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CORRESPONDENCE: Trond.Slagstad@ngu.no

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Magma-driven, high-grade metamorphism in the Sveconorwegian Province, southwest Norway, during the terminal stages of Fennoscandian Shield evolution

Trond Slagstad[1], Nick M.W. Roberts[2], Nolwenn Coint[1], Ingjerd Høy[1,3], Simone Sauer[1,4], Christopher L. Kirkland[5], Mogens Marker[1], Torkil S. Røhr[1], Iain H.C. Henderson[1], Martin A. Stormoen[1,6], Øyvind Skår[1], Bjørn Eske Sørensen[6], and Grant Bybee[7]

[1]Geological Survey of Norway, PO Box 6315 Sluppen, 7491 Trondheim, Norway
[2]NERC (Natural Environment Research Council) Isotope Geosciences Laboratory, British Geological Survey, Nottingham, NG12 5GG, UK
[3]Department of Geosciences, University of Oslo, PO Box 1047 Blindern, 0316 Oslo, Norway
[4]IFREMER (Institut Français de Recherche pour l'Exploitation de la Mer) Unité de Recherche Géosciences Marines, 29280 Plouzané, France
[5]Centre for Exploration Targeting–Curtin Node, Department of Applied Geology, Western Australian School of Mines, Curtin University, Perth, WA 6102, Australia
[6]Department of Geoscience and Petroleum, Norwegian University of Science and Technology, Sem Sælands veg 1, N-7491 Trondheim, Norway
[7]School of Geosciences, University of the Witwatersrand, Private Bag 3, 2050 Wits, Johannesburg, South Africa

ABSTRACT

Recently it has been argued that the Sveconorwegian orogeny in southwest Fennoscandia comprised a series of accretionary events between 1140 and 920 Ma, behind a long-lived, active continental margin characterized by voluminous magmatism and high-grade metamorphism. Voluminous magnesian granitic magmatism is recorded between 1070 and 1010 Ma (Sirdal Magmatic Belt, SMB), with an apparent drop in activity ca. 1010–1000 Ma. Granitic magmatism resumed ca. 1000–990 Ma, but with more ferroan (A type) compositions (hornblende-biotite granites). This ferroan granitic magmatism was continuous until 920 Ma, and included emplacement of an AMCG (anorthosite-mangerite-charnockite-granite) complex (Rogaland Igneous Complex). Mafic rocks with ages corresponding to the spatially associated granites suggest that heat from underplated mafic magma was the main driving force for lower crustal melting and long-lived granitic magmatism. The change from magnesian to ferroan compositions may reflect an increasingly depleted and dehydrated lower crustal source. High-grade metamorphic rocks more than ~20 km away from the Rogaland Igneous Complex yield metamorphic ages of 1070–1015 Ma, corresponding to SMB magmatism, whereas similar rocks closer to the Rogaland Igneous Complex yield ages between 1100 and 920 Ma, with an apparent age peak ca. 1000 Ma. Ti-in-zircon temperatures from these rocks increase from ~760 to 820 °C ca. 970 Ma, well before the inferred emplacement age of the Rogaland Igneous Complex (930 Ma), suggesting that long-lived, high-grade metamorphism was not directly linked to the emplacement of the latter, but rather to the same mafic underplating that was driving lower crustal melting. Structural data suggest that the present-day regional distribution of high- and low-grade rocks reflects late-stage orogenic doming.

INTRODUCTION

The Sveconorwegian orogen (Fig. 1) is the result of a series of geographically and tectonically discrete events on the southwest margin of Fennoscandia between ca. 1140 and 920 Ma (Slagstad et al., 2017). These events apparently took place along an active margin that underwent a long-lived magmatic, metamorphic and deformational evolution that is the topic of this paper. The long-duration, high- to ultrahigh-temperature (HT-UHT) metamorphic evolution in Rogaland, southwest Norway, during the Sveconorwegian orogeny has been extensively studied since the 1970s (e.g., Hermans et al., 1975; Tobi et al., 1985), and the extreme metamorphic conditions (to 1000 °C at 5–7 kbar) determined in these early studies have largely been confirmed by more recent investigations (Blereau et al., 2017; Drüppel et al., 2013). Studies of the timing of metamorphism have produced a wide range of U-Pb zircon and monazite ages between ca. 1050 and 920 Ma (Bingen et al., 2008a; Bingen and van Breemen, 1998; Drüppel et al., 2013; Laurent et al., 2016; Möller et al., 2002, 2003; Tomkins et al., 2005), suggesting that this region underwent >100 m.y., and possibly as much as 150 m.y., of HT-UHT metamorphism. The metamorphic evolution in Rogaland has typically been ascribed to a two-phase tectonic history, involving Himalayan-type continent-continent collision ca. 1050 Ma or earlier, with resulting regional, high-grade metamorphism from ca. 1035 to 970 Ma (Bingen et al., 2008a; Drüppel et al., 2013), followed by decompression between ca. 970 and 950 Ma (Tomkins et al., 2005), and finally widespread contact metamorphism ca. 930 Ma related to emplacement of the Rogaland Igneous Complex (Blereau et al., 2017; Westphal et al., 2003) during extensional collapse of the orogen.

Although the general metamorphic evolution is broadly agreed upon, the underlying tectonic interpretation has been questioned, with many now advo-

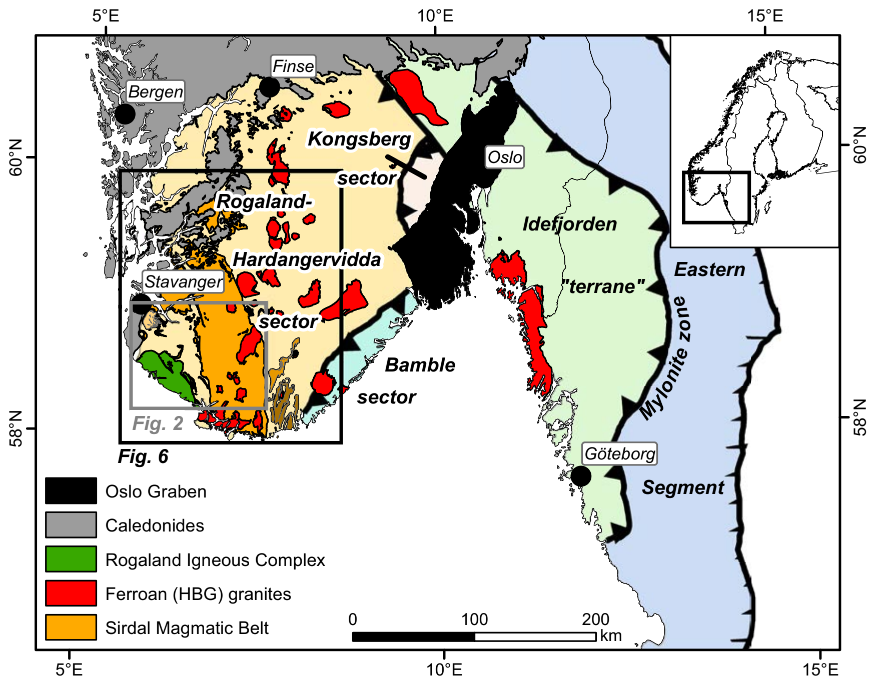


Figure 1. Simplified geologic map of the Sveconorwegian Province outlining major tectonic units. The areas covered by maps in Figures 2 and 6 are indicated by squares. The quotation marks in Idefjorden terrane have been added to signify that this tectonic unit may not fulfill all criteria for a geologic terrane as defined by Jones et al. (1983). HBG—hornblende-biotite granite.

cating a general accretionary setting (Blereau et al., 2017; Bybee et al., 2014; Coint et al., 2015; Slagstad et al., 2013a) as an alternative to Himalayan-type collision, although this model is not universally accepted (Möller et al., 2015, 2013). A major reason for invoking accretionary processes is the recognition that magmatism played a much greater role in the evolution of the western Sveconorwegian Province than previously known. Few studies have, however, attempted to reconcile the magmatic and metamorphic evolution into a single tectonic model.

Here we present a new comprehensive U-Pb zircon geochronology dataset for magmatic and metamorphic events, new regional structural and aeromagnetic data, as well as some critical field relationships from southwest Norway, that, taken together, point to a tectonic evolution involving (1) construction of a continental-margin arc between 1070 and 1010 Ma, (2) decreased magmatic activity and coeval peak in growth of metamorphic zircon between 1010 and 1000 Ma, possibly a result of flat-slab subduction, (3) renewed arc magmatism starting at 1000 Ma, followed by extension and doming and/or uplift at 970–950 Ma, and (4) final emplacement of the Rogaland Igneous Complex at 950–920 Ma in the core of this dome. We also show how the geochronologic record is highly dependent on the reactivity (i.e., composition) of the rocks in which the chronometer (zircon) grew, and that, depending on lithology, different rocks may preserve evidence for all, some, or none of the geodynamic events that shaped the region. Regional geological interpretations therefore need to be founded on multiple, lithologically varied samples.

REGIONAL BACKGROUND

The rocks investigated in this study span ~600 m.y., from 1.5 to 0.9 Ga. Although the focus is on the period from 1070 to 920 Ma, the preceding history is important as it allows the identification of suites of rocks and because it may have primed some rocks prior to Sveconorwegian high-grade metamorphism (cf. Clark et al., 2011). Priming in this case refers to an earlier phase of high-grade metamorphism that leaves the rocks dehydrated. The dehydration makes the rocks less susceptible to melting during later high-grade metamorphism, thereby reducing their capacity to buffer increased temperatures by undergoing melting. We therefore briefly describe the main lithologic units in the study area, subdivided according to age. Figure 2 shows a simplified map of the area, based on a compilation of recent mapping by the Geological Survey of Norway (NGU, Norges Geologiske Undersøkelse) and the Mandal 1:250,000 map sheet (Falkum, 1982).

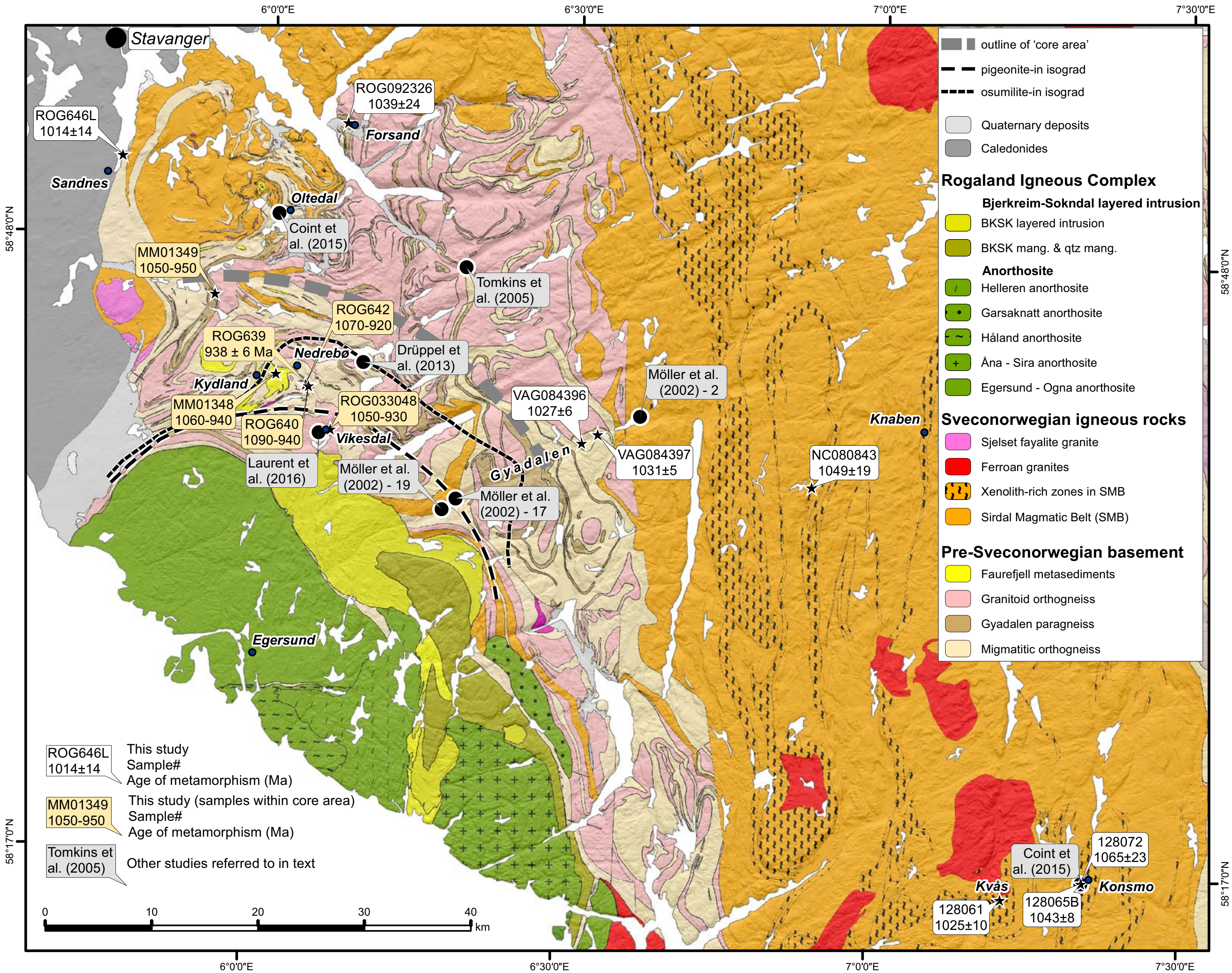


Figure 2. Detailed geologic map of the main study area in southwest Norway. The map is compiled from the 1:250,000-scale Mandal map sheet (Falkum, 1982), the 1:75,000-scale map of the Rogaland Anorthosite Province (Marker et al., 2003), and mapping by the Geological Survey of Norway (NGU, Norges Geologiske Undersøkelse) since the early 2000s (Marker, 2013, 2018; Marker and Slagstad, 2018a, 2018b, 2018c; Marker et al., 2012). Also shown on the map are locations of samples dated in this study that have a bearing on the metamorphic evolution of the area, as well as previously published samples that are discussed in the text. The geographic extent of the core area is loosely defined to encompass samples that contain evidence of long-lived continuous or semicontinuous growth and regrowth of zircon between ca. 1080 and 920 Ma, covering the entire age range of granitic magmatism in the southwest Sveconorwegian Province. BKSK—Bjerkreim-Sokndal; mang. & qtz mang.—mangerite and quartz mangerite.

Pre-Sveconorwegian Rocks

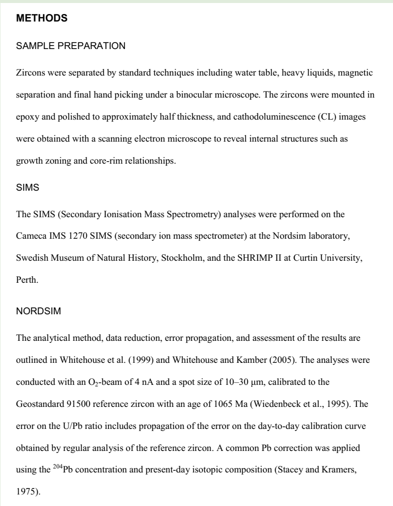
The oldest rocks in the southwestern part of the Sveconorwegian Province are ca. 1.5 Ga orthogneisses, with plutonic and volcanic protoliths (Bingen et al., 2005; Roberts et al., 2013). The 1.5 Ga orthogneisses have been variably metamorphosed, and in southwest Rogaland are generally migmatitic (Fig. 3A). These rocks are therefore referred to as migmatitic orthogneisses herein. Mafic sheets and dikes are commonly observed in these rocks, suggesting a later mafic magmatic event that has not yet been dated directly. Although original field relationships are obscured by strong deformation, the 1.5 Ga orthogneisses appear to have formed the basement for pelitic or semipelitic sediments, now metamorphosed at high grade (Figs. 3B, 3C, 3D) (Blereau et al., 2017; Tomkins et al., 2005). These supracrustal rocks are referred to as the Gyadalen paragneisses (Gyadal garnetiferous migmatites of Hermans et al., 1975). The age of deposition of the Gyadalen paragneisses is poorly constrained. Detrital zircons display a clear ca. 1.5 Ga peak (Drüppel et al., 2013; Tomkins et al., 2005; this study), with a few grains extending down to ca. 1.25 Ga, but extracting a reliable younger than 1.5 Ga maximum age of deposition is difficult. Like the migmatitic orthogneisses, the Gyadalen paragneisses also contain sheets of mafic rock, similarly with primary contacts obscured by deformation. The migmatitic orthogneisses and Gyadalen paragneisses were intruded by voluminous granitoid orthogneisses (Fig. 3E) dated at 1.23–1.20 Ga (Slagstad et al., this study), with ages as old as 1.28 Ga farther east in Setesdalen (Pedersen et al., 2008). Unlike the older migmatitic orthogneisses and Gyadalen paragneisses, the ca. 1.23–1.20 Ga granitoid orthogneisses (and younger rocks, described in the following) do not contain mafic dikes, providing a minimum age for this mafic magmatism. The last known pre-Sveconorwegian event to take place in southwest Norway was deposition of the Faurefjell metasediments (Hermans et al., 1975), consisting of quartzite, impure marble, calc-silicate gneiss (Fig. 3F), and distinct oxide-rich layers (Fig. 3G) that have been interpreted to represent metamorphosed laterite (Bol et al., 1989).

Sveconorwegian Magmatism

The oldest Sveconorwegian rocks in the study area compose the Sirdal Magmatic Belt (SMB), a ca. 1070–1020 Ma magnesian, calc-alkaline granite batholith, mainly consisting of K-feldspar porphyritic granites (Fig. 4A) and described in detail in Slagstad et al. (2013a) and Coint et al. (2015). The SMB is ~50 km wide and extends almost 150 km northward, from the southern tip of Norway, before it swings west and disappears under the Paleozoic Caledonian nappes. The SMB has been interpreted to represent a continental arc on the southwest margin of Fennoscandia. Magmatism in the SMB ceased ca. 1020 Ma, and was succeeded by widespread, long-lived granitic and bimodal magmatism between ca. 990 and 920 Ma (e.g., Jensen and Corfu, 2016; Slagstad et al., 2013a; Vander Auwera et al., 2003). The granites of the latter age range constitute relatively large, seemingly isolated bodies throughout the orogen, except farthest east, in the Eastern Segment. They are typically grouped into the so-called HBG suite (hornblende-biotite granites, Fig. 4B), and their ferroan (A-type like) compositions have led most workers to favor formation in an extensional setting, either as a result of postorogenic collapse (Vander Auwera et al., 2003) or extension behind an active continental margin (Slagstad et al., 2013a). The Rogaland Igneous Complex comprises two distinctly different units, the Rogaland Anorthosite Province and the Bjerkreim-Sokndal layered intrusion. The Rogaland Anorthosite Province is a massif-type anorthosite comprising three separate units: Egersund-Ogna, Håland-Helleren, and Åna-Sira. The Bjerkreim-Sokndal layered intrusion is a large, layered intrusion consisting of anorthosite through norite to quartz mangerite (Charlier et al., 2010; Wilson et al., 1996). The age of the Bjerkreim-Sokndal layered intrusion is well constrained at ca. 930 Ma (Vander Auwera et al., 2011), and the generally agreed-upon interpretation has been that the Rogaland Anorthosite Province is the same age (Schärer et al., 1996; Westphal et al., 2003). The anorthosites only contain zircon in association with locally abundant high-Al orthopyroxene megacrysts (Fig. 4C); however, the relationship (e.g., exsolution, xenocryst, or growth from late-stage melt pockets) between zircon and high-Al orthopyroxene megacrysts has never been directly observed, leaving the significance of the ca. 930 Ma ages from the anorthosites (Schärer et al., 1996) uncertain. This uncertainty was reinforced in a study (Bybee et al., 2014) that yielded a 1041 ± 17 Ma Sm-Nd isochron age for the high-Al orthopyroxene megacrysts from the Egersund-Ogna anorthosite massif, which they interpreted to reflect growth from the basaltic magma that gave rise to the anorthosite. In the same study, exsolution lamellae of plagioclase from their high-Al orthopyroxene megacryst hosts, orthopyroxene, and bulk high-Al orthopyroxene megacrysts yielded an Sm-Nd isochron age of 968 ± 43 Ma, which was interpreted (Bybee et al., 2014) as reflecting decompression during emplacement of the anorthosite. This age overlaps with the 930 Ma age from Schärer et al. (1996), and supports their suggestion that the zircons may have grown from more evolved melts, crystallized relatively late in the magmatic evolution of the anorthosites.

U-Pb ZIRCON ISOTOPE AND TRACE ELEMENT ZIRCON DATA AND DATA TREATMENT

The analytical methods employed in the U-Pb zircon study are presented in Supplemental Item A[1], and the analytical data are collected in several supplements to relate closely to the text. Supplemental Item B contains a sample-by-sample data description of new key samples for the discussion on metamorphic evolution of the southwest Sveconorwegian Province, and an overview of the data from the nonmetamorphosed magmatic rocks. Supplemental Item C includes new U-Pb zircon data from syn-Sveconorwegian (1060–920 Ma) granites. These samples have comparatively simple U-Pb age



METHODS

SAMPLE PREPARATION

Zircons were separated by standard techniques including water table, heavy liquids, magnetic separation and final hand picking under a binocular microscope. The zircons were mounted in epoxy and polished to approximately half thickness, and cathodoluminescence (CL) images were obtained with a scanning electron microscope to reveal internal structures such as growth zoning and core-rim relationships.

SIMS

The SIMS (Secondary Ionisation Mass Spectrometry) analyses were performed on the Cameca IMS 1270 SIMS (secondary ion mass spectrometer) at the Nordsim laboratory, Swedish Museum of Natural History, Stockholm, and the SHRIMP II at Curtin University, Perth.

NORDSIM

The analytical method, data reduction, error propagation, and assessment of the results are outlined in Whitehouse et al. (1999) and Whitehouse and Kamber (2005). The analyses were conducted with an O_2-beam of 4 nA and a spot size of 10–30 µm, calibrated to the Geostandard 91500 reference zircon with an age of 1065 Ma (Wiedenbeck et al., 1995). The error on the U/Pb ratio includes propagation of the error on the day-to-day calibration curve obtained by regular analysis of the reference zircon. A common Pb correction was applied using the ^{204}Pb concentration and present-day isotopic composition (Stacey and Kramers, 1975).

[1]Supplemental Items. A: Methods. B: Description of dated samples. C: Geochronological data, SMB and ferroan HBG granites. D: Geochronological data, other igneous rocks. E: Geochronological data, metamorphic rocks. F: Geochronological data ca. 1.25 Ga granitoids. G: Sample coordinates and age data summary. Please visit http://doi.org/10.1130/GES01565.S1 or the full-text article on www.gsapubs.org to view the Supplemental Items.

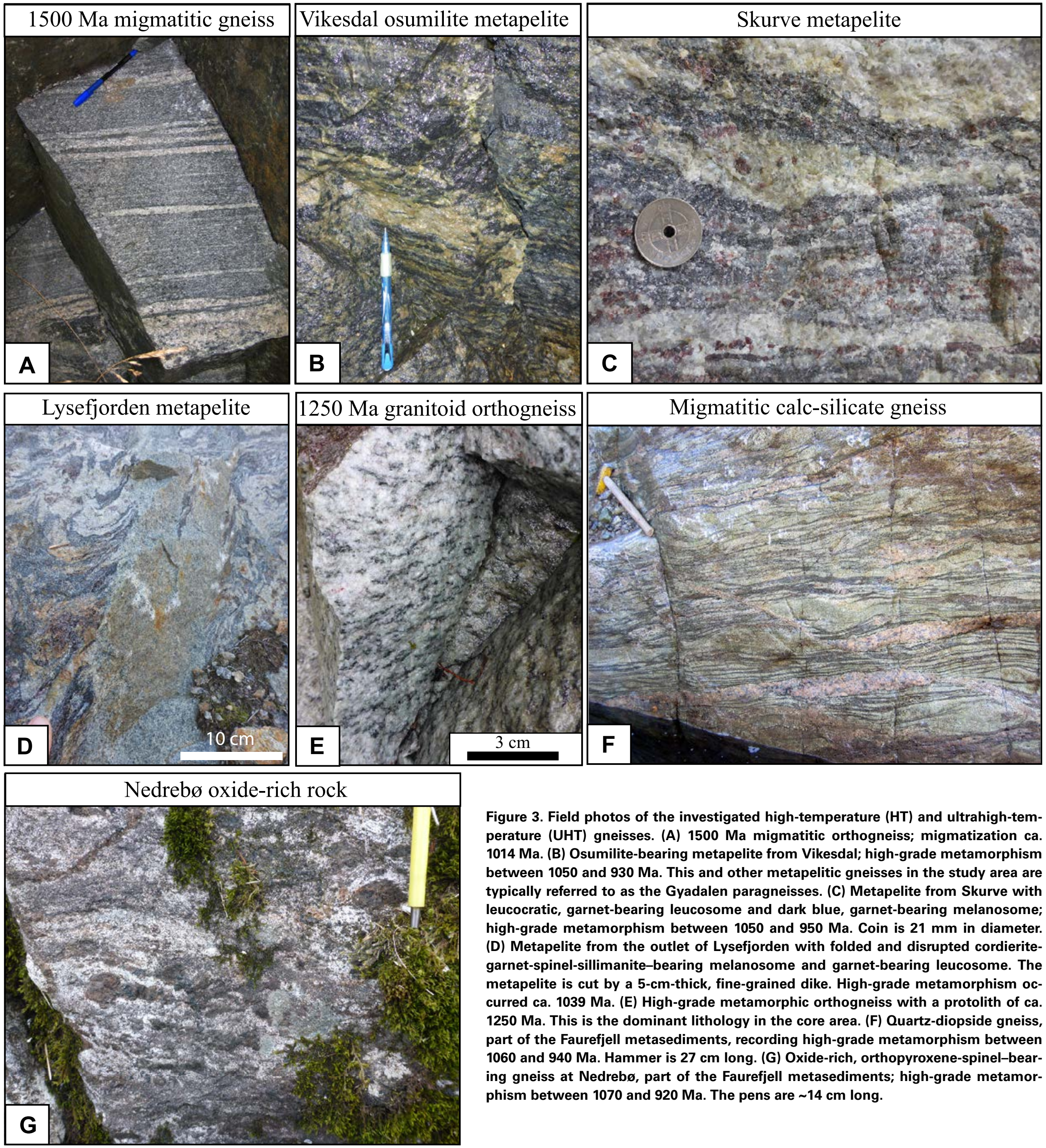


Figure 3. Field photos of the investigated high-temperature (HT) and ultrahigh-temperature (UHT) gneisses. (A) 1500 Ma migmatitic orthogneiss; migmatization ca. 1014 Ma. (B) Osumilite-bearing metapelite from Vikesdal; high-grade metamorphism between 1050 and 930 Ma. This and other metapelitic gneisses in the study area are typically referred to as the Gyadalen paragneisses. (C) Metapelite from Skurve with leucocratic, garnet-bearing leucosome and dark blue, garnet-bearing melanosome; high-grade metamorphism between 1050 and 950 Ma. Coin is 21 mm in diameter. (D) Metapelite from the outlet of Lysefjorden with folded and disrupted cordierite-garnet-spinel-sillimanite–bearing melanosome and garnet-bearing leucosome. The metapelite is cut by a 5-cm-thick, fine-grained dike. High-grade metamorphism occurred ca. 1039 Ma. (E) High-grade metamorphic orthogneiss with a protolith of ca. 1250 Ma. This is the dominant lithology in the core area. (F) Quartz-diopside gneiss, part of the Faurefjell metasediments, recording high-grade metamorphism between 1060 and 940 Ma. Hammer is 27 cm long. (G) Oxide-rich, orthopyroxene-spinel–bearing gneiss at Nedrebø, part of the Faurefjell metasediments; high-grade metamorphism between 1070 and 920 Ma. The pens are ~14 cm long.

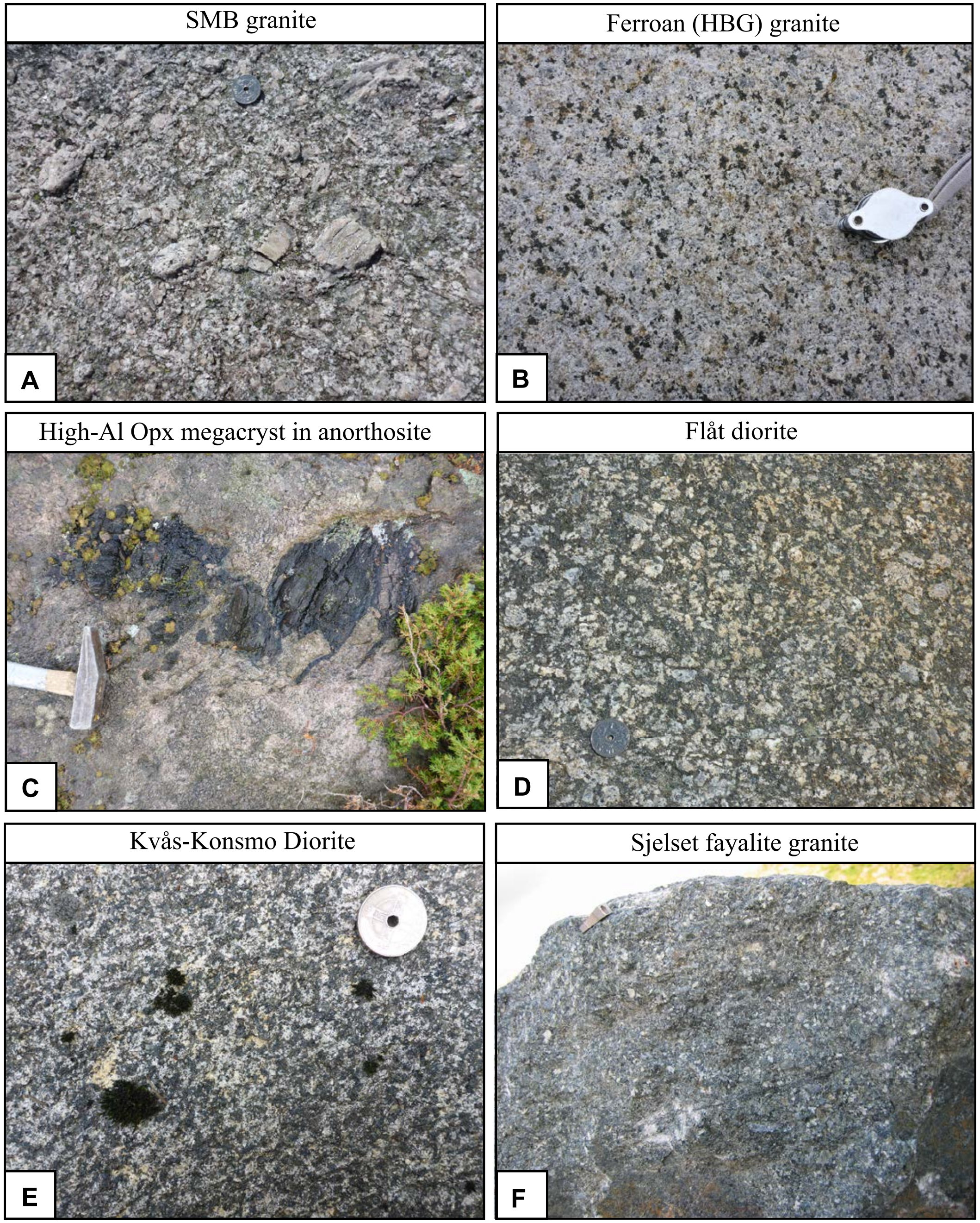


Figure 4. Field photos of Sveconorwegian-age magmatic rocks. (A) Porphyritic granite from the Sirdal Magmatic Belt (SMB). Coin is 21 mm in diameter. (B) Hornblende-phyric granite from the ferroan hornblende-biotite granite (HBG) suite. Hand lens is ~3 cm across. (C) High-alumina orthopyroxene megacryst from the Egersund-Ogna anorthosite. Hammer head is ~14 cm long. (D) Flåt diorite, Ni-bearing diorite. Coin is 21 mm in diameter. (E) Kvås-Konsmo diorite, intruding the SMB. Coin is 21 mm in diameter. (F) Sjelset fayalite granite. Pencil sharpener for scale; ~2 cm long.

systematics and are not discussed on an individual basis in Supplemental Item B. Supplemental Item D includes new U-Pb zircon data from Sveconorwegian-age metaigneous rocks that are presented and discussed individually in Supplemental Item B, as they have particular significance for understanding the geologic evolution. Supplemental Item E includes new U-Pb zircon data from the high-grade metamorphic rocks.

This study includes U-Pb zircon data from 58 samples, ranging from polymetamorphic metapelites with very complicated zircon textures, reflecting an equally complex history of growth and resorption, to granites with relatively simple crystallization histories. In addition, the data have been obtained on a range of instruments, including three laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) systems [at NGU, NIGL (National Environmental Research Council Isotope Geosciences Laboratory and the University of Oslo)] and two secondary-ion mass spectrometry (SIMS) systems (1280 IMS, Nordsim laboratory, Stockholm, Sweden, and the SHRIMP II at Curtin University, Perth, Australia). The amount of data and range of investigated lithologies and analytical tools pose some challenges with respect to objective and consistent data treatment. The approach adopted here is to (1) calculate a concordia age using all available data, only excluding clear statistical outliers that are mixtures, influenced by radiogenic Pb loss, or have sampled inherited domains. (2) If datasets fail to yield a statistically acceptable concordia age for a single population, we have proceeded to calculate a weighted mean $^{207}Pb/^{206}Pb$ age, excluding analyses that are >10% discordant or clear statistical outliers as previously defined. In this study, most dates so determined are ca. 1000 Ma, which may warrant the use of weighted mean $^{206}Pb/^{238}U$ ages (e.g., Spencer et al., 2016); however, in most cases, this approach results in similar ages but with greater uncertainty, possibly as a result of elemental fractionation during LA-ICP-MS analysis. Moreover, the effect of recent Pb loss will not have affected the preferred $^{207}Pb/^{206}Pb$ age. (3) A few samples plot on discordia lines that trend toward an ancient radiogenic Pb-loss event. In these cases, the reported age is derived by regression through the data. (4) Some of the SIMS analyses show a strong correlation between ^{204}Pb-corrected age and common Pb, implying that the ^{204}Pb correction may be imprecise or unreliable. In this case we have anchored a discordia at a $^{207}Pb/^{206}Pb$ composition of 0.93 (equivalent to contemporaneous, regional common Pb) and calculated a lower intercept age, which avoids the need for an explicit measurement of the common Pb composition.

For maximum transparency, Supplemental Item G (footnote 1) presents ages and statistics for all the samples, calculated by different means, including weighted mean and concordia ages. For each sample, we have highlighted the preferred method for the final age, following the criteria outlined here. It is important to note that for a particular sample, the different methods typically yield ages that are within uncertainty of each other, and therefore do not directly influence our interpretations. Furthermore, neither the form nor the application of a common Pb correction results in significant differences to the calculated ages.

All ages and uncertainties are quoted at the 2σ level, and all ellipses shown in figures represent 2σ uncertainties.

SVECONORWEGIAN MAGMATISM

Granitic Magmatism

We have acquired 40 new U-Pb zircon crystallization ages from southwest Norway, approximately one-half from the SMB, one-half from the ferroan hornblende-biotite granites, and a small number from mafic rocks (discussed in more detail in the following). The ages are presented in Table 1, Figure 5, and Supplemental Items B and C (footnote 1), and their locations are shown in Figure 6.

The two oldest SMB granites in this study are both 1057 ± 14 Ma, and the youngest is 1014 ± 8 Ma. The oldest ferroan hornblende-biotite granite is 1000 ± 8 Ma, and overlaps within uncertainty with several of the youngest SMB granites. These new, along with previously published (Coint et al., 2015; Slagstad et al., 2013a), age data show that granitic magmatism was continuous, within the resolution of the data, between ca. 1070 and 920 Ma (Fig. 5). In light of the new data, the conjecture that there was a magmatic hiatus between ca. 1020 and 990 Ma (Coint et al., 2015; Slagstad et al., 2013a) appears to be incorrect (Fig. 5). However, the period appears to represent reduced magmatic activity.

Mafic Magmatism

Although much work has focused on the granitic and anorthositic magmatism in the Sveconorwegian orogen, cryptic mafic magmatism in the lower crust has been invoked as a source of heat and material in several studies (Bybee et al., 2014; Slagstad et al., 2013a; Vander Auwera et al., 2003). However, no mafic magmatism with the same age as the SMB and the high-Al orthopyroxene megacrysts (ca. 1040 Ma) has been identified. Younger mafic magmatism, ca. 930 Ma, in association with the Rogaland Igneous Complex (Vander Auwera et al., 2011) and in the northern part of the Sveconorwegian Province, near Finse, ca. 986 Ma, in association with ferroan hornblende-biotite granites (Jensen and Corfu, 2016) has been documented. The new data presented here (Table 1; Fig. 5; Supplemental Items B, C, D; see footnote 1) show evidence of mafic magmatism ca. 1030 Ma, both east and west of the SMB, as well as intrusive into the SMB ca. 990 Ma, attesting to geographically widespread mafic magmatism during both SMB and ferroan granitic magmatism.

Although mafic rocks related to the SMB and ferroan hornblende-biotite suite are unlikely to be voluminous at the present crustal level, the facts that they are geographically widespread and formed over a long period of time support ideas involving long-lived, mantle-derived heating through underplating and ponding of mafic magmas (e.g., Stephens and Andersson, 2015), discussed in detail in the following.

TABLE 1. SUMMARY OF NEW GEOCHRONOLOGIC DATA FROM IGNEOUS AND HIGH-GRADE ROCKS IN THE SOUTHWEST SVECONORWEGIAN PROVINCE

Sample	UTM_E32	UTM_N32	Rock type	Unit	Interpretation	Age	2σ	MSWD	Prob.
Sirdal Magmatic Belt									
ROG033099	301619	6600203	Granite	SMB	Crystallization	1014	8	3.4	0.065
MM057969	396208	6545608	Granite	SMB	Crystallization	1021	21	1.3	0.25
128074	398980	6460798	Granite	SMB	Crystallization	1026	26	2.2	0.029
VAG064841	415289	6486019	Granite	SMB	Crystallization	1029	8	0.0076	0.93
VAG084396	356102	6502652	Granitic gneiss	SMB	Cores, protolith	1030	15	1.6	0.069
VAG084399	359380	6504171	Granite	SMB	Crystallization	1037	7	0.0105	0.92
NC84327	396281	6489549	Granite	SMB	Crystallization	1039	17	0.66	0.71
128068	400453	6465027	Granite	SMB	Crystallization	1042	14	1.00	0.44
VAG098769	390185	6489735	Granite	SMB	Crystallization	1044	23	2.3	0.008
NC84331	394968	6485976	Granite	SMB	Crystallization	1047	18	0.15	0.997
128070	401687	6465410	Granite	SMB	Crystallization	1050	31	2.9	0.005
073168	396521	6472930	Granite	SMB	Crystallization	1051	13	0.41	0.95
073199	396456	6547682	Granite	SMB	Crystallization	1051	14	0.24	0.992
VAG098886	389446	6506651	Granite	SMB	Crystallization	1054	23	1.04	0.40
NC84341	397397	6485014	Granite	SMB	Crystallization	1055	9	0.33	0.57
073173	398432	6480150	Granite	SMB	Crystallization	1057	14	0.86	0.56
VAG098765	390410	6492381	Granite	SMB	Crystallization	1057	14	1.01	0.43
Ferroan hornblende-biotite granite suite, fayalite granite and HAOM									
VAG098766	392699	6494486	Granite	Svofjell	Crystallization	922	7	8.1	0.004
ROG147 LA-ICP-MS	314651	6510490	Fayalite granite	Sjelset	Crystallization	926	6	0.29	0.996
ROG641	330123	6508272	Granite	n.a.	Crystallization	927	4	1.11	0.35
ROG151 SIMS	313768	6516623	Fayalite granite	Sjelset	Crystallization	931	9	3.1	0.077
OPX033146 LA-ICP-MS	315757	6489628	Anorthosite, HAOM	Eg.–Og	Zircon crystallization	932	6	0.36	0.994
ROG147 SIMS	314651	6510490	Fayalite granite	Sjelset	Crystallization	935	8	1.3	0.16
OPX033146 SIMS	315757	6489628	Anorthosite, HAOM	Eg.–Og	Zircon crystallization	941	14	1.8	0.057
073172	391971	6477720	Granite	n.a.	Crystallization	941	6	2.3	0.13
ROG076444	384563	6482786	Granite	n.a.	Crystallization	944	6	1.8	0.18
073158	391041	6446995	Granite	Kleivan	Crystallization	945	19	1.07	0.38
073189	457526	6481977	Granite	Herefoss	Crystallization	949	6	0.27	0.60
073152	364170	6450412	Charnockite	Farsund	Crystallization	950	9	0.71	0.84
073169	395016	6467770	Granite	n.a.	Crystallization	951	7	1.14	0.28
073191	471555	6534121	Granite	Treungen	Crystallization	952	5	2.6	0.11
073186	477173	6469118	Granite	Grimstad	Crystallization	957	12	1.3	0.16
073198	396593	6547802	Granite	Rustfjellet	Crystallization	970	7	5.3	0.0221
073194	415673	6562055	Granite	Valle	Crystallization	972	9	0.6	0.90
073155	374387	6441896	Granite	Lyngdal	Crystallization	984	10	1.3	0.12
073200	406653	6498944	Granite	n.a.	Crystallization	1000	8	3.2	0.072
Mafic magmatic rocks									
IH128058	397261	6459640	Diorite	Kv–Kons	Crystallization	990	12	0.47	0.92
FLAAT089953	434362	6495898	Diorite	Flåt diorite	Crystallization	1025	13	0.58	0.84
VAG084397	357625	6503463	Meta-quartz diorite	Gya qtz-di	Combined rims/cores	1031	5	0.104	0.75

(*continued*)

TABLE 1. SUMMARY OF NEW GEOCHRONOLOGIC DATA FROM IGNEOUS AND HIGH-GRADE ROCKS IN THE SOUTHWEST SVECONORWEGIAN PROVINCE (*continued*)

Sample	UTM_E32	UTM_N32	Rock type	Unit	Interpretation	Age	2σ	MSWD	Prob.
High-grade rocks outside core area									
ROG646L	312996	6529819	Leucosome	Migm. ogn	Cores, protolith	1483	39	1.9	0.11
	312996	6529819	Leucosome	Migm. ogn	Rims 1	1277	23	0.44	0.64
	312996	6529819	Leucosome	Migm. ogn	Rims 2, metamorph.	1014	14	1.2	0.25
IH128061	395379	6459660	Metapelite	Gyad. paragn.	Sveconor. rims, metamorphosed	1025	10	0.27	0.60
VAG084396	356102	6502652	Granitic gneiss	SMB	Rims, metamorphosed	1027	6	4.2	0.039
VAG084397	357625	6503463	Meta-quartz diorite	Gya qtz-di	Combined rims and cores	1031	5	0.104	0.75
ROG092326	334213	6532806	Metapelite	Gyad. paragn.	Rims, metamorphosed	1039	24	1.4	0.20
IH128065B	403000	6461183	Migmatitic orthogneiss, leucosome	Migm. ogn	Migmatization	1043	8	1.6	0.2
NC080843	377725	6498443	Hornblende-biotite gneiss	SMB xenolith	Rims, metamorphosed	1049	19	0.35	0.84
IH128072	402995	6461179	Migmatitic orthogneiss	Migm. ogn	Rims, metamorphosed	1065	23	0.38	0.54
High-grade rocks in core area									
ROG640	330427	6508051	Oxide-rich rock	Faurefj. m.s.	Long-lived metamorphism	1080–920			
ROG642	330139	6508171	Oxide-rich rock	Faurefj. m.s.	Long-lived metamorphism	1080–920			
ROG033048	332476	6503968	Osumilite-bearing gneiss	Gyad. paragn.	Long-lived metamorphism	1130–950			
MM01349	321646	6516751	Metapelite	Gyad. paragn.	Long-lived metamorphism	1050–940			
MM01348	327364	6509224	Calc-silicate gneiss	Faurefj. m.s.	Long-lived metamorphism	1060–960			
ROG639	330427	6508051	Quartzite	Faurefj. m.s.	Detrital cores	1340–1930			
	330427	6508051	Quartzite	Faurefj. m.s.	High-grade metamorphism	938	6	0.66	0.42
Granitoid gneisses with 1.20–1.27 Ga protolith ages									
MM01341	323579	6519386	Granitoid gneiss		Crystallization	1227	11	5.3	
	323579	6519386	Granitoid gneiss		Metamorphism	1012	11	3.6	
Ro-98B-2001	342200	6517000	Granitoid gneiss		Crystallization	1233	42	3.1	
073196	410205	6597419	Granitoid gneiss		Crystallization	1267	8	1.09	
MM02278	328646	6526886	Granitoid gneiss		Crystallization	1200	21	1.05	
VAG064842	423498	6489068	Granitoid gneiss		Crystallization	1248	9	0.82	
VAG064843	422876	6490270	Granitoid gneiss		Crystallization	1246	9	0.22	
NC084308	386594	6486093	Granitoid gneiss		Crystallization	1243	11	2.1	
	386594	6486093	Granitoid gneiss		Metamorphism	1019	11	1.11	
MM026214	316049	6508373	Granitoid gneiss		Crystallization	1233	10	3.3	

Note: Some samples are listed twice where they provide relevant information on both magmatism and metamorphism. Abbreviations: MSWD—mean square of weighted deviates; Prob.—probability; SMB—Sirdal Magmatic Belt; HAOM—high-Al orthopyroxene megacrysts; Eg.–Og—Egersund–Ogna anorthosite; Kv–Kons—Kvås–Konsmo diorite; Gya qtz-di—Gyadalen quartz diorite; Migm. ogn—migmatitic orthogneiss with inferred ca. 1.5 Ga igneous protolith; Gyad. paragn.—Gyadalen paragneiss; Faurefj. m.s.—Faurfjell metasediments; n.a.—no given name; Sveconor.—Sveconorwegian.

SVECONORWEGIAN HIGH-GRADE METAMORPHISM

Analyses from 15 samples of high-grade rocks have been carried out to constrain the timing and duration of high-grade metamorphism in the southwest Sveconorwegian Province. The results are summarized in Table 1, described on a sample-by-sample basis in Supplemental Item B, and presented in Supplemental Items D and E (footnote 1).

Samples outside the core area, defined in Figure 2, yield ages of metamorphism between ca. 1070 and 1015 Ma, with peaks at ca. 1050 and 1035 Ma, coeval with the peaks in the SMB age dataset. The samples within the core area yield apparently continuous ages between ca. 1080 and 920 Ma, with a peak at ca. 1000 Ma, coincident with the lull in magmatic activity.

Calculated Ti-in-zircon temperatures for two samples within the core area are between 740 and 780 °C for most of the recorded time interval, but increase to ~800–840 °C from ca. 970 Ma onward. In contrast, samples from outside the core area yield calculated Ti-in-zircon temperatures between 700 and 780 °C, similar to the pre–970 Ma temperatures from within the core area.

REGIONAL GEOPHYSICAL DATA

The ferroan granites are typically thought of as isolated bodies scattered throughout the orogen, although inferred by some to define an approximately north-south trend (e.g., Sigmond, 1985; Vander Auwera et al., 2003). The SMB granites generally have lower magnetic susceptibilities than the ferroan gran-

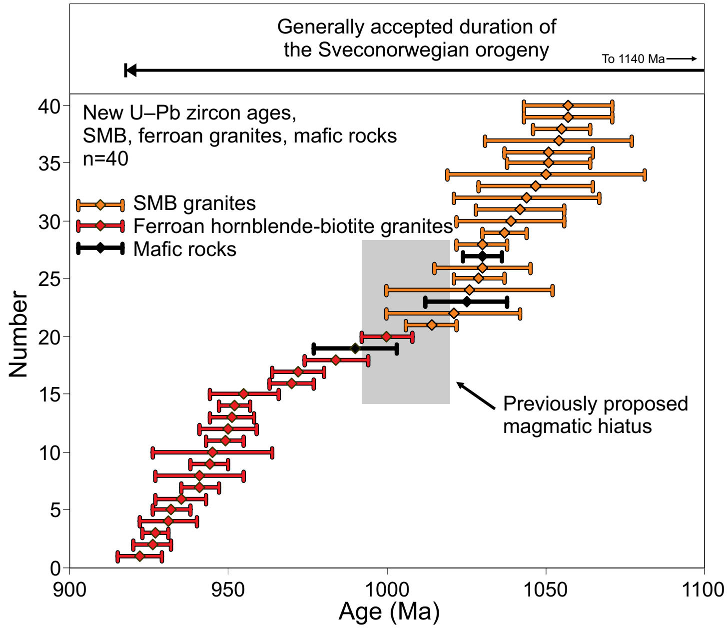


Figure 5. Plot of 40 new crystallization ages from the Sirdal Magmatic Belt (SMB) and ferroan hornblende-biotite granite suite, presented here. The ages range from 1060 to 920 Ma, and although there is a lower frequency of ages at ca. 1000 Ma, there is no clear gap indicating a truly amagmatic period between the end of SMB magmatism and the onset of ferroan hornblende-biotite granite magmatism. The bar on top shows the generally accepted duration of Sveconorwegian orogenic processes, from 1140 to 920 Ma, as discussed in the text.

ites, and the aeromagnetic data (Fig. 6) seem to suggest that the real volume of ferroan granite is much higher than indicated from the geological maps, and that some of these granites may, at least partially, be concealed at shallow depths in the crust. Figure 6 presents an interpretation of the aeromagnetic data, with a projection onto the surface of the inferred subsurface extents of individual bodies of the high-magnetic ferroan granites. We emphasize that this interpretation does not involve any geophysical modeling, but is a purely visual exercise, and is made difficult by the fact that the SMB granites locally also have rather high magnetic susceptibilities, approaching those of the ferroan granites. In general, however, the SMB granites yield lower magnetic anomalies, with a texture characterized by a distinct north-south–oriented banding, changing to a more east-west direction in the northern part of the SMB. This banding typically corresponds to xenolith-rich zones in the SMB that separate sheet-like intrusions of SMB granite with different magnetic properties (Coint et al., 2015; Stormoen, 2015). The interpretation of the aeromagnetic data in Figure 6 shows that many of the bodies of ferroan granite may be interconnected, and, when coupled with the new geochronological data indicating their range in crystallization ages, suggests that they may define granitic batholiths in their own right. It is also clear from the map that rather than forming elongate, sheet-like bodies as observed in the SMB, the ferroan granite intrusions define nearly equidimensional bodies, suggesting that they were emplaced in a different stress regime.

REGIONAL STRUCTURAL DATA

The southwestern parts of the Sveconorwegian orogen have not been subjected to detailed structural study and analysis, in contrast to the eastern and central parts of the orogen (cf. Henderson and Ihlen, 2004; Viola et al., 2011). An exception is the work around Knaben (Stormoen, 2015), in the central parts of the SMB (Fig. 2). However, regional mapping by NGU over the past ~10 yr has amassed abundant structural and map data that allow us to make some first-order interpretations regarding the overall structure of the Sveconorwegian Province in southwest Norway.

A major difference in structural style and metamorphic grade is observed between the high-grade, strongly deformed gneisses in the UHT core area and the little-deformed, nonmetamorphic SMB to the east and north, and the Rogaland Igneous Complex to the southwest (Fig. 2). Although the margins of the Rogaland Igneous Complex tend to be strongly deformed, with host-rock fabrics generally parallel to the shape of the intrusive bodies, this is likely to be an effect of emplacement rather than post-intrusion tectonic deformation (e.g., Barnichon et al., 1999; Bolle et al., 2002).

Figures 7 and 8 present detailed maps and ~550 foliation measurements from the central (Knaben area) and northern (mouth of Lysefjorden) parts of the SMB. The foliations include both tectonic foliations and foliations interpreted to represent magma flow, in order to provide an impression of the overall geometry of the region. In both areas, the lithological makeup of the SMB is characterized by screens and xenolith-rich zones (Coint et al., 2015) surrounded and intruded by SMB granites. In the Knaben area, these sheet-like zones consistently dip ~30° to the east, whereas they are subhorizontal farther north, around Lysefjorden (Fig. 8). The overall structure of the SMB therefore appears to outline the flanks of a large dome-like structure with the high-grade metamorphic rocks and Rogaland Igneous Complex in the core.

DISCUSSION

Magmatic and Metamorphic Evidence of a Long-Lived, Evolving Continental-Margin Arc

Although much work remains to be done to fully elucidate the orogenic evolution of the western part of the Sveconorwegian orogen, the currently available data allow some primary interpretations. Table 1 summarizes all the new U-Pb zircon data presented in this study, organized into: (1) SMB granites, (2) ferroan hornblende-biotite granites, fayalite granite, and zircon ages associated with high-Al orthopyroxene megacrysts, (3) mafic magmatic rocks, (4) high-grade rocks outside the core area, and (5) high-grade rocks within

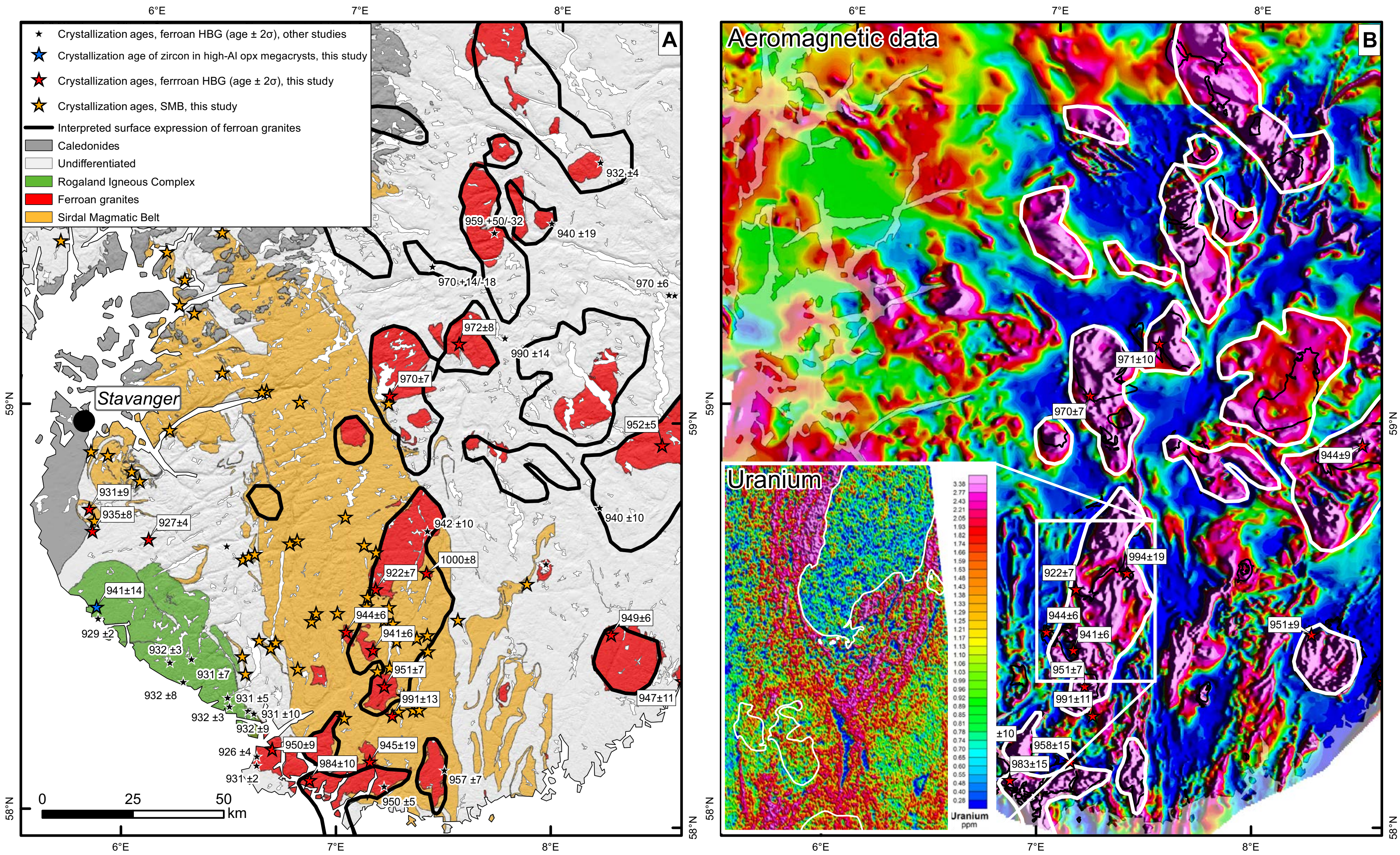


Figure 6. (A) Simplified geological map of southwest Norway, highlighting the Sveconorwegian magmatic rocks. Also shown are the sample locations of Sirdal Magmatic Belt (SMB), ferroan hornblende-biotite granite (HBG), and high-Al orthopyroxene (opx) megacryst samples investigated here, along with previously published age data from these rocks, taken from an updated version of the database presented by Bingen and Solli (2009). The ages of particular SMB samples are not shown. The black polygons outline the anomalies. (B) Aeromagnetic data from the same region. The positive magnetic anomalies generally correspond to ferroan hornblende-biotite granites, and appear to suggest that these bodies are much larger than indicated on the geological map. The white polygons outline the anomalies. Inset shows uranium concentrations from a recent airborne radiometric survey (Ofstad, 2015).

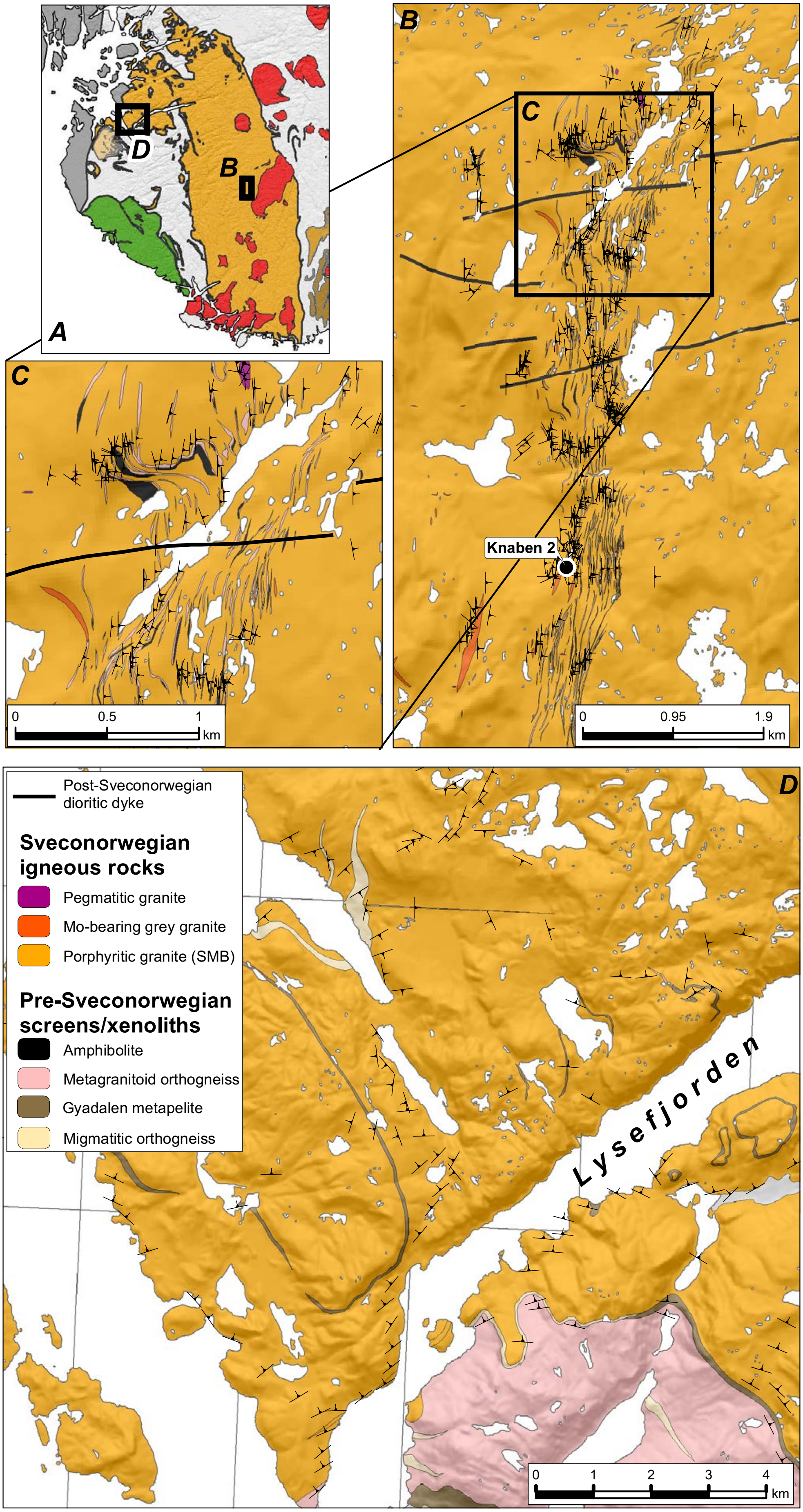


Figure 7. Detailed geological maps with structural data of the Knaben area (Stormoen, 2015) and the Lysefjorden area (Marker and Slagstad, 2018c; Marker et al., 2012). SMB—Sirdal Magmatic Belt.

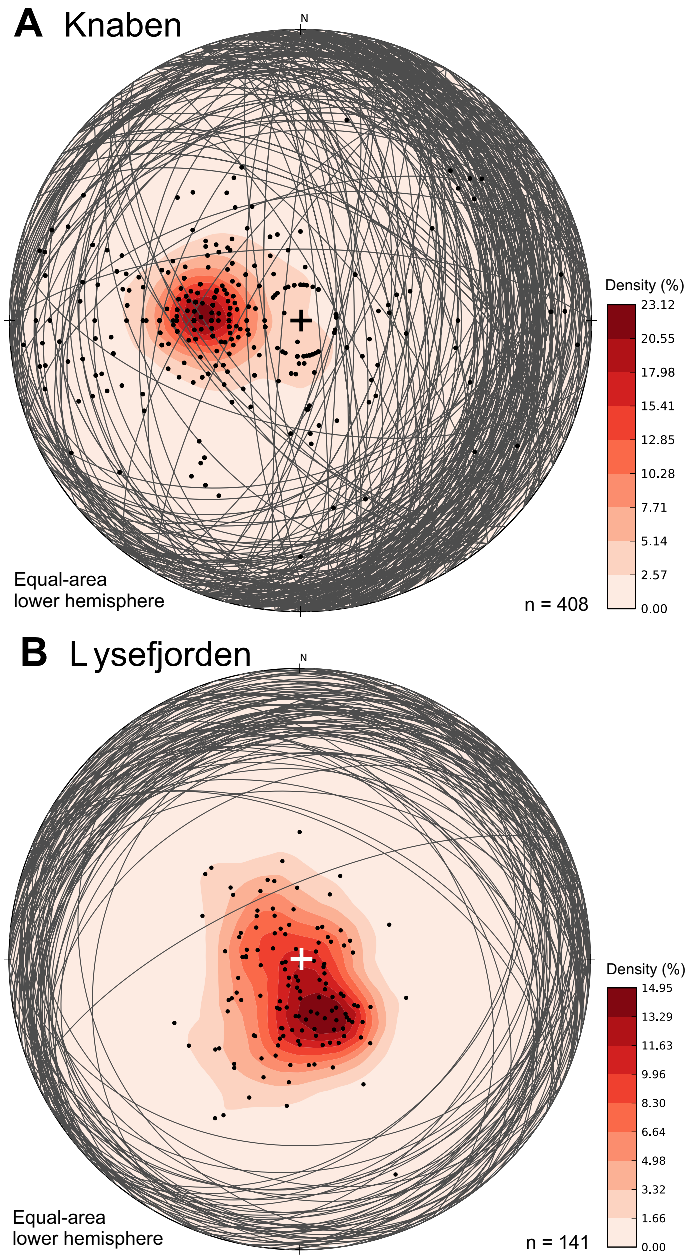


Figure 8. Lower-hemisphere stereoplots presenting foliation measurements (tectonic and magmatic) from the areas shown in Figure 7.

the core area, as defined in Figure 2. Figure 9 summarizes relevant published magmatic crystallization ages, including those presented here, grouped into SMB and ferroan hornblende-biotite granite based on age and composition. The figure also presents available age data from mafic magmatic rocks and the Rogaland Igneous Complex. Figure 9 also presents a summary of the new metamorphic age data presented here, and these data are discussed in relation to previously published data in the following. The figure also includes a generalized tectonic illustration of the main stages of arc evolution, discussed in detail in the following.

The SMB started forming ca. 1070 Ma, and possibly somewhat earlier, although evidence of this early onset is scant. There are peaks in the geochronologic dataset ca. 1050 and 1035 Ma, but it is not clear if they signify more voluminous magmatism at these times. The youngest SMB age is 1014 ± 8 Ma, signaling the end of SMB magmatism. The onset of ferroan hornblende-biotite granite magmatism was at 1000 ± 8 Ma, overlapping with the youngest SMB magmatism. Some of us (Slagstad et al., 2013a; Coint et al., 2015) argued for a magmatic hiatus between ca. 1020 and 990 Ma; however, given the overlapping ages for cessation of SMB and onset of ferroan hornblende-biotite granite magmatism, such a hiatus may not exist. The period does, however, appear to represent a period of less magmatic activity, at least as recorded at the current crustal level. Ferroan hornblende-biotite granite magmatism lasted until ca. 920 Ma, overlapping in age with the emplacement of the Rogaland Igneous Complex.

In our study we recorded mafic magmatism at 1031 ± 5, 1025 ± 13 and 990 ± 12 Ma; along with earlier studies (Jensen and Corfu, 2016; Vander Auwera et al., 2011; Wiest et al., 2018), this suggests geographically widespread mafic magmatism during Sveconorwegian orogenesis.

The new U-Pb data from metamorphic zircon show unequivocally that the region underwent continuous (within resolution of the data) growth and regrowth of zircon, probably related to continuous or semicontinuous high-grade metamorphism, between ca. 1080 and 920 Ma. The high-grade rocks can be divided into two groups based on geography and age: one consisting of high-grade xenoliths within the SMB and host rocks close to the SMB, and another within the core area, closer to the Rogaland Igneous Complex (outlined in Fig. 2). The geographic extent of the core area is loosely defined to encompass samples that contain evidence of long-lived, semicontinuous growth and regrowth of zircon between ca. 1080 and 920 Ma, covering the entire age range of granitic magmatism in the southwest Sveconorwegian Province. A peak in zircon growth or recrystallization at ca. 1000 Ma is recorded, coinciding in time with the apparent lull in magmatic activity. Some samples preserve ages spanning the entire 160 m.y. age range, whereas others only display part of the range. The high-grade rocks outside the core zone display zircon U-Pb geochronological evidence of high-grade metamorphism between ca. 1070 and 1015 Ma, with peaks at ca. 1050 and 1035 Ma, coinciding in duration and intensity (similar age peaks) with the SMB. Calculated Ti-in-zircon temperatures from two samples from the core area show an increase in temperature starting ca. 970 Ma; no such variation is observed for samples outside the core area.

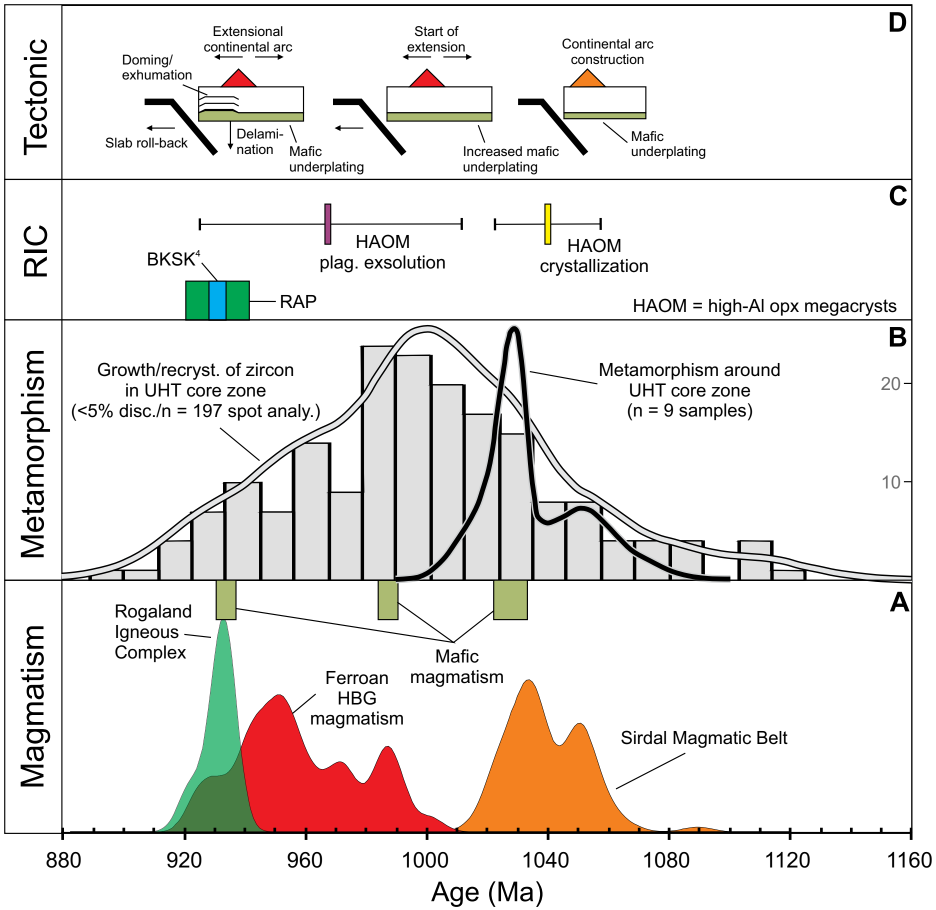


Figure 9. Plots summarizing available geochronological data. (A) Data pertaining to Sveconorwegian magmatism. HBG—hornblende-biotite granite. (B) Data pertaining to high-grade metamorphism in core area. UHT—ultrahigh temperature; recryst.—recrystallization; disc.—discordance; analy.—analyses. (C) Data pertaining to Rogaland Igneous Complex (RIC) magmatism (plag—plagioclase; opx—orthopyroxene). HAOM—high-alumina orthopyroxene megacrysts; BKSK—Bjerkreim-Sokndal layered intrusion; RAP—Rogaland Anorthosite Province. (D) Sketches illustrating the interpreted tectonic evolution of the southwest Sveconorwegian Province.

We discuss a tectonic model that is interpreted based on, and consistent with, the available data presented herein. The tectonic model is generally similar to the accretionary model presented in Slagstad et al. (2017, 2013a) and Coint et al. (2015), and the inferred geodynamic processes are fundamentally different from those of the continent-continent collisional model previously advocated for the Sveconorwegian orogeny (Bingen et al., 2008b; Möller et al., 2015).

Establishment of a Magmatic Arc, 1070–1010 Ma

Formation of the SMB in a continental-margin magmatic arc may have started as early as ca. 1070 Ma, with apparently continuous granitic magmatism until ca. 1010 Ma. SMB magmatism was associated with high-grade metamorphism, including migmatization (Coint et al., 2015; this study), with apparent peaks in magmatic activity during the middle to later stages of its history, ca. 1050 and 1035 Ma. However, the SMB largely escaped later high-grade overprinting except for its southwest part, closest to the Rogaland Igneous Complex (Coint et al., 2015). In Slagstad et al. (2013a) it was suggested that the heat source for SMB magmatism could be a basaltic underplate, an interpretation corroborated in Bybee et al. (2014), who argued that the high-Al orthopyroxene megacrysts in the Rogaland Igneous Complex crystallized from underplated, mantle-derived basaltic magma at 1041 ± 17 Ma. The new geochronological data from the diorite at Flåt (1025 ± 13 Ma) and the high-grade diorite in Gyadalen (1031 ± 5 Ma) provide the first direct evidence of syn-SMB mafic magmatism, and give further credence to this hypothesis.

Changing Arc, ca. 1010–1000 Ma

Although most earlier workers have argued for a long-lived metamorphic event (termed M1) between ca. 1035 and 970 Ma (Bingen et al., 2008b; Möller et al., 2002), a change in tectonic regime is clearly recorded at ca. 1010 Ma by the cessation of SMB magmatism, interpreted to reflect a shallowing of the subducting slab (Slagstad et al., 2013a). Such a shallowing is likely to result in compression and crustal thickening (Collins, 2002), although, as discussed herein, there is no evidence to suggest significant crustal thickening in southwest Norway at that time. The high-grade rocks in the core area record continuous growth of zircon at that time, and there were arguments for different metamorphic peaks: 1006 ± 11 Ma (Möller et al., 2002), 1035 ± 9 Ma (Tomkins et al., 2005), 1010 ± 7 and 1006 ± 4 Ma (Drüppel et al., 2013), and 1032 ± 5, 1002 ± 7, and 990 ± 8 Ma (Bingen et al., 2008a). Figure 9B shows a probability plot of our zircon U-Pb geochronologic data from the core area; 197 spot analyses, all <5% discordant, reveal a range of ages from ca. 1100 to 900 Ma, with a clear peak ca. 1000 Ma. Although we cannot exclude the possibility of inadvertent sampling bias, we note that the samples represent different lithologies from different parts of the core area, and that the ca. 1000 Ma peak coincides closely with the majority of previously published U-Pb zircon ages from the area (Bingen et al., 2008a; Drüppel et al., 2013; Möller et al., 2002). We therefore argue that this peak bears some geologic significance, outlined in the following. The pre–1020 Ma dates probably record the same event as observed in high-grade xenoliths in the SMB and its immediate host rock outside the core area, whereas the peak at 1000 Ma is significantly younger, coinciding with the transition between SMB and ferroan hornblende-biotite granite magmatism. We interpret this peak in metamorphic ages ca. 1000 Ma to reflect increased dissolution and growth of metamorphic zircon. It is somewhat surprising that this event is not recorded in our zircon data from the SMB, suggesting that it remained at relatively high crustal levels during this event. The significance of these features is discussed in more detail in the following.

Renewed Ferroan Magmatism and Continued High-Grade Metamorphism, 1000–920 Ma

The earliest trace of renewed magmatism following the magmatic minimum is at 1000 ± 8 Ma, marking the onset of emplacement of the ferroan hornblende-biotite granite suite, which appears to have lasted until ca. 920 Ma and affected the entire orogen apart from the easternmost Eastern Segment (Vander Auwera et al., 2003, 2011). The ferroan, A-type composition of the hornblende-biotite granite suite has led most to infer formation in an extensional setting, related to gravitationally driven extension of the orogen (e.g., Bingen et al., 2006; Vander Auwera et al., 2011); however, as discussed in Slagstad et al. (2013a), such a driving mechanism is difficult to reconcile with the long duration of magmatism and the very large time gap (>100 m.y.) between high-grade metamorphism, which would reflect crustal thickening, and purported extension in parts of the orogen (e.g., Bamble sector, Fig. 1). The lack of high-pressure metamorphism and major compressional structures in western and central parts of the orogen is also inherently problematic, because crustal and lithospheric thickening is a key requirement in such a model. In Slagstad et al. (2013a) it was suggested instead that the ferroan hornblende-biotite granite suite formed during continental backarc extension as a result of slab steepening and rollback. This interpretation posits that all the granites currently labeled ferroan hornblende-biotite granite (Fig. 1) formed the same way, which is not necessarily correct. For example, the easternmost granites formed quite late, ca. 930–920 Ma (Bingen et al., 2008b; Eliasson and Schöberg, 1991), and are strongly enriched in heat-producing elements (Slagstad, 2008). In contrast, the granites intruding the SMB, much farther west in the orogen, are comparatively low in heat-producing elements (Fig. 6B). Other elements also show distinct differences between granites east and west in the orogen (our data), and it is possible that there is more than one group of ferroan hornblende-biotite granite, as suggested by Andersen et al. (2002) and Brueckner (2009). The eastern granites are neither ferroan nor do they contain hornblende, clearly warranting a geological distinction.

Brueckner (2009) suggested that the late orogenic magmatism in the Sveconorwegian Province was caused by partial melting of subducted continental crust (crustal tongues). In this model, the easternmost, high heat–producing granites formed by partial melting of upper continental crust, which was tectonically peeled off during subduction, leaving behind a more mafic lower crust that could melt to form both the westernmost ferroan hornblende-biotite granites and the Rogaland Anorthosite Province. Although this interpretation needs to be tested chemically and isotopically, the general idea is consistent with the suggestion in Slagstad et al. (2017) that the Sveconorwegian orogen formed by a series of accretionary events between ca. 1140 and 920 Ma. These events may have variably modified and refertilized the lower crust and upper mantle (Slagstad et al., 2018) and given rise to magmas of different composition.

On the basis of available data, it appears likely that there is more than one group of ferroan hornblende-biotite granite, and that they may reflect variations in either source or tectonic setting from east to west and/or through time. Thus far, most work has focused on the ferroan hornblende-biotite granites in western parts of the orogen, where existing data suggest isotopically similar sources to the SMB (Vander Auwera et al., 2003, 2011). Our tentative interpretation is that the ferroan hornblende-biotite granites formed by high-temperature melting of the same lower crustal source as the SMB, with the added heat derived from basaltic underplating during extension of the arc (cf. Landenberger and Collins, 1996; Slagstad et al., 2004; Zhao et al., 2016). This differs somewhat from the interpretation proposed in Slagstad et al. (2013a), in that it argues that the position of the arc remained largely unchanged since formation of the SMB. This modified interpretation makes the westernmost ferroan hornblende-biotite granites arc related, rather than backarc related, although the overall tectonic setting is similar. Formation in an arc is also supported by aeromagnetic data, which show that the ferroan hornblende-biotite granites constitute a major, linear granite batholith, more similar to the SMB than previously recognized. To some extent, the distinction between arc and backarc in this respect is semantic, as active magmatic arcs are expected to migrate back (extension) and forth (compression) over time. If the magmatic record from southwest Norway is as complete as it appears, it is possible that further work, involving high-precision U-Pb geochronology, can identify and elucidate such a cyclic evolution.

We are currently investigating the ferroan hornblende-biotite granites to see if there are temporal and/or geographical trends in chemical and isotopic composition that can elucidate and distinguish the different interpretations. We have already mentioned the east-to-west compositional variations, and when excluding the easternmost ferroan granites (Idefjord-Bohus and Flå granites), the remaining granites cover a much narrower age range than the generally accepted 990–920 Ma period. Disregarding ages from the Rogaland Igneous Complex and associated charnockites and fayalite granites, the ferroan granites mainly yield ages between 970 and 940 Ma (Fig. 9A), which we interpret to represent peak activity in a reestablished active continental margin arc. By this time, we infer that the entire orogen was in extension, with mafic dikes intruding the eastern parts of the orogen and its foreland (Söderlund et al., 2005), and development of major extensional shear zones (Viola et al., 2011). Although crustal thickening at 990–970 Ma in the eastern part of the orogen may have resulted in gravitationally driven extension a few tens of millions of years later (Viola et al., 2011), this process is unlikely to have caused extension and mafic magmatism in the orogenic foreland that did not undergo crustal thickening. This process is not likely to have caused extension in more western parts of the orogen, where thickening (as inferred from ages of high-grade metamorphism) in many cases preceded extension by nearly 100 m.y. We therefore suggest that extension driven by slab rollback in a retreating subduction zone is the most likely candidate, and moreover provides a plausible scenario in which the lower arc crust beneath the SMB, depleted and dehydrated by the extraction of the SMB granitic melts, may have melted anew to give rise to granites of more ferroan compositions. Recent work suggests that this long-lived, overall extensional phase may have given way to intermittent compression (Bolle et al., 2018), consistent with our proposed accretionary setting.

Anorthosite Emplacement Related to Doming in an Extending Continental Arc

As discussed herein, the regional structural data from the southwest Sveconorwegian Province suggest that the SMB east and north of the high-grade rocks outline a large, dome-like structure (Figs. 7 and 8). Pressure estimates from close to the Rogaland Igneous Complex indicate contact metamorphism at ~5 kbar, with slightly lower pressure at 3–4 kbar ~10 km away from the contact (Blereau et al., 2017), similar to the pressure of emplacement of the SMB (Coint et al., 2015). The pressure estimates for contact metamorphism are consistent with emplacement or exhumation of the Rogaland Igneous Complex in a broad, dome-like structure, as suggested by the general geometry of the area, which also constrains the timing of doming to be coeval with or after anorthosite emplacement. U-Pb zircon ages from cordierite coronas around garnet suggest decompression at 955 ± 8 Ma (Tomkins et al., 2005), within uncertainty of the 968 ± 43 Ma age constraining decompression exsolution of plagioclase from high-Al orthopyroxene megacrysts in the Rogaland Anorthosite Province, interpreted in Bybee et al. (2014) to reflect ascent and emplacement of the anorthosite, and the 941 ± 14 Ma SHRIMP (sensitive high-resolution ion microprobe) age obtained here for zircon crystallization in high-Al orthopyroxene megacrysts. Although the significance of the zircons associated with the high-Al orthopyroxene megacrysts is unclear, the spatial relationship with the high-Al orthopyroxene megacrysts suggests that they did not grow directly from the anorthositic melt (or, more appropriately, anorthositic mush), or the parental melt to the anorthosites, but are somehow related to the presence of the high-Al orthopyroxene megacrysts. At present, we therefore favor the hypothesis by Schärer et al. (1996) that the zircons most likely grew from small amounts of evolved melt trapped between high-Al orthopyroxene megacryst aggregates. Although this evolved, residual melt would initially have been homogeneously distributed in the crystallizing anorthosite mush, the presence of the rigid high-Al orthopyroxene megacrysts might have created local low-pressure sites into which the residual melt migrated and concentrated, as has been observed in some migmatites (e.g., Slagstad et al., 2015). We therefore interpret the zircons to have grown from the last vestiges of melt in the anorthosite. An alternative hypothesis is that these zircon-forming vestiges represent partial melting related to heating from later, adjacent anorthosite bodies, which implies an incremental emplacement history for the anorthosites. Field relationships and mapping, for example, show that ductile fabrics related to emplacement of the Egersund-Ogna anorthosite are cut by the later Helleren anorthosite (Marker et al., 2003), suggesting that the system had time to cool between emplacement of the two anorthosite bodies.

The ages of zircons from the Egersund-Ogna massif range from 930 Ma (Schärer et al., 1996) to 941 Ma (this study), but overlap within uncertainty. Blereau et al. (2017) suggested that melts may have been present in the host rocks to the Rogaland Igneous Complex for as long as 100 m.y., consistent with the zircon data presented here. It is therefore possible that the range of ages obtained from the Egersund-Ogna massif reflects crystallization of residual melts in the anorthosite over a long period of time, or even that the younger ages reflect remelting during a late-stage thermal event, such as emplacement of the Bjerkreim-Sokndal layered intrusion. Also noteworthy in this regard is the observed increase in Ti-in-zircon temperature in metamorphic zircon in the Rogaland Igneous Complex contact aureole ca. 970 Ma, well before the inferred emplacement age of the Rogaland Igneous Complex. This observation suggests that if UHT metamorphism was related to emplacement of the Rogaland Igneous Complex, this event took place significantly earlier than 930 Ma. Based on regional structural, metamorphic, and geochronological data, we interpret the anorthosites to have been emplaced ca. 950–930 Ma, i.e., over a protracted period of time, during doming of the crust. As discussed in Slagstad et al. (2017, 2013a), the entire Sveconorwegian orogen appears to have been in extension at that time; this offers a plausible stress regime for doming (e.g., Yin, 2004). At present, we are unable to document the existence of large-scale extensional structures that one would expect to accompany such doming. We note, however, the existence of a several-hundred-meter-wide zone of east-west–trending, high-strain rocks with extensional structures north of Nedrebø (Marker and Slagstad, 2018a, 2018b) that appears to separate high-grade rocks, with an apparently simple metamorphic history on the flank of the purported dome, from high-grade rocks that preserve evidence of long-lived metamorphic overprinting toward the core of the dome (core area). This high-strain zone may therefore represent part of an otherwise cryptic tectonic contact within the high-grade area, further disrupting an already complex geology. However, further structural and geochronological work is needed to prove the existence of such a structure. A ca. 950 Ma age of extension and doming suggests that these processes were coeval with voluminous ferroan magmatism in the region, and possibly genetically related.

Calculated Ti-in-zircon temperatures for samples ROG092326 and IH128061, both from outside the core area, are similar to those from ROG642 and ROG033048 (within the core area) prior to ca. 1000 Ma, corroborating the idea that these extreme crustal conditions are only recorded within the core area, and that the present-day distribution of high-grade, low-grade, and lower grade rocks in the region results from juxtaposition of different crustal levels during doming.

The doming suggested here is different from that proposed by Bingen et al. (2006) in some respects. Our proposed dome is significantly smaller, limited to the southwest part of the Rogaland-Vest-Agder sector, and delineated by nonmetamorphosed SMB rocks to the east and north that show no evidence of high-grade metamorphism after 1035 Ma. Furthermore, we argue that the doming is related to externally driven orogenic extension rather than postorogenic collapse, an interpretation made less likely by the lack of evidence arguing for significant crustal thickening, discussed herein. It follows from our interpretation that the presence of Sveconorwegian UHT rocks and anorthosite may be much more widespread than hitherto known, and that their confinement to southwest Norway is related to exposure limited to this relatively late orogenic dome. Anorthosite and jotunite-mangerite, dated at 965 and 951 Ma, respectively (Lundmark and Corfu, 2008; Lundmark et al., 2007), in the Caledonian Lindås and Jotun nappes north of our study area are associated with rocks that preserve evidence of high-grade metamorphism from before 954 Ma to 930 Ma (Lundmark et al., 2007). To the east of the study area, geographically widespread pegmatites were emplaced as late as ca. 910 Ma (Seydoux-Guillaume et al., 2012), possibly as a result of ongoing mafic underplating (Müller et al., 2015). We interpret these data in terms of the widespread and protracted nature of arc and/or backarc processes (e.g., Currie and Hyndman, 2006). The mechanism (e.g., metamorphic core complex; Vanderhaeghe et al., 2003) behind the inferred doming is largely unconstrained, and future investigations, involving structural mapping and thermochronology, are needed to shed light on this issue.

Long-Lived High-Grade Metamorphism in the Lower to Middle Crust of a Magmatic Arc and the Quality of Different Lithological Recorders

If our interpretation that the westernmost part of the Sveconorwegian orogen was an active continental-margin arc between ca. 1070 and 920 Ma is correct, and that the later part of this period was characterized by doming and uplift, it allows a privileged view into the lower to middle crust of a long-lived Mesoproterozoic continental-margin arc. The complexities that such a setting entail are reflected in the variety of interpretations proposed, in particular regarding the timing and conditions of high-grade metamorphism, pressure-temperature paths, and tectonic significance. This variety is to be expected considering the duration of high-grade conditions, and the highly variable ability of different lithologies to record these conditions. For example, the ca. 1.2 Ga granitoid orthogneisses yield only a limited number of early Sveconorwegian (ca. 1030 Ma) metamorphic ages (Table 1; Supplemental Item F [footnote 1]), and reveal little about the duration of Sveconorwegian metamorphism. The ca. 1.5 Ga migmatitic orthogneisses have ca. 1050–1015 Ma leucosomes, but also appear to have been poor recorders of the following 100 m.y. of high-grade metamorphism. The calc-silicate– and oxide-rich rocks of the Faurefjell metasediments and the metapelitic rocks of the Gyadalen paragneisses, however, have recorded virtually the entire metamorphic evolution close to the core of the dome-like structure, although a single sample does not necessarily record the entire history (e.g., Möller et al., 2002). Contrasts in different lithologies' ability to react, and thus record metamorphic events, must therefore be taken into account when assessing the significance of geochronological data.

A prime example of different lithologies' ability to record metamorphic events comes from the Nedrebø locality, where two samples have been investigated (Supplemental Item B; see footnote 1). Sample ROG640 is from an orthopyroxene–spinel–FeTi-oxide gneiss with a granoblastic texture, containing abundant orthopyroxene, plagioclase and opaque minerals, and accessory apatite, zircon, green spinel, and orange-brown biotite. Its peculiar composition, characterized by apparent enrichment in Fe, Ti, P, Al, and some trace elements have led some to suggest a premetamorphic lateritic protolith (Bol et al., 1989). This highly reactive lithology resulted in growth and/or regrowth of zircon, yielding a semicontinuous range of concordant ages between ca. 1080 and 920 Ma. Sample ROG639 is from an ~1.5-m-thick quartzitic layer directly overlying the oxide-rich gneiss, taken <20 m from sample ROG640, and arguably with a similar metamorphic history. In contrast to ROG640, however, the zircons from ROG639 are mainly detrital with apparently only one rather tightly constrained metamorphic age of 938 ± 6 Ma. It is clear that data from only one or the other lithology would result in widely different interpretations regarding metamorphic evolution. Progress in understanding the evolution of areas such as the southwest Sveconorwegian Province therefore requires data from a relatively large number of samples, reflecting different lithologies and compositions.

High-Grade Metamorphism as a Result of Crustal Thickening?

The metamorphic framework outlined here supersedes traditional interpretations in which the metamorphic evolution can be understood within the framework of initial crustal thickening followed by consequent orogenic collapse, and finally contact metamorphism related to emplacement of the Rogaland Igneous Complex. The available metamorphic age data presented here and elsewhere do not allow distinct metamorphic episodes to be identified. Granted, one particular sample may show distinct episodic growth of metamorphic zircon or monazite (e.g., Laurent et al., 2016; Möller et al., 2003); however, when a larger number of samples is considered, these events tend to blur. A major challenge for further work on the metamorphic evolution of the Sveconorwegian orogen in southwest Norway is therefore to distinguish between events recorded by a particular sample that bear a regional, tectonic significance, and events that simply reflect the conduciveness of that sample to react, producing datable minerals, at particular times. The latter may be more closely related to local conditions, such as the availability of fluids (e.g., melt, H_2O, CO_2), than to regional orogenic processes. Laurent et al. (2016) linked monazite age data to sulfate concentrations; this may be one such venue whereby metamorphic ages can be correlated over a larger area.

Long-lived (>100 m.y.) UHT metamorphism has become increasingly recognized as a common feature of Precambrian orogenic terrains (e.g., Korhonen et al., 2013; Walsh et al., 2015), and identifying the mechanisms and tectonic settings responsible for attaining and sustaining such conditions has been a prime target for numerous recent studies. Clark et al. (2011) suggested that long-lived UHT conditions could result from orogenic thickening of highly radioactive crust, and this mechanism was proposed by Drüppel et al. (2013) to explain long-lived UHT metamorphism in southwest Norway. The crust in southwest Norway is, however, not particularly radioactive (Slagstad, 2008; Slagstad et al., 2009a), and there is no geological evidence of significant crustal thickening. For example, the ca. 1040 Ma age for high-alumina orthopyroxene megacrysts in the Rogaland Anorthosite Province (Bybee et al., 2014) constrains crustal thickness to <33 km (i.e., <11 kbar; Charlier et al., 2010) at that time, and no major structures that could accommodate later major crustal thickening have been identified. Granitic magmatism is continuous through the period of high-grade metamorphism, and although more work is needed to determine the exact source or sources of this magmatism, there is no evidence that garnet was ever stable in the inferred lower crustal source (Demaiffe et al., 1990; Slagstad et al., 2013a), limiting crustal thickness to ~10 kbar (Behn and Kelemen, 2006). In Slagstad et al. (2013a, 2013b) it was argued that an apparent magmatic hiatus between ca. 1020 and 990 Ma could be interpreted to reflect flat-slab subduction and compression, which could have triggered crustal thickening and high-grade metamorphism. The new data presented here suggest, however, that there was not a real hiatus, but rather a lull in magmatic activity, concomitant with widespread high-grade metamorphism. This change may be related to a shallowing of the angle of the subducting slab, to a more compressional arc, but there is no evidence that compression resulted in significant crustal thickening. We therefore conclude that long-lived UHT metamorphism in southwest Norway is unlikely to have resulted from abnormal crustal thickening.

SMB High-Grade Metamorphism as a Result of Repeated Basaltic Underplating?

Most studies on the mechanisms of long-lived UHT metamorphism invoke mantle-derived heat in the form of mafic magmatism resulting from extensional and/or compressional tectonics in arc and/or backarc settings (e.g., Brown, 2006; Currie and Hyndman, 2006; Walsh et al., 2015), similar to the tectonic-switching model of Collins (2002). Emerging evidence from the Sveconorwegian Province, including this study and that of Wiest et al. (2018) near Bergen, and Jensen and Corfu (2016) at Finse, show that the Sveconorwegian orogeny was accompanied by geographically widespread mafic magmatism. A marked increase in Ti-in-zircon temperatures ca. 970 Ma in the UHT core area coincides with emplacement of voluminous ferroan granites farther east, consistent with widespread extension and ponding of mafic magma at the base of relatively thin (or at least not significantly thickened) crust. We also note the coincidence in duration of SMB magmatism and high-grade metamorphism outside the core area, between ca. 1070 and 1015 Ma, as well as the coeval peaks in the age datasets for SMB magmatism and high-grade metamorphism. A likely interpretation is that influx of basaltic magma at the base of the crust triggered lower crustal melting and an increased heat flux recorded as a peak in metamorphic activity and zircon crystallization or recrystallization.

A range of possible scenarios exists that could explain the observed ca. 1010–1000 Ma lull in magmatic activity and coeval increase in metamorphic activity. Ridge subduction, for example, will result in a flattening of the subduction zone and decrease in arc magmatic activity, while at the same time increasing crustal heat flow by allowing asthenospheric material to impinge on the lower crust (Li and Li, 2007; Santosh and Kusky, 2010). Flattening from subduction of an oceanic plateau, or simply a change in convergence rate, may achieve the same effect if preceded by extension and basaltic underplating (Collins, 2002), and would be consistent with evidence for significant mafic underplating both before and after ca. 1000 Ma (i.e., tectonic switching). A third possibility is that the lower crustal source of SMB magmatism was becoming infertile by ca. 1010 Ma, explaining the magmatic lull, and that a steepening of the subduction zone ca. 1000 Ma resulted in increased basaltic underplating, allowing the less fertile source to be remelted and the heat flow to increase. This latter model does not require compression.

A long line of evidence from the Sveconorwegian Province points toward protracted accretionary processes, with marked differences in tectonic style in different parts of the orogen (Roberts and Slagstad, 2015; Slagstad et al., 2017). These repeated accretionary events are likely to have had a stop-and-go effect on subduction along the plate margin by affecting the movements of the upper, continental plate, which would have been ideal for the type of extension-compression tectonic regime inferred to be responsible for sustained

UHT conditions. This coupling between seemingly unrelated tectonic events in different parts of the orogen provides a framework in which orogen-scale variations in metamorphic and magmatic evolution can be understood.

In detail, several different interpretations can explain the observed magmatic and metamorphic evolution of southwest Norway, including ridge subduction, tectonic switching, or compositional changes in the lower crust. More work is required to evaluate these and other possibilities, but in all cases mounting evidence suggests that long-lived, repeated basaltic underplating was the main driving force for sustained high-grade conditions and continuous crustal melting and magmatism for ~150 m.y. at the southwest margin of Fennoscandia.

Accretion or Collision; Two Opposing, Mutually Excluding Models or Simply a Question of Semantics?

It was suggested (Slagstad et al., 2013a) that the Sveconorwegian orogeny did not result from continent-continent collision with an unknown landmass to the west, as had been the consensus since the mid-1990s (see discussion in Coint et al., 2015). This consensus view, however, was contrary to that of geologists working in south Norway in the 1970s and 1980s, who viewed the orogen as Cordilleran-type accretionary in nature (Falkum and Petersen, 1980; Torske, 1977). As far as we can tell, the change from accretion to collision as the preferred tectonic setting was based on correlation with the roughly time-correlative collisional Grenville Province in eastern Canada as well as very scant paleomagnetic data suggesting that Amazonia was located outboard of southwest Norway, both lines of argument suggesting that the two continents may have collided (see discussion in Coint et al., 2015). To our knowledge, however, there are no geologic data from the Sveconorwegian Province that strictly require indenter-style collision; although, in fairness, no data strictly ruled out such a model. Progress in science is based on rejecting hypotheses and theories that do not satisfactorily explain currently available data. The accretionary model outlined here can explain all of the currently available data regarding the overall orogenic evolution of the Sveconorwegian Province, whereas the collisional model cannot. Although the accretionary model is likely to be modified as more data become available, the correct scientific approach is to reject the collisional model. We explain where the collisional model fails.

As documented in this and several recent papers, the Sveconorwegian orogenic evolution in the western part of the province was characterized by continuous or semicontinuous granitic and mafic magmatism between ca. 1070 and 920 Ma (Coint et al., 2015; Slagstad et al., 2013a; this work), and coeval and equally long-lived, continuous or semicontinuous HT-UHT metamorphism (Blereau et al., 2017; this work). Overall, Sveconorwegian orogenic processes appear to have been sustained for ~220 m.y. (1140–920 Ma), but not everywhere at once (Slagstad et al., 2017). Collision with a major continental landmass to the west ca. 1140 Ma (Bingen et al., 2008b), with resulting Himalayan-type orogenic processes, cannot explain this long-lived orogenic evolution.

1. Asthenosphere-derived heat, transported to the lower crust via basaltic magmas, is required to explain the recorded UHT conditions and lower crustal melting, expressed as continuous granitic magmatism, in the southwest Sveconorwegian Province. The Himalayas are nearly unique as far as continent-continent collisions go, but would need to continue for another 100 m.y. to match the duration of Sveconorwegian orogenesis observed in the western part of the province. Moreover, if we include the entire orogeny (1140–920 Ma), the Himalayas fall short in duration by ~170 m.y.

2. The 1000–920 Ma ferroan hornblende-biotite granite suite is typically interpreted to have formed during the postcollisional stage of the Sveconorwegian orogeny, driven by underplating of basaltic magma, possibly accompanied by partial melting of the lower crust (Vander Auwera et al., 2003, 2014). Although we agree in general with this petrogenetic interpretation, it is unclear how a postcollisional setting can account for the 80 m.y. duration of magmatism. Delamination of overthickened orogenic crust is one viable option, but is unlikely to cause more than a relatively short burst of magmatism; this process definitely cannot explain how magmatism related to delamination starting at 1000 Ma could have peaked at 930 Ma with emplacement of the Rogaland Igneous Complex. In addition, evidence for significant crustal thickening prior to 1000 Ma, a prerequisite for overthickening and later delamination, is missing in the southwest Sveconorwegian Province. Extension at or behind active continental margins has been proposed in other cases to explain long-lived ferroan granitic magmatism, for example, in the Mesoproterozoic Granite-Rhyolite Province in North America (Bickford et al., 2015; Slagstad et al., 2009b).

3. There are major differences in tectonic evolution between the western and eastern parts of the Sveconorwegian Province. The magmatic and metamorphic evolution in the western part is characterized by long-lived HT-UHT metamorphism and magmatism between 1070 and 920 Ma (Blereau et al., 2017; Bybee et al., 2014; Coint et al., 2015; Slagstad et al., 2013a), whereas coeval (ca. 980 Ma), but relatively short-lived high-pressure metamorphism is equally well documented in the eastern part (Möller et al., 2015; Tual et al., 2017). In contrast, central parts of the orogen display few, if any, signs of orogenesis after 1080 Ma, although one could argue that these rocks were situated at a high crustal level during most of this period (orogenic lid). A continent-continent collision cannot account for duration, geographical distribution, or differences in tectonic style between different parts of the orogen.

In recent years the ca. 980 Ma eclogite facies event in southwest Sweden has been cited as a key piece of evidence for a Himalayan-type collisional orogeny (Möller et al., 2015; Tual et al., 2017). These eclogites were subducted to great depth (to 19 kbar) and heated quickly (to 900 °C), before being rapidly exhumed (Möller, 1998; Tual et al., 2017), and it was argued that the most likely explanation for this rapid exhumation is by extrusion from the lower parts of overthickened, hot orogenic crust. However, unlike the eclogites in southwest Sweden, the eastern Himalayan eclogite counterparts formed rather late in the Himalayan orogenic evolution (Grujic et al., 2011), well after the establishment of thick orogenic crust, whereas no evidence exists to suggest that such an overthickened orogenic crust existed in the eastern Sveconorwegian Province prior to the time of eclogite facies metamorphism. Regardless, the eclogites in the Eastern Segment almost certainly formed as a result of subduction beneath the Idefjorden terrane (Möller et al., 2015; Tual et al., 2017), which by then formed a single crustal entity with more western parts of the province (Slagstad et al., 2017). Thus, if one regards this ~300-km-wide tract of continental crust (roughly the width of Japan) to be a continent, then the eclogites in southwest Sweden may be said to result from continent-continent collision. However, if one defines a continent as being surrounded by oceanic crust on all sides, then a continent-continent collision must always be preceded by oceanic subduction. The current data do not support such a scenario (i.e., no trace of arc magmatism in the eastern Sveconorwegian Province), and a more plausible scenario is that previously thinned continental crust, separating the Eastern Segment and Idefjorden terrane, subducted westward, pulling the Eastern Segment with it (Slagstad et al., 2017). This process is equivalent to amalgamating Japan back onto Asia, subducting the thinned continental (and minor oceanic) crust between the two entities. If this is what is meant by collision, it needs to be clearly stated. However, although one may argue that the ca. 980 Ma eclogite facies event in the eastern Sveconorwegian Province is a continent-continent collision, it certainly is not a Himalayan-type collision, and is clearly incapable of explaining the overall Sveconorwegian evolution between 1140 and 920 Ma.

A collisional model for the Sveconorwegian orogenic evolution needs to explain how collision, initiated ca. 1140 Ma, was sustained for 220 m.y., and why this collision resulted in widely different tectonic processes operating in different parts of the orogen at different times. Although the accretionary model presented here and in earlier publications is likely to be modified as more data become available from other parts of the orogen, the currently available data appear to rule out the Himalayan-type continent-continent collision hypothesis, or at least render it unlikely.

CONCLUSIONS

The new geochronologic data on magmatic rocks suggest nearly continuous granitic magmatism, with emplacement of the calc-alkaline SMB between 1070 and 1010 Ma, followed by ferroan hornblende-biotite granite magmatism between 1000 and 920 Ma, and emplacement of anorthosite poorly constrained as 950–920 Ma.

Aeromagnetic data suggest that the ferroan hornblende-biotite granites form a major north-south–trending batholith rather than relatively small, isolated plutons as previously inferred, consistent with formation in a magmatic arc. Geochemical and isotopic data, discussed elsewhere, suggest that the ferroan hornblende-biotite granites may have formed by remelting of the depleted and dehydrated SMB lower crustal source during extension of the arc, resulting in mafic underplating.

We consider the Sveconorwegian orogenic evolution to reflect a series of accretionary events, with repeated extension in the southwest part of the orogen. There is no evidence of significant crustal thickening in the study area, and mantle-derived heat appears to have been the main driver, resulting in widespread lower crustal melting and high-grade metamorphic conditions that were sustained for ~150 m.y. The Sveconorwegian orogenic evolution cannot be explained by Himalaya-type continent-continent collision, which fails to explain the duration of orogenesis (220 m.y.), as well as the major differences in style and timing of orogenic events in different parts of the orogen.

Long-lived Sveconorwegian HT-UHT metamorphism and coeval mafic and felsic magmatism between ca. 1070 and 920 Ma in the southwest Sveconorwegian Province mark final assembly and stabilization of the Fennoscandian Shield.

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