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.

Tom Andersen and Arild Andresen, Department of Geology, University of Oslo, PO Box 1047 Blindern, N-0316 Oslo, Norway. Arthur G. Sylvester, Department of Geological Sciences, University of California, Santa Barbara, CA 93106-9630, USA.

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

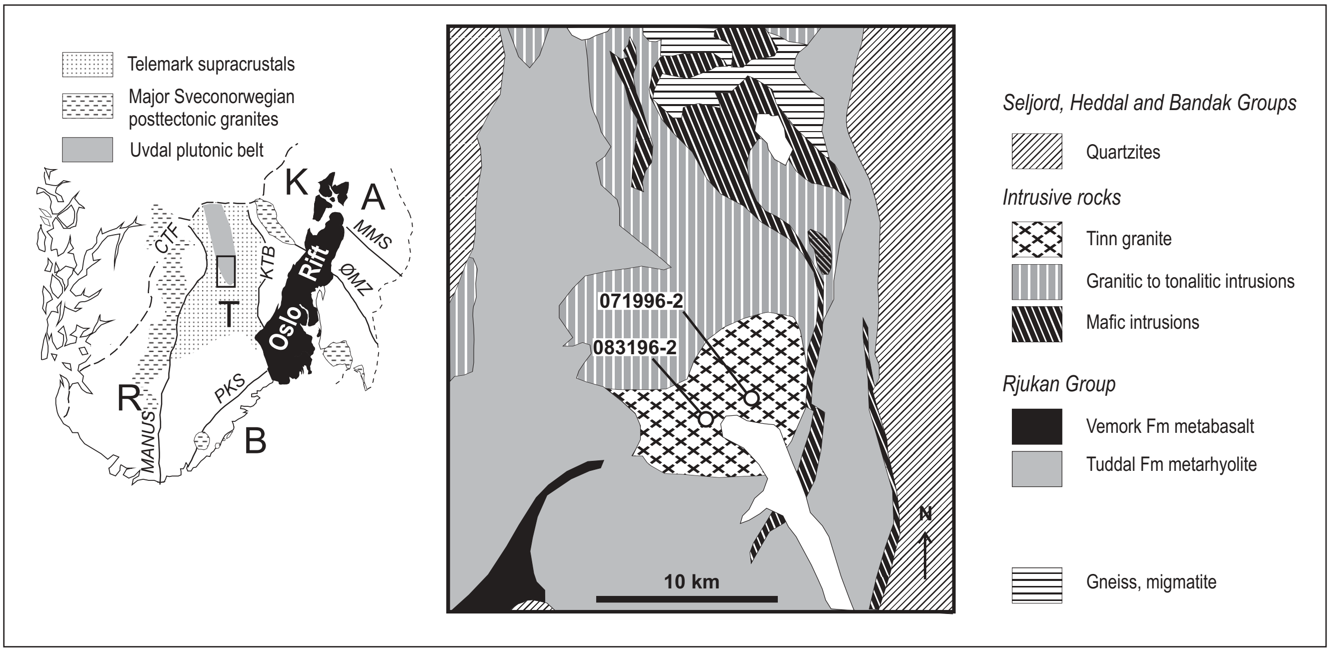

Fig. 1. Simplified geologic map of the Tinn granite and surrounding rocks, compiled from Dons & Jorde (1978) and Sigmond (1998). Overview map: Simplified map of South Norway showing regional subdivision used in this study: A: Østfold-Akershus sector; K: Kongsberg sector; T: Telemark sector; B: Bamble 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

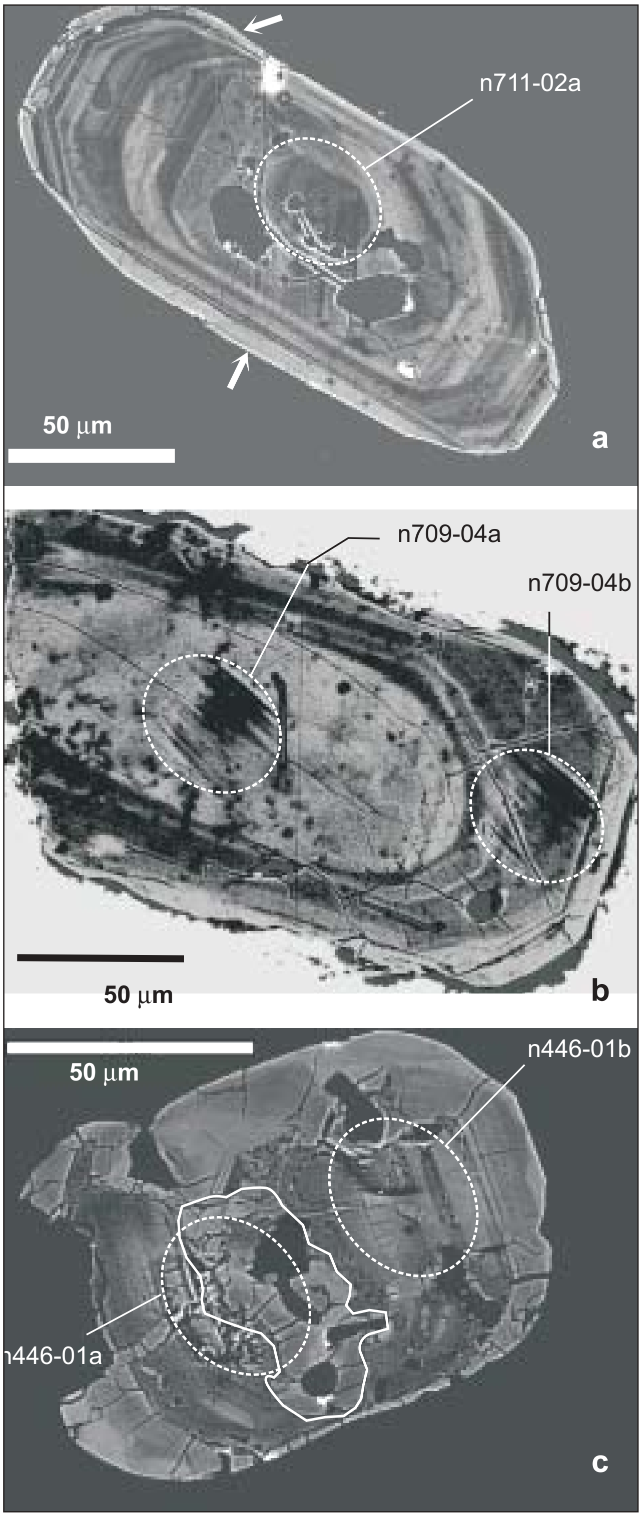

Fig. 2. Backscatter electron images of zircons from the Tinn granite. Location of SIMS analytical spots indicated (stippled ellipses) with numbers referring to Table 1. Images made after analysis, using a Cameca SX-50 electron microprobe at GEMOC National Key Centre, Macquarie University, Australia. 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

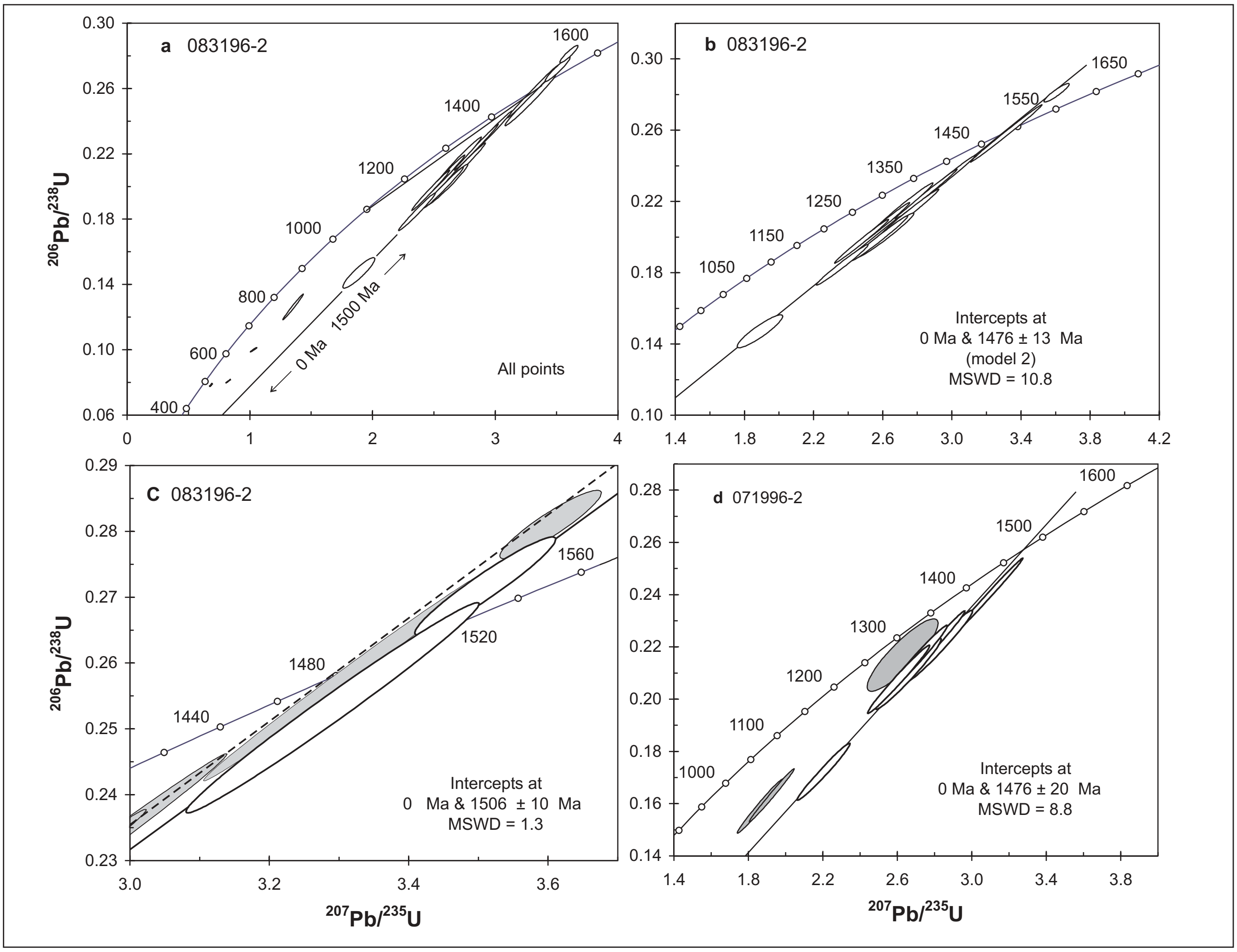

Fig. 3. Concordia diagrams of SIMS U-Pb data for the Tinn granite (Table 1). Data-points are shown with 1s error ellipses. a: All points, with a reference recent lead-loss line drawn to 1500 Ma. Note widely discordant grains, plotting between the reference line and concordia. These grains lost radiogenic 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

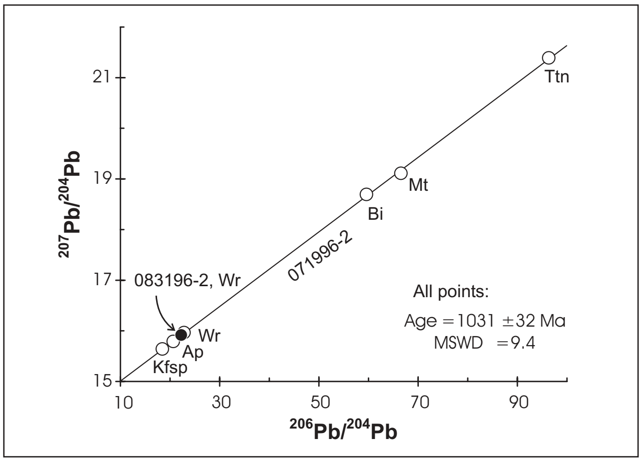

Fig. 4. Lead isotope data for whole rocks and minerals from the Tinn granite; data from Table 2. The age represents an event of Sveconorwegian lead isotope homogenization, whose timing is comparable to other events in South Norway (e.g. Andersen & Munz 1995). Abbreviations: Ttn: titanite; Mt: magnetite;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

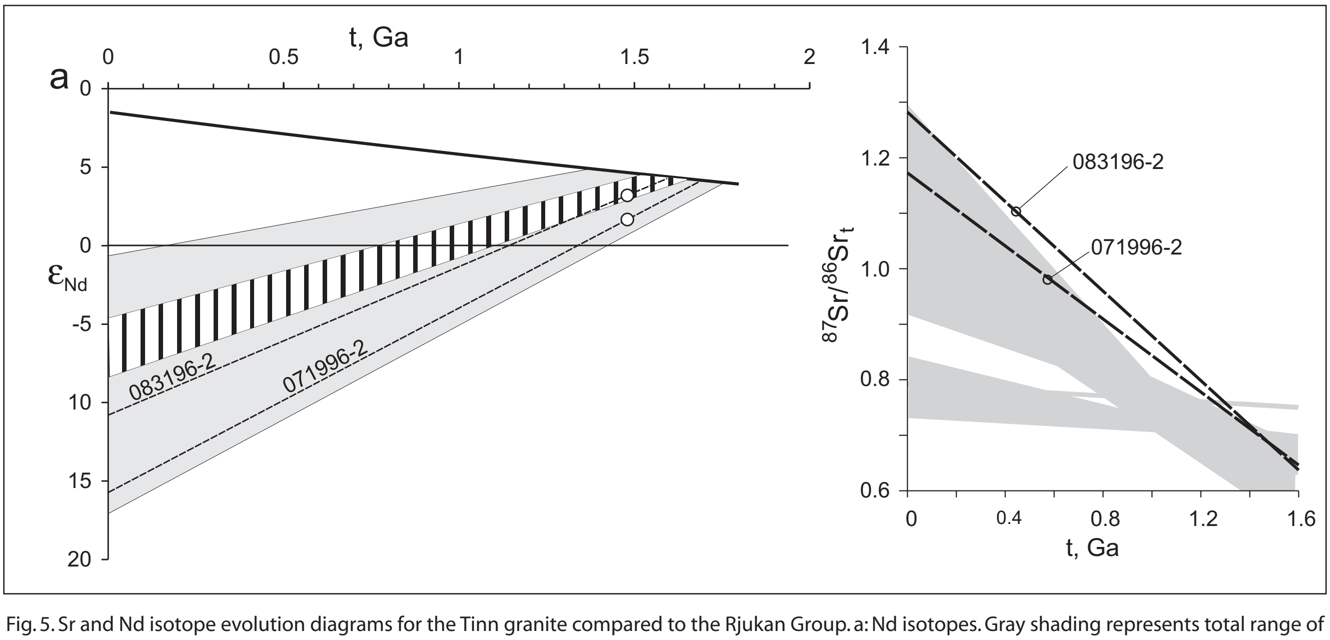

Fig. 5. Sr and Nd isotope evolution diagrams for the Tinn granite compared to the Rjukan Group. a: Nd isotopes. Gray shading represents total range of Tuddal Formation; striped field is corresponding range of Vemork Formation (data from Brewer & Menuge 1998). b: Sr isotopes. Gray shading represents total 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.

Acknowledgments

This study has been supported by the Norwegian Research Council (grants no.1105774/410 and 128157/410 to T.A., and a visiting scientist fellowship to A.G.S.) and by a travelling grant to TA from the Faculty of Mathematics and Natural Sciences, University of Oslo. The Nordsim laboratory is funded by the joint Nordic research councils and the Swedish Museum of Natural History. Analytical assistance by Gunborg Bye-Fjeld, Turid Winje and Norman J. Pearson is gratefully acknowledged. T.A. wants to express special thanks to professors W.L. Griffin and S.Y. O'Reilly, Macquarie University, Sydney, Australia, for an invitation to visit the GEMOC Key Centre, during which a first draft was produced. Thanks to Bernard Bingen, Øystein Nordgulen and Ellen Sigmond for their critical comments on the manuscript. Nordsim contribution no. 74.

References

Andersen, T. & Knudsen, T.-L. 2000: Crustal contaminants in the Permian Oslo Rift, South Norway: Constraints from Precambrian geochemistry. *Lithos 53*, 247-264.

Andersen, T. & Munz, I.A. 1995: Radiogenic whole-rock lead in Precambrian metasedimentary gneisses from South Norway: Evidence for LILE mobility. *Norsk Geologisk Tidsskrift 75*, 156-168.

Andersen, T., Andresen, A. & Sylvester, A.G. 2001: Nature and distribution of deep crustal reservoirs in the southwestern part of the Baltic Shield: Evidence from Nd, Sr and Pb isotope data on late Sveconorwegian granites. *Journal of the Geological Society, London 158*, 253-267.

Bingen, B., Birkeland, A., Nordgulen, Ø. & Sigmond, E.M.O. 2001: Correlation of supracrustal sequences and origin of terranes in the Sveconorwegian orogen of SW Scandinavia: SIMS data on zircon in clastic metasediments. *Precambrian Research 108*, 293-318.

Brewer, T.S. & Menuge, J.F. 1998: Metamorphic overprinting of Sm-Nd isotopic systems in volcanic rocks: the Telemark Supergroup, Southern Norway. *Chemical Geology 45*, 1-16.

Byers, F.M., Carr, W.J. & Orkild, P.P. 1989: Volcanic centers of southwestern Nevada: Evolution of understanding, 1960–1988. *Journal of Geophysical Research 94,* 5908–5924.

Dahlgren, S.H., Heaman, L. & Krogh, T.E. 1990: Geological evolution and U-Pb geochronology of the Proterozoic central Telemark area, Norway (abstract). *Geonytt 17 (1)*, 38.

De Paolo, D. 1981: Trace element and isotopic effects of combined wall-rock assimilation and fractional crystallization. *Earth and Planetary Science Letters 5*3, 189-202.

Dons, J.A. 1960: Telemark supracrustals and associated rocks. *Norges geologiske undersøkelse 208*, 49-58.

Dons, J.A. & Jorde K. 1978: *Geologisk kart over Norge, Berggrunnsgeologisk kart SKIEN, M 1:250 000*. Norges geologiske undersøkelse, Trondheim.

Elders, W. A. 1963: On the form and mode of emplacement of the Herefoss granite. *Norges geologiske undersøkelse 214*, 1-52.

Haas, G.J.L.M de., Andersen, T. & Vestin, J. 1999: Application of detrital zircon geochronology to assembly of a Proterozoic terrain – an example from the Baltic Shield. *Journal of Geology 107,* 569-586.

Heaman, L.M. & Smalley, P.C. 1994: A U-Pb study of the Morkheia Complex and associated gneisses, southern Norway: implications for disturbed Rb-Sr systems and the temporal evolution of Mesoproterozoic magmatism in Laurentia. *Geochimica et Cosmochimica Acta 58*, 1899-1911.

Johannes, W. & Holtz, F. 1996: *Petrogenesis and experimental petrology of granitic rocks*. Springer-Verlag, Berlin, 335 pp.

Kleppe, A.V. 1980: *Geologiske undersøkelser fra sentrale deler av Telemark med hovedvekt på geokronologi, og kontaktrelasjonen mellom granittisk gneiss og suprakrustalene.* Cand. Real. thesis, University of Oslo, 157 pp.

Knudsen, T.-L., Andersen, T. Whitehouse, M.J. & Vestin, J. 1997: Detrital zircon ages from Southern Norway - implications for the Proterozoic evolution of the Southwestern part of the Baltic Shield. *Contributions to Mineralogy and Petrology 130,* 47-58.

Laajoki, K., Corfu, F. & Andersen, T. 2000: U-Pb zircon dating of the Mesoproterozoic Brunkeberg Formation and its bearing on the stratigraphy of Telemark Supracrustals, South Norway. *Goldschmidt 2000, Oxford UK, Journal of Conference Abstracts 5(2)*, 611.

Lipman, P.W. 1992: Magmatism in the Cordilleran United States: Progress and problems. *The Geology of North America, Vol. G-3: The Cordilleran Orogen: Conterminous U.S.* The Geological Society of America.

Ludwig, K. R. 2000: Isoplot/Ex, version 2.3. A Geochronological Toolkit for Microsoft Excel. *Berkeley Geochronology Center, Special Publication No. 1*.

Menuge, J.F. & Brewer, T.S. 1996: Mesoproterozoic anorogenic magmatism in southern Norway. In Brewer, T.S. (ed.): *Precambrian Crustal Evolution in the North Atlantic Region*. Geological Society Special Publication 112, 275-295.

Perry, F. V., Crowe, B.M., Valentine, G.A. & Bowker, L.M. 1998: Volcanism Studies: Final Report for the Yucca Mountain Project. Los Alamos National Laboratory Report LA-13478.

Ploquin, A. 1972: Le granite acide d'Åmli, Norvège méridonale: Transformation des laves acides du Tuddal (Formation inférieure des Séries de roches supracrustales du Telemark). Sciences de la Terre 17, 81-95.

Rollinson, H. 1993: Using Geochemical Data. Longman Scientific & Technical. Harlow, United Kingdom, 352 pp.

Sigmond, E.M.O. 1998: Geologisk kart over Norge, Berggrunnsgeologisk kart ODDA, M 1:250 000. Norges geologiske undersøkelse, Trondheim.

Sigmond, E. M.O., Gjelle, S. & Solli, A. 1997: Telemarksgruppens oppbygning fra det sentrale Telemark nordover til de kaledonske dekken, - Rjukangruppens underlag og bergartene i Uvdalsplutonbeltet (abstract). *Geonytt 24 (1)*, 82.

Simonsen, S.L. 1997: *Radiogenic isotope systematics of charnockite and retrograded augen gneiss in the Gjeving Complex, South Norway.* Cand. Scient. Thesis, University of Oslo, 60 pp.

Sylvester, A. G. 1998: Magma mixing, structure, and re-evaluation of the emplacement mechanism of Vrådal pluton, central Telemark, southern Norway. *Norsk Geologisk Tidsskrift 78*, 259-276.

Verschure, R.H., Maijer, C. & Andriessen, P.A.M. 1990: Isotopic age determinations in South Norway: 2. The problem of errorchron ages from Telemark rhyolites. *Norges geologiske undersøkelse Bulletin 418*, 47-60.

Whitehouse, M.J., Claesson, S., Sunde, T. & Vestin, J. 1997: Ion-microprobe U-Pb zircon geochronology of the Archaean gneisses from the Gruinard Bay Lewisian Complex, north-west Scotland. *Geochimica et Cosmochimica Acta 61*, 4429-4438.

Whitehouse, M.J., Kamber, B.S. & Moorbath, S. 1999: Age significance of U-Th-Pb zircon data from early Archaean rocks of west Greenland – a reassessment based on combined ion-microprobe and imaging studies. *Chemical Geology 160*, 201-224.