

Mineralogy and geological setting of allanite-(Ce)-pegmatites in western Hurrungane, Jotun Nappe Complex, Norway: an EMP and ID-TIMS study

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The 950 ± 1 Ma Berdalsbandet Pegmatite Swarm (BPS) in Hurrungane, in the Jotun Nappe Complex, SW Norway, consists of ≤ 2 m wide subparallel granitic dykes emplaced in a sinistral non-coaxial shear zone, constraining the age of local Sveconorwegian deformation. The abyssal class dyke swarm is interpreted to be related to local Sveconorwegian anatexis; mineralogy and REE characteristics indicate low degree partial melting. The pegmatites are symmetrically zoned and typically consist of an outer zone of feldspar + allanite-(Ce) + biotite, a transitional zone of feldspar + quartz + schorl + Fe-Ti-oxides, an inner zone of quartz + schorl + garnet + beryl and a core of hydrothermal quartz. A metasomatic rim of biotite is locally developed along contacts to the amphibolite wallrock. Secondary zeolites are suggested to relate to late- to post-Caledonian alteration. The dykes contain various rare-earth minerals, predominantly allanite-(Ce). Its composition, determined by electron microprobe (EMP) and expressed by representative analyses from three samples, is: $(\text{Ca}_{0.86}\text{REE}_{0.80}\text{Mn}_{0.09}\text{Th}_{0.01}\text{Na}_{0.03})_{\Sigma 1.79}(\text{Al}_{1.79}\text{Fe}^{2+}_{0.99}\text{Mg}_{0.16}\text{Ti}_{0.10})_{\Sigma 3.04}(\text{Si}_{3.17}\text{P}_{0.01})_{\Sigma 3.18}\text{O}_{12}(\text{OH})$, $(\text{Ca}_{1.12}\text{REE}_{0.72}\text{Mn}_{0.03}\text{Th}_{0.01}\text{Na}_{0.02})_{\Sigma 1.90}(\text{Al}_{1.73}\text{Fe}^{2+}_{0.66}\text{Fe}^{3+}_{0.33}\text{Mg}_{0.26}\text{Ti}_{0.03})_{\Sigma 3.01}\text{O}_{12}(\text{OH})$ and $(\text{Ca}_{1.10}\text{REE}_{0.74}\text{Mn}_{0.02}\text{Th}_{0.01}\text{Na}_{0.01})_{\Sigma 1.88}(\text{Al}_{1.75}\text{Fe}^{2+}_{0.69}\text{Fe}^{3+}_{0.26}\text{Mg}_{0.30}\text{Ti}_{0.04})_{\Sigma 3.04}\text{O}_{12}(\text{OH})$. The ratio of $\text{Fe}^{3+}/(\text{Fe}^{3+}+\text{Fe}^{2+})$ is variable. Dominant substitution mechanisms are $\text{Ca}^{2+} + \text{M}^{3+} = \text{REE}^{3+} + \text{M}^{2+}$ and $3\text{Ca}^{2+} = 2\text{REE}^{3+}$. Isotope dilution thermal ionisation mass spectrometry (ID-TIMS) documents excess ^{206}Pb reflecting high Th/U ratios, showing that the allanite largely retains its magmatic Pb isotopic signature despite the regional Caledonian upper greenschist- to lower amphibolite facies metamorphism. Local compositional heterogeneities in the allanite crystals are suggested to reflect Caledonian annealing of metamict domains accompanied by minor Pb-loss.

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Introduction

During Sveconorwegian collisional tectonics, igneous rocks of late Paleoproterozoic to Mesoproterozoic age, now exposed in the Upper Jotun Nappe, SW Norway, were metamorphosed and deformed under granulite- to amphibolite facies conditions (Fig. 1). This led to at least two episodes of anatexis and the emplacement of several generations of granitic pegmatites (Lundmark et al. 2007). During Caledonian thrusting of the Upper Jotun Nappe, renewed magmatism led to the emplacement of 427 ± 1 Ma, synkinematic granitic dykes, the Årdal dyke complex (Lundmark & Corfu 2007). The different generations of granitic pegmatites and dykes offer a reliable way of constraining the relative, and sometimes absolute ages of different phases of deformation and metamorphism in this polydeformed and polymetamorphic area. Detailed studies of mineral textures and compositions may allow conclusions on the specific conditions leading to generation and emplacement of the dykes at well constrained times to be drawn.

This study focuses on the 950 ± 1 Ma (Lundmark et al. 2007), allanite-(Ce)-bearing Berdalsbandet Pegmatite Swarm (BPS) in the western Hurrungane region, part of the Upper Jotun Nappe (Fig. 1). It presents the first detailed mineralogical description of the BPS as well as mineral data from its amphibolite host rock together with a description of the local geological setting. The BPS displays a mineral spectrum dominated by quartz, alkali feldspar, plagioclase and allanite-(Ce). Allanite (Aln), $\text{CaREEAl}_2\text{Fe}^{2+}(\text{SiO}_4)_3\text{OH}$, is of particular interest since it is one of the main repositories for light rare earth elements (LREE) in the crust, and can be used in U-Th-Pb geochronology (e.g. Oberli et al. 2004, Gieré & Sorensen 2004). To investigate the response of the BPS allanites to late-Sveconorwegian and Caledonian metamorphic events, and their potential for U-Pb geochronology, isotope dilution thermal ionisation mass spectrometry (ID-TIMS) analyses were performed on parts of two grains. The mineral chemistry of allanite was determined by electron microprobe (EMP) to complement the limited published compositional data record on pegmatitic allanite. The results are used to

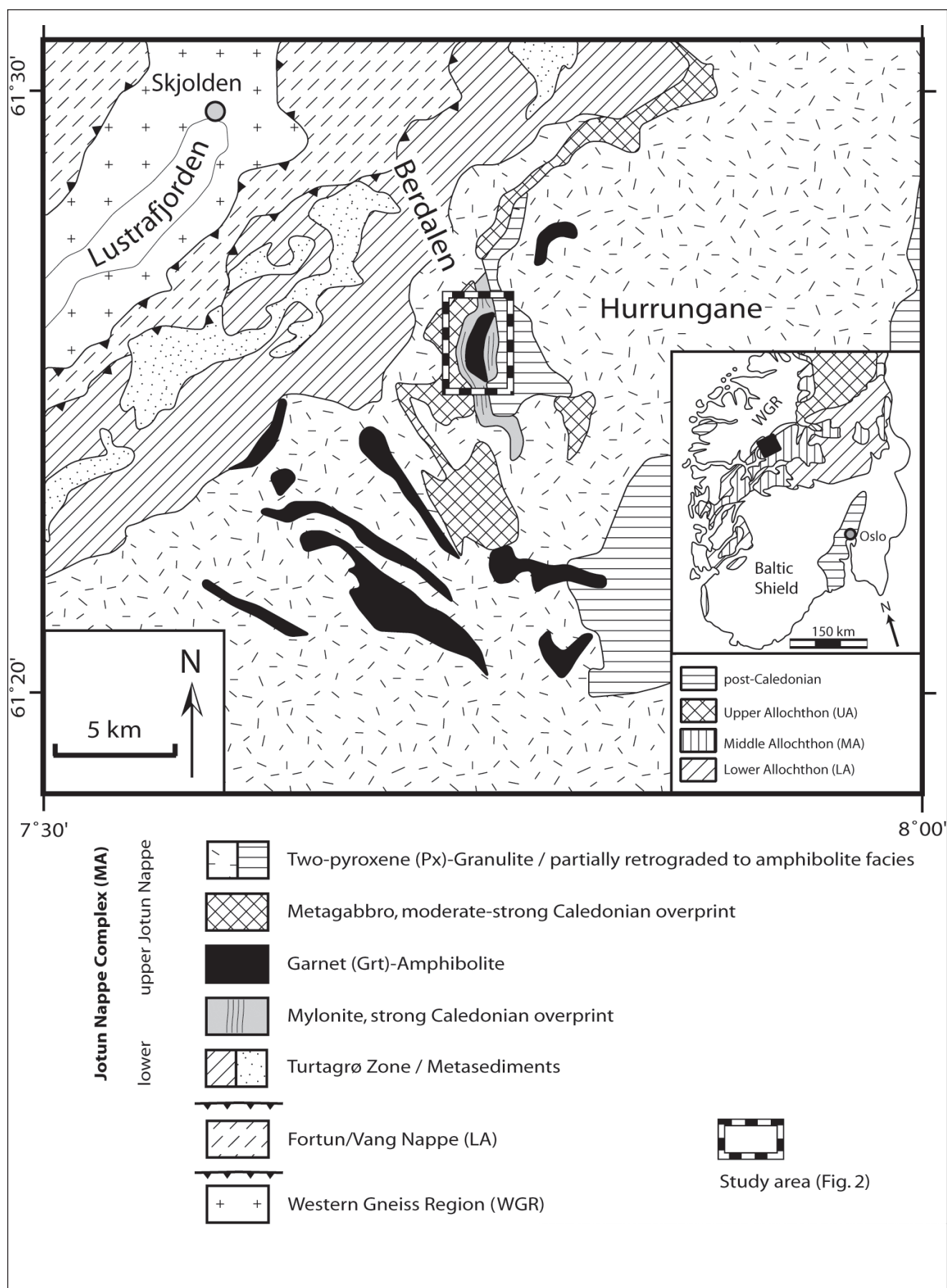


Fig. 1. Simplified geological map of Hurrungane, showing the NW margin of the Jotun Nappe Complex and the location of the study area (modified from Lutro & Tveten 1996 and Koestler 1983). Inset: Tectonostratigraphy of southern Norway (simplified from Roberts & Gee 1985). WGR = Western Gneiss Region. Black square in inset marks map section.

discuss and constrain the origin and metamorphic evolution of the pegmatites and the Hurrungane region.

Geological background

Southern Norway (Fig. 1) is traditionally divided into three main geological units. (I) The autochthonous and parautochthonous Precambrian crystalline basement (Baltic Shield and Western Gneiss Region, WGR). (II) The overlying nappes of the Caledonian orogen, subdivided into the Lower Allochthon (LA), Middle Allochthon (MA), Upper Allochthon (UA) and Uppermost Allochthon (UMA), the latter absent in southern Norway (Roberts & Gee 1985). (III) Post-Caledonian formations comprising Devonian molasse sediments, preserved in basins on the west coast of Norway and at Røragen, and the Permocarboneous Oslo Graben with related sediments and igneous rocks. The study area is located in the Jotun Nappe Complex (Bryhni & Sturt 1985), which makes up the largest part of the MA in central southern Norway.

The Jotun Nappe Complex

The medium- to high-grade Jotun Nappe Complex derives from Caledonian imbrication and thrusting of the Baltoscandian margin (e.g., Lundmark et al. 2007; Fig. 1). Caledonian strain is largely concentrated in the basal thrust zone of the nappe complex, the Turtagrø Zone, and in the phyllites of the Fortun/Vang nappes (Lower Allochthon) that underlie the Jotun Nappe Complex (Milnes & Koestler 1985; Lutro & Tveten 1996). Caledonian top-to-southeast thrusting and late-orogenic, top-to-northwest extension have been dated at ca. 410 Ma and 400 Ma, respectively, in the phyllites (^{40}Ar - ^{39}Ar cooling ages, Fossen & Dunlap 1998). Caledonian metamorphism in the Jotun Nappe Complex reached upper greenschist- to lower amphibolite facies conditions (Schärer 1980; Koestler 1983; Bryhni & Andréasson 1985; Milnes et al. 1997), and Caledonian metamorphic overprinting is extensive in the Turtagrø Zone, which comprises tonalitic to granitic mylonites and blastomylonites, metasediments and slices of crystalline lithologies (Milnes & Koestler 1985). In contrast, nappe-internal Caledonian deformation above the basal thrust zone is typically limited to distinct shear zones (Bryhni & Sturt 1985), except for areas of intense Caledonian magmatism and hydration, associated with locally extensive deformation (Lundmark & Corfu 2008a).

The Jotun Nappe Complex can be subdivided into two main units distinguished by their different metamorphic histories, which were juxtaposed during the Silurian south-eastwards translation over the WGR (Lundmark & Corfu 2008b). The medium grade Lower Jotun Nappe is exposed along the margins of the complex. It consists

mainly of syenitic to monzonitic Gothian igneous rocks crystallised between 1694 ± 20 Ma (Schärer 1980) and ca. 1600 Ma (Bryhni & Sturt 1985), and intruded by gabbros at 1252 ± 28 Ma (Schärer 1980). Amphibolite facies metamorphism was dated by Schärer (1980) to 909 ± 16 Ma. The Lower Jotun Nappe is separated by mylonitic shear zones from the overlying, variably retrograded granulite facies rocks of the Upper Jotun Nappe, where 1660 ± 2 to 1634 ± 5 Ma felsic gneisses and 1253 ± 15 Ma monzonite and gabbro conform to a similar pattern of protolith ages (Lundmark et al. 2007). The central and south-western parts of the Upper Jotun Nappe are dominated by an anorthosite-gabbro-troctolite suite. A 965 ± 4 Ma U-Pb zircon age was interpreted as its protolith age (Lundmark and Corfu 2007a). The minimum age of one event of Sveconorwegian granulite facies metamorphism in the Upper Jotun Nappe was constrained by 954 ± 3 Ma granitic anatexite in felsic country rock, whereas a second high-grade event, also associated with local anatexis, was dated to 934 ± 3 Ma (Lundmark et al. 2007).

The Upper Jotun Nappe is transected by numerous dykes, and several types have been mapped and described in the north-western Hurrungane region (Koestler 1989), where the study area is located. Granitic pegmatites are common, and different generations have been dated to 950 ± 1 Ma (the BPS), 942 ± 3 Ma and 927 ± 3 Ma, respectively (Lundmark et al. 2007); they exhibit preferred orientations on the outcrop/local scale, but the presence and possible significance of a regional-scale geometric pattern is so far uncertain. The ages disprove a previous suggestion that the pegmatites were cogenetic with the granulite host rocks and intruded during a late phase of their solidification under high grade metamorphic conditions (Koestler 1983). Also present are 1-10 m wide metadolerites of unknown age, which locally contain primary ophitic plagioclase-clinopyroxene-orthopyroxene assemblages and metamorphic garnet (e.g., Battey & McRitchie 1973; Koestler 1983). The youngest dykes belong to the 427 ± 1 Ma synkinematic Årdal dyke complex (Lundmark & Corfu 2007).

Geological setting of the study area

The study area is located in the Upper Jotun Nappe on the western flank of the Hurrungane mountains (Fig. 2). The dominant lithologies are felsic to mafic two-pyroxene granulites, mesocratic gabbroids, granitic gneisses and subordinate garnet amphibolite and spinel-bearing two-pyroxene peridotites. A N-S striking shear zone, defined by a wide zone of mylonitic gneisses, divides the area into an eastern and a western part, and encloses a lens of garnet amphibolite. The eastern part is dominated by variably deformed pyroxene granulites. Growth of hornblende blasts at the expense of orthopyroxene and clinopyroxene testifies to local

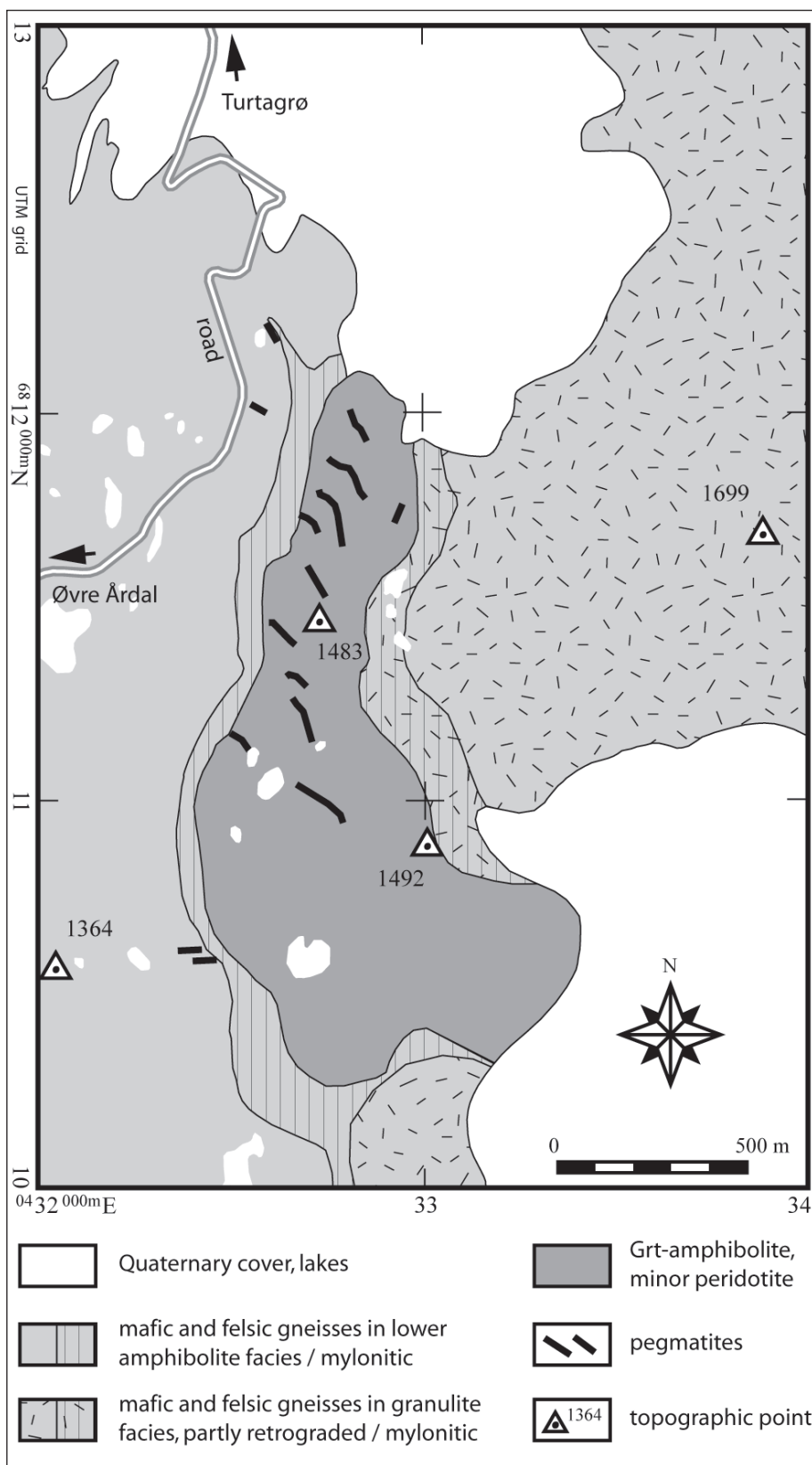


Fig. 2. Major lithological units at Berdalsbandet, Hurrungane.

retrogression under amphibolite facies conditions. Two different types of corona structures are observed in the garnet amphibolite lens and in related spinel-bearing pyroxene peridotites, respectively. In the garnet amphibolite, cores of rounded plagioclase are rimmed by garnet against clinopyroxene. Reaction of garnet with

clinopyroxene in the presence of H_2O has produced hornblende. In the pyroxene peridotites, olivine grains are surrounded by orthopyroxene against clinopyroxene. The metamorphic assemblages are compatible with a common metamorphic history for these rock types. West of the N-S striking shear zone, gabbroic and granitic

Table 1									
Alkali feldspar ^{a)}	++	p	Heulandite-(Ca), -(Sr) ^{c)}	-	s	Pyrite	--	s	
Allanite-(Ce)	++	p	Ilmenite	++	p	Quartz	++	p	
Apatite-(F)	+	p	Magnetite	++		Rutile	-	p	
Biotite	++	p	Malachite	-	s	Stilbite	-	s	
Beryl	--	p	Monazite-(Ce)	-	p	Thorite	--	p	
Chabazite	-	s	Muscovite	+	p	Titanite	+	p	
Chalcopyrite	-	s	Parisite-(Ce)	--	s	Tourmaline ^{e)}	++	p	
Epidote	+	s	Plagioclase ^{d)}	++	p	Xenotime-(Y)	+	p	
Fluorite	--	p	Polycrase-(Y)	--	p	Zircon	+	p	
Garnet ^{b)}	+	p	Prehnite	-	s				

relative abundance: ++ = very frequent, + = frequent, - = rare, -- = very rare

p = primary, s = secondary

^{a)} Orthoclase and microcline; ^{b)} Spessartine; ^{c)} analyses in Nordrum et al. (2005); ^{d)} Oligoclase; ^{e)} Schörl

Table 1. List of minerals found in the Berdalsbandet pegmatites and their relative abundance.

gneisses dominate. Caledonian upper greenschist-lower amphibolite facies metamorphism (see below) overprints and obscures mineral assemblages representative of earlier metamorphic cycles. However, a minimum estimate of pre-Caledonian p-T conditions from relict plagioclase + clinopyroxene + garnet assemblages points to an event reaching at least upper amphibolite facies grade. Evidence of partial melting is found in rocks both inside (garnet amphibolite lens) and west (gabbroic and granitic gneiss) of the shear zone.

The N-S striking shear zone displays the highest strain immediately west of the garnet amphibolite lens (Fig. 2), where it exhibits a sharp contact to the garnet amphibolite. The assemblage epidote + green hornblende + chlorite indicates upper greenschist- to amphibolite facies conditions, typical of the Caledonian in the Jotun Nappe Complex, suggesting either a Caledonian age of the shear zone, or Caledonian reactivation and/or overprinting of an older shear zone.

The Berdalsbandet Pegmatite Swarm

The mineralised pegmatites that make up the BPS occur in an area of approximately 1 km², and are mainly concentrated in a garnet amphibolite lens enclosed by steeply dipping, N-S striking mylonitic shear zones (Fig. 2). The medium-grained, olivine-normative amphibolite mainly contains faintly zoned to unzoned, tschermakitic to Mg-hastingsitic hornblende, equal amounts of plagioclase (An₈₈) and garnet (Prp₄₄Alm₅₅Sps₀₁) as well as clinopyroxene (En₃₈Fs₁₅Wo₄₇) >> orthopyroxene and ore (representative analyses in Table 2). Reactions between the BPS granitic melt and the amphibolite can locally be seen as cm-thick metasomatic biotite rims developed in the country rock along the contacts with the pegmatites (Fig. 3). The mean width of the felsic pegmatites is ca. 50 cm, the maximum width ca. 2 m, and they can typically be traced for some tens of metres

in the strike direction. The BPS forms a vein-like, close-meshed network of steeply dipping dykes with a predominant strike direction of 300-320°. The total volume of the BPS is estimated to ≤5x10⁵ m³, assuming ≤50 dykes of tabular shape (100 x 100 x 1 m). A small number of pegmatites belonging to the BPS is also found in gabbros to the west of the garnet amphibolite lens. The mylonitic shear zones that enclose the garnet amphibolite body cross-cut all other structures in the field area.

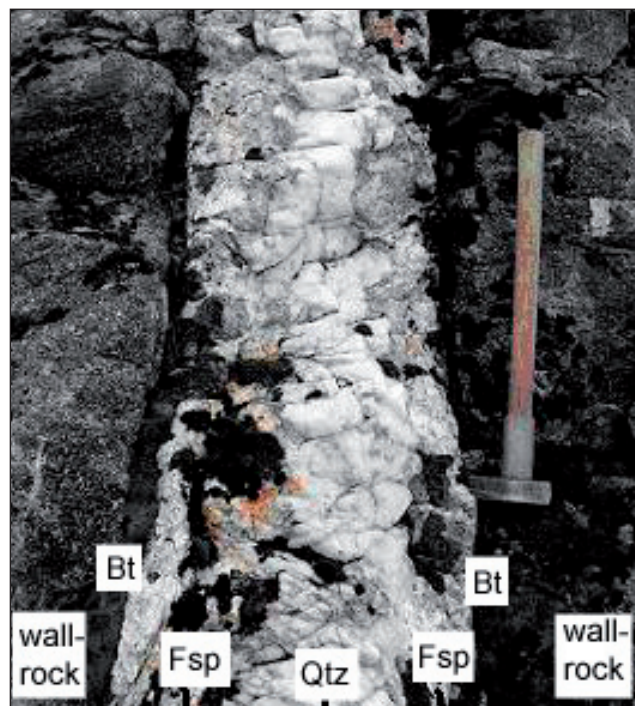


Fig. 3. Symmetrically zoned, ca. 0.5 m wide pegmatite in garnet amphibolite. Fsp: feldspar-rich zone, Qtz: quartz core. The biotite rim (Bt) towards the wall rock is covered with moss (black).

Table 2

wt%	hornblende		plagioclase		clinopyroxene		garnet
	am4 Ts	am7 MgHs	pl1 An ₉₂	pl5 An ₈₈	cpx1 En ₄₁ Wo ₄₈	cpx5 En ₃₈ Wo ₄₇	grt1 Prp ₄₄ Alm ₅₅
SiO ₂	43.19	42.13	44.54	45.89	51.30	50.45	39.63
TiO ₂	0.97	1.12	--	0.03	0.19	0.23	0.03
Al ₂ O ₃	13.51	14.30	35.68	34.90	3.63	4.78	22.33
Cr ₂ O ₃	0.06	0.03	--	--	--	0.02	0.03
FeO ⁱ	12.49	12.89	0.20	0.17	6.59	8.82	22.05
MnO	0.02	0.03	--	--	0.03	0.21	0.47
MgO	13.61	13.31	--	--	14.52	12.63	9.89
CaO	12.20	11.84	18.66	17.74	23.26	22.19	6.68
Na ₂ O	1.66	1.92	0.85	1.37	0.44	0.58	--
K ₂ O	0.83	0.86	0.02	0.03	--	--	--
Total	98.55	98.44	99.97	100.13	99.99	99.90	101.13
O	12.5	12.5	8	8	6	6	12
Si	6.30	6.17	2.06	2.11	1.90	1.89	2.98
Ti	0.11	0.12			0.01	0.01	
Al	2.32	2.47	1.94	1.89	0.16	0.21	1.98
Cr	0.01						
Fe ²⁺	1.52	1.58	0.01	0.01	0.20	0.28	1.39
Mn						0.01	0.03
Mg	2.96	2.91			0.80	0.70	1.11
Ca	1.91	1.86	0.92	0.87	0.92	0.89	0.54
Na	0.47	0.55	0.08	0.12	0.03	0.04	
K	0.15	0.16					
Sum	15.74	15.82	5.01	5.01	4.03	4.02	8.03

-- = below detection limit. Ts = tschermakite, MgHs = magnesiohastingsite

Table 2. Representative EMP analyses of major minerals from the garnet amphibolite hosting the BPS. Oxygen equivalents for formula calculation are indicated (O).

Mineral content of the BPS

Major minerals are quartz, perthitic orthoclase/microcline and oligoclase, minor constituents are allanite-(Ce), biotite and Fe-Ti-oxides. The pegmatites also contain muscovite, zircon, tourmaline (schorl), garnet (spessartine-dominated), thorite, phosphates and other accessory minerals, along with the first reported occurrence of beryl in the Jotun Nappe Complex. Other REE phases beside allanite-(Ce) are xenotime-(Y), monazite-(Ce), polycrase-(Y) and parisite-(Ce), the latter occurring as an alteration product of allanite (Table 1). However, the modal proportions of the minerals are highly variable since the pegmatites typically are made up of largely barren quartzofeldspathic parts with interspersed mineralised parts hosting most of the accessories.

Some of the pegmatites are symmetrically zoned along the main axis (Fig. 3). They show a progression in the mineralogy (Fig. 4) from feldspar + allanite + biotite at the margin to feldspar + quartz + tourmaline (schorl) + Fe-Ti-oxide and quartz + tourmaline + garnet + beryl in the interior parts. Graphic intergrowth of quartz and feldspar occurs locally in the larger dykes. The centres of zoned dykes consist of hydrothermal quartz, locally with a core of rose quartz. Zones are not continuous for more than a few metres, and their boundaries are undulating. Secondary, typically euhedral minerals occur in fissures

in the rock and as crusts covering the primary minerals described above. The secondary minerals include epidote, prehnite, chabazite, stilbite, heulandite-(Ca), heulandite-(Sr) (both analysed by Nordrum et al. 2005) and the Cu-minerals malachite, chalcopyrite along with unidentified mixtures of various phases, probably cuprite/tenorite.

Analytical methods

EMP

Mineral analyses were performed on a CAMECA SX100 electron microprobe at the Geochemical-Mineralogical Institute, University of Freiburg, Germany. The EMP is equipped with five WD spectrometers and one ED detector. Natural and synthetic standards were used for calibration, and the data were corrected for background using a mean atomic number procedure. Operating conditions for all measured elements, except REE, were 20 nA and 15 kV, counting time 20 s for K α -lines and 30 s for Th Ma-line. Sodium was analysed first to minimise loss of the element. The REE (including Y) La-lines (L β for Sm) were calibrated on phosphate standards and counted for 50 s with 100 nA and 25 kV. All elements detected in the ED spectra were analysed quantitatively. In total 21 elements were measured, including 10 REEs.

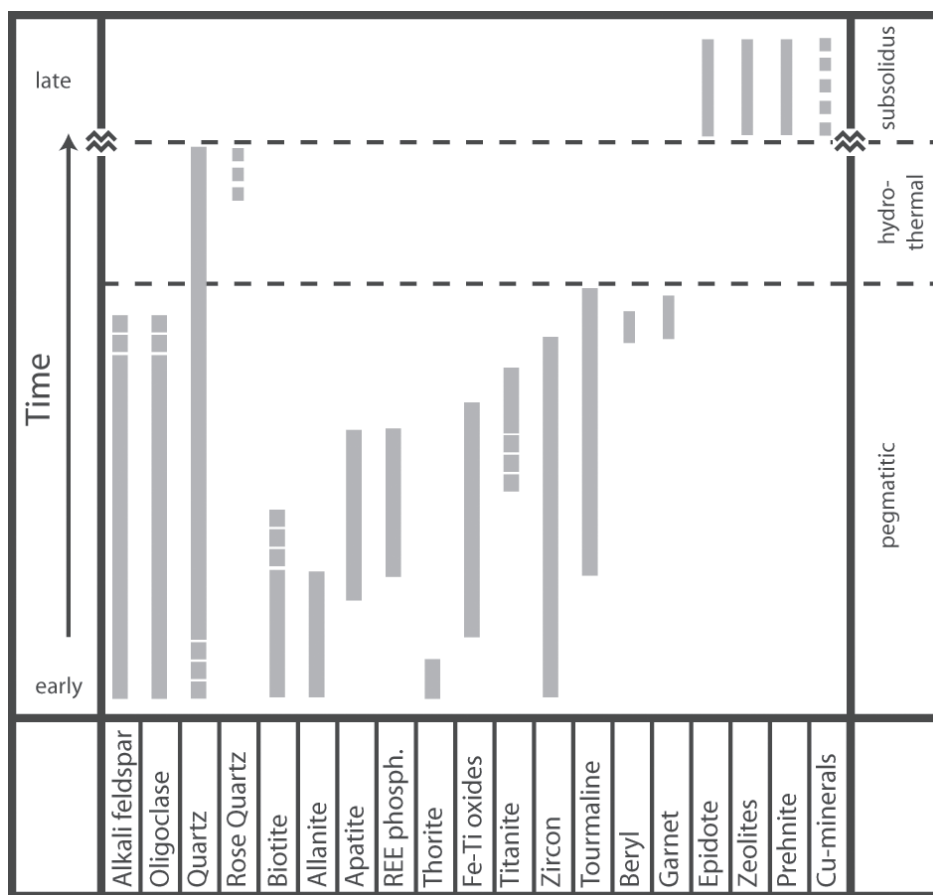


Fig. 4. Growth relations between some pegmatite minerals, as deduced from textures.

XRD

Crystalline materials (e.g., allanite and other pegmatite minerals) were identified using a BRUKER axs D8 Advance X-ray powder diffractometer (XRD) equipped with a ceramic X-ray tube (2.2 kW) and a Cu-K α -anode, combined with DIFFRACplus v5.0 software for evaluation of the data.

ID-TIMS

Uranium-lead ID-TIMS analyses were performed on fragments mechanically broken off from larger crystals. The fragments were air abraded (Krogh 1982), spiked with a $^{202}\text{Pb}/^{205}\text{Pb}/^{235}\text{U}$ tracer and dissolved in Savillex vials on a hot-plate. The U and Pb from each fragment were purified using one-stage HCl or HCl - HBr, or two stage HBr and HNO_3 ion exchange separation (see below), loaded together or separately on Re filaments and analysed on a MAT262 mass-spectrometer at Oslo University. The data were corrected using fractionation factors determined with the $^{202}\text{Pb}/^{205}\text{Pb}$ spike. Blank correction was 2 pg Pb and 0.1 pg U. The results were plotted and regressed using the Isoplot software (v. 3.00) of Ludwig (2003). The decay constants are those of Jaffey et al. (1971). For other details of the analytical procedure see Lundmark and Corfu (2007) and references therein. Uncertainties in the isotopic ratios and the ages are given and plotted at 2σ .

Allanite-(Ce) in the BPS

Allanite, $\text{CaREEAl}_2\text{Fe}^{2+}(\text{SiO}_4)_3\text{OH}$, is the most abundant rare-earth element (REE)-bearing mineral of the epidote group (Armbruster et al. 2006). It is a common accessory in many types of evolved igneous rocks, as well as in a variety of metamorphic rocks of both igneous and sedimentary origin. Magmatic allanite shows a strong preference for light rare-earth elements (LREE) over heavy rare-earth elements (HREE), while the REE distribution in metamorphic allanite is highly variable and ultimately depends on the specific REE activity in the precursor rock (Gieré & Sorensen 2004 and references therein). Due to the presence of minor amounts of radioactive elements, typically Th and U, (partial) metamictisation of allanite is a common phenomenon. This results in an increase of unit cell volume and cracking of neighbouring minerals, and diminished sharpness of diffraction patterns (Ueda 1957).

The Ce-dominated variety of allanite, allanite-(Ce), is locally the major constituent of the pegmatites in the BPS after quartz and feldspars. In some mineralised pockets it makes up >50 vol% of the rock, and its frequency in less-mineralised, non-barren areas of the rock is gauged to approximately 1 vol%. The allanite-(Ce) typically occurs as euhedral tabular or thin-tabular crystals of 1 to 3 cm, but can be up to 15 x 8 x 3 cm large. Usually the allanite-(Ce) is associated with biotite, which grows

as flakes on the crystal faces or in the interstices between allanite grains. Other REE-rich minerals typically found with the allanite are parisite-(Ce), monazite-(Ce), polycrase-(Y), thorite and zircon. Common fracturing of the surrounding minerals suggests an advanced state of metamictisation, but the samples are all optically anisotropic and XRD peaks are detectable (though diffuse), which would not be expected from completely non-crystalline materials (Ueda 1957).

Mineral chemistry of allanite-(Ce)

Analytical results from EMP measurements of allanite were recalculated on the basis of 8 cations per formula unit (apfu). Iron was measured as FeO^{tot} , and the $\text{Fe}^{2+}/\text{Fe}^{3+}$ ratio was calculated by balancing 25 negative charges. All other components were assumed to occur in one oxidation state only.

The monoclinic structure of allanite can absorb many elements by different substitution mechanisms and allocates them to $^{\text{IX}}\text{A}(1)$, $^{\text{XI}}\text{A}(2)$, $^{\text{VI}}\text{M}(1-3)$ and $^{\text{IV}}\text{T}$ positions. Following the suggestions in the recent compilations of Gieré & Sorensen (2004) and Ercit (2002) the various ions in the allanite are distributed in the following way:

- The A(1) position is fully occupied by Ca^{2+}
- A(2) contains a broad assortment of larger ions: all REE^{3+} , remaining Ca^{2+} (if present), Th^{4+} , Mn^{2+} , Na^{+} and K^{+}
- M(1) is filled with Al^{3+} (excess from M(2)), Ti^{4+} and Mg^{2+}
- M(2) contains Al^{3+}
- M(3) is the position of ferrous and ferric iron
- The T position holds Si^{4+} , P^{5+} ; excess Al^{3+} , which can be located in T, was not found in this study.

The composition of allanite-(Ce) from the Berdalsbandet pegmatites (Table 3 and 4) meets the criteria for belonging to the allanite subgroup of the epidote group (Armbruster et al. 2006). The EMP measurements also reveal internal compositional variations distributed in a patchwork-like pattern within the grains, also visible by microscopical investigation due to different optical properties of these domains (Fig. 5a).

Table 3

	RB 1		RB 6		RB 15	
(wt%)	min.	max.	min.	max.	min.	max.
SiO_2	31.20	32.31	31.17	33.76	31.02	32.83
TiO_2	0.90	1.29	0.32	0.62	0.51	0.69
ThO_2	0.40	0.62	0.40	0.70	0.36	1.07
Al_2O_3	14.20	15.10	15.64	16.93	14.91	16.35
Y_2O_3	0.03	0.06	0.06	0.17	0.04	0.10
La_2O_3	6.03	6.40	4.91	6.69	6.62	6.95
Ce_2O_3	11.58	12.47	9.02	12.01	11.61	14.11
Pr_2O_3	1.05	1.21	0.83	1.07	0.96	1.15
Nd_2O_3	1.42	1.76	0.83	1.52	1.05	1.24
Sm_2O_3	0.21	0.29	0.16	0.27	0.17	0.32
Eu_2O_3	--	0.13	0.06	0.12	0.07	0.13
Gd_2O_3	0.06	0.09	0.07	0.15	0.06	0.11
Dy_2O_3	--	--	--	--	--	0.02
Yb_2O_3	--	--	--	0.02	--	--
FeO^{t}	11.34	13.45	10.13	13.37	9.66	12.90
MnO	0.29	1.05	0.25	0.53	0.23	0.32
MgO	0.99	1.83	1.04	2.09	1.08	2.19
CaO	7.99	11.32	10.47	14.13	7.20	11.08
Na_2O	0.05	0.18	0.03	0.11	0.06	0.12
K_2O	--	0.07	--	0.03	--	0.09
P_2O_5	--	0.46	--	0.02	--	0.05
Total	91.57	96.24	90.20	97.31	91.06	96.96

-- = below detection limit.

Table 3. Maximum and minimum values (wt%) for allanite-(Ce) analyses from samples RB 1, RB 6 and RB 15.

Analytical totals of EMP measurements are relatively low (< 97.2 wt%). Since no other significant components have been found in XRF or in EMP spectra, this deficit may reflect additional incorporation of OH^- groups or H_2O molecules into the semi-metamict structure (Poitrasson 2002).

Representative analyses from the allanite samples provide an average calculated formula composition of $(\text{Ca}_{1.11}\text{RE}_{0.72}\text{Mn}_{0.03}\text{Th}_{0.01}\text{Na}_{0.01})_{\Sigma=1.89}(\text{Al}_{1.76}\text{Fe}^{2+}_{0.54}\text{Fe}^{3+}_{0.43}\text{Mg}_{0.24}\text{Ti}_{0.05})_{\Sigma=3.02}\text{Si}_{3.08}\text{O}_{12}(\text{OH})$. As the A(1) site only contains Ca, the dominant cations on A(2) are the REEs and Ca which fill approximately 70-90 % of A(2) space (Fig. 6a). Mn and

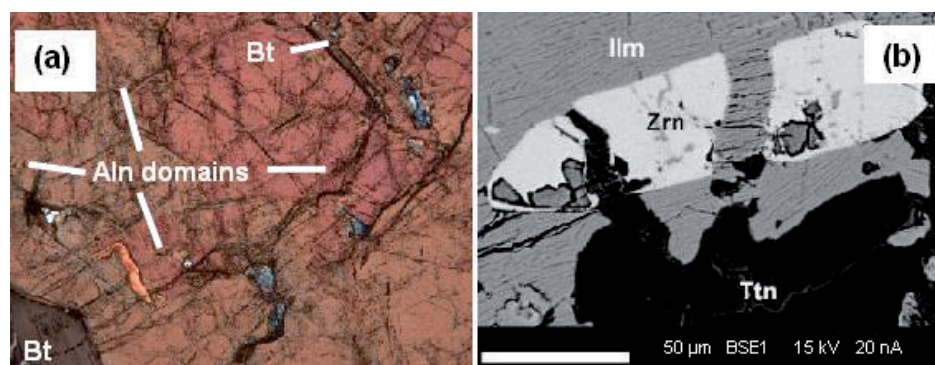


Fig. 5. Textures of pegmatite minerals, optical and backscatter electron (BSE) images. (a) Allanite-(Ce) crystal (Aln) with patchy domains (brown and reddish-brown) and biotite (Bt) inclusions. Image width 3.7 mm. (b) Fractured zircon (Zrn) in pegmatite, healed by large, pegmatitic crystals of titanite (Ttn) and ilmenite (Ilm).

Table 4																																
wt%	RB 1					RB 6										RB 15										→						
	←	2-R	3-C	4-C	5-R	→	←	6-R	7-B	8-D	9-D	10-B	11-B	12-B	13-B	14-D	15-D	16-B	→	←	17	18	19	20	21			22	23	24	25	26
SiO ₂	32.13	32.31	31.51	31.60	31.20	32.55	31.76	31.54	33.32	32.93	33.59	32.93	32.93	31.17	33.76	32.88	32.72	32.35	32.78	32.69	31.37	32.83	32.57	31.02	32.22	32.57	32.57	32.57	32.57	32.57	32.54	32.52
TiO ₂	1.04	0.90	0.92	1.29	1.22	0.62	0.35	0.33	0.50	0.56	0.50	0.45	0.32	0.35	0.60	0.64	0.58	0.57	0.54	0.62	0.69	0.60	0.56	0.54	0.51	0.60	0.60	0.65	0.60	0.60	0.65	
ThO ₂	0.52	0.40	0.46	0.53	0.62	0.70	0.50	0.46	0.59	0.48	0.58	0.65	0.57	0.40	0.61	0.47	0.55	0.54	0.56	0.56	1.07	0.55	0.36	0.38	0.65	0.57	0.62	0.60	0.61	0.62	0.61	
Al ₂ O ₃	14.95	14.82	14.20	15.08	15.10	15.64	16.44	16.53	16.27	15.67	15.90	15.75	15.70	16.93	15.64	15.67	15.66	16.35	15.85	15.38	16.19	15.78	14.91	16.16	15.66	15.88	15.72	15.72	15.72	15.72	15.72	
Y ₂ O ₃	0.06	0.03	0.03	0.05	0.04	0.06	0.07	0.08	0.11	0.17	0.11	0.07	0.09	0.09	0.06	0.07	0.07	0.08	0.09	0.10	0.07	0.04	0.04	0.06	0.07	0.07	0.05	0.05	0.05	0.05	0.05	
La ₂ O ₃	6.28	6.34	6.40	6.12	6.03	6.69	5.06	4.91	5.91	6.26	5.86	6.61	6.13	5.00	6.62	6.87	6.82	6.91	6.83	6.94	6.87	6.95	6.81	6.62	6.87	6.87	6.86	6.86	6.86	6.87	6.86	
Ce ₂ O ₃	12.47	11.66	11.58	12.45	12.33	11.92	9.32	9.07	10.82	11.44	11.02	11.81	11.16	9.02	12.01	11.88	11.88	11.90	11.84	14.11	11.86	11.83	11.61	11.73	11.86	11.83	11.82	11.82	11.82	11.82	11.82	
Pr ₂ O ₃	1.21	1.12	1.09	1.08	1.05	1.07	0.83	0.83	0.94	1.03	0.95	1.05	1.01	0.86	1.02	1.09	1.00	1.06	1.00	1.15	1.03	1.03	1.03	1.06	0.96	1.03	1.01	1.01	1.01	1.01	1.01	
Nd ₂ O ₃	1.75	1.76	1.66	1.59	1.42	1.19	0.83	0.86	1.17	1.52	1.28	1.17	1.19	1.02	1.29	1.14	1.13	1.17	1.16	1.24	1.14	1.15	1.05	1.07	1.14	1.10	1.21	1.21	1.21	1.21	1.21	
Sm ₂ O ₃	0.29	0.24	0.27	0.28	0.21	0.27	0.19	0.21	0.24	0.26	0.22	0.21	0.22	0.16	0.17	0.23	0.25	0.23	0.24	0.32	0.23	0.20	0.17	0.22	0.21	0.20	0.23	0.23	0.23	0.23	0.23	
Eu ₂ O ₃	0.08	0.14	0.13	--	--	0.09	0.08	0.08	0.07	0.06	0.09	0.08	0.08	0.10	0.12	0.07	0.09	0.10	0.08	0.13	0.10	0.12	0.08	0.13	0.12	0.12	0.12	0.12	0.12	0.12	0.12	
Gd ₂ O ₃	0.08	0.06	0.08	0.09	0.07	0.10	0.09	0.10	0.08	0.15	0.09	0.08	0.09	0.09	0.09	0.07	0.06	0.11	0.11	0.09	0.11	0.07	0.06	0.07	0.09	0.06	0.07	0.07	0.07	0.07	0.07	
Dy ₂ O ₃	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	0.02	--	--	--	--	--	--	--	--	--	--	--	
Yb ₂ O ₃	--	--	--	--	--	--	--	--	--	0.02	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	--	
FeO ¹	11.91	13.45	13.41	11.78	11.34	12.51	13.37	13.19	12.80	12.48	12.63	12.69	10.13	12.97	12.24	11.72	12.24	11.72	12.24	12.12	9.66	12.16	11.75	11.11	12.90	12.00	12.12	12.04	12.04	12.04	12.04	
MnO	0.77	0.29	0.32	1.02	1.05	0.33	0.25	0.32	0.47	0.53	0.36	0.41	0.30	0.41	0.37	0.28	0.26	0.31	0.28	0.32	0.32	0.24	0.25	0.23	0.23	0.24	0.32	0.26	0.26	0.26	0.26	
MgO	1.09	1.57	1.83	1.06	0.99	2.09	1.16	1.16	1.63	1.84	1.49	1.88	1.50	1.04	1.98	2.16	2.15	1.84	2.13	1.08	1.84	2.10	2.00	1.47	2.14	2.16	2.19	2.19	2.19	2.19	2.19	
CaO	8.68	11.09	11.32	7.99	8.19	10.90	13.33	13.66	12.28	11.36	11.76	11.22	10.47	14.13	11.13	10.93	10.99	10.89	11.00	7.20	11.08	10.97	10.21	11.05	10.83	10.78	10.80	10.80	10.80	10.80	10.80	
Na ₂ O	0.14	0.05	0.07	0.16	0.18	0.10	0.03	0.05	0.07	0.07	0.04	0.11	0.05	0.05	0.11	0.07	0.06	0.09	0.08	0.06	0.06	0.06	0.06	0.12	0.08	0.06	0.08	0.08	0.08	0.08	0.08	
K ₂ O	0.02	0.02	--	--	0.07	--	--	--	0.03	--	--	--	--	--	0.02	--	--	--	--	--	0.09	--	--	--	--	--	--	--	--	--	--	--
P ₂ O ₅	0.08	--	--	0.09	0.46	--	--	--	0.02	--	--	--	--	--	0.02	--	0.04	--	0.02	0.02	0.05	0.05	0.03	0.04	--	--	0.05	--	--	--	--	
Total	93.54	96.24	95.27	92.29	91.57	96.84	93.65	93.38	97.31	96.84	96.46	97.17	90.20	96.39	96.92	96.13	96.18	96.67	96.69	91.06	96.96	95.83	91.34	96.31	95.99	96.42	96.28	96.28	96.28	96.28	96.28	
CaA(1)	Ca	0.92	1.00	1.00	0.86	0.89	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	0.80	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	
CaA(2)	Ca	0.13	0.16	--	--	0.10	0.34	0.38	0.21	0.14	0.17	0.12	0.13	0.37	0.12	0.11	0.11	0.10	0.11	0.09	0.12	0.11	0.09	0.12	0.10	0.09	0.09	0.09	0.09	0.09	0.09	
REE	0.81	0.74	0.75	0.80	0.78	0.74	0.57	0.56	0.65	0.72	0.67	0.72	0.74	0.54	0.74	0.74	0.74	0.74	0.73	0.92	0.73	0.74	0.76	0.73	0.74	0.73	0.74	0.74	0.74	0.74	0.74	
Th	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.03	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	
Mn	0.06	0.02	0.03	0.09	0.09	0.03	0.02	0.03	0.04	0.04	0.03	0.03	0.03	0.03	0.03	0.02	0.02	0.02	0.02	0.02	0.03	0.02	0.02	0.02	0.02	0.02	0.03	0.02	0.02	0.02	0.02	
Na	0.03	0.01	0.01	0.03	0.04	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	
K	--	--	--	--	--	0.01	--	--	--	--	--	--	--	--	--	--	--	--	--	0.01	--	--	--	--	--	--	--	--	--	--	--	
Σ _A	1.83	1.91	1.96	1.79	1.82	1.90	1.95	1.99	1.92	1.92	1.89	1.90	1.92	1.96	1.92	1.89	1.89	1.89	1.88	1.80	1.88	1.89	1.89	1.90	1.88	1.87	1.87	1.87	1.87	1.87	1.87	
M(3)	Fe ²⁺	0.99	0.74	0.63	0.99	0.96	0.66	0.48	0.42	0.61	0.67	0.76	0.66	0.77	0.53	0.67	0.69	0.65	0.72	0.84	0.73	0.68	0.72	0.70	0.69	0.67	0.67	0.67	0.67	0.67	0.67	
Fe ³⁺	--	0.33	0.45	--	--	0.32	0.57	0.62	0.37	0.31	0.23	0.33	0.33	0.08	0.45	0.29	0.24	0.32	0.20	0.28	--	0.23	0.25	0.21	0.32	0.26	0.27	0.28	0.28	0.28	0.28	
M(2)	Al	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	
M(1)	Al	0.75	0.66	0.61	0.79	0.80	0.73	0.82	0.83	0.76	0.71	0.75	0.73	0.86	0.81	0.73	0.74	0.74	0.81	0.75	0.88	0.79	0.76	0.75	0.80	0.75	0.76	0.75	0.75			

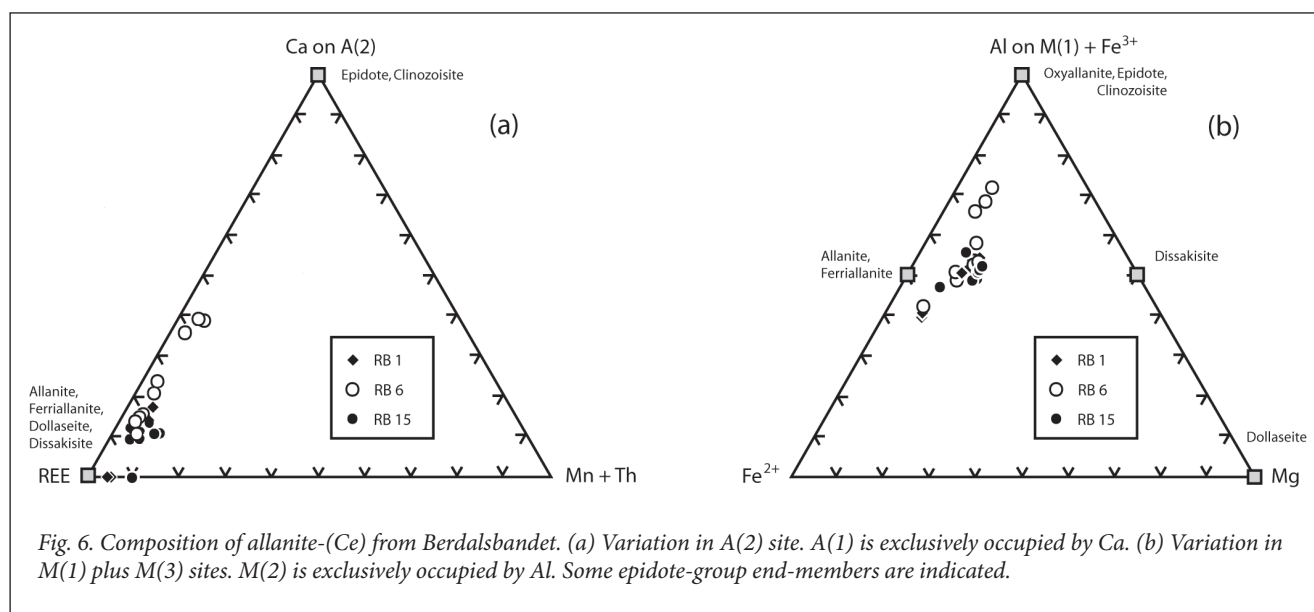


Fig. 6. Composition of allanite-(Ce) from Berdalsbandet. (a) Variation in A(2) site. A(1) is exclusively occupied by Ca. (b) Variation in M(1) plus M(3) sites. M(2) is exclusively occupied by Al. Some epidote-group end-members are indicated.

Th are present to a lesser degree. Data points cluster on the REE-rich end of the Ca-REE connecting line of the A(2)-site diagram, which is expected in pure allanite. Fig. 6b shows the variability on the M(1)- and M(3)-sites, whereas M(2) is fully occupied by Al (not plotted in the diagram). A large spread in distribution is found between Fe²⁺ and trivalent cations, whereas incorporation of the Mg-component (dollaseite) is homogeneous with an average of ca. 10–15 %. Again, most data points are close to the allanite end-member composition with equal amounts of Fe²⁺ and Al³⁺ in M(3) and M(1), respectively.

The most significant compositional difference between the allanite-(Ce) samples is found in the $Fe_{ox} = Fe^{3+}/(Fe^{3+} + Fe^{2+})$ value. Fe_{ox} is elevated in sample RB 6 (0.36) compared to RB 1 (0.14) and RB 15 (0.25). In general, RB 6 shows the widest compositional range of all analysed allanite-(Ce) samples. RB 6 data points are nearly in the same relative order in Fig. 6a and b: measurements resulting in high Ca values on A(2) also show high Al+Fe³⁺, whereas REE-rich ones yield high Fe²⁺.

REE distribution in BPS allanite-(Ce)

The structure of allanite is strongly selective for LREE. The Berdalsbandet allanite-(Ce) has a total REE-content between 0.56 and 0.92 apfu (average 0.72 apfu). Ce (average 0.40 apfu) is dominant, followed by La (0.22 apfu) and the other LREE. The HREE are commonly below the detection limit of the EMP. C1-chondrite normalised REE patterns in Fig. 7 (C1 data from Evensen et al. 1978) show a distinct enrichment of LREE by 10⁵ with a rapid decrease with higher atomic number (sample/C1 ≈ 10³). This is comparable to data reported from other igneous allanite (Gromet & Silver 1983; Peterson & MacFarlane 1993; Wood & Ricketts 2000; Gieré & Sorensen 2004).

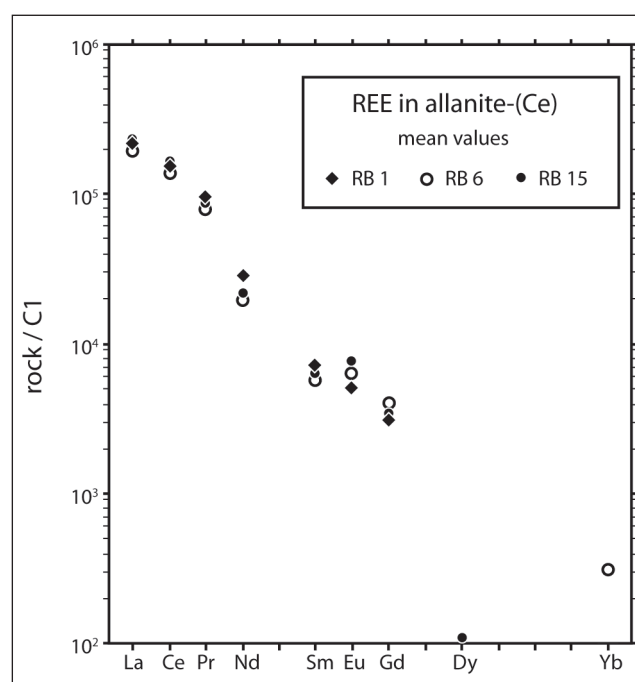
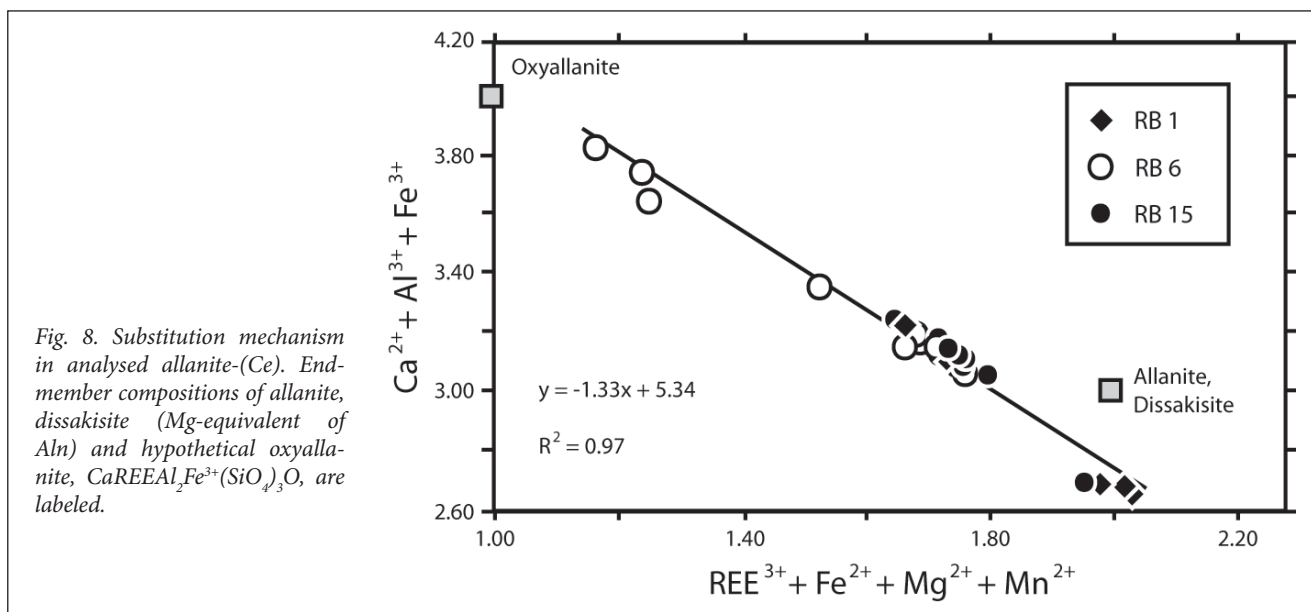


Fig. 7. C1-normalised REE pattern of allanite-(Ce) from the Berdalsbandet pegmatites. C1 data from Evensen et al. 1978.

Geochronology

The allanites are texturally part of the pegmatite mineral assemblage, and their crystallisation age is thus constrained by a 950 ± 1 Ma zircon ID-TIMS age (Lundmark et al. 2007). To test the resilience of the allanites to resetting, and potentially to assign an age to the annealing of metamict domains revealed by BSE imaging (see below), eight analyses were obtained from fragments of two cm-sized allanite crystals, RB15 and SS97 (Table 5). The analyses display a generally low analytical precision that mainly reflects the uncertainty



introduced by the correction for initial Pb, the result of both a relatively low radiogenic Pb content due to U concentrations of only 18 to 38 ppm, and a high initial Pb content of 6 to 16 ppm (Table 5). Additionally, incomplete purification using one-stage HCl (5, 8; Table 5), or one stage HCl-HBr (1; Table 5) separation led to weak ion beams. The remaining analyses were obtained using two-stage HBr and HNO_3 separation, and the purified U and Pb were loaded on separate Re filaments which improved the precision of the measurements. Correction for initial Pb using the Stacey and Kramers (1975) two-stage Pb evolution model yields reversely discordant analyses (but still overlapping the Concordia curve). Initial common Pb with lower $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ ratios than predicted by the Stacey and Kramers (1975) model could result from long-standing depletion of U following high-grade metamorphism

of the precursor (-s) to the pegmatites. Use of a more primitive initial Pb correction corresponding to a Depleted Mantle composition (Neymark 1990) shifts the data more firmly onto Concordia, but also towards ages that are distinctly older than the 950 ± 1 Ma age of the pegmatite. Given the large, pegmatitic size of the allanite crystals it seems unlikely that the data reflect inheritance. Also, no reverse discordance indicative of a more primitive initial Pb has been observed in titanite data from other samples in the Upper Jotun Nappe (Lundmark et al. 2007; Lundmark and Corfu 2008a). Therefore, a more likely reason for the reverse discordance is the presence of excess ^{206}Pb , an effect commonly observed in minerals with high Th/U ratios such as monazite and allanite (Schärer 1984). Incorporation of excess ^{230}Th , an intermediate daughter nuclide in the ^{238}U decay series, during magmatic

Table 5

Nr (grain)	Weight ^a	U ^a	Th/U ^b	Pbc ^{a,c}	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$ ^d	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$ ^d	2 sigma	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$ ^e	2 sigma	rho ^f	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$ ^e	2 sigma	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$ ^e	2 sigma	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$ ^e	2 sigma	Disc
	[ug]	[ppm]		[ppm]			[abs]		[abs]			[abs]	[Ma]	[abs]	[Ma]	[abs]	[%]
1 (SS97)	80	31.2	138.8	15.53	37.2	1.54	0.19	0.1643	0.0031	0.06	0.0682	0.0083	981	17	948	72	-13
2 (SS97)	13	25.9	168.6	7.70	49.5	1.45	0.11	0.1603	0.0017	0.00	0.0655	0.0049	958	10	908	44	-23
3 (SS97)	15	38.0	148.9	12.28	46.7	1.46	0.13	0.1583	0.0019	0.11	0.0668	0.0059	948	11	913	53	-15
4 (SS97)	29	34.7	163.0	11.22	46.8	1.45	0.11	0.1583	0.0018	0.00	0.0663	0.0053	947	10	909	47	-17
5 (RB15)	57	18.1	255.8	5.81	46.8	1.39	0.55	0.1572	0.0095	0.74	0.064	0.023	941	53	887	210	-27
6 (SS97)	17	33.2	161.3	8.62	53.1	1.421	0.093	0.1549	0.0015	0.00	0.0665	0.0044	928.6	8.8	898	39-14	
7 (SS97)	24	31.4	163.7	7.26	57.1	1.407	0.083	0.1532	0.0013	0.00	0.0666	0.0040	918.7	7.8	892	35	-12
8 (SS97)	2	37.4	169.5	11.92	44.2	1.36	0.12	0.1523	0.0019	0.02	0.0649	0.0057	914	11	873	50	-20

^a) weight and concentrations are known to about 10% for samples larger than 5 ug and not better than 50% for smaller fractions

^b) Th/U model ratio inferred from 208/206 ratio and age of sample

^c) Pbc = initial common Pb in sample (corrected for spike, fractionation and blank)

^d) raw data corrected for fractionation

^e) corrected for fractionation, spike, blank and initial common Pb; error calculated by propagating the main sources of uncertainty

^f) the low error correlations reflect the large proportion of common Pb

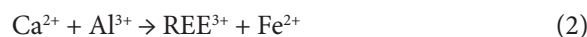
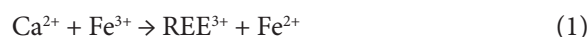
crystallisation leads to formation of ^{206}Pb that is not related to the ^{238}U content of the mineral, resulting in too high $^{206}\text{Pb}/^{238}\text{U}$ ratios and hence reverse discordance. The data may be corrected if the Th/U ratio of the host rock is known, and assuming that it mirrors the Th/U ratio of the melt from which the allanites grew (Schärer, 1984). In our case, excess ^{206}Pb is supported by the very high Th/U ratios of 139–256 (Table 1). Unfortunately it is difficult to correct the data for initial disequilibrium due to the mineralogical inhomogeneity of the pegmatite. Even so, the presence of excess ^{206}Pb testifies to a Pb isotopic system that has undergone but very minor post-magmatic resetting (e.g., von Blanckenburg 1992).

Excess ^{206}Pb does not affect the $^{207}\text{Pb}/^{235}\text{U}$ ages. Corrected with crustal Pb (Stacey and Kramers 1975) they range from 948 to 873 Ma, overlapping with or slightly younger than the 950 ± 1 Ma zircon age. This is most easily explained by minor post-magmatic Pb loss, which fits with observations supporting recrystallisation of partly metamict allanite (see below). The analytical uncertainties in the data do not allow us to completely resolve the contributions from the initial Pb composition, excess ^{206}Pb , and subsequent Pb loss. Nevertheless, the data firmly support the first order conclusion that the allanite fragments largely retain magmatic Pb isotopic signatures, with only marginal resetting of the isotopic systems during subsequent events.

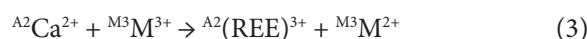
Discussion

Exchange mechanisms in allanite-(Ce)

Ideal allanite is connected to epidote and to clinoziosite by REE-incorporating coupled substitutions (1) and (2), respectively (Gieré & Sorensen 2004):



To maintain charge balance in natural allanite, Ca, Al and Fe^{3+} may be replaced by REE^{3+} and by various divalent M- and A-site species. The universal substitution is presented in Peterson & MacFarlane (1993) and Armbruster et al. (2006):



with Al^{3+} and Fe^{3+} for M^{3+} and Fe^{2+} , Mg^{2+} and Mn^{2+} for M^{2+} . Substitution (3) is prominent in the analysed allanite-(Ce). The linear relationship for the BPS allanite is apparent in Fig. 9.

Another mechanism of charge balancing, described by Peterson & MacFarlane (1993) and shown in Fig. 10, is restricted to the main A-site cations and leaves one position unfilled:



Allanite-(Ce) and bulk rock REE in the BPS

Allanite-(Ce) is the major REE-bearing phase in the BPS, dominating the LREE spectrum of the rock. Monazite-(Ce), the second most abundant LREE-rich mineral, is far less common than allanite-(Ce). It shows C1-normalised LREE patterns (Cresbron 1989; Möller 1989) that generally follow the same trend as in allanite-(Ce). The LREE concentrations in major minerals in the pegmatites beside allanite-(Ce) (quartz, K-feldspar, plagioclase) are negligible (Fig. 11) (Rollinson 1993). The HREE spectrum is likely controlled by other minerals than allanite-(Ce), such as zircon (Brooks et al. 1981)

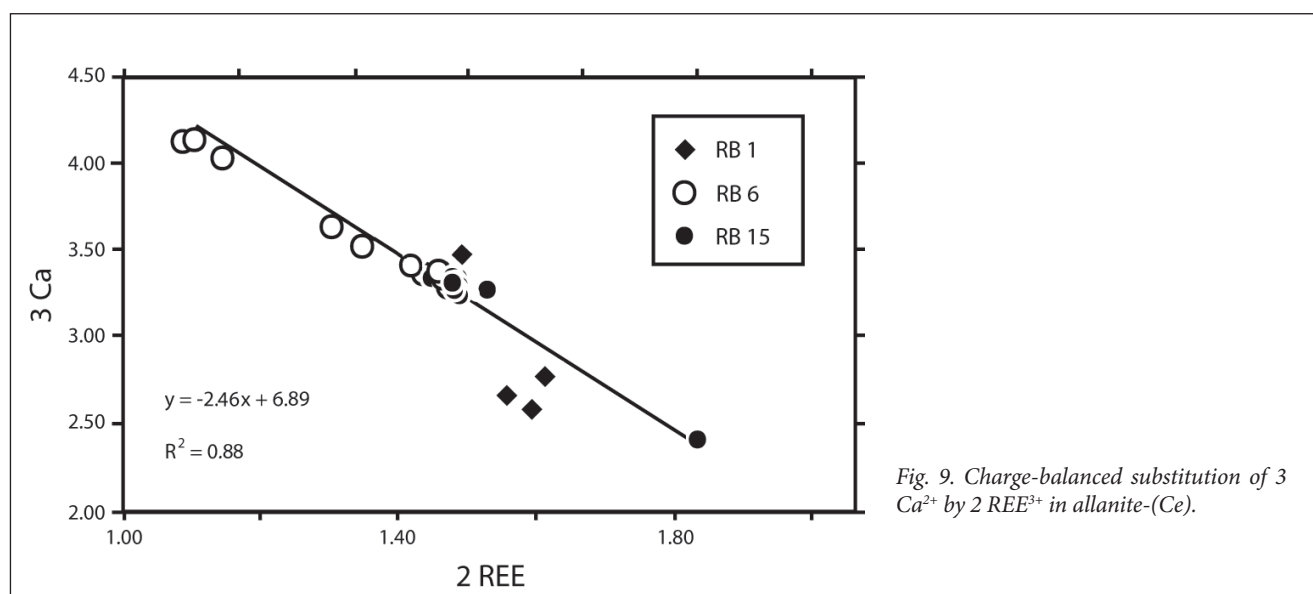


Fig. 9. Charge-balanced substitution of 3 Ca^{2+} by 2 REE^{3+} in allanite-(Ce).

and xenotime-(Y). Therefore, it can be assumed that the LREE (La-Sm) content of allanite-(Ce) represents a good approximation of the bulk rock LREE content. The LREE content for mineralised areas in the pegmatites can then be estimated by the linear function

$$\text{LREE}_{\text{N,rock}} = \text{LREE}_{\text{N,Aln}} X_{\text{Aln,rock}} \quad (5)$$

with $\text{LREE}_{\text{N,rock}}$ representing the C1-normalised LREE content in the bulk rock, $\text{LREE}_{\text{N,Aln}}$ the normalised LREE content in analysed allanite-(Ce), and $X_{\text{Aln,rock}}$ the modal abundance of allanite-(Ce) in the rock, in this case estimated to 1% in mineralised parts (given the precision of this estimate vol% is assumed equal to wt%). It is evident that the shape of a C1-normalised curve is identical to that of allanite-(Ce) (Fig. 7) diluted to 1%. Compared to the chondritic value, and depending on the chosen constraints, the mineralised pegmatite batches are enriched in La by a factor of 2000 and in Sm by 100.

Recrystallisation of allanite-(Ce) in the BPS

Magmatic zoning in allanite is typically characterised by straight compositional boundaries parallel to the crystal faces, and oscillatory zoning, whereas irregular zoning with different brightness in backscatter electron (BSE) imaging is indicative of secondary alteration (e.g., Petrák et al. 1995; Poitrasson 2002; Hoshino et al. 2006). The patchwork-like, irregular compositional variations observed in this study are suggestive of late- or post-

magmatic fluid flow, known to cause alteration along grain-internal cracks (Petrák et al. 1995). This typically produces sectors with elevated Fe_{ox} , while Fe_{ox} in original igneous allanite is buffered by biotite and remains close to 0.4 (Petrák et al. 1995). However, the analysed samples are hardly oxidised, in fact, they are mildly reduced compared to the biotite buffer (Table 3). Only a few domains in RB 6 show increased Fe_{ox} -values of 0.57 and 0.60 (Table 4). Also, there is typically no direct correlation between cracks in the grains and compositional domains; in fact, most compositional domains are independent of the density and orientation of cracks (Fig. 5a). Alternatively, allanite may re-equilibrate and its components may be redistributed into domains of more stable composition during post-magmatic p-T-changes. Minor recrystallisation of metamict allanite is reported to start at temperatures as low as 200–300°C, with major annealing occurring in two stages at 500–600°C (greenschist-amphibolite facies transition) and at 700–800°C (amphibolite-granulite facies transition), reflecting different annealing mechanisms (Janeczek & Eby 1993; Paulmann et al. 2000; Paulmann & Bismayer 2001).

The annealing of metamict domains observed as zoning in BSE and the minor Pb-loss indicated by the ID-TIMS data in this study are unlikely to reflect the late-Sveconorwegian magmatic / metamorphic events reported by Lundmark et al. (2007), since the allanite would not have had time to accumulate significant radiation damage at this time. Therefore, we attribute the observed textures to Caledonian annealing, at which time the Hurrungane rocks were exposed to upper greenschist-lower amphibolite facies conditions. The fact that the allanite fragments have largely retained their magmatic Pb isotopic signatures through Caledonian metamorphism when the allanites were already more than 500 m. y. old is remarkable, given the locally strong metamorphic overprint affecting the Upper Jotun Nappe (Lundmark et al. 2007; Lundmark and Corfu 2008a) and the tendency of allanite to become metamict (e.g., Poitrasson 2002 and references therein).

The origin and emplacement of the BPS

Plutonic activity in the Hurrungane area at ca. 1660 to 1630 Ma is distinctly older than the 950 ± 1 Ma pegmatites (Lundmark et al. 2007), indicating that large-scale plutonism is not the source of the evolved, REE-enriched melt manifested in the pegmatites. The presence of 954 ± 3 Ma syntectonic anatexite in the Hurrungane area and ca. 1600 Ma strongly discordant inherited zircons in the BPS led Lundmark et al. (2007) to suggest that the pegmatites are related to partial melting in the country rock. When classified according to the scheme of Černý & Ercit (2005) and Černý (1990, 1992), the Berdalsbandet pegmatites belong to the LREE-enriched abyssal class (AB-LREE). These are known to form under upper amphibolite- to granulite-facies conditions in anatectic terranes and typically contain

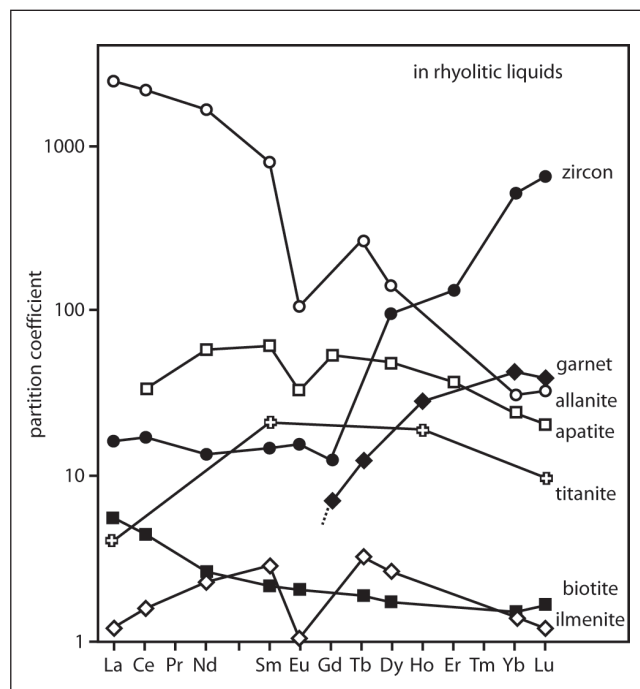


Fig. 10. Partition coefficients (D) for REE between rhyolitic melt and some of the minerals occurring in the Berdalsbandet pegmatites. Data from Rollinson (1993). LREE partition chiefly in allanite, HREE in zircon.

minerals such as allanite, monazite and thorite. The high REE content of the pegmatites (modelled from allanite-(Ce)) indicates either a low degree of partial melting, concentrating incompatible elements from a large volume of country rock into a small volume of melt, or fractionation processes in the anatectic melt batches; the small volumes of anatexite observed in the country rock in the extended field area supports the first alternative.

The generally subparallel pegmatites (Figure 12b) are oblique to the shear zones surrounding the garnet amphibolite lens (Figure 12a). The geometry is consistent with emplacement of the dykes normal to the minimum compressive stress (σ_3) in an active sinistral non-coaxial shear zone (Figure 12c). This is further supported by the concentration of the dykes in the sheared, lens-shaped garnet amphibolite. Furthermore, shear deformation during solidification of the dykes is indicated by fractured pegmatitic minerals (e.g., zircon, beryl) that were healed by later pegmatitic phases (titanite, quartz) (Fig. 5b), and by dynamic high-T recrystallisation of quartz and biotite near deformation zones. Therefore, we conclude that the BPS is syntectonic, and that the age of the pegmatites dates Sveconorwegian movements along the shear zones surrounding the garnet amphibolite lens. It follows that the Caledonian deformation observed in the mylonites constitutes reactivation of Sveconorwegian structures.

Field observations of a small number of BPS type pegmatites in medium-grained gabbros on the western

side of the amphibolite lens suggest a limited amount of Caledonian displacement along this part of the N-S stretching mylonitic shear zone, despite the pervasive Caledonian overprint.

Although zeolites may exist as primary phases in pegmatites (Gottardi 1989, London 2008), the p-T sensitive zeolites in the BPS cannot be older than the Caledonian greenschist-facies event. Their growth is likely related to late- to post-Caledonian retrogression from greenschist- to zeolite facies conditions. The copper mineralisation is clearly a subsolidus event, but its age remains uncertain.

Conclusions

Four stages of mineral growth are distinguished in the Berdalsbandet pegmatites: (1) an initial pegmatitic stage with mineral growth from the melt and formation of a metasomatic biotite rim by interaction of the pegmatitic melt with the wallrock; (2) a hydrothermal stage, leading to the formation of massive quartz cores during the final dyke solidification; (3) a stage of allanite-(Ce) recrystallisation (4) a subsolidus, late- or post-Caledonian stage, where local zeolitisation of the dykes took place. The age of a secondary copper mineralisation remains uncertain.

Allanite-(Ce) constitutes an important part of the BPS mineralogy. Its composition, expressed by representative analyses of grains from three samples, is: $(\text{Ca}_{0.86}\text{R}$

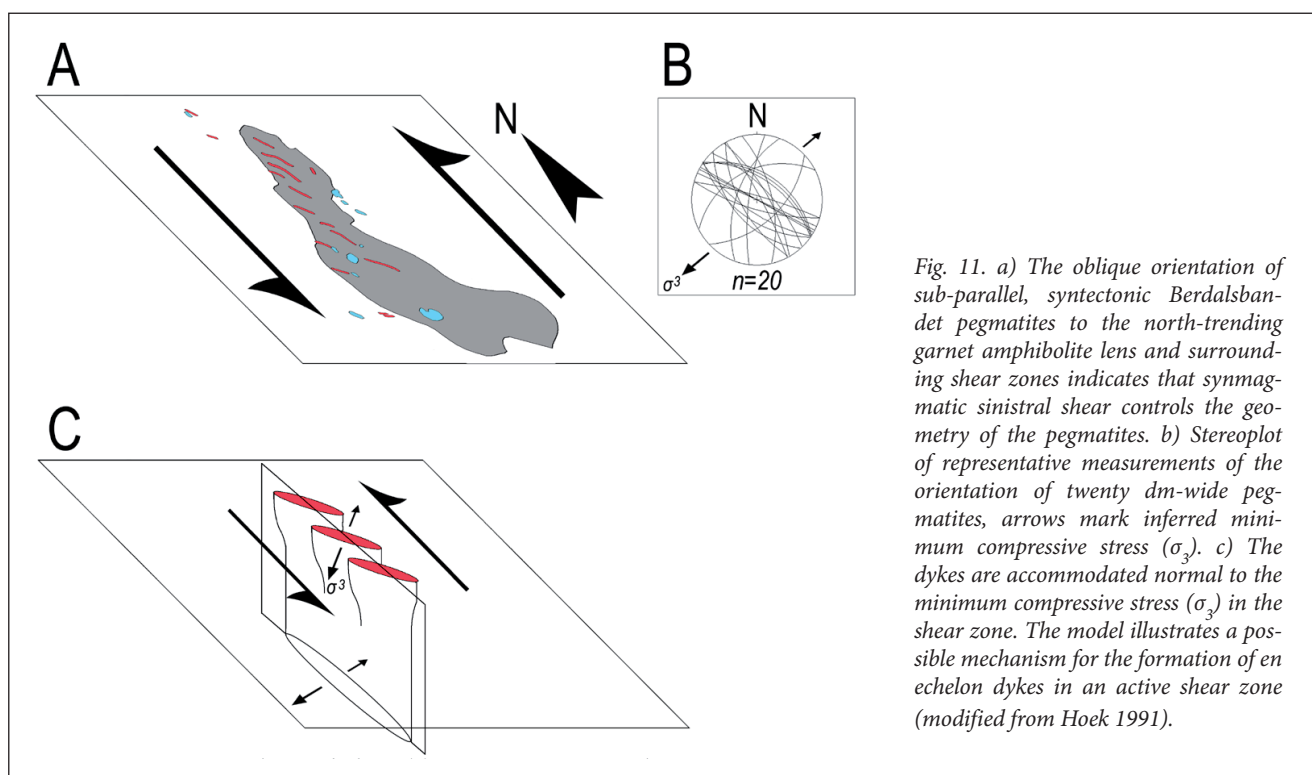


Fig. 11. a) The oblique orientation of sub-parallel, syntectonic Berdalsbandet pegmatites to the north-trending garnet amphibolite lens and surrounding shear zones indicates that synmagmatic sinistral shear controls the geometry of the pegmatites. b) Stereoplot of representative measurements of the orientation of twenty dm-wide pegmatites, arrows mark inferred minimum compressive stress (σ_3). c) The dykes are accommodated normal to the minimum compressive stress (σ_3) in the shear zone. The model illustrates a possible mechanism for the formation of en echelon dykes in an active shear zone (modified from Hoek 1991).

EE_{0.80}Mn_{0.09}Th_{0.01}Na_{0.03})_{Σ1.79}(Al_{1.79}Fe²⁺_{0.99}Mg_{0.16}Ti_{0.10})_{Σ3.04}(Si_{3.17}P_{0.01}O₁₂(OH), (Ca_{1.12}REE_{0.72}Mn_{0.03}Th_{0.01}Na_{0.02})_{Σ1.90}(Al_{1.73}Fe²⁺_{0.66}Fe³⁺_{0.33}Mg_{0.26}Ti_{0.03})_{Σ3.01}Si_{3.07}O₁₂(OH) and (Ca_{1.10}REE_{0.74}Mn_{0.02}Th_{0.01}Na_{0.01})_{Σ1.88}(Al_{1.75}Fe²⁺_{0.69}Fe³⁺_{0.26}Mg_{0.30}Ti_{0.04})_{Σ3.04}Si_{3.08}O₁₂(OH). The dominant substitution mechanisms are Ca²⁺ + M³⁺ = REE³⁺ + M²⁺ and 3Ca²⁺ = 2REE³⁺. A pattern of distinct LREE enrichment is comparable to data reported from other igneous allanites. Additionally, a series of accessory minerals are identified that contain High Field Strength Elements.

A range of ²⁰⁷Pb/²³⁵U ages from 948 to 873 Ma and the presence of excess ²⁰⁶Pb show that allanites in this study have largely retained their magmatic Pb isotopic signatures despite the locally strong Caledonian upper greenschist-lower amphibolite facies metamorphic overprint in the field area. A patchwork-like distribution of compositional variations in the allanite-(Ce) grains is thought to reflect Caledonian annealing of metamict allanite, associated with minor Pb loss.

The BPS belongs to the abyssal class of pegmatites (AB-LREE), supporting a genetic link to coeval, granitic anatexite present in the field area. A strong enrichment in LREE content of the BPS, inferred from allanite-(Ce) compositions and relative abundance in the BPS, indicates a low degree of partial melting consistent with field observations.

The geometric relationship of the BPS to its amphibolite host rock and to the surrounding shear zones, along with detailed observations of minerals in the BPS, lead us to conclude that the BPS was emplaced in an active sinistral, non-coaxial shear zone, and that the pegmatites date Sveconorwegian shearing to 950 ± 1 Ma.

It is further proposed that a strong Caledonian overprint in the area surrounding the amphibolite lens hosting the BPS constitutes reactivation of Sveconorwegian shear zones, though field observations point to a limited amount of displacement despite the locally pervasive metamorphic overprint.

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