,700 - 21,000 ^{14}C-yr BP. LGM 2
8 7	5	till				40,025 +/- 965	Ice cover; ice advance after 21,000 ^{14}C-yr BP. LGM 1 or LGM 2
	6	G-S S G-S	*Nuculana pernula, Macoma calcarea, Mya truncata, Arctica islandica**, Hiatella arctica*	0.100 +/- 0.007	0.273 +/- 0.004	*20,880 +/- 130 T 32,100 + 2600 /-2000	Glacio-/marine; resedimented, temperate**, interstadial
5	7	till				*28,355 +/- 430	Ice cover; ice advance after 28,400 ^{14}C-yr BP; LGM 1?
4	8	G-S	As in unit 6, except Nuculana pernula			*35,940 + 1455	As for unit 6
	9	boulders				/-1230	Shore deposit?

* AMS - ^{14}C dating; ** Temperate sea-water indicator; T = Terrestrial material (bulk sample).

Fig. 5: Bogneset I stratigraphy – sedimentary units, shells, AAR ratios and dates, including environmental interpretations. Location: c. 66°47' N. Lithology: S - sand; G - gravel.

m a.s.l.	Lithostratigraphy		Dinocysts/ markers	Foraminifera	Pollen	Palaeomilieu, - climate and time
16	1	stones				For units 1 - 5, 7 and 9, see Fig. 5
15 14	2	till				Ice covered; Older Dryas ice advance
13	3	S G-S				Bølling interstadial
12 11 10 9	4	till				Ice covered; LGM 2
8 7	5	till				Ice covered; LGM 1 or LGM 2
6	6	G-S S G-S	*4/385; Protoperidinium spp.: 4* *49/410; ----- " ------ : 38* *94/915; Operculodinium centrocarpum: 94*	7 observed 53 benthic obs. > 10 observed	only spores observed resedimented pollen very few pollen obs.	Interstadial, marine, cold (?) Marine, resedimented, temperate Marine, resedimented, temperate
5	7	till				Ice covered; LGM 1 (?)
4	8	G-S				Marine, resedimented, temperate
	9	boulders				

Fig. 6: Bogneset I stratigraphy – sedimentary units and microfossils, including environmental interpretations. Lithology: S - sand; G - gravel.

marine units and till units spanning an age interval from at least 40 ka BP to the Younger Dryas Chron. Shells of the marine mollusc *Arctica islandica*, which is an indicator of temperate Atlantic water (Peacock 1989), are found in the lower parts of the Bogneset I stratigraphy. AMS-^{14}C dating supported by amino acid measurements indicate that the northerly cool – temperate Atlantic water reached this area at c. 32–38 ka BP (Fig. 5). The preliminary analyses of dinoflagellates, which indicate a strong dominance of the species *Operculodinium centrocarpum* and *Protoperidinium spp. (cf. P. conicoides)*, also support this interpretation (Fig. 6).

Four ice advances between c. 28.4 ka BP and the Allerød Chron are represented by till units on Åmøya (Figs. 3 & 7). Correlation between the units and sites indicates that the youngest of these advances occurred during the Older Dryas phase after 12,420 +/- 105 ^{14}C-years BP (Table 2, T-10541), and this ice advance reached only 2–3 km farther west on the island (Fig. 2; Rasmussen 1981, Gjelle et al. 1995). The subsequent Younger Dryas Scandinavian ice-sheet readvance did not reach farther west than to the innermost parts of the fjord areas dealt with here, i.e. some 10–15 km east of Åmøya. The upper diamict unit at Bogneset II (Fig. 6) is inferred to represent glaciomarine/-iceberg-moulded sediments, or sediments influenced by a partly grounded shelf ice from the Younger Dryas interval. This diamict unit is less compacted than, for example, the upper diamict unit thought to be a till at the Stamnes site. Another difference between these diamict units is that the diamicton at Stamnes contains fragments of broken marine shells, whereas no shells are found in the diamicton at Bogneset II. The glaciomarine unit underlying the diamicton at Bogneset II contains, however, an abundance of unbroken shells, one of which is dated to 11,165 +/- 105 (Table 2, T-11784).

Glaciofluvial and glaciomarine sediments with redeposited marine sediments are interbedded with the tills, and also underlie the tills at Bogneset I (Fig. 7), indicating ice-retreat phases between the ice advances. It is thought that the tills and glacial diamictons in the Bogneset I stratigraphy represent several ice advances during the last glacial maximum (LGM), as well as one during the lateglacial period (the Older Dryas ice advance). The preliminary investigation provides a moderate precision for the age-range of these ice advances, which occurred between ice-retreat phases at c.

Table 4: ^{14}C dates associated with the Older Dryas ice advance in the area NW of Svartisen. Maximum (1-6) and minimum (7-12) ages. Ice advance between 12.0 and 12.2 ka BP.

No.	Lab. no. / reference	Ice advance vs. shelldates ^{14}C-yr BP; +/- 1 sd	Location (site)	Host material
1	UtC - 8309	< 12,600 +/- 60	Sandvika	Subtill glaciomarine sed.
2	TUa - 744	< 12,520 +/- 85	Åsmoen	Till
3	T - 11785	< 12,520 +/- 205	Neverdalen	Tillised glaciomarine sed.
4	T - 10797	< 12,450 +/- 195	Vargvika	Till
5	T - 10541	< 12,420 +/- 105	Stamnes	Till
6	UtC - 8314	< 12,200 +/- 60	Stigen	Till
7	UtC - 5465	> 11,990 +/- 60	Fonndalen	Endmoraine (YD)
8	T - 12567	> 11,975 +/- 155	Nattmålsåga	Glaciofluvial fan
9	T - 10798	> 11,865 +/- 60	Skavik	Glaciomarine sediment
10	Rasmussen (1981)	> 11,740 +/- 100	Bratsberg	--------- " ---------
11	------- " -------	> 11,720 +/- 200	Engavågen	--------- " ---------
12	------- " -------	> 11,700 +/- 150	Ågskaret	--------- " ---------

24.9*- 28.4, 17.7*- 20.9, 12.2*- 12.6* and 11.1- 12.0* ka BP (* dates from other sites in the study area). These results improve the precision of the age-range estimation of the last glacial maximum in this area by making the interval at least 7,000 years shorter compared to previous estimations (Rasmussen 1981, Andersen et al. 1981). In addition, the LGM interval is no longer regarded as a long interval of continuos ice cover, but an interval of alternating phases of ice advances and retreats, occasionally with considerable ice-free areas included. The extent of the ice advances, however, can only be found through correlations with formations closer to the ice margin on the continental shelf, but that topic is beyond the scope of this article.

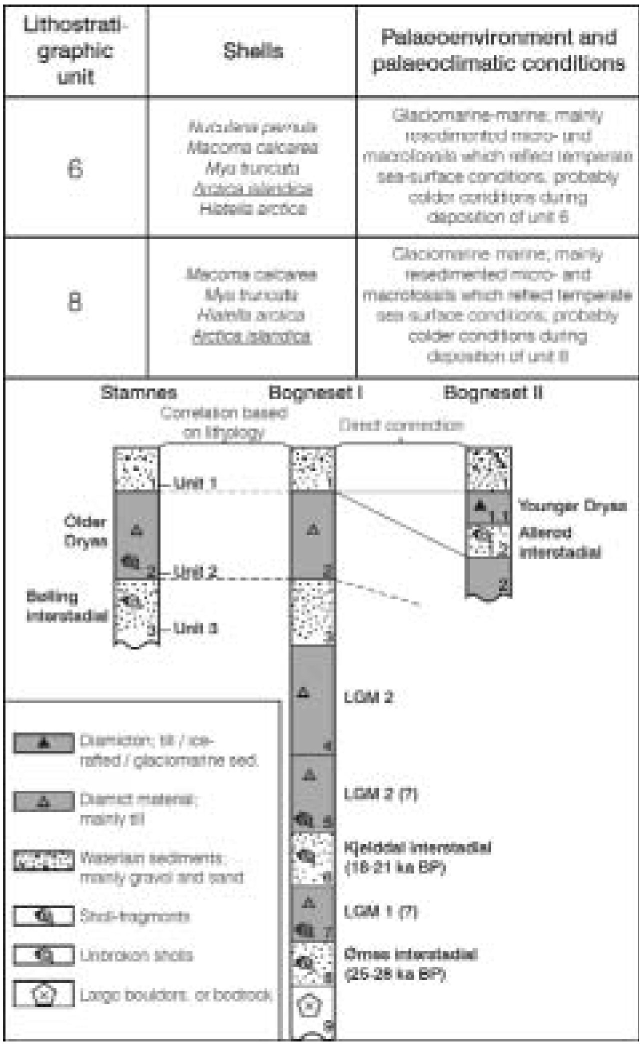

Fig. 7: Correlation chart – Stamnes, Bogneset I & II.

Dating and stratigraphy at *Aspåsen, Oldra* and *Kjelddal* (sites 17–19, Figs. 2–3) indicate an ice-marginal retreat during the interval c. 32–37 ka BP, as well as during the younger ice-retreat phases c. 25 and 19 ka BP, of similar size as, or even larger than during the lateglacial Allerød interstadial. The intercalated waterlain sediments and the upper till units at Kjelddal include some plant fragments, redeposited shells and foraminifera which indicate a harsh climate during mainly Arctic conditions (e.g., Table 5). AMS-^{14}C dates of shells and foraminifera from the upper till yielded ages of c. 33-36 ka BP (Table 2, UtC-8311, 8312 & 10100), whereas dates of the organic material (> 90–95 % plant remains and algae) of bulk sediment samples from the intermediate and upper sub-units of the upper till yielded ages of 24,858 +/- 161 and 18,880 +/- 100 ^{14}C-yr BP, respectively (Table 2, UtC-8313 & 8457). The Kjelddal stratigraphy is inferred to represent an ice advance (lower till; subunit C2) possibly older than 33–36 ka BP (40 ka BP?), a subsequent interstadial (C1) followed by a readvance of the ice after 33 ka BP (upper till; B4), a new interstadial c. 25 ka BP (upper till; diamict material B3), another ice advance (LGM 1; B2) and then another interstadial c. 19 ka BP (intercalated waterlain sediments in subunit B1), followed by the last ice advance (LGM 2), which is represented by the diamictons in the uppermost part (B1) of the upper till (Olsen et al. 2001c). In this reconstruction the Kjelddal stratigraphy seems to comprise all major glacial fluctuations during the LGM interval 15 – 30 (40?) ka BP.

The Holandsfjorden area

Dating of shells from a glaciofluvial – marine fan deposit at *Nattmålsåga* and from the end moraine complex at the outlet of *Fonndalen* (sites 20 & 21, Fig. 2) indicate an early deglaciation of the entire Holandsfjorden area. The ice retreat after the Older Dryas ice advance seems to have reached these sites before c. 11,975 +/- 155 and 11,990 +/- 60 ^{14}C-years BP (Table 2, T-12567 and UtC-5465), which gives a minimum age for the Older Dryas phase in this area.

The age of the Vassdal event/ Substage A, northwest of Svartisen

The oldest, lateglacial, ice-margin position in Nordland, represented by the Vega moraines and named *Substage A* after

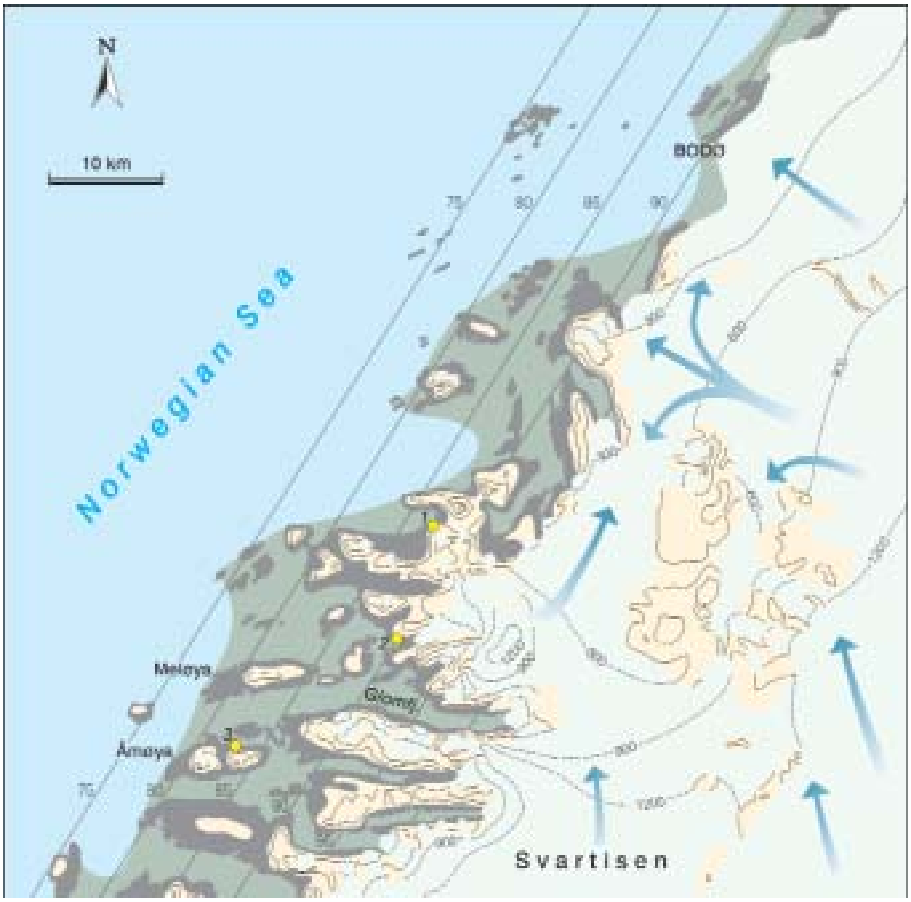

Fig. 8: Suggested extension of shelf ice (light grey) and the inland ice (white colour on land areas) during the Younger Dryas interval. The locations of sites (1 - Storvika, 2 - Djupvika, 3 - Furumo) with associated stratigraphic information are also indicated in Fig. 2, but with different loc. nos. (1, 5 & 8, respectively). Present land areas submerged by the sea during the YD interval is also indicated (black colour). Isobases for the YD sea-level are indicated as well (m a.s.l.). Arrows indicate some of the main YD ice flow directions.

Table 5: Results from foraminifera analyses of samples from the Kjelddal I-II stratigraphy. Sample no. 1A, B-7/8-98 (from unit B1) and no. 2A, B-7/8-98 (from unit B3).

Species	Sample no. 1A-7/8-98		Sample no. 1B-7/8-98		Sample no. 2A-7/8-98		Sample no. 2B-7/8-98	
	No.	%	No.	%	No.	%	No.	%
Elphidium excavatum	184	53.8	164	48.8	80	24.5	62	19.9
Cassidulina reniforme	71	20.8	94	28.0	64	19.6	67	21.5
Islandiella helenae	28	8.2	24	7.1	47	14.4	46	14.7
Protelphidium albiumbidicatum	10	2.9	4	1.2	32	9.8	41	13.1
Buccella tenerrima	6	1.8			30	9.2	25	8.0
Islandiella norcrossi	10	2.9	6	1.8	26	8.0	32	10.3
Elphidium subarcticum					14	4.3	+	
Nonionella auricula	+		1	0.3	9	2.8	6	1.9
Astrononion gallowayi	+		1	0.3	3	0.9	4	1.3
Elphidium asklundi	6	1.8	7	2.1	3	0.9	+	
Nonion labradoricum					3	0.9		
Buccella frigida	9	2.6	19	5.7	2	0.6	7	2.2
Cibicides lobatulus	+		2	0.6	2	0.6	8	2.6
Protelphidium orbiculare	14	4.1	4	1.2	2	0.6	9	2.9
Trifarina fluens					2	0.6	1	0.3
Protelphidium magellanicum			1	0.3	1	0.3	2	0.6
Stainforthia loeblichi	3	0.9	6	1.8	1	0.3	1	0.3
Stainforthia skagerakensis	1	0.3						
Fissurina laevigata	+							
Guttulina glacialis	+							
Guttulina lactea	+		1	0.3			1	0.3
Islandiella islandica	+							
Nonion barleeanum	+							
Oolina borealis	+							
Neogloboquadrina pachyderma					+(D)			
Globorotalia sp.					+			
Indetermined species			2	0.6	5	1.5		
Sum	342	c.100	336	c.100	326	c.100	312	c.100
No. of species	20		17		19		17	

Andersen et al. (1981), has traditionally been correlated with the Skarpnes Substage farther north, and would therefore be expected to have an age of c. 12.2 ka BP (Andersen et al. 1995). The Vassdal event/ Substage A is represented by the Skogreina and Risneset moraines in the area northwest of Svartisen (Rasmussen 1981). Based on new radiocarbon dates of shells from sediments underlying and overlying till which represents the ice readvance during Substage A (Figs. 2–3, and Table 4), the age of this ice readvance is c. 12.0–12.2 ka BP. This result seems to support quite strongly both an Older Dryas age and a correlation with the Skarpnes Substage of Substage A northwest of Svartisen.

Discussion and regional correlations

The glacier fluctuations as inferred from the stratigraphical successions in Fig. 3 indicate several ice advances and retreats in the interval 10–40 ka BP (Figs. 9–11). The fluctuations may have been of variable extent, but seem to correspond well in number and time with the regional ice-sheet fluctuations described for the western part of Fennoscandia (Olsen 1997, Olsen et al. 2001a, b, c). However, problems of chronological character, such as a lack of shell dates, but several dates of terrestrial material in the interval 13 to 28 ka BP and ^{14}C dating of sediments with a low organic content, which were thoroughly discussed by Olsen et al. 2001a, are still issues which must be

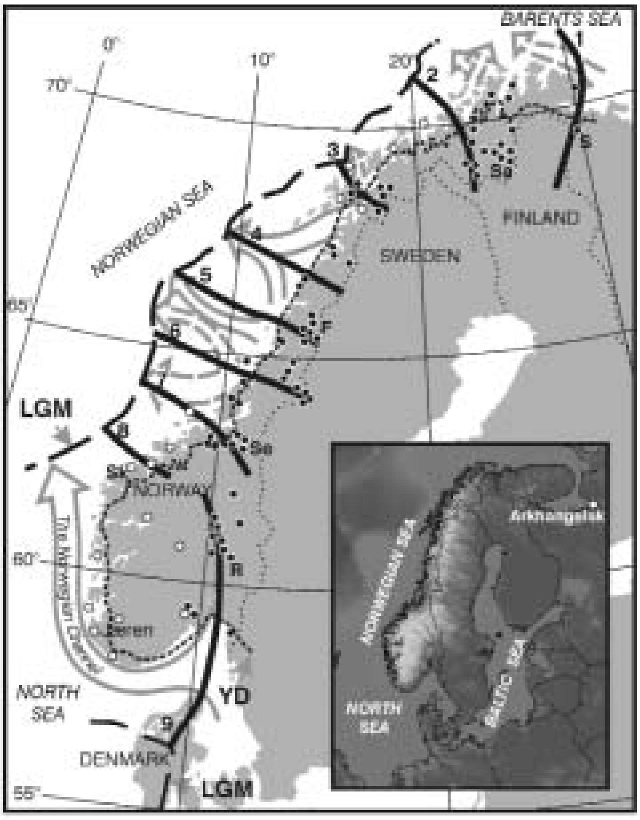

Fig. 9: Map of Fennoscandia with transects (1-9) of glaciation curves of which three are indicated in Fig. 10. Inset map: Topography of Fennoscandia and adjacent areas. Modified from Olsen 1997.

considered during future studies of ice-sheet fluctuations and their timing.

Regional ice advance c. 40 ka BP. – This ice advance, which is recorded in other parts of Norway (e.g., western Norway: Larsen et al. 1987; central Norway: Olsen et al. 2001a, b, c; southeastern Norway: Olsen 2001e, Olsen et al. 2001a), is not clearly represented in the available data from the area northwest of Svartisen. Sedimentary units (glacial diamictons) which could possibly represent candidates for this phase, as e.g. the lower strata at Skogreina and Kjelddal, may

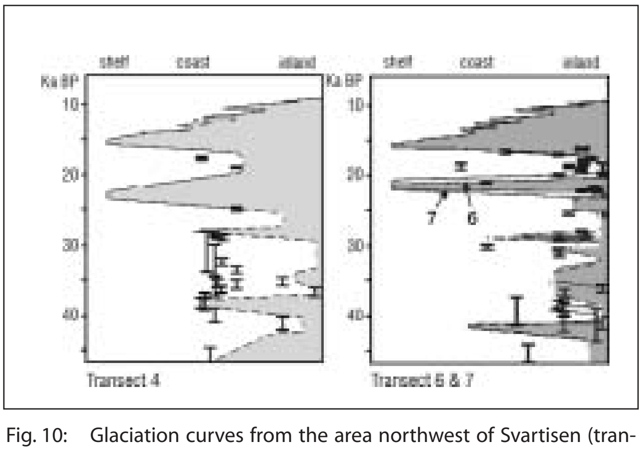

Fig. 10: Glaciation curves from the area northwest of Svartisen (transect 4) and central Norway (transects 6 & 7). Ice covered areas are indicated with grey shading. Modified from Olsen et al 2001c.

well represent other, older or even younger phases. However, based on the shell dates from these diamictons the ice advance is thought to have occurred at c. 38–40 ka BP (Fig. 10).

Regional ice advance c. 30 ka BP. – After an interval of almost totally ice-free conditions and a relatively favourable climate for growth of vegetation and other terrestrial and marine organisms (e.g., Mangerud et al. 1981, Larsen et al. 1987), a regional ice advance occurred at c. 28–30 ka BP. Ice-margin fluctuations representing this phase are dated to 28–29 ka BP at Åsmoen I–II, 28.3 ka BP at Ytresjøen and to 28.4 ka BP at Bogneset I (Fig. 3). This ice advance is represented by till and bracketed in age between 33 and 25 ka BP at Kjelddal I–II.

LGM 1, regional ice advance c. 22 ka BP. – Ice retreat in the fjord areas after the 28–30 ka ice advance was followed by the LGM 1 advance (Olsen 1997, Olsen et al. 2001a, b, c). This ice advance is represented by tills at Bogneset I and Kjelddal I–II, and bracketed in age between 25 and 19 ka BP at the latter site (Fig. 3).

LGM 2, regional ice advance c. 16 ka BP. – This ice advance followed a phase of considerable ice retreat and extensive ice-free areas, c. 17–21 ka BP (Olsen 1997, Olsen et al. 2001a, b, c). LGM 2 is represented by tills at Meløya, Bogneset and Kjelddal I–II, and probably at numerous other sites. It is bracketed in age between 17.7–21 ka BP and 12.6 ka BP at the sites illustrated with stratigraphic logs in Fig. 3.

Older Dryas (OD), regional ice advance c. 12.2 ka BP (Fig. 11). – The oldest regional ice advance during the lateglacial period is well represented and bracketed in age between c. 12.0 and 12.2 ka BP in this area (Table 4). The minimum length of this advance was at least 10 km from Åsmoen I (site 4B) to the OD ice margin at Skogreina (site 2) in the northwest and at least 15 km from Neverdal (site 6) to the ice margin at Meløya in the west (Olsen 2001d).

The OD ice advance is represented by the Repparfjord Substage in Finnmark county (Marthinussen 1960, 1961, 1962, 1974, Sollid et al. 1973), the Skarpnes Substage in Troms (Marthinussen 1962, Andersen 1968), the Vassdal event in the area northwest of Svartisen (Rasmussen 1981), the Substage A (Vega Moraines) in the coastal areas of Nordland as a whole (Andersen et al. 1981, 1982), the Uran and Osen events in Nord-Trøndelag (Olsen & Sveian 1994, Olsen & Riiber, unpublished material 1997), the 'Outer Coastal' Substage in Sør-Trøndelag (Reite 1994), the Tingvoll Substage in northern Møre & Romsdal (Follestad 1985), and so on, with additional correlatives farther south along the coast. Reite (1994) has reported marginal moraines representing this substage in the fjord areas of Sør-Trøndelag, but he could not find correlative moraines in the land areas between the fjords. Therefore, he suggested that this glacial event may have been a result of the unstable physical conditions along the glacial margins of the ice itself, during deglaciation, and not a result of a climatic variation. However, as the OD ice advance is recorded in all parts of

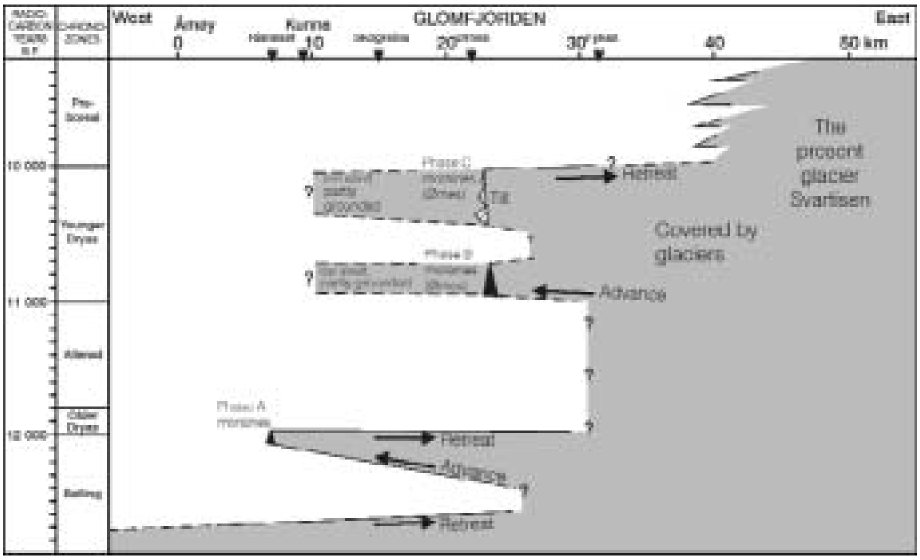

Fig. 11: Glaciation curve for the lateglacial period in the area northwest of Svartisen. Modified from Olsen et al. 2001d.

Norway (e.g., Marthinussen 1962, Andersen 1968, 1979, Mangerud 1970, 1977, Anundsen 1977, Vorren & Elvsborg 1979, Mangerud 1980, Andersen et al. 1981, 1982, Sørensen 1983, Follestad 1985, Sørensen 1992, Reite 1994, Bergstrøm 1999), it is thought by most authors to be linked to a period of climatic deterioration. This is supported by vegetation data from some areas. For example, a short interval at c. 12.0 – 12.2 ka BP of slightly colder conditions than the mean temperature for the Bølling – Allerød interval (13–11 ka BP) as a whole is documented in pollen records from some basins along the coast of western Norway (Paus 1988, 1989, 1990, Birks et al. 1994). The OD climatic fluctuation at c. 12 ka BP is also documented in marine records from the North Sea and the Norwegian Sea areas (e.g., Koc Karpuz & Jansen 1992, Koc & Jansen 1994), and may also be reflected farther north in a significant ice advance of approximately the same age at the western margin of the Svalbard-Barents Sea ice sheet (e.g., Svendsen et al. 1996).

The documented length of the OD ice advance is normally only some few km, but reached a length similar to that recorded northwest of Svartisen (> 10 km) in at least one more area, which is situated along the coast of Nord-Trøndelag (Sveian & Olsen 1991, Olsen & Sveian 1994).

Younger Dryas (YD), regional ice advances c. 10–11 ka BP (Fig. 11). – The main ice advance event northwest of Svartisen during this interval is recorded and named the Glomfjord event by Rasmussen (1981). This event is correlated with the Tjøtta Substage, which occurred in two phases at c. 10.8–10.9 and 10.5–10.6 ka BP, and which represent the main YD ice advance phases for Nordland county as a whole (Andersen et al. 1982, 1995). The new information reviewed here does not require significant changes for the reconstructed position of the ice margin during the Glomfjord event, which is generally situated at the end of the fjord-valleys in this area (Fig. 8; Rasmussen 1981, Olsen 2001d). However, a change with a considerably more extensive YD ice (> 5–15 km) than reported earlier is recorded north of Bodø, during the corresponding Straumøy event (Olsen 2001b).

The glacial conditions and ice-sheet fluctuations seem to have been more complex and therefore more difficult to reconstruct during the later part of the YD interval. During the Glomvasshaug event (Rasmussen 1981), which probably corresponds to the Nordli Substage at 10.1–10.2 ka BP for Nordland as a whole (Andersen et al. 1995), the ice advance in the area north of Glomfjorden may, in places, have reached a position distally to the main YD moraines. Similar ice extensions may have occurred during an early part of the YD interval. However, this suggestion is not based on the occurrence of ice-marginal moraines, but on shell dates of late Allerød to mid-YD age in tills /glacial diamictons from sites which are located distally to the early YD Glomfjord

Fig. 12: Photograph of a cirque moraine. The glacier crossed the Allerød – early YD shoreline (in bedrock) at the mouth of Tenøyrsdalen, south of Nordfjorden (the eastern part of Holandsfjorden).

event moraines (e.g., at Storvika (1) east of the Skogreina Moraine, at Djupvika (2), Ørnes, and at Furumo (3), Åmøya). Diamictons of this type may have been produced by thick sea-ice, icebergs or grounded ice shelves. The altitude of the recorded localities with these diamictons ranges from just below the YD sea level to just above the present sea level, which indicates that thick sea-ice is not a likely producer of the diamictons. Icebergs are a better alternative, because the production of icebergs must have been prolific during the intensive calving and rapid ice retreat that occurred just before and after the YD ice advances, and icebergs of different sizes could be grounded and reach the bottom or sides of the fjords at any depth in many places.

Clast fabrics in the diamictons trending parallel to the main ice-flow directions may be an argument against icebergs and for a grounded ice shelf, but the number of recorded localities is so far not high enough to exclude icebergs as a producer of the diamictons. Numerous icebergs floating together may freeze and grow together during favourable climatic conditions, as was the case during parts of the YD interval. An iceberg complex of this type can probably move like a fjord glacier in some areas, and where grounded it can probably produce diamictons similar to the type we have recorded for the YD interval. However, it is thought, as a working hypothesis, that an ice shelf developed in this area during the YD interval (Fig. 8; Olsen 2001d, Olsen et al. 2001d). This shelf ice was grounded in places and remoulded and redeposited shell-bearing diamictons as tills, e.g. at Djupvika and Storvika. A similar glacial development with shelf ice in contact with land-based ice is also suggested for the late YD interval in Nord-Trøndelag (Olsen & Sveian 1994), where the reconstructed ice extension is also based on stratigraphical evidence (shell-bearing tills with a clast fabric trending parallel to the main ice-flow direction) and not ice-marginal deposits. During such glacial conditions no ordinary ice-marginal moraines would have been deposited, but grounding-line moraines may have accumulated where suitable ice-grounding conditions occurred. The glacial diamicton recorded at Spilderneset and Korsneset in the outer part of the small bay of Spildervika adjacent to the Djupvika till site, may represent such a grounding-line moraine from the late YD interval (Olsen 2001d).

Several cirque glaciers occurred distally to the inland ice sheet during the YD interval in the area northwest of Svartisen (Rasmussen 1981). It is not known whether small cirque glaciers also existed during the Allerød interstadial in some of these locations. Shorelines developed during the Allerød and early YD intervals were crossed by different cirque glaciers (e.g., Fig. 12), which indicate that such glaciers existed during both the early and the late parts of the YD interval (Rasmussen 1981, Gjelle et al. 1995).

Conclusions

The main conclusions of this compilation are as follows:-

1. The ice-sheet fluctuations during the 10–40 ka BP interval in the area northwest of Svartisen follow closely, in number and time, all but one of the major regional, west Scandinavian, ice-sheet fluctuations (stadials and interstadials) of this interval (Olsen et al. 2001a, b, c). The regional ice advance at c. 40 ka BP is probably represented, but not well documented, in the available stratigraphic record from this area (Fig. 10).
2. Temperate Atlantic water, inferred from the presence of marine fossils which indicate temperate sea-surface conditions (e.g., *Arctica islandica;* underlined in Fig. 7), reached at least to latitude 67° N during periods of ice-free coasts in the interval 28–38 ka BP. Similar data from the coast farther north, at Grytåga, Fauske, c. 67° 20' N, at Grytøya, Harstad, c. 68° 48' N and at Slettaelva, Tromsø, c. 69° N, for an older ice-free period at c. 41–42 ka BP have previously been reported by Olsen et al. (2001c), Olsen & Grøsfjeld (1999) and Vorren et al. (1981), respectively.
3. The lateglacial Substage A (Andersen et al. 1981), which is represented by, e.g., the Risneset and Skogreina Moraines (Fig. 1) and tills at several sites (Stigen, Åsmoen, Vargvika, Sandvika & Stamnes), is ^{14}C-dated to 12.0 – 12.2 ka BP (Table 4). It is, therefore, indeed an Older Dryas glacial advance event, as suggested but not well documented by dates previously reported from this region. The length of this advance is at least 10–15 km in this area (Fig. 11).
4. The glacial history with ice extension and glacial development during the Younger Dryas interval is characterised by rapid ice-margin fluctuations, intensive calving with iceberg production and the inferred development of shelf ice in contact with land-based ice, and glacial diamictons (tills), but with few ordinary, late YD ice-marginal moraines in the coastal zone (Figs. 8 & 11; Olsen 2001d, Olsen et al. 2001d). This is a similar situation to that recorded in the coastal area of Nord-Trøndelag (Olsen & Sveian 1994).

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Instructions to authors – NGU Bulletin

First – a bit of history

The Geological Survey of Norway, which was founded in 1858, began publishing collections of papers on Norwegian geology in 1890. During these early years, and in the first part of the 20th century, almost all papers were in Norwegian, although the very first article in English appeared in 1905. Papers in the English language have dominated from around 1960. The publications, which were numbered consecutively, had no formal title until as late as 1972 when two separate series were introduced – the ***Bulletin***, in English, and the ***Skrifter***, in Norwegian. The consecutive numbering was maintained, however, with a secondary number for each series, but in 1984 the *Bulletin* formally took over the main numbering system that had then existed for over 90 years. At the start of the new millennium, after 110 years of publishing, we had reached *Bulletin 435*.

The Bulletin today

The *Bulletin* publication series of the Geological Survey of Norway comprises contributions to the Earth Sciences with special reference to Norwegian geology, geophysics, geochemistry, hydrogeology and certain aspects of environmental geology. Manuscripts on particular topics of geoscience research from neighbouring countries may, in certain cases, also be considered, providing that the subject matter is of direct interest to Norwegian geology.

In general, *Bulletin* numbers consist of several research papers on diverse topics. From time to time, a collection of papers may appear as a thematic *Bulletin* number; for example, on the geology of one particular region, or on one particular aspect of geoscientific research.

The *Bulletin* does <u>not</u> appear on a regular basis, as with most other journals. On average, we publish two Bulletins per year. For technical reasons, we wait until we have enough final, accepted, manuscript copy to make up a minimum of 60 printed pages.

General

For a general idea on the layout of the journal, please try to consult one of our recent issues, i.e. those printed in A4 format starting with *Bulletin 431* from 1996. An exception to this rule is *Bulletin 433*, which is a collection of extended abstracts. As there have been some slight changes in layout, form of references, etc., over the years, it is advisible to brouse through one of our very latest issues.

Three copies of the manuscript should be sent to the Editor. To save time, and to allow us to contact potential reviewers, you may send an advance copy of your manuscript, and digitally drawn figures, via e-mail. We will then distribute two of the submitted copies to reviewers, who generally have about 4–5 weeks in which to produce their verdict on the ms.

The language used for manuscripts in our *Bulletin* series is English. We accept manuscripts in both conventional British English and American English, but the syntax, spelling, etc., should, of course, be consistent throughout any one ms. As Norway is a part of Europe (although some Norwegians may dispute this!), Norwegian authors should try to follow British English. An exception to this might be where a Norwegian first author is coauthoring with an American colleague.

Authors are requested to have their ms read by at least one colleague familiar with the area or subject-matter before submission. The name of this person, or persons, should be included in the Acknowledgements. To authors whose mother-tongue is not English, please try to have your English text checked for possible linguistic errors before submission. A well-written manuscript makes life easier and simpler for the reviewers. Whatever the case, all final, accepted manuscripts will be carefully controlled, from the linguistic and technical points of view, here at NGU.

An author will generally receive only one set of proofs. These should be read carefully and returned promptly to the Editor. No changes to the text, other than correction of errors, will be allowed at this stage. A reprint order form will be sent out with the proofs. Twenty-five reprints will be sent to the first author free of charge. Additional reprints can be ordered on the form provided when returning the proofs. The cost of reprints will vary according to the number of pages in the printed article.

Main text

Manuscripts should be typewritten on one side of standard, A4-size paper, with an ample margin on the left-hand side, and should be double-spaced throughout. All pages should be numbered. Figures and Tables should appear, and be numbered, in accordance with the sequence of their citation in the text. Manuscripts should be arranged in the following order:-

1. Title in small case, 16-point font.
2. Names of the authors, in capitals; 12-point font. First name followed by surname.
3. Abstract. This is preceded by authors' surnames and initials, year, title of the manuscript and journal name (the last, in italics) – look at any recent *Bulletin* for the correct style. The names and addresses of the authors should be placed below the abstract, in italics. If you wish, add an e-mail address for the corresponding author.
4. The main text of the manuscript.

5. Acknowledgements.
6. References.
7. List of figure captions.
8. Figures and tables.

(1) The ***title*** should be brief, but informative. Begin with a word useful in indexing and information retrieval.

(3) The ***abstract*** should be no more than 300 words. An abstract should be a synopsis of the results, interpretations and conclusions. It should not discuss the subjects covered or the methods employed. References, figures or tables should not be cited in an abstract.

(4) For the general layout of the main text, please consult any recent issue of the *Bulletin*. Arrange the headings according to rank, i.e., first-order, second-order, etc. First-order – small case, 14-point bold; second-order – 12-point bold; third-order. 12-point normal, italics.

All geographical names mentioned in the text should appear on one or other of the accompanying figures. Exceptions are place names cited which are far outside the confines of the area discussed in the manuscript. Points of the compass should be abbreviated where they pertain to structural trends or directions, e.g., " the fold axes trend NE-SW"…. or "the NW-plunging lineation…" Otherwise, use small case, e.g., "in the southwest of the area…". For specific directions, etc., write, e.g., …065°…., not …N65°E.

References to authors in the text should be written as follows: "As Carstens (1960) has shown,…", or ….."…high-grade metamorphism has been reported (Zwaan 1975, Black & White 1980, Glenfiddich et al. 1988)." For personal communications, write initials, name and year, e.g., "…fossils have recently been reported (D.L.Bruton, pers. comm. 1994)."

Avoid using footnotes. Such incidental information should be incorporated into the main text, or included in parentheses. Excessively long tables of analytical and other data should be avoided.

Figures and Tables should be cited as follows: Fig. 2, Figs. 6 & 7, Table 3. In the rare cases where a map will exceed full page size, these illustrations are called Plates. Such Plates are inserted in a pocket on the inside back cover of the Bulletin.

Any manuscript, whatever its length, should end with a short Conclusions. Give a clear, explicit statement of the main results of the study and the interpretations.

(5) The ***Acknowledgements*** should be kept as short as possible, thanking the pre-submission readers and the reviewers, as well as sources of funding, etc..

(6) For the style of ***References***, please consult any recent Bulletin. Note that all journal names are written out in full, in italics. Italics are also used for the titles of doctoral or other theses, and for book titles. Indent the second and subsequent lines of all references.

Check that all references cited in the text are included in the list of references. Conversely, all references in the list should be cited somewhere in the main text or in the Figure/Table captions.

(7) Figure ***captions*** should be collected together on a separate page or pages at the end of the manuscript.

(8) All ***figures*** should be reducible to a maximum page width of 17.5 cm, or maximum page height of 24 cm. Many figures can be reduced to single-column width, 8.3 cm. In some cases, we print figures at 10-12 cm width, and place the figure caption to the side of the figure. Figures should be submitted either at the preferable, final size, or slightly larger than the final size.

When planning the drafting of maps and other line-drawings, think ahead to the final, printed version; and please ensure that all lettering and numbers will be a minimum of 1 mm in height after reduction. Publishing any journal nowadays is a costly business, so try to arrange your figures, legends, etc., in as space-saving a way as possible.

Map figures should preferably be oriented with north at the top, and a north arrow should be included (unless latitude/longitude marks along the figure frame make this unnecessary). Use a bar scale on all maps. Check that the spelling of all place names conforms with that appearing on the official 1:50,000 topographic map series. All map figures should be enclosed in a frame.

Regarding ornament on maps and diagrams, avoid using too many grey tones, or fine stipples, as these do not reproduce well. Very pale grey tones should definitely be avoided. All lines should have line widths of at least 0.5 mm.

Some figures may be reproduced in colour. This applies especially to maps with a complex and detailed geology, geophysical anomaly maps or particular types of figure. Selected field photographs may also be printed in colour. However, we try to be restrictive with the use of colour, as such printing adds appreciably to the overall cost of producing the *Bulletin*.

Photographic figures of field motifs may appear either as individual figures or as a montage, A, B, C, etc., which together constitute one figure. With today's technology, such figures can be constructed by scanning slides/dias, and ultimately transferring these figures electronically to NGU. However, in some cases the printer may wish to borrow your slides.

All photographs should carry some form of scale, preferably placed discretely below or to the side of the subject matter. Photos with a large hammer in the middle of the picture are likely to be rejected. All photomicrographs should carry a bar-scale in the bottom right-hand corner.

As well as a form of presenting essential data, ***tables*** are, in some cases, meant to supplement or replace lengthy text, but *not* to duplicate it. Present tables on separate sheets. Avoid using horizontal and vertical lines, except those which are absolutely necessary, e.g., below the column headings or separating distinct groups of analyses. Eliminate wasteful space between columns – many tables can be reduced to single-column width simply by carefully thinking ahead to the final print size.

Other points

Final, revised manuscripts: When a manuscript has been finally accepted, the author will be asked to supply us with a paper copy of the final word-processed text, plus figures, tables, etc. In addition, the manuscript, figures and tables should be tranferred to the Editor electronically (ms in Word format), either as e-mail attachments, or on a standard disc/diskett or CD-rom.

New lithostratigraphic names: Authors writing manuscripts in which new lithostratigraphic or structural names are introduced or proposed are asked to mention this in their cover letter when submitting. We can then check if these names are acceptable, following the rules of the Norwegian Committee for Stratigraphy, while the ms is being reviewed.

Short manuscripts: As well as normal-length manuscripts, shorter contributions that will take up just 3 or 4 printed pages are also welcome. Abstracts to such short papers should not exceed 150 words.

Publication costs: In cases where large coloured maps (Plates) are involved, we may be forced to ask authors to provide a contribution to these extra printing costs.

Reviewing procedure, etc.: While every effort is made to shorten the process of refereeing and editing of mss, there will always be delays during the summer period, May to September, because of fieldwork commitments, university vacations, and holidays generally.

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