The U–Pb age of the Finse batholith, a composite bimodal Sveconorwegian intrusion

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The Sveconorwegian autochthonous basement in the Finse area, South Norway, comprises a c. 900 km² bimodal batholith, dominated by K-feldspar megacrystic granite. The granite is locally intermingling with gabbroic rocks, and cut by different generations of pegmatite and aplite dykes. Zircon U–Pb geochronological data obtained by ID–TIMS demonstrate that granite, granodiorite and gabbro are coeval, as they define a common upper intercept age of 985.6 ± 1.6 Ma, dating emplacement of the batholith. Titanite is variously disturbed by later Sveconorwegian events. Pegmatites cross-cutting all structures yield zircon and titanite ages of 976 ± 8 Ma, 939 ± 2 , and possibly 958 ± 2 Ma, reflecting the latest stages of the evolution of the batholith. The Finse batholith is one of the earliest members of the widespread ferroan hornblende-biotite-granitoid (HBG) suite, in the western part of the Sveconorwegian orogen.

Keywords: batholith, bimodal magmatism, Sveconorwegian, U–Pb geochronology, zircon, titanite

Received 14. November 2015 / Accepted 13. April 2016 / Published online 3. May 2016

Introduction

The Precambrian basement of southwestern Norway formed between 1500 and 900 Ma, an evolution concluded by the widespread emplacement of intrusive rocks across the orogen (e.g., Bingen et al., 2008; Slagstad et al., 2013a). The latter are grouped into different suites with different ages, compositional characteristics, and geneeses, as elaborated below. To obtain a clear picture of the evolution of this crust, and the relationships to other terranes in Baltica and other cratons, it is important to characterise and date these plutons.

The area underneath Hardangerjøkulen in Finse, Hallingdal, comprises a prominent coarse-grained and mostly massive granite. The rock has also played an important role as construction material for housing and bridges during the establishment of the railway some 100 years ago. Mapping of the area as part of a

MSc thesis (Jensen, 2012) has permitted observation of the relationships between this locally dominant granitic facies and other felsic and mafic phases and their structural state. In this paper we report some of the key observations, and U–Pb ages for the main phases of the batholith and for one late pegmatite suite. The results are then discussed in the context of the tectonic relationships and inferred processes responsible for the development of the Sveconorwegian orogen in the region.

The Finse magmatic complex

Finse lies in the Hardangervidda sector of the Telemarkia Terrane, west of the Mandal–Ustaoset Fault and Shear Zone, which separates it from the Telemark sector (Fig. 1; Bingen et al., 2008). The autochthonous basement

Jensen, E. & Corfu, F. 2016: The U–Pb age of the Finse batholith, a composite bimodal Sveconorwegian intrusion. *Norwegian Journal of Geology* 96, 171–178. <http://dx.doi.org/10.17850/njg96-3-01>.

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Figure 1. Left side: Large-scale map of the Sveconorwegian orogen showing the main subdivisions (after Bingen et al., 2005), the approximate distribution of the main plutonic suites (SMB and HBG), and location of the Finse area. The Mandal–Ustaoet Fault divides Telemarkia into the Telemark sector to the east and the Hardangervidda sector to the west. Right side: Geological map of the Finse area with sample locations (simplified from Sigmond, 1998; Roffeis & Corfu, 2014). Abbreviation: WGR – Western Gneiss Region.

exposed around Finse, underneath the Palaeozoic sedimentary cover and Caledonian nappes, is dominated by intrusive rocks with subordinate occurrences of supracrustal rocks (Sigmond, 1998; Henriksen, 2000; Askvik, 2008). The most characteristic type of rock of the intrusive complex is a very coarse-grained feldspar-megacrystic granite, with 1–3 cm-long feldspar phenocrysts (Fig. 2A). The mineralogical composition of the granite and its retrogression were described in detail in Goldschmidt (1912), who also listed a major element analysis. This coarse facies ranges locally into equigranular, medium- to fine-grained granite, the feldspar phenocrysts becoming sparse or fully disappearing. Other parts of the complex, however, show a high degree of complexity involving fine-grained gabbro (Fig. 2B–E), locally with sparse feldspar phenocrysts (Fig. 2D), and various types of fine-grained felsic rocks and pegmatites. There is local evidence of gabbro intruding granite and vice versa, of mingling of mafic and felsic magmas (Fig. 2C), and sedimentary structures reflecting magma chamber convection (Fig. 2F). Mafic rocks are mostly fine grained, but coarse gabbro is also observed locally in an area west of Finse Lake where it is associated with medium-grained granodiorite. There is no sharp boundary between gabbro and granodiorite suggesting that the transition is due to fractional crystallisation, but the granodiorite also contains mafic xenoliths.

The main megacrystic granite facies is commonly massive and without oriented structures, but locally there is also evidence for deformation and oriented emplacement of dykes (Fig. 2G–I). There are several generations of pegmatite dykes, the earliest one highly deformed (Fig. 2H–I), whereas the youngest dykes are straight and without evidence of deformation.

The extent and geometry of the batholith is not yet well known, but the specific granitic units can be traced for several tens of kilometres in the area of Hardangerjøkulen suggesting a size of over 900 km² (Sigmond, 1998; Henriksen, 2000; Askvik, 2008).

The basement was eventually uplifted, eroded down, forming a regolith which is locally preserved as a zone of weathering and alteration a few metres thick, with typical rustiness, and the local development of a basal conglomerate (Goldschmidt, 1912; Gabrielsen et al., 2015).

Analytical procedure

The analyses were carried out by the ID–TIMS U–Pb technique (Krogh, 1973). Zircon and titanite were separated by crushing, pulverising, Wilfley table,

Figure 2. (A) Typical coarse-grained megacrystic granite, (B) net-veined fine-grained gabbro, (C–D) mingling of mafic and felsic components, (E) feldspar-phyric veined gabbro; late granitic veins in gabbro, (F) primary cross-bedding in granite, (G) mafic xenoliths and late felsic veins, (H) boudinaged felsic dykes in gabbro, (I) folded felsic dyke in gabbro.

magnetic separation and heavy liquids. Suitable grains were subjected to either air abrasion (zircon and titanite; Krogh, 1982) or chemical abrasion (zircon; Mattinson, 2005). The grains were dissolved in HF, after addition of a mixed ^{202}Pb – ^{205}Pb – ^{235}U spike, and processed through ion exchange resin separation and solid source mass spectrometry. Details are described in Corfu (2004). The data are calculated with the decay constants of Jaffey et al. (1971) and plot with the program of Ludwig (2009). Initial common Pb has been corrected with compositions estimated with the model of Stacey & Kramers (1975).

Samples

Four samples were investigated in this study. Sample C-08-4 represents the typical coarse-grained megacrystic granite (Figs. 1, 2A), C-11-5 is granodiorite, and C-08-5 a coarse-grained gabbro. Sample C-11-8 represents late pegmatites which cross-cut all visible structures and phases and are themselves essentially undeformed. The sample consisted of pieces collected from different dykes and dykelets in the same outcrop.

Zircon characteristics and U–Pb data

The granite contains abundant zircon crystals, dominantly brownish and translucent to opaque, and prismatic ($l/w = 2$ to >5). The crystal faces are of variable type, but well developed $\{100\}$ and $\{101\}$ faces are most common. Cores are visible in some of the zircon grains. Five analyses carried out on selections of the grains, either air abraded or chemically abraded, are 1 to 7% discordant (Table 1, Fig. 3). They define a discordia line with intercept ages of 368 ± 26 and 986.7 ± 1.9 Ma (MSWD = 0.38). The lower intercept age is consistent with Pb loss during the Caledonian orogeny.

The zircon population in the granodiorite sample is also abundant and prismatic as in the granite, but less strongly coloured. The morphology is dominated by $\{100\}$ and $\{101\}$ faces with euhedral to, in part, nearly anhedral outlines. Inclusions of other minerals are ubiquitous. Four of the analyses, both air and chemically abraded, are about 1% discordant and a fifth 4.5% discordant. Together they define a line with intercepts of 36 ± 84 and 980.1 ± 1.6 Ma (MSWD = 0.91). The trajectory, however, is strongly influenced by the more discordant analysis (x in Fig. 3), the other four points fitting on the same line as the granite results, suggesting some superimposed younger Pb loss or an inherited component affecting the discordant fraction.

Figure 3. Concordia plots with zircon and titanite data for the various phases of the Finse batholith. Ellipses indicate 2 sigma uncertainty. Analysis x not included in the main regression line (G) yielding the main age of emplacement of the granite-granodiorite-gabbro complex at 985.6 ± 1.6 Ma. Lines P1 and P2 are calculated from subsets of data from the pegmatites; see text for details.

Table 1. U–Pb data.

Characteristics	Weight [μg]	U [ppm]	Th/U (3)	Pbi [ppm]	Pbc [pg]	$^{206}\text{Pb}/^{204}\text{Pb}$ (5)	$^{207}\text{Pb}/^{235}\text{U}$ (6)	$\pm 2\sigma$ [abs]	$^{206}\text{Pb}/^{238}\text{U}$ (6)	$\pm 2\sigma$ [abs]	ρ_{ho}	$^{207}\text{Pb}/^{206}\text{Pb}$ (6)	$\pm 2\sigma$ [abs]	$^{206}\text{Pb}/^{238}\text{U}$ [age in Ma]	$\pm 2\sigma$ (6)	$^{207}\text{Pb}/^{235}\text{U}$ [age in Ma]	$\pm 2\sigma$ (6)	$^{207}\text{Pb}/^{206}\text{Pb}$ [age in Ma]	$\pm 2\sigma$ (6)	Disc. [%]
Granite - C-08-4, 60°35.899' / 7°27.957' (60.59832 / 7.46595)																				
Zeup [13] CA	13	392	0.60	0.00	1.5	35079	1.6144	0.0036	0.16290	0.00032	0.96	0.07188	0.00005	972.9	1.8	975.8	1.4	982.4	1.3	1.0
Zeusp [3] CA	41	446	0.47	0.15	8.3	21828	1.5643	0.0037	0.15848	0.00033	0.96	0.07159	0.00005	948.3	1.8	956.2	1.5	974.2	1.3	2.9
Zeup [11] AA	4	695	0.86	1.12	6.6	4088	1.5239	0.0039	0.15485	0.00033	0.93	0.07138	0.00007	928.1	1.8	939.2	1.6	965.4	2.0	4.2
Zlp-fr [4] AA	7	426	0.69	0.80	7.8	3633	1.4715	0.0038	0.15030	0.00032	0.92	0.07101	0.00007	902.7	1.8	918.0	1.6	954.9	2.1	5.9
Zeup br-pk [7] AA	14	505	0.48	2.63	39.8	1661	1.4434	0.0043	0.14775	0.00033	0.83	0.07085	0.00012	888.3	1.8	906.3	1.8	950.4	3.5	7.0
Tbr [10] AA	69	171	2.68	3.59	256	484	1.5860	0.0110	0.16144	0.00041	0.51	0.07125	0.00043	964.8	2.3	963.9	4.3	962	12	-0.3
Tbr-r [5] AA	45	157	3.19	3.26	153	481	1.5708	0.0101	0.16001	0.00039	0.50	0.07120	0.00040	956.8	2.2	957.9	4.0	960	11	0.4
Granodiorite - C-11-5, 60°35.838' / 7°26.062' (60.5973 / 7.4344)																				
Zeusp [1] CA	2	502	0.77	0.00	1.9	5467	1.6170	0.0041	0.16323	0.00032	0.86	0.07185	0.00009	974.7	1.8	976.8	1.6	981.6	2.7	0.8
Zlp [3] CA	5	125	0.98	0.18	3.0	2184	1.6163	0.0049	0.16322	0.00034	0.79	0.07182	0.00014	974.6	1.9	976.6	1.9	980.8	3.9	0.7
Zeusp [8] CA	21	190	0.83	0.02	2.4	16683	1.6032	0.0036	0.16203	0.00032	0.95	0.07176	0.00005	968.1	1.8	971.5	1.4	979.2	1.4	1.2
Zsb [3] AA	1	589	0.85	0.00	0.9	6724	1.5980	0.0052	0.16154	0.00043	0.88	0.07174	0.00011	965.3	2.4	969.4	2.0	978.7	3.1	1.5
Zeup [8] CA	12	149	0.85	0.01	2.1	8392	1.5503	0.0038	0.15669	0.00032	0.93	0.07176	0.00007	938.3	1.8	950.6	1.5	979.1	1.9	4.5
Tfr br [2] AA	34	124	2.88	3.43	122	356	1.524	0.011	0.15652	0.00036	0.33	0.07062	0.00049	937.4	2.0	940.1	4.5	946	14	1.0
Tfr br [1] AA	200	123	2.90	8.67	1789	148	1.456	0.034	0.15287	0.00103	0.36	0.06908	0.00153	917.0	5.9	912.4	14.2	901	45	-1.9
Gabbro - C-08-5, 60°35.856' / 7°26.321' (60.5976 / 7.4387)																				
Zanpk [1] AA	11	260	0.90	0.11	3.3	8932	1.6229	0.0039	0.16354	0.00033	0.93	0.07197	0.00007	976.4	1.8	979.1	1.5	985.1	1.8	1.0
Zan [2] AA	19	183	0.83	0.12	4.3	8286	1.6160	0.0039	0.16322	0.00033	0.95	0.07181	0.00006	974.6	1.8	976.5	1.5	980.6	1.6	0.7
Zanpk [1] AA	27	258	0.74	0.63	19.7	3643	1.6181	0.0039	0.16319	0.00033	0.93	0.07191	0.00007	974.5	1.9	977.3	1.5	983.5	1.9	1.0
Zan [10] AA	23	163	0.91	0.09	4.1	9370	1.6064	0.0037	0.16230	0.00032	0.95	0.07179	0.00006	969.5	1.8	972.7	1.5	979.9	1.6	1.1
Pegmatite - C-11-8, 60°35.622' / 7°25.456' (60.5937 / 7.4243)																				
Zpk AA [1]	1	518	0.83	0.00	1.4	3649	1.5740	0.0075	0.15976	0.00062	0.86	0.07145	0.00018	955.5	3.5	960.0	2.9	970.5	5.0	1.7
Zeusp AA [1]	1	277	0.45	0.00	0.9	2916	1.3976	0.0054	0.14576	0.00037	0.81	0.06955	0.00016	877.1	2.1	887.9	2.3	914.9	4.8	4.4
Zfr AA [1]	1	250	0.50	0.32	2.3	988	1.4000	0.0083	0.14463	0.00050	0.67	0.07020	0.00031	870.8	2.8	888.9	3.5	934.3	9.1	7.3
Zpk AA [1]	1	696	0.29	0.00	1.3	4023	1.0702	0.0039	0.11572	0.00034	0.86	0.06708	0.00012	705.9	1.9	738.9	1.9	840.1	3.8	16.9
Tbr AA [1]	2	331	2.68	7.5	17.5	397	1.5732	0.0103	0.16050	0.00038	0.40	0.07109	0.00043	959.6	2.2	959.7	4.1	960.0	12.2	0.0
Tbr AA [1]	20	180	2.45	5.2	108.0	352	1.5629	0.0115	0.16001	0.00040	0.36	0.07084	0.00049	956.8	2.3	955.6	4.6	952.8	14.0	-0.5
Tbr AA [1]	1	316	3.41	7.8	10.0	328	1.5162	0.0120	0.15713	0.00042	0.38	0.06998	0.00051	940.8	2.4	936.9	4.9	927.8	14.9	-1.5
Tbr AA [1]	13	284	2.60	3.4	46.8	792	1.5186	0.0061	0.15677	0.00035	0.64	0.07025	0.00022	938.8	2.0	937.9	2.5	935.7	6.4	-0.4
Tbr AA [1]	2	363	3.93	11.0	24.6	306	1.5171	0.0125	0.15667	0.00036	0.29	0.07023	0.00055	938.2	2.1	937.3	5.1	935.2	16.0	-0.4

(1) Z = zircon; T = titanite; eu = euhedral; sb = subhedral; an = anhedral; lp = long prismatic; sp = short prismatic; fr = fragment; br = brown; r = red; pk = pink; AA = air abrasion; CA = chemical abrasion; [1] number of grains in fractions

(2) concentration better than 10%, except for grains of 1–2 μg where uncertainty is up to 50%

(3) Th/U model ratio inferred from 208/206 ratio and age of sample

(4) Pbi: initial common Pb; Pbc = total amount of common Pb (initial + blank)

(5) raw data corrected for fractionation

(6) corrected for fractionation, spike, blank and initial common Pb; error calculated by propagating the main sources of uncertainty; initial common Pb corrected using Stacey & Kramers (1975) model compositions.

The zircon grains from the gabbro are mostly anhedral and in part irregular. The four analyses are clustered together and when regressed from a Caledonian lower intercept age of 400 ± 40 Ma yield an upper intercept age of 986.4 ± 4.3 Ma (MSWD = 4.4).

Regression of all the thirteen zircon data from the three samples together, except for the discordant analysis in the granodiorite, yields a line with intercepts at 358 ± 32 and 985.6 ± 1.6 Ma (MSWD = 1.8; line G in Fig. 3).

The pegmatite sample contains relatively few, predominantly metamict and zoned zircon grains together with brown-red titanite. Four zircon analyses were carried out on air abraded euhedral crystals and fragments. The results are variously discordant; three of the analyses define a discordia line with intercepts at 976 ± 8 Ma and 409 ± 14 Ma (MSWD = 0.08; line P₁ in Fig. 3), interpreted as indicating the time of crystallisation and of Caledonian Pb loss. Another zircon data point, however, deviates to the left of this line. The titanites are rich in U (up to 360 ppm) and Th (Th/U up to 3.9), and also in common Pb (3–11 ppm). The analyses plot in two groups, two grains indicating 958 ± 2 Ma and three others 939 ± 2 Ma. The latter group coincides with a titanite analysis from the gabbro, but that could be a coincidence. More importantly, the titanite data fit on a line through the deviating zircon analysis mentioned above and projecting to a Caledonian lower intercept age (line P₂ in Fig. 3). These relationships suggest that 939 Ma is the age of intrusion of one of the pegmatite dykes. The significance of the 958 Ma titanite date is more uncertain, and may reflect partial resetting. This is evident in two other titanite analyses of the granite and the gabbro that clearly postdate the initial magmatic crystallisation. The effect of a superimposed Caledonian Pb loss on the most reset analysis is possible but would have been of restricted magnitude.

Discussion

The concluding period of the Sveconorwegian orogeny in the Telemarkia Terrane was accompanied by widespread plutonic activity lasting close to 150 m.y. Three main plutonic associations have been recognised: (1) the Sirdal Magmatic Belt (SMB), including the Feda–Fennefoss suite, formed between 1060 and 1020 Ma (Bingen et al., 2008, 2015; Vander Auwera et al., 2011; Slagstad et al., 2013a; Coint et al., 2015); (2) the hornblende-biotite-granitoid suite (HBG), formed between 990 and 920 Ma (Bingen et al., 2008; Vander Auwera et al., 2011); and (3) the anorthosite-mangerite-charnockite suite formed between 950 and 920 Ma (Schärer et al., 1996; Vander Auwera et al., 2011). The SMB is a calc-alkaline, magnesian association whereas the HBG is more alkaline and ferroan (Slagstad et al., 2013a).

The chemical analysis available for the Finse granite (Goldschmidt, 1912) indicates a ferroan affinity confirming a link between the batholith and the HBG suite. The age of 985.6 ± 1.6 Ma places it as one of the oldest of the suite as the age is within error of the 989 ± 9 Ma age of the Grimstad granite (Kullerød & Machado, 1991) and 990 ± 14 Ma of the Torsdalfjell granite (Andersen et al., 2007). Interestingly, granites with a similar appearance in the more immediate regions at Flå, Øye and Eidsfjord are much younger at about 930 Ma (Priem et al., 1976 (Rb–Sr whole-rock age); Corfu, 1980; Bingen et al., 2008). These younger intrusions, however, seem to lack the mafic components typical of the Finse batholith.

Contrasting models have been proposed for the tectonic setting and causes of this magmatic progression. Bingen et al. (2008) view the SMB as the synorogenic products of either the closing stages of a continental arc or crustal thickening related to an oblique continent-continent collision. Burial during this event caused the regional metamorphism, reaching locally granulite-grade conditions. The HBG suite is then interpreted as a function of the subsequent relaxation with gravitational collapse and upwelling of asthenospheric mantle, the shorter-lived anorthosite-mangerite-charnockite plutonism indicating melting of dry and reduced lower crust whereas the HBG is attributed to melting of wet and oxidized lower crust (Vander Auwera et al., 2011). The contrasting view advanced by Slagstad et al. (2013a) considers the SMB suite instead as a Cordilleran-type, subduction-related, plutonic complex. The post-SMB lull in magmatic activity is attributed to flattening of the slab, with renewed steep subduction and presumably stepping back of the slab causing some delamination and back-arc activity which then produces the HBG suite.

Both models have merits and problematic aspects. Möller et al. (2013) pointed out that the eastern parts of the orogen exhibit much evidence for collisional tectonics, including thick and extensive mylonite zones indicating SE transport and subsequent extension. Crustal imbrication and deep burial creating high-pressure rocks are also indicative of convergent processes; and they also emphasise the difficulty of achieving ultrahigh-temperature metamorphism, such as is observed in Rogaland at about 1010 Ma, in a 'cold' subduction environment. Coint et al. (2015), however, concluded that these ultrahigh-temperature processes were likely of local rather than of regional extent. Slagstad et al. (2013a, b) argued, furthermore, that the magmatic and metamorphic record of the western part of the Sveconorwegian orogen is most compatible with long-lived accretionary processes in an Andean-type subduction setting, stressing in particular that there is no direct evidence in that part of the orogen for crustal thickening and collision with another continent. The difference in apparent tectonic regimes between the eastern and western domains of the Sveconorwegian

orogen may reflect relatively late amalgamation of crustal blocks (Coint et al., 2015). A separate evolution between Telemarkia and the rest of the orogen, and their eventual collision, has been considered before (e.g., Bingen et al., 2005). Although controversial (e.g., Andersen et al., 2004), the concept may yet provide the best explanation for the conflicting tectonic features observed across the Sveconorwegian orogen.

Some relevant information on the Sveconorwegian tectonic processes is also available in the overlying Caledonian allochthons. The Hardangervidda–Ryfylke Nappe Complex, which in the Silurian was thrust over autochthonous basement and its cover, includes the Kvitenut and Dyrskard nappes. The Kvitenut Nappe has a Gothian affinity, such as the crust in the Sveconorwegian foreland to the west and north, and was thrust on top of the Dyrskard Nappe, which has a Telemarkian affinity (Roffeis et al., 2013). Thrusting occurred at around 1000 Ma and the suture was stitched by an essentially coeval synorogenic granite. The pre-thrusting position of the nappe complex is uncertain, but it is reasonable to assume that it was located in the projection of the present boundary between the two terranes. The important point in this context is that the time at around 1000 Ma corresponded, at least locally, to convergence and terrane juxtapositions, combined with metamorphism and some magmatism. Such a process would seem to fit better a continent collisional setting, but terrane aggregation during accretional processes would also be a viable solution.

Two important observations in the Finse batholith are (1) the bimodal nature of the intrusion and (2) the evidence for the oriented emplacement of mafic and felsic dykes associated with the batholith (Jensen, 2012). The first point implies the availability of sources for the mafic material, most likely the mantle with basaltic magma rising and mingling with crustal melts from the lower crust. Future isotope analyses should contribute to verify this inference. The contribution from mantle and lower crust is consistent with both the back-arc and the extension-delamination hypothesis. It is also consistent with the isotope geochemical characteristics of the HBG suite (Andersen et al., 2009). Vander Auwera et al. (2011), however, postulated a provenance of the more juvenile component by melting of underplated mafic crust. The second point, the orientation of small intrusive bodies and their local deformation indicating emplacement in a stress regime, is indicative of a compressional setting, a feature already evident by the general N–S arrangement of the plutons on the larger scale of the craton (Coint et al., 2015). These features are in principle compatible with aspects of both models presently under discussion.

Conclusions

The Finse batholith is a bimodal magmatic complex emplaced at 985.6 ± 1.6 Ma in Telemarkian crust. It was followed by discrete events of pegmatite emplacement, at least two of which are reflected by the data in this paper at 976 ± 8 Ma and 939 ± 2 Ma. The bimodal composition implies mixing of crustal and mantle derived magmas. Other features of the batholith also suggest emplacement in a compressional regime.

Acknowledgements. We are grateful to Gunborg Bye Fjell for help in the preparation of the samples. Fieldwork benefited from financial support from Småforsk grants. Reviews by Bernard Bingen and editor Trond Slagstad are gratefully acknowledged.

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