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The Sveconorwegian orogeny – Reamalgamation of the fragmented southwestern margin of Fennoscandia

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ABSTRACT

The Sveconorwegian orogeny encompasses magmatic, metamorphic and deformational events between ca. 1140 and 920 Ma at the southwestern margin of Fennoscandia. In recent years, the tectonic setting of this nearly 200 Myr-long evolution has been debated, with some workers arguing for collision with an unknown continent off the present-day southwest coast of Norway, and others advocating accretionary processes inboard of an active margin. Recently, it has been suggested that orogeny may have been gravity-driven by delamination and foundering of heavy subcontinental lithospheric mantle in an intraplate setting, in some ways similar to proposed sagduction processes in the Archaean. Resolving the tectonic setting of the Sveconorwegian orogen has implications for correlation with other orogens and Rodinia supercontinent reconstructions and for assessments of the evolution of plate tectonics on Earth, from the Archaean to the present. Here, we present new mapping and geochronological data from the Bamble and Telemark lithotectonic units in the central and western Sveconorwegian orogen – the former representing a critical region separating western parts of the orogen that underwent long-lived high- to ultrahigh-temperature metamorphism and magmatism from parts closer to the orogenic foreland that underwent episodic high-pressure events. The data show that the units constituting the Sveconorwegian orogen most likely formed at the southwestern margin of Fennoscandia between ca. 1800 and 1480 Ma, followed by fragmentation during widespread extension between ca. 1340 and 1100 Ma marked by bimodal magmatism and sedimentation. A summary of Sveconorwegian magmatic, metamorphic and depositional events in the different units shows disparate histories prior to their assembly with adjacent units. The most likely interpretation of this record seems to be that episodic, Sveconorwegian metamorphic and deformational events in the central and eastern parts of the orogen represent accretion and assembly of these units. This process most likely took place behind an active margin to the southwest that sustained mafic underplating in the proximal back-arc, resulting in high- to ultrahigh-temperature metamorphism in the western parts. In this interpretation, all features of the Sveconorwegian orogen are readily explained by modern-style plate tectonic processes and hypotheses involving some form of vertical, intraplate tectonics are not supported.

1. Introduction

It is generally acknowledged that present-day tectonic processes, characterised by horizontal movements of rigid plates, are unlikely to have operated in the same manner or magnitude in the early Earth. The tectonic regime prior to the current tectonic mode is not well constrained, but many argue for some form of vertical tectonics where mafic magmatism led to thick basaltic crustal piles that delaminated and sank back into the mantle (sagduction) where they underwent melting to form TTG (tonalite–trondhjemite–granodiorite) rocks – or some variation thereof (Johnson et al., 2017). These processes are suggested to have formed the characteristic TTG–greenstone complexes observed in Archaean cratons such as the Pilbara (Smithies and Champion, 2000; Van Kranendonk et al., 2004), Superior (Bédard,

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2006) and Barberton (Van Kranendonk, 2011). Although there is general acceptance that the Earth's tectonic regime has evolved through time, there is currently no consensus on the timing, duration, and transition mechanisms from a ’primordial’ tectonic mode to the modern plate tectonic mode. Brown and Johnson (2018) and Smithies et al. (2005), citing preservation of ‘paired’ high- and intermediate dT/dP metamorphism and magmatic evidence, argued that plate tectonics was established as early as the Mesoarchaean, and most workers agree that plate tectonic processes akin in principle to present day were operative from the Palaeoproterozoic onwards (see summary in Roberts et al., 2015). Although plate tectonics may have operated locally, it is still unclear when these processes became a global phenomenon. Some workers have argued that present-day, global plate tectonics did not operate until the Neoproterozoic (Hamilton, 2011; Piper, 2013; Stern, 2005), citing both palaeomagnetic and geological/petrological evidence including a general lack of ophiolites and blueschist-facies/ultrahigh-pressure metamorphism in pre-Phanerozoic terrains. Most arguments for the timing of onset of present-day tectonics rely on consideration of global datasets, which to some extent lack geological context. For example, ‘paired’ metamorphic belts, albeit coeval, are not necessarily paired, but may have been juxtaposed during later processes (e.g., Brown, 2010). Based on patterns of Hf isotopes through time constrained by zircon crystals, Hartnady and Kirkland (2019) considered the change to modern subduction-style tectonics to have developed progressively over a transitional period between 3.2 and 2.7 Ga. This transitional period was characterised by an exponential increase in the proportion of zircon with radiogenic Hf isotopic signatures, reflecting the time period over which the source region for granitoid (zircon-forming) magmas gradually switched from more chondritic to more depleted reservoirs (Hartnady and Kirkland, 2019). This gradual change into modern style tectonics was envisaged to have also occurred over geographically restricted belts before becoming a global phenomenon. Hence, geologically constrained tectonic models for Precambrian orogens provide the primary data with which to constrain their (non)similarity to present-day tectonic processes and the evolution of the planet's crustal-recycling mechanism.

Prior to the 1990s, most models of Sveconorwegian orogenesis argued for accretionary tectonic processes without the need for a continental indenter (e.g., Falkum and Petersen, 1980; Gower, 1985). Since then, the ca. 1140–920 Ma Sveconorwegian orogeny has generally been interpreted to represent a Himalayan-type continent–continent collision between SW Fennoscandia and an unknown continent, often inferred to be Amazonia (Bingen et al., 2008b; Möller et al., 2015; Romer, 1996; Tual et al., 2017). In the last few years, however, a number of workers have advocated a tectonic model involving processes at or behind an active, convergent continental margin (Blereau et al., 2017; Bybee et al., 2014; Slagstad et al., 2018b, 2017, 2013), generally similar to the pre-1990 models. Although consensus has yet to be reached on this matter, we note that the proposed tectonic processes accounting for the geodynamic evolution of the Sveconorwegian orogen have always been viewed in the adage of ‘The present is the key to the past’ (attributed to James Hutton). However, in a recent break with this present-day plate tectonic view, Bingen and Viola (2018) and Bingen et al. (2018) argued that the Sveconorwegian orogenic evolution is better explained as being driven by a series of long-lived delamination events, with delaminating and foundering subcontinental mantle lithosphere (SLCM) pulling the overlying crust down with it, a process that in some ways may be analogous to Archaean sagduction (Arndt, 2014). In their model, simply referred to here as the ‘sagduction model’, gravitational instability was the main driving force of orogeny, not the other way around, with the latter commonly observed in modern orogens with overthickened crust (e.g., the Himalayan–Tibetan system). If correct, this interpretation of Sveconorwegian orogenesis could imply that present-day plate tectonic processes did not operate on a global scale until after 1000 Ma. This model would also have major implications for supercontinent reconstructions, where coeval orogens of apparently similar tectonic style are typically, albeit sometimes erroneously, linked with one another (Slagstad et al., 2019).

Magmatic, metamorphic and deformational events between ca. 1140 and 920 Ma, at the southwestern margin of Fennoscandian, are generally attributed to the Sveconorwegian orogeny. In this contribution, we present new field and geochronological data, with relevance to the pre-Sveconorwegian configuration of the SW Fennoscandian margin and early Sveconorwegian orogenesis in the central Sveconorwegian Province, between ca. 1140 and 1050 Ma. Based on these and previously published data, we present an interpretation of the evolution of the SW Fennoscandian margin between ca. 1500 and 900 Ma. We also review the arguments for sagduction and discuss whether gravity-driven vertical tectonics are indeed needed to explain Sveconorwegian orogenic processes, or if modern-style, horizontal plate tectonics can provide a more actualistic alternative.

2. Regional geology

The 1700–1500 Ma tectonic evolution of the SW margin of Fennoscandia, represented by the various units making up the Sveconorwegian Province, is relatively well known and appears to have involved oceanward growth along a retreating active continental margin (Åhäll and Connelly, 2008; Andersen et al., 2004; Bingen et al., 2005; Petersson et al., 2015b; Roberts and Slagstad, 2015; Roberts et al., 2013). The succeeding period was tectonically quiescent until the emplacement of the ca. 1340 Ma bimodal Kungsbacka suite in the Idefjorden lithotectonic unit (LU; Hegardt et al., 2007) and similar magmatism farther west in the Province (Pedersen et al., 2008). As discussed in more detail below, this bimodal magmatism – typically interpreted to reflect an extensional tectonic regime – was active until 1300 Ma in the Idefjorden LU and 1150 Ma in the Telemark and Bamble LUs. In the Telemark LU, this long-lived extension was accompanied by deposition of geographically extensive, thick sedimentary sequences – the Telemark supracrustals – until ca. 1110 Ma (Bingen et al., 2003; Lamminen, 2011; Spencer et al., 2014), i.e., outlasting high-grade metamorphism in the Bamble LU at 1140 Ma by several tens of million years.

The Sveconorwegian orogen (Fig. 1) has been intensely studied over several decades and, although the tectonic settings and driving forces are still debated, the disparate tectonic styles between the western (hinterland) and eastern (foreland) parts of the orogen are well established (Falkum and Petersen, 1980). To the east, the tectonic evolution is characterised by tectonic juxtaposition of units with distinctly different Sveconorwegian geological histories, recorded by several medium- to high-pressure events at 1140 Ma (11.5 kbar, > 850 °C, Bamble LU), 1050 and 1025 Ma (10–15 kbar, 700–740 °C, Idefjorden LU) and 990 Ma (16.5–19 kbar, 850–900 °C, Eastern Segment) (e.g., Bingen et al., 2008a; Möller et al., 2015; Söderlund et al., 2008). Magmatic activity in the eastern part is limited to the waning stages of the orogeny. In contrast, the western parts of the orogen are characterised by long-lived granitic magmatism between ca. 1070 and 920 Ma, and associated high- to ultrahigh-temperature (850–1000 °C), low- to medium-pressure (6–8 kbar) metamorphism (Blereau et al., 2019, 2017; Drüppel et al., 2013; Laurent et al., 2018a,b; Slagstad et al., 2018b). Although metamorphic history and grade vary significantly in the Telemark LU, with some regions only reaching greenschist-facies conditions (Brewer and Menuge, 1998), persistent granitic magmatism with a dominantly lower-crustal source (Granseth et al., 2020) suggests generally high-temperature conditions at depth, with the current distribution of metamorphic facies reflecting late-orogenic differential exhumation (Slagstad et al., 2018b). These contrasts have been emphasised in recent papers and interpreted as an indication of an accretionary rather than continent–continent collisional tectonic setting for the Sveconorwegian orogen (e.g., Slagstad et al., 2017). The Sveconorwegian evolution is at least partly consistent with a hot back-arc setting, as envisaged by Hyndman (2019). Although no true arc has

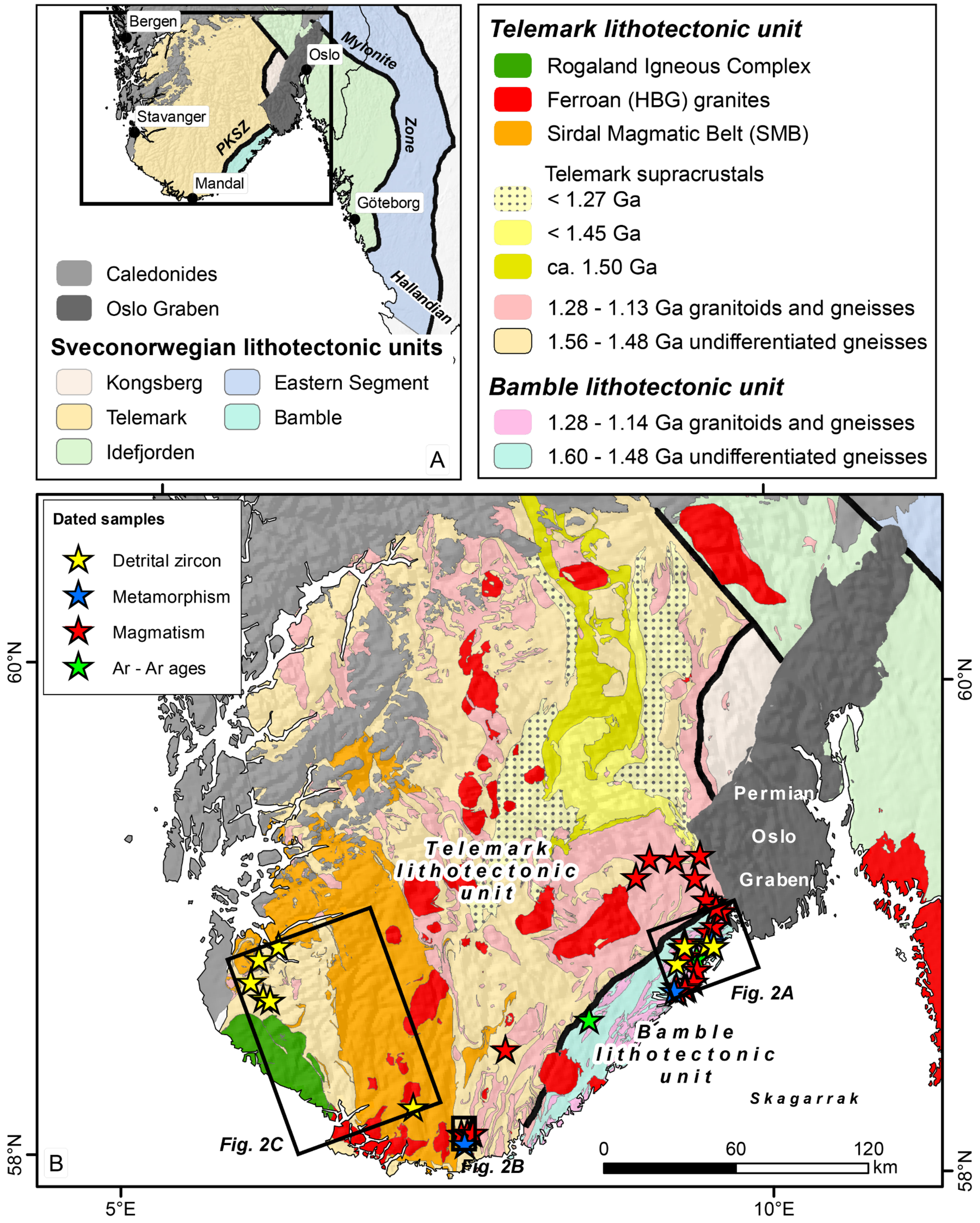

Fig. 1. Geological map draped over digital elevation model showing the main features of the Sveconorwegian orogen, dated locations and outlines of areas investigated in detail during this study.

been identified in Precambrian basement in southwest Fennoscandia so far, Corfu (2019) recently identified ca. 985–950 Ma allochthonous metavolcanic and -plutonic rocks with arc chemical affinities in Caledonian nappes in southwest Norway, most likely derived from the pre-Caledonian southwestern margin of Fennoscandia, consistent with a back-arc setting for at least the western parts of the Sveconorwegian orogen.

3. Geology of the study areas

The main study area for this contribution is in the Bamble LU. Nijland et al. (2014) presented a comprehensive review of the geology of the Bamble LU that, along with recent mapping by the Geological Survey of Norway (Marker et al., 2019), forms the basis for the summary presented here. The study areas in the western and central Telemark LU have been described in detail (Møkkelgjerd, 2019; Slagstad et al., 2018b), and are only briefly discussed in this contribution.

3.1. Bamble lithotectonic unit

The Bamble LU consists of amphibolite- to granulite-facies gneisses with an overall NE–SW strike and moderate to steep dip, and steeply to

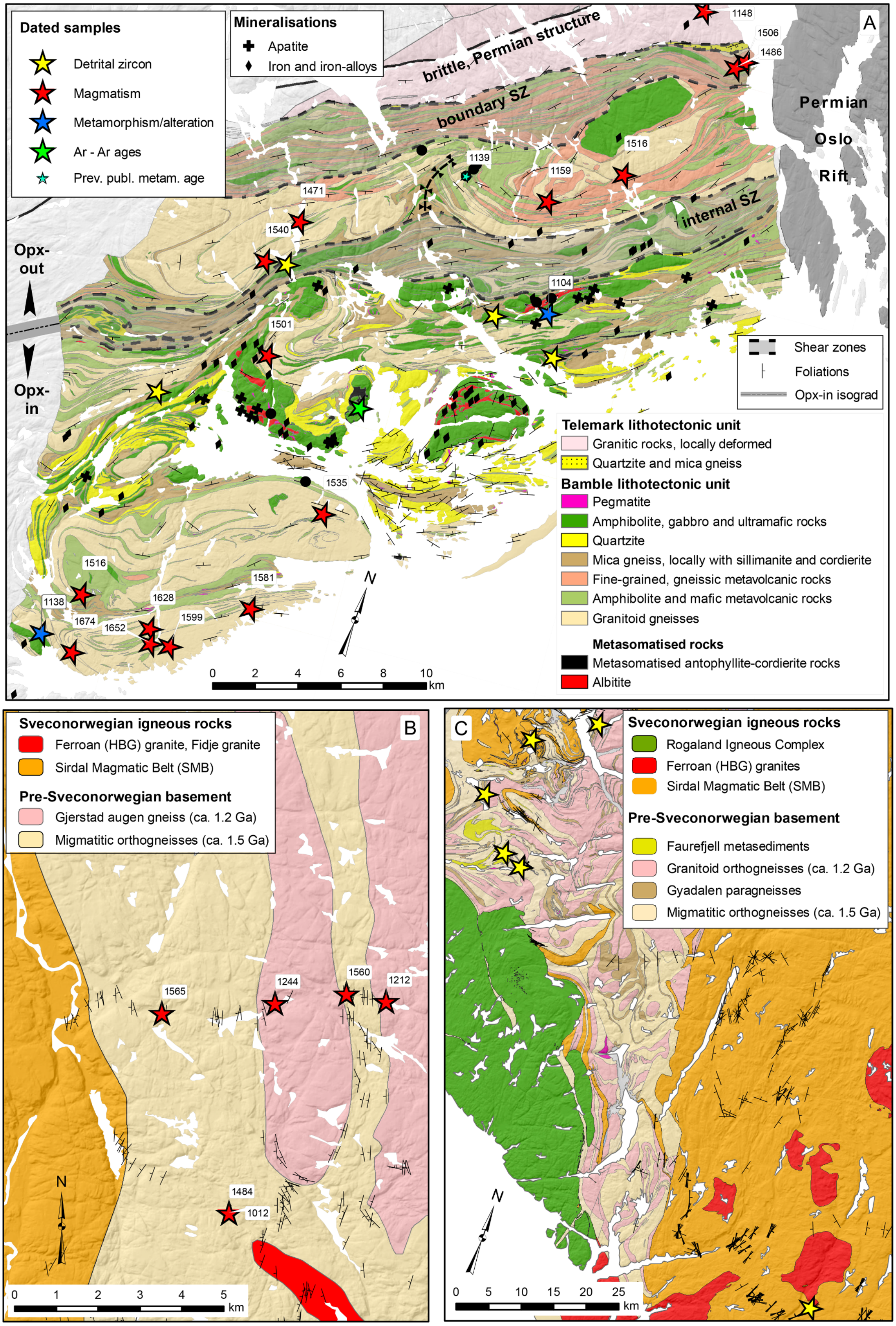

Fig. 2. Detailed geological maps draped over digital elevation models of the study areas, based on our own mapping in the Bamble lithotectonic unit (A) and partly on the 1:250,000 map sheet Mandal of Falkum (1982) in Vest-Agder (B) and Rogaland (C).

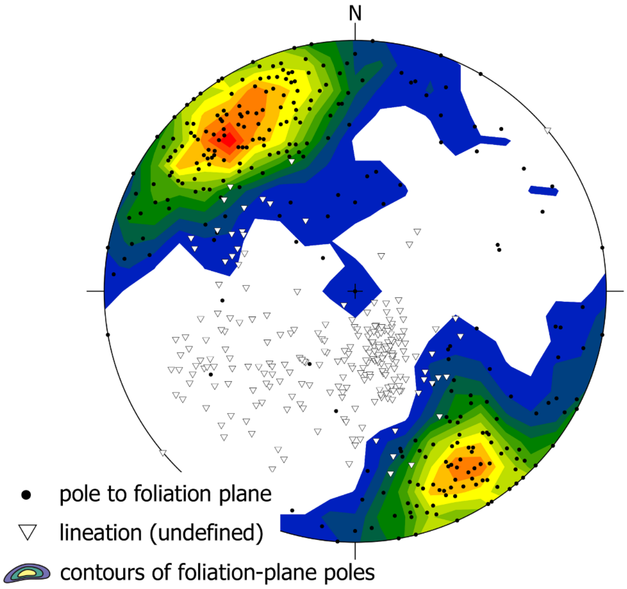

Fig. 3. Equal-area projection, lower-hemisphere stereoplot of 280 foliation and 238 lineation measurements from the study area in the northeastern part of the Bamble lithotectonic unit. Contour diagram of poles of foliation planes.

moderately, S- and SW-plunging lineations (Figs. 2A; 3). The dominant lithologies include: (1) variably migmatitic granitic orthogneisses (Fig. 4A) with protolith ages between 1600 and 1520 Ma and major- and trace-element compositions suggestive of formation in a continental magmatic arc (Andersen et al., 2004); (2) quartz-rich and metapelitic metasedimentary rocks (Fig. 4B) reflecting shallow-water and turbiditic deposits with poorly constrained maximum depositional ages around 1500–1400 Ma (de Haas et al., 1999); and (3) younger intrusive granitic and gabbroic orthogneisses with igneous crystallisation ages between 1235 and 1135 Ma (Heaman and Smalley, 1994). The youngest intrusions in the Bamble LU are the ca. 950 Ma Herefoss and Grimstad granites that effectively stitch the Bamble and Telemark LUs.

Geochronological and petrological evidence suggest that the Bamble LU rocks underwent upper amphibolite- to granulite-facies metamorphism at ca. 11.5 kbar and > 850 °C between 1145 and 1120 Ma (Bingen et al., 2008a; Engvik et al., 2016). In addition, cross-cutting relationships indicate the presence of an earlier high-grade metamorphic event (Gupta and Johannes, 1982; Nijland and Senior, 1991; Fig. 4C); however, the timing and nature of this event is unknown. Following the ca. 1140 Ma high-grade metamorphic event, which is generally taken to represent initiation of compression and Sveconorwegian orogeny (Bingen et al., 2008a), large tracts of the Bamble LU were metasomatically altered, with widespread scapolitisation, albitisation and associated development of hydrothermal apatite, rutile and Fe deposits (Engvik and Austrheim, 2010; Engvik et al., 2018; Fig. 4D). U–Pb dating of rutile from scapolite metagabbro and albitite has yielded ages of 1090–1084 Ma and Rb–Sr dating of phlogopite from scapolite metagabbro and a phlogopite–apatite vein has yielded ages between 1070 and 1040 Ma (Engvik et al., 2011). Metasomatism is interpreted to have taken place at pressures between 2 and 4 kbar and temperatures of 350–700 °C (Engvik et al., 2011; Nijland and Touret, 2001). Citing the relatively low closure temperatures of the rutile and phlogopite systems (discussed further later in this paper), Engvik et al. (2011) argued that these ages represent regional cooling following the metasomatic activity. These ages are consistent with growth of monazite between 1101 and 1078 Ma, interpreted to have grown during hydrothermal activity (Engvik et al., 2017), thereby allowing for a protracted period of metasomatism. Work on metasomatic mineral replacement reactions by Engvik et al. (2014, 2011) suggests that the fluid must have been enriched in Na, K, Cl, Mg, B and P, indicative of brines associated with seawater, which may have been released during prograde metamorphism of sedimentary rocks and/or evaporites.

The boundary between the Bamble and Telemark LUs is typically taken to be the Porsgrunn–Kristiansand Fault or Shear Zone (PKSZ). The PKSZ is a steeply E-dipping, ductile shear zone showing top-to-NW thrusting (Henderson and Ihlen, 2004), and reactivated as a Permian, extensional brittle structure that continues into the Oslo Graben (Starmer, 1993). The brittle structure forms a strong linear feature, which in the study area is located several kilometres into the Telemark LU (Fig. 2A). Recent mapping in the northeastern Bamble LU (Marker et al., 2019) shows that the PKSZ there may consist of two separate, SW–NE-oriented, km-thick shear zones (Figs. 2A, 4E). Although a detailed structural study of these shear zones has not been conducted, they do preserve evidence of top-to-the-NW thrusting, including C–S fabrics and rotated porphyroclasts (Fig. 4F). As discussed below, it is not entirely clear which of these shear zones separates the Bamble from the Telemark LU. The southeasternmost, structurally higher shear zone was labelled the Sannidal–Stokke Shear Zone and the northwesternmost, structurally lowest shear zone the Rørholt–Asdal Shear Zone (RASZ) by Marker et al. (2019). For simplicity, the Sannidal–Stokke Shear Zone, which is located close to the generally accepted boundary between the Telemark and Bamble LUs, is referred to here as the *boundary shear zone*, whereas the Rørholt–Asdal Shear Zone, located internal to the Bamble LU, is referred to as the *internal shear zone*.

There are some notable lithological contrasts across the two shear zones (Fig. 2A); in particular, ca. 1200–1140 Ma granitic rocks, including the Tråk rapakivi granite, extend across the boundary shear zone and are deformed by it, with sparse evidence of coeval magmatism across the structurally higher internal shear zone (e.g., Engvik et al., 2011). Conversely, quartzite and evidence of hydrothermal alteration and associated iron and apatite ores are restricted to rocks above the internal shear zone. The early, ca. 1140 Ma high-grade metamorphism that is characteristic of the Bamble LU has been detected both structurally above and below the internal shear zone (Engvik et al., 2016; this study); however, the amount of data from either side of the shear zone is limited making it difficult to assess whether there is a contrasting metamorphic history between rocks separated by the shear zone. We note that the one reported occurrence of ca. 1140 Ma high-grade metamorphism apparently below the internal shear zone is located close to a N–S-trending synform in which folded, juxtaposed rocks from the hanging wall and footwall of the internal shear zones are preserved (Fig. 2A); thus, ca. 1140 Ma granulite-facies metamorphism has not yet been unequivocally observed in the footwall. It is also notable that the internal shear zone coincides with the regionally defined opx-in isograd (Padget and Brekke, 1996; Fig. 2A), separating granulite-facies rocks in the hanging wall from upper amphibolite-facies rocks in the footwall. The boundary shear zone marks a contrast in metamorphic grade from upper amphibolite- to granulite-facies in the Bamble LU hanging wall to greenschist to amphibolite-facies in the Telemark LU footwall, but the lithologies are otherwise similar on either side. Thus, although more work is needed to elucidate the significance of these variations, it is tempting to interpret the Bamble LU as a complex thrust stack rather than a coherent LU, maybe along the similar lines as suggested by Bingen et al. (2005).

On a regional scale, the PKSZ displays evidence of top-to-the-NW thrusting (Henderson and Ihlen, 2004). The age of thrusting is unknown; one apparently *syn*-tectonic pegmatite yielded an age of ca. 1060 Ma (Baadsgaard et al., 1984), whereas Henderson and Ihlen (2004) suggested that shear was protracted, but as yet of unknown duration. Seismic data from Skagerrak have been interpreted to show a Moho offset along the PKSZ (Andersson et al., 1996), which, along with evidence of a protracted tectonic history, makes the PKSZ a dominant structural feature in the Sveconorwegian orogen.

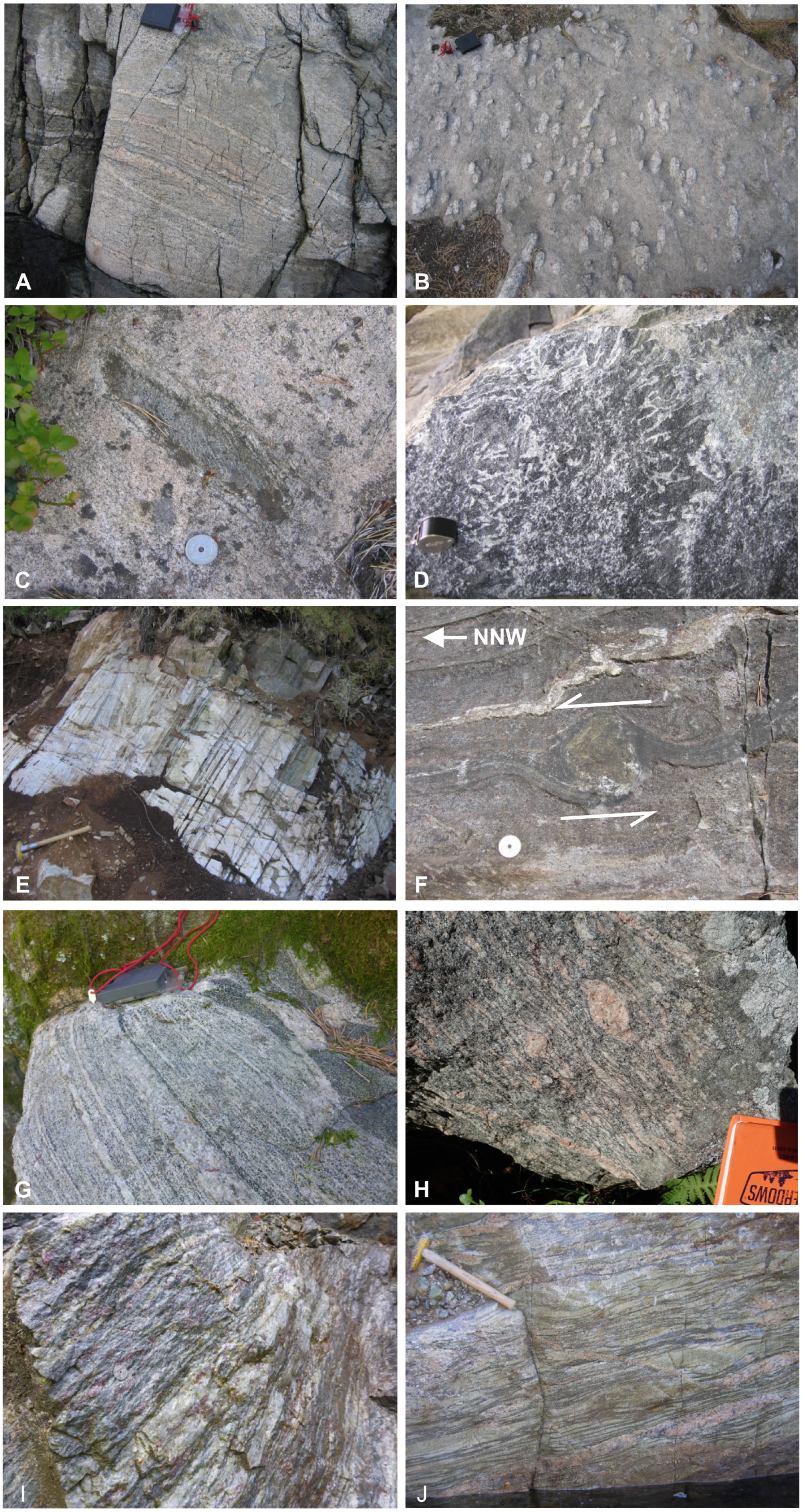

Fig. 4. Field photos of some of the rocks discussed in the text. (A) Migmatitic granitic orthogneiss, ca. 1550 Ma, Bamble lithotectonic unit (BAM143–1). (B) Quartzitic gneiss with cm-size quartz-sillimanite nodules, Bamble lithotectonic unit (BAM85–1). (C) Gneissic xenolith in ca. 1160 Ma Hørsfjell granite, Bamble lithotectonic unit (BAM285). (D) Scapolitised gabbro, Bamble lithotectonic unit (BAM110–2). (E) Straight gneiss from the boundary shear zone, Bamble lithotectonic unit (BAM248–2). (F) Rotated calc-silicate delta clast showing top-to-NNW movement, from boundary shear zone, Bamble lithotectonic unit (BAM280-1, photo taken looking east). (G) Migmatitic granitic orthogneiss, ca. 1550 Ma, Telemark lithotectonic unit (VAG128018). (H) Porphyritic granitic gneiss, ca. 1200 Ma, Telemark lithotectonic unit. (I) Metapelitic gneiss, Gyadalen metapelite, Telemark lithotectonic unit (MM01349). (J) Quartz-rich, calc-silicate gneiss, Faurefjell metasediments, Telemark lithotectonic unit (MM01348).

3.2. Western, central and eastern Telemark lithotectonic unit

The western part of the Telemark LU (Fig. 2B) has been extensively studied over several decades, whereas the central and eastern parts (Fig. 2C) have received much less attention. The main reason for this difference is the prominent, high- to ultrahigh-temperature metamorphic history observed in the west. In the study areas in the Telemark LU, the dominant rock types are variably metamorphosed and deformed igneous rocks with protolith ages of ca. 1.5 and 1.32–1.15 Ga (Fig. 4G, H; Bingen et al., 2005; Brewer et al., 2004; Coint et al., 2015; Pedersen et al., 2008). The tectonic setting of the 1.5 Ga rocks is generally inferred to have been an active margin (e.g., Roberts et al., 2013), whereas the protoliths to the 1.32–1.15 Ga gneisses were formed as part of an extensive bimodal suite during widespread crustal extension. The magmatism has been ascribed to repeated episodes of continental-margin extension starting at ca. 1.3 Ga and lasting until the onset of the Sveconorwegian orogeny traditionally defined to be at ca. 1140 Ma (Bingen et al., 2003; Brewer et al., 2004). Deposition of sediments took place during both periods, with the older, Rjukan rift basin receiving detritus from local 1.5 Ga sources as well as subordinate < 2 Ga and sparse Archaean sources, consistent with derivation from the SW Fennoscandian Shield (Lamminen and Köykkä, 2010). The younger Telemark supracrustals (Fig. 1) received detritus from similar sources, in addition to younger, probably active-margin, continental-arc sources, with the youngest detrital zircons suggesting deposition at or after ca. 1110 Ma (Lamminen, 2011; Spencer et al., 2014).

Although original field relationships in the western field area in Rogaland (Fig. 2C) are obscured by strong deformation and high- to ultrahigh-temperature metamorphism, the 1.5 Ga orthogneisses appear to have formed the basement for (semi)pelitic sediments, now metamorphosed at high grade (Blereau et al., 2017; Drüppel et al., 2013; Tomkins et al., 2005). These supracrustal rocks are referred to as the Gyadalen paragneisses (Fig. 4I; Coint et al., 2015; Hermans et al., 1975). The age of deposition of the Gyadalen paragneisses is poorly constrained, but field relationships show that they were intruded by voluminous granitoid orthogneisses at 1.23–1.20 Ga (Slagstad et al., 2018b), providing a minimum depositional age. The last known pre-Sveconorwegian event to take place in the western study areas, was deposition of the Faurefjell metasediments (Fig. 4J; Hermans et al., 1975), consisting of quartzite, impure marble, calc-silicate gneiss, and distinct oxide-rich layers that have been interpreted to represent metamorphosed laterite (Bol et al., 1989). The Faurefjell metasediments probably were deposited after ca. 1.2 Ga magmatism and before the onset of Sveconorwegian high-grade metamorphism in the western Telemark LU at around 1080 Ma.

4. Results

In addition to the new mapping data outlined above, the main new results presented here are U–Pb zircon and Ar–Ar age data. Although most of the data are consistent with previously published ages, the new data expand some age ranges implying overlap and continuity of various processes that have implications for the nature and evolution of Sveconorwegian orogenesis. Importantly, however, the new mapping data provide a greater contextual control on new and published data, which allows novel interpretations. To facilitate later discussion, the new geochronological data have been sorted based on age, geography and significance (magmatism, metamorphism, cooling, deposition). Table 1 provides descriptions of the dated samples and a summary of the results. Methods are described in Electronic Supplement 1 and the U–Pb zircon and Ar–Ar data are presented in Electronic Supplements 2 and 3, respectively.

4.1. 1650–1450 Ma magmatism, Bamble and Telemark lithotectonic units

Thirteen U–Pb zircon analyses of orthogneisses in the northeastern part of the Bamble LU yield rather imprecise ages between 1674 ± 71 and 1471 ± 30 Ma, interpreted to represent protolith crystallisation ages that reflect long-lived magmatic activity (Fig. 5, Table 1). Age uncertainties (2σ) are typically several tens of million years for most of the samples. Nearly all the samples form imprecise discordias with lower intercept ages in the range 500–200 Ma, and a rather large spread in discordance close to the upper intercept. Such complex U–Pb systematics are consistent with well-documented high-grade metamorphism affecting these rocks at ca. 1140 Ma, followed by hydrothermal activity between ca. 1100 and 1080 Ma (Bingen et al., 2008a; Engvik et al., 2017; see discussion below). In addition, many of the rocks reveal a noticeable younger thermal overprint characterised by paleomagnetic directions of Permian age (Piper, 2009; Kulakov et al., unpublished data), reflecting secondary overprinting associated with the formation of the Oslo Rift. We have therefore anchored many of the samples at lower intercepts of 400 ± 100 Ma, which reflect a dominant Permian resetting, but allow for a multitude of other possibilities. In addition, several analyses plot on discordia trends with lower intercepts around 1080 Ma, when anchored at the inferred protolith age. These analyses may record dominantly ca. 1140 Ma high-grade metamorphism, but with some younger overprinting. In addition to these two thermal events, cross-cutting field relationships (references in Nijland et al., 2014; see below) suggest an older, possibly Gothian (1.7–1.5 Ga) high-grade metamorphic event, including migmatisation. Despite the complexities in the dataset, the range of ages is similar to that previously shown for the Bamble LU (Bingen and Viola, 2018 and references therein), and intermediate to, but overlapping with generally younger Telemarkian ages to the west (ca. 1520–1480 Ma; Bingen et al., 2005; Roberts et al., 2013) and older, Gothian ages to the east (ca. 1690–1555 Ma; Åhäll and Connelly, 2008; Åhäll et al., 2000; Petersson et al., 2015a,b).

Three samples of mesosome from high-grade, migmatitic orthogneisses from the south-central Telemark LU yield ages of 1565 ± 11, 1560 ± 19 and 1484 ± 20 Ma (Fig. 6, Table 1), interpreted to reflect protolith crystallisation ages. The two older ages are significantly older than most previously published ages from the Telemark LU, which are generally between ca. 1520 and 1480 Ma (Bingen et al., 2005; Roberts et al., 2013), and overlap with ages from the Bamble LU. The increasing number of late Paleo- to early Mesoproterozoic ages available from the Sveconorwegian Province seems to suggest complete overlap from east to west, but with generally decreasing ages moving westwards. Leucosome from one of the samples yields an age of 1012 ± 9 Ma (Fig. 6C), comparable to ages of migmatisation farther west in the orogen (e.g., Coint et al., 2015; Slagstad et al., 2018b).

4.2. ca. 1300 Ma magmatism, Telemark lithotectonic unit

Magmatic activity in the period 1480–1280 Ma was comparatively sparse, but bimodal magmatism is well documented in the Idefjorden, Bamble and Telemark LUs. A sample of a Ni-bearing diorite in the central Telemark LU yielded a Concordia age of 1303 ± 9 Ma (Fig. 7A, Table 1), interpreted to represent the age of mafic magmatism. Several older, concordant analyses between ca. 1370 and 1480 Ma may represent inherited zircons. The 1303 Ma age contributes to previously published data (e.g., Pedersen et al., 2008) on bimodal magmatism in the central Telemark LU, overlapping with ages of bimodal magmatism in the Idefjorden and Bamble LUs.

4.3. 1240–1140 Ma magmatism, Bamble and Telemark lithotectonic units

Ten samples of granitoid rocks and gneisses from the Telemark LU yielded ages between 1244 ± 22 and 1135 ± 22 Ma (Fig. 7, Table 1), interpreted to reflect crystallisation. This range is consistent with previously published igneous crystallisation ages from the Telemark LU, which is dominated by felsic lithologies, but with sparse mafic rocks, possibly reflecting a sampling bias towards easier-to-date felsic rocks.

Table 1
Samples and summary of U-Pb zircon and Ar-Ar geochronological data.

Sample	UTM_E	UTM_N	Method	Rock description	Zircon morphology	Age	2σ	Interpretation
Pre-Sveconorwegian magmatism, Bamble								
MM36496	518,590	6,534,785	NGU	Red, variably foliated, porphyritic[a] bt-bearing Farsjø granitic gneiss, locally with thick amphibolite layers	100–250 µm, equidimensional with poorly developed crystal faces, CL-dark with irregular oscillatory zoning	1471	30	Upper intercept, protolith crystallisation
MM36632	535,298	6,548,522	NGU	Fine-grained, red, foliated granitic gneiss, mylonitic	100–200 µm, stubby to prismatic with well-developed oscillatory zoning	1486	24	Upper intercept, protolith crystallisation
BAM417	535,645	6,548,948	NGU	Migmatitic, fine-grained granitic gneiss, mylonitic	150–200 µm, stubby prismatic with well-developed oscillatory zoning	1494	16	Weighted average 207/206 age, protolith crystallisation
BAM108	519,330	6,528,391	BGS MC	Hbl-bt-granite, reddish, medium-grained, sparsely prophyritic, variably foliated Helle granitic gneiss	50–200 µm, equidimensional to prismatic, oscillatory zoned with some CL-dark grains and mantles	1501	11	Upper intercept, protolith crystallisation
BAM358	514,950	6,514,897	NGU	Reddish grey, foliated, hbl-bt-bearing, porphyritic Levang granite	200–250 µm, prismatic, CL-dark with well-developed oscillatory zoning	1516	20	Upper intercept, protolith crystallisation
MM36616	532,088	6,542,040	NGU	Dark grey, foliated granitic gneiss, mylonitic with fine-grained grt	100–200 µm, rounded stubby to prismatic with well-developed oscillatory zoning	1516	25	Upper intercept, protolith crystallisation
MM36736	524,291	6,522,301	NGU	Reddish grey, foliated, hbl-bt-bearing, porphyritic Levang granite	100–250 µm, prismatic with oscillatory and sector zoning	1535	39	Upper intercept, protolith crystallisation
BAM107	517,689	6,532,496	BGS MC	Red, variably foliated, porphyritic bt-bearing Farsjø granitic gneiss, locally with thick amphibolite layers	50–150 µm, equidimensional to prismatic, oscillatory zoned with some CL-dark grains and mantles	1540	30	Upper intercept, protolith crystallisation
MM36743	522,616	6,516,963	NGU	Grey, medium-grained, dispersely bt-hbl striped, foliated qtz-diorite with thin amphibolite layers interpreted as former mafic dykes	100–250 µm, rounded equidimensional to prismatic with oscillatory and sector zoning	1581	52	Upper intercept, protolith crystallisation
MM57818	519,641	6,513,998	NGU	Grey, medium-grained, dispersely bt-hbl striped, foliated qtz-diorite with thin amphibolite layers interpreted as former mafic dykes	50–150 µm, rounded equidimensional to prismatic with irregular, variably developed zoning	1599	37	Upper intercept, protolith crystallisation
MM57819	518,499	6,514,465	NGU	Grey, medium-grained, dispersely bt-hbl striped, foliated qtz-diorite with thin amphibolite layers interpreted as former mafic dykes	50–150 µm, irregular equidimensional to prismatic with oscillatory zoning	1628	40	Upper intercept, protolith crystallisation
MM57820	518,754	6,513,812	NGU	Grey, medium-grained, dispersely bt-hbl striped, foliated qtz-diorite with thin amphibolite layers interpreted as former mafic dykes	50–150 µm, rounded equidimensional to prismatic, relatively CL-dark with oscillatory zoning	1652	52	Upper intercept, protolith crystallisation
MM57821	515,440	6,512,178	NGU	Light reddish grey, medium-grained, weakly foliated bt-bearing granodiorite	50–200 µm, stubby to prismatic, relatively CL-dark with well-developed oscillatory zoning	1674	71	Upper intercept, protolith crystallisation
Early Sveconorwegian magmatism, Bamble								
BAM285	529,184	6,539,643	NGU	Medium-grained, weakly foliated, bt-bearing Hørsfjell granite	150–400 µm, prismatic with well-developed oscillatory zoning	1159	14	Weighted average 207/206 age, protolith crystallisation
Early Sveconorwegian metamorphism/metasomatism, Bamble								
BAM110A	530,990	6,534,713	BGS MC	Scapolite-metasomatised gabbro, medium grained and weakly foliated, locally with grt	100–150 µm, equidimensional, CL-dark with faint irregular zoning	1104	7	Concordia age, scapolitisation
MM57815	513,774	6,512,526	NGU	Brownish grey, medium-grained, bt-gedrite-bearing (qtz) diorite, weakly foliated	100–150 µm, equidimensional, CL-dark with faint irregular zoning	1138	13	Weighted average 207/206 age, high-grade metamorphism
Pre-Sveconorwegian magmatism, Telemark								
HAL193073	418,910	6,443,145	SHRIMP II	Migmatitic orthogneiss with fine-grained, bt-bearing granitic mesosome (protolith) and medium-grained, hbl-bearing granitic leucosomes	100–250 µm, prismatic, CL-bright, oscillatory-zoned cores with thick, CL-dark, featureless mantles	1484	20	Concordia age from CL-bright, oscillatory-zone rims, protolih crystallisation
VAG128018	421,718	6,448,384	Curtin QQQ	Migmatitic orthogneiss with fine-grained, bt-bearing granitic mesosome and medium-grained, hbl-bearing granitic leucosomes	100–200 µm, (short) prismatic, with well-developed oscillatory zoningm some grains with CL-bright, thin rims	1560	19	Weighted average 207/206 age, protolith crystallisation
HAL193062	417,304	6,447,923	SHRIMP II	Migmatitic orthogneiss with fine- to medium-grained granitic mesosome (protolith) and sparse granitic leucosomes	100–200 µm, prismatic with well-developed oscillatory zoning	1565	11	Concordia age, protolih crystallisation
Early Sveconorwegian magmatism, Telemark								
85,152	528,372	6,554,297	NGU	Grey porphyritic granite	200–300 µm, prismatic with somewhat faint oscillatory and sector zoning	1135	22	Weighted average 207/206 age, protolith crystallisation

(*continued on next page*)

Table 1 (*continued*)

Sample	UTM_E	UTM_N	Method	Rock description	Zircon morphology	Age	2σ	Interpretation
85,151	531,262	6,552,800	NGU	Dark porphyritic granite	150–200 µm, prismatic with oscillatory and sector zoning	1138	9	Concordia age, protolih crystallisation
MM36639	533,042	6,550,526	NGU	Red porphyritic rapakivi granite, Tråk granite	150–300 µm, prismatic with oscillatory and sector zoning	1148	8	Concordia age, protolih crystallisation
85,153	525,520	6,574,415	NGU	Fine-grained, pink granitic/rhyolitic gneiss	200–250 µm, prismatic with oscillatory and sector zoning	1152	31	Upper intercept, protolith crystallisation.
85,159	523,332	6,563,959	NGU	Medium-grained, red granite	200–250 µm, prismatic with oscillatory and sector zoning	1193	8	Concordia age, protolih crystallisation
85,154	514,306	6,572,346	NGU	Dark porphyritic granite	100–200 µm, prismatic with oscillatory and sector zoning	1200	8	Concordia age, protolih crystallisation
85,156	496,615	6,564,397	NGU	Small-porphyritic granitic/rhyolitic gneiss	200–300 µm, prismatic with oscillatory and sector zoning	1201	9	Concordia age, protolih crystallisation
VAG128019	422,671	6,448,199	Curtin QQQ	Fine- to medium-grained, coarse flaser-textured hbl-bt granitic gneiss	100–250 µm, stubby to prismatic with variable internal zoning including oscillatory, sector and irregular	1212	25	Weighted average 207/206 age, protolith crystallisation
85,155	502,609	6,572,876	NGU	Porphyritic granite	150–300 µm, prismatic with oscillatory and sector zoning	1232	19	Weighted average 207/206 age, protolith crystallisation
VAG128016	420,005	6,448,159	Curtin QQQ	Fine- to medium-grained, coarse-textured hbl-bt granitic gneiss, the coarse texture probably a result of recrystallised kfs phenocrysts	100–250 µm, stubby to prismatic with variable internal zoning including oscillatory, sector and irregular	1244	22	Weighted average 207/206 age, protolith crystallisation
AUG131322	437,292	6,486,154	NGU	Medium-grained metadiorite	100 µm, equidimensional , irregular, rounded, with comlpex interiors incl. oscillatory zoned rims and featureless rims	1303	9	Concordia age, protolih crystallisation
Sveconorwegian metamorphism, Telemark								
HAL193073	418,910	6,443,145	SHRIMP II	Migmatitic orthogneiss with fine-grained, bt-bearing granitic mesosome (protolith) and medium-grained, hbl-bearing granitic leucosomes	100–250 µm, prismatic, CL-bright, oscillatory-zoned cores with thick, CL-dark, featureless mantles	1012	9	Concordia age from CL-dark mantles, high-grade metamorphism
Detrital zircon, Bamble								
BAM109, n2315	528,666	6,533,763	NORDSIM	Qtz-rich, sil-mu-bt gneiss, migmatitic, fine to medium grained, locally with crd and grt, locally with qtz-sil nodules	ca. 50 µm, rounded equidimensional with oscillatory to irregular zoning	Major peak at 1.49 Ga, scattered L. Palaeoprot. and one Archaean grain. Max. dep. age = 1.49 Ga.		
BAM112, n2318	515,041	6,525,028	NORDSIM	Qtz-rich, sil-mu-bt gneiss, migmatitic, fine to medium grained, locally with crd and grt, locally with qtz-sil nodules	ca. 100 µm, rounded irregular, CL-dark with faint irregular zoning	Major peak at 1.51 Ga, one < 5% disc. analysis at 1.35 Ga. Max. dep. age = 1.51 Ga		
BAM111, n2317	531,911	6,532,847	NORDSIM	Qtz-rich, sil-mu-bt gneiss, migmatitic, fine to medium grained, locally with crd and grt, locally with qtz-sil nodules	50–100 µm, rounded equidimensional with oscillatory, sector and irregular zoning, some CL-dark mantles	Major peak at 1.50 Ma, subsidiary peaks between 1.67 and 1.97 Ga. Max. dep. age = 1.45 Ga		
MM36466, n2319	518,700	6,532,684	NORDSIM	Grt-sil-bt metapelitic gneiss, migmatitic, rich in pink grt, locally with crd and layers of amphibolite	50–100 µm, rounded equidimensional with oscillatory, sector and irregular zoning, some CL-dark mantles	Major peak at 1.59 Ga, subsidiary peaks between 1.72 and 1.89 Ga. Scattered Archaean and E. Palaeoprot. grains. Max. dep. age = 1.59 Ma		
Detrital zircon, Rogaland, Gyadalen metapelites								
MM1349, n2311[b]	321,646	6,516,751	NORDSIM, BGS Attom	Crd-plag-bt-qtz-grt-spinel ± sil metapelitic gneiss, layers of amphibolite	50–100 µm, rounded equidimensional with oscillatory, sector and irregular zoning, some CL-dark mantles and grains	Broad peak between 1.45 and 1.65 Ga, subsidiary peaks at 1.8 and 1.9 Ga. Max. dep. age = 1.45 Ga. A few scattered 1.45–1.22 Ga, significance unknown		
ROG92326[b]	334,213	6,532,806	NGU	Crd-plag-bt-qtz-grt-sil metapelitic gneiss, tightly folded with a patchy migmatitic texture, sparse amphibolite	100–150 µm in size, with rather simple internal zoning characterised by oscillatory zoned grains or cores overgrown by CL-dark rims and mantles	Broad peak between 1.45 and 1.65 Ga, subsidiary peaks at 1.72 and 1.89 Ga. Max. dep. age = 1.45 Ga.		
IH128061[b]	395,379	6,459,660	NGU	Crd-plag-bt-qtz-grt-sil metapelitic gneiss, tightly folded with a patchy migmatitic texture, sparse amphibolite. The sample comes from a large xenolith within the SMB granites	100 µm, elongate and irregular, with irregularly zoned cores transected by slightly darker, irregularly zoned rims and mantles	Broad peak between 1.45 and 1.60 Ga. Max. dep. age = 1.45 Ga. A few scattered 1.45–1.22 Ga, significance unknown		
Detrital zircon, Rogaland, Faurefjell metasediments								
MM1348, n2312[b]	327,364	6,509,224	NORDSIM	Granoblastic diopside gneiss with light-green pyroxene, plag, qtz and up to 5% disseminated opaque minerals	50–100 µm, rounded equidimensional with oscillatory, sector and irregular zoning, some CL-dark mantles and grains	Dominant 1.35–1.60 Ga ages, subsidiary peaks at 1.68 and 1.75 Ga. Max dep. age = 1.18 Ga. Few analyses available		

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Table 1 (*continued*)

Sample	UTM_E	UTM_N	Method	Rock description	Zircon morphology	Age	2σ	Interpretation
ROG644	325,436	6,527,008	NGU	Medium-grained quartzite	100 µm, equidimensional, rounded, oscillatory to complex zoning, CL-dark mantles	Major peaks at 1.45 and 1.6–1.8 Ga, subsidiary peaks at 2.0 Ga and between 1.40 and 1.17 Ga. Max. dep. age = 1.17 Ga		
ROG639[b], n3001	330,452	6,508,055	NORDSIM	Medium-grained quartzite, associated with qtz-diopside gneiss	100–200 µm, rounded with partially resorbed, oscillatory-zoned cores and CL-darker, diffusely zoned mantles and grains	Dominant 1.32–1.50 and 1.68–1.95 Ga ages. Max. dep. age = 1.32 Ga. Few analyses available		
Ar-Ar samples, Bamble								
BM18	524,209	6,527,545	NGU	Amphibolitised metagabbro	Internal in Bamble lithotectonic unit	Amph 1101 ± 4 Ma, Bt 1041 ± 1 Ma, Plag 876 ± 38 Ma		
BM43	475,268	6,499,600	NGU	Amphibolite	In high-strain zone separating Bamble and Telemark lithotectonic units	Amph 1016 ± 3 Ma, Bt 920 ± 2 Ma		

Abbreviations: bi - biotite, crd - cordierite, grt - garnet, hbl - hornblende, kfs - K-feldspar, mu - muscovite, qtz - quartz, sil – sillimanite

[a] Phenocrysts are K-feldspar, unless otherwise stated.

[b] Data from metamorphic components of these zircons were presented and discussed by Slagstad et al. (2018).

In the Bamble LU, an undeformed medium-grained granite cutting the high-grade metamorphic fabric of a migmatitic gneiss xenolith (Fig. 4C) yielded an age of 1159 ± 14 Ma (Fig. 8, Table 1). Although the uncertainty on this age prohibits firm conclusions, the temporal relationship suggests that the rocks in the region experienced a phase of high-grade metamorphism prior to the 1140 Ma high-grade event identified by Bingen et al. (Bingen et al., 2008a), as suggested by Nijland et al. (2014) and references therein.

The new ages from the Bamble and Telemark LUs are similar to previously published ages from the Bamble and Kongsberg LUs (Bingen and Viola, 2018 and references therein), as well as from the Telemark LU (Bingen et al., 2002; Brewer et al., 2004; Slagstad et al., 2018b), showing that the period between ca. 1280 and 1130 Ma was characterised by orogen-wide bimodal magmatism. Noteworthy here is that coeval magmatism was all but absent in the Idefjorden LU.

4.4. *Early Sveconorwegian metamorphism and metasomatism (scapolitisation), Bamble lithotectonic unit*

U–Pb zircon dating of an orthogneiss in the southwestern part of the main study area in the Bamble LU yielded an age of 1138 ± 14 Ma (Fig. 9A, Table 1). The zircons from this sample are CL-dark, featureless with Th/U ratios < 1, and the age is interpreted to reflect high-grade metamorphism. This age is similar to previously determined ages for high-grade metamorphism in the Bamble LU (Bingen et al., 2008a; Bingen and Viola, 2018). The dated sample was collected structurally above the internal shear zone.

The Bamble LU is particularly well known for widespread and locally highly pervasive hydrothermal alteration. One sample from a scapolite-bearing gabbro contains equidimensional, somewhat irregular zircons that are CL-dark with faint, irregular zoning indicative of metamorphic growth. These zircon crystals yield an age of 1104 ± 7 Ma (Fig. 9B), significantly younger than the ca. 1140 Ma high-grade event recorded in the Bamble LU, but similar to previously published ages of hydrothermal alteration (Engvik et al., 2011). We interpret the zircon age of this scapolite gabbro to reflect hydrothermal alteration and scapolitisation.

4.5. *Ar–Ar hornblende, biotite and plagioclase ages*

Ar–Ar isotopic compositions from hornblende, biotite and plagioclase from two samples were analysed (Figs. 1 and 2 for locations). Sample BM18 was taken from a metagabbro near Kragerø, in the central part of the Bamble LU, an area well-known for widespread metasomatism. The sample was dated in order to constrain the timing of cooling of the Bamble LU. Sample BM43 was taken from an amphibolite in immediate vicinity of the shear zone separating the Bamble and underlying Telemark LUs, but outside of the main study area in the northeastern Bamble LU. BM18 yielded a hornblende plateau age of 1101 ± 4 Ma, biotite is 1041 ± 1 Ma, and plagioclase is 876 ± 38 Ma (Fig. 10A, Table 1). BM43 yields significantly younger ages, with hornblende at 1016 ± 3 Ma; a mini plateau was obtained from biotite at 920 ± 2 Ma (Fig. 10B, Table 1). The latter ages are significantly younger than those reported for the internal NE segment of the PKSZ, suggesting long-lived tectonic activity and thrusting between the Bamble and Telemark LUs.

4.6. *Detrital zircon age data*

Detrital zircon U–Pb ages have been obtained from metasedimentary rocks in the Bamble and Telemark LUs. The sampled units have all undergone high-grade metamorphism, with associated recrystallisation and growth of zircon. The distinction between detrital and metamorphic zircon has been made mainly on textural and compositional grounds, with oscillatory-zoned cores and grains with relatively high Th/U (> 0.1) interpreted to be detrital and CL-dark rims and grains

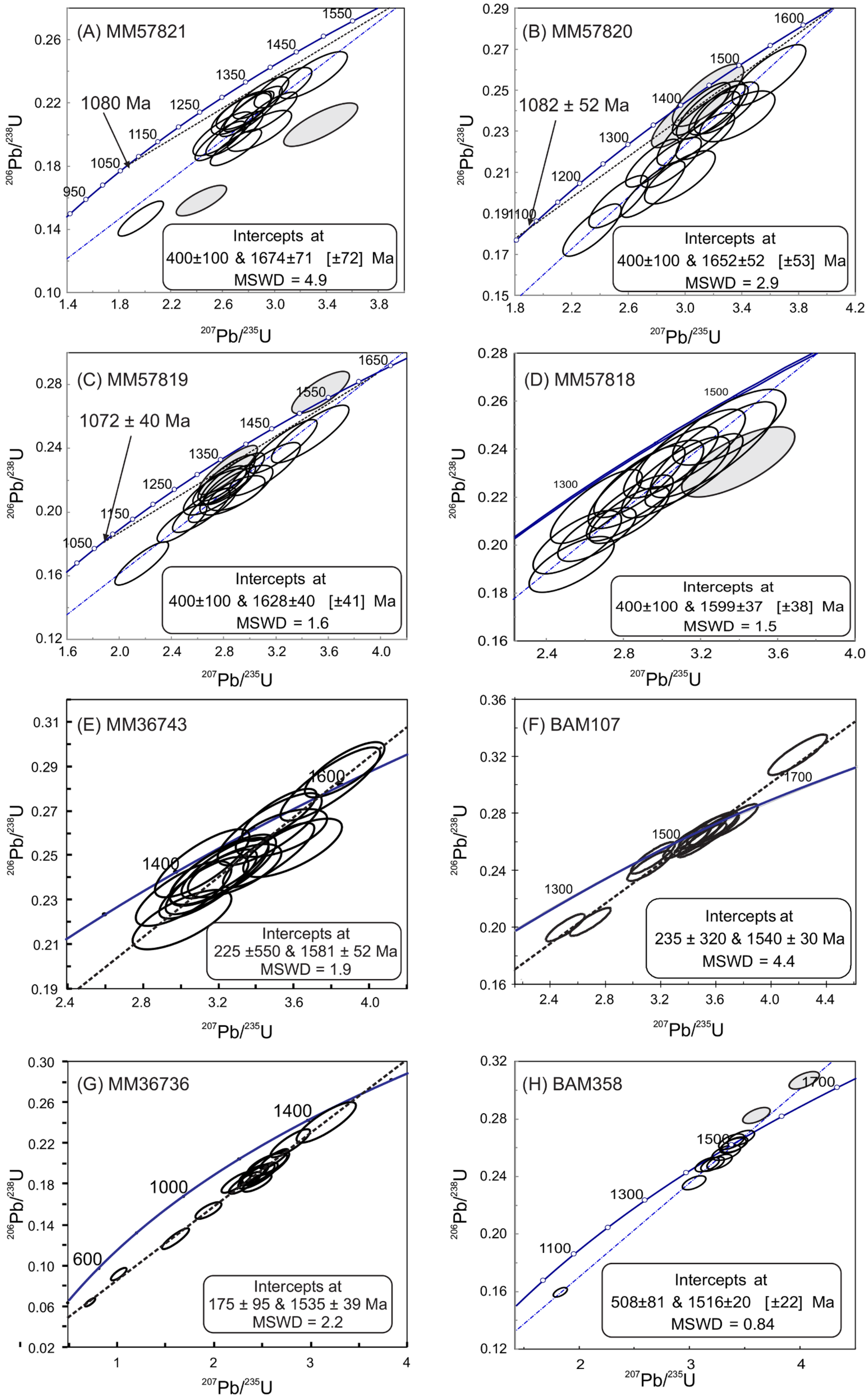

Fig. 5. U–Pb zircon geochronology of pre-Sveconorwegian (ca. 1650–1500 Ma) magmatic events in the Bamble lithotectonic unit. Light-grey ellipses not included in age determinations.

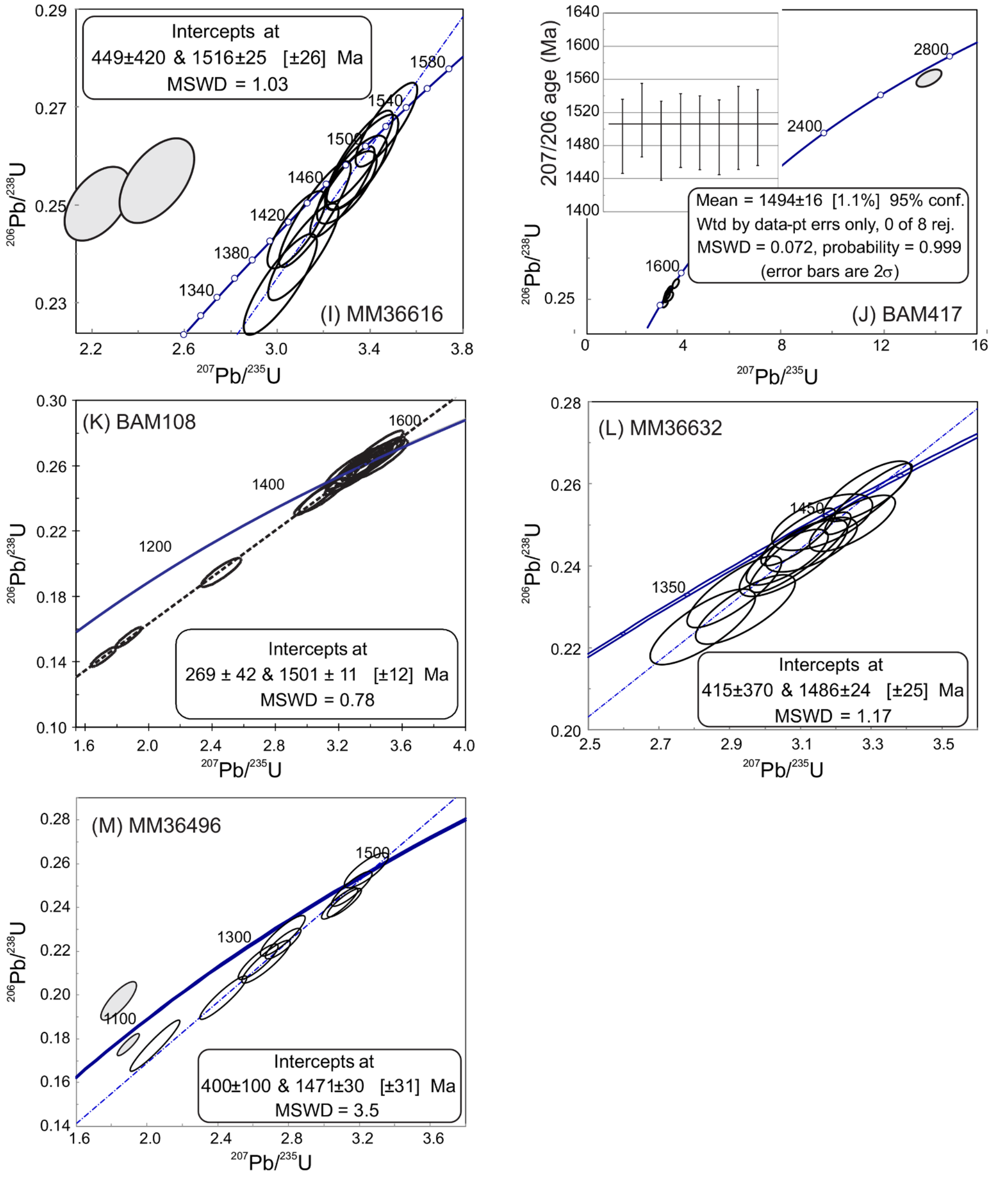

Fig. 5. (*continued*)

with low Th/U (< 0.1) interpreted to be metamorphic. These textural differences and their age significance have been documented previously (Coint et al., 2015; Slagstad et al., 2018b).

The data and cross-cutting relationships observed in the field (Slagstad et al., 2018b) suggest that the metasedimentary units can be subdivided in two main groups: 1) an older group consisting of metapelites in both the Telemark and Bamble LUs and quartz-rich, nodular sillimanite gneisses in the Bamble LU with dominant ca. 1500 Ma populations and sparse grains as old as ca. 2000 Ma, and 2) a younger group consisting of the Faurefjell metasediments in Rogaland that yielded detrital zircon ages ranging from ca. 2000 to 1150 Ma (Fig. 11 Table 1). The older group of metasedimentary rocks is consistent with widespread sedimentation during and following Gothian and Telemarkian orogenesis, characterised by a long period of magmatic quiescence until ca 1280 Ma. The younger group of metasedimentary rocks in Rogaland has a similar age distribution as the much better studied Telemark supracrustals (Bingen et al., 2003; Lamminen, 2011; Spencer et al., 2014), suggesting widespread sedimentation in units west of the Bamble LU.

5. Discussion

5.1. Pre-Sveconorwegian configuration of the Fennoscandian margin

A key factor for understanding the evolution of any orogen is the pre-orogenic distribution of the various units that make up the orogenic belt. This is typically achieved by considering the pre-orogenic magmatic, metamorphic and depositional histories of the various units in the system. Although some workers have suggested that the westernmost units in the Sveconorwegian Province may be exotic to Baltica (Bingen et al., 2008b, 2005; Möller et al., 2015), new data presented here from orthogneisses in the Bamble LU (ca. 1650–1500 Ma), as well as recent data from the correlative Kongsberg LU (Bingen and Viola, 2018), appear to fill the age gap between the ca. 1800–1550 Ma rocks that make up the autochthon and overlying Idefjorden LU in the east of the prvince (Åhäll and Connelly, 2008), and the ca. 1560–1480 Ma rocks of the Telemarkian orogen in the west of the province (Bingen et al., 2005; this study; Roberts et al., 2013).

Fig. 12 shows a compilation of nearly 600 U–Pb zircon ages interpreted to reflect igneous crystallisation, and about 80 U–Pb zircon and

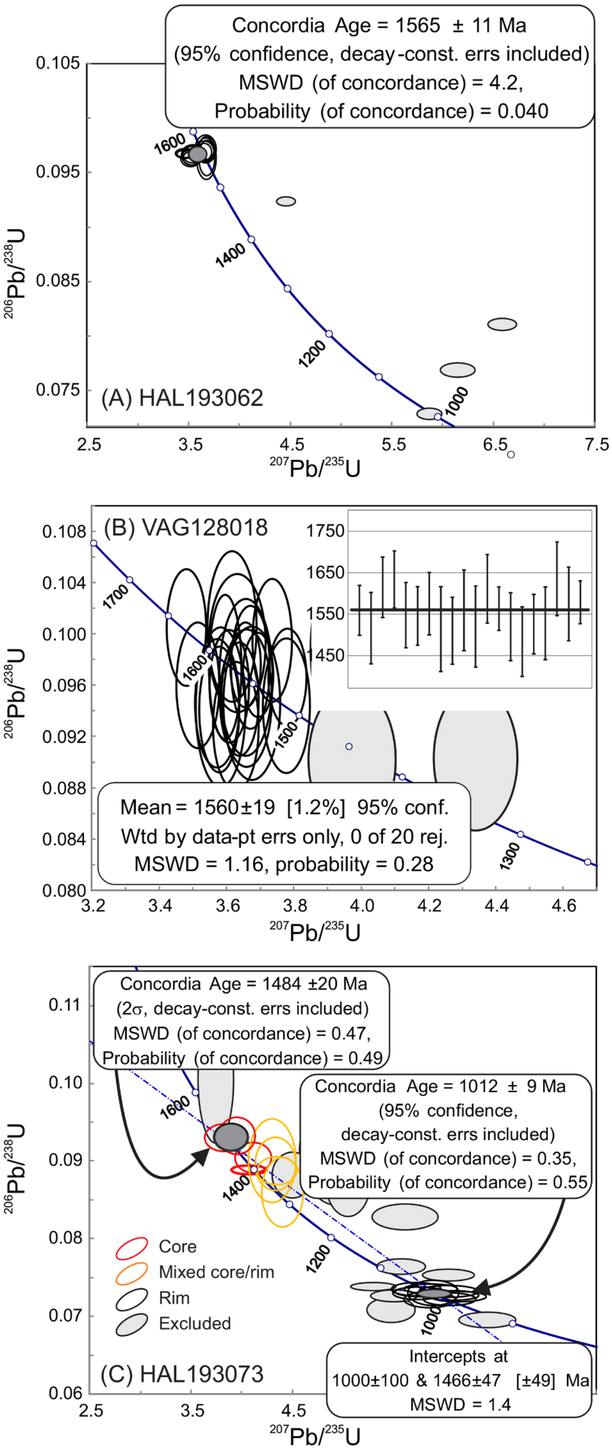

Fig. 6. U–Pb zircon geochronology of pre-Sveconorwegian (ca. 1650–1480 Ma) magmatic events in the Telemark lithotectonic unit, and age data from leucosome from one sample (C). Light-grey ellipses not included in age determinations.

monazite ages interpreted to reflect metamorphism, from east to west across the orogen. The oldest rocks, up to ca. 1800 Ma, are found in the Eastern Segment, but all parts of the orogen show apparently continuous activity until ca. 1500 Ma, with younger ages dominating westwards. A possible interpretation is, therefore, that the bulk of the rocks that make up the Province formed at the southwestward-growing Fennoscandian margin before ca. 1480 Ma (Åhäll and Connelly, 2008; Bingen and Viola, 2018; Roberts and Slagstad, 2015; Fig. 13 A). After ca. 1480 Ma, magmatic and metamorphic activity ceased in units west of and structurally above the west-dipping Mylonite Zone but continued in the Eastern Segment and the autochthon until ca. 1420 Ma, in what is generally known as the Hallandian orogeny. The Hallandian is characterised by apparently subduction-related magmatism (Brander and Söderlund, 2009) and migmatisation at peak pressure–temperature conditions of 6 kbar and 720–750 °C down to 4 kbar and 675 °C (Bogdanova et al., 2014). Although the Hallandian orogeny remains poorly understood in terms of geodynamic setting (Brander and Söderlund, 2009; Möller et al., 2007; Roberts and Slagstad, 2015), the style of magmatism and metamorphism led Wahlgren and Stephens (2020) to suggest a non-collisional, accretionary orogenic system. The fact that coeval orogenic events are not recorded in domains to the west has been used to argue that the crust west of the Mylonite Zone was not adjacent to the Eastern Segment prior to the Sveconorwegian orogeny (Andersson et al., 2002), apparently at odds with the pre-1500 Ma continuity between the domains. The N–S continuity of the late Palaeoproterozoic belts means that the units presently exposed west of the Mylonite Zone could have been transported laterally for some distance, consistent with sinistral transpression along the Mylonite Zone (Stephens et al., 1996; Viola et al., 2011). Considering evidence that the Telemarkian orogeny terminated with extension at ca. 1500 Ma (Menuge and Brewer, 1996; Slagstad et al., 2009) and that Hallandian orogeny initiated with extension at ca. 1470 Ma (Brander, 2011), it is possible that the Idefjorden and more westerly lithotectonic units separated from the Eastern Segment during this long-lived extensional event. In that case, the Hallandian event may reflect continued active-margin processes following a reorganisation or reorientation of the subduction zone (Roberts and Slagstad, 2015; Wahlgren and Stephens, 2020), or alternatively, a re-accretion of the Idefjorden and more westerly lithotectonic units onto the Fennoscandian margin. Although, further work is needed to unravel the nature and significance of the Hallandian event, current interpretations largely agree that it formed as part of the active, Palaeo- through Mesoproterozoic evolution of the Fennoscandian margin.

The Hallandian period was a magmatically quiescent period in units west of the Mylonite Zone, until the onset of bimodal magmatism at ca. 1350 Ma, which lasted until ca. 1280 Ma in the Idefjorden LU, 1140 Ma in the Bamble LU and 1130 Ma in the Telemark LU (Fig. 13B, C, D). Emplacement of mafic dykes farther northeast in Fennoscandia between ca. 1280 and 1230 Ma suggests geographically widespread extension at this time, with several authors arguing for a back-arc setting (Brewer et al., 2004; Söderlund et al., 2006). Termination of magmatic activity in the Idefjorden LU at 1280 Ma may reflect separation of the Idefjorden LU from the Bamble and Telemark LUs at that time (Fig. 13B). Adding credence to this interpretation is the identification of the ca. 1200 Ma Tromøy arc in the Bamble LU (Knudsen and Andersen, 1999), which suggests the presence of at least some oceanic crust between the two units. We note, however, that the recent failure to reproduce this age (Bingen and Viola, 2018) warrants some caution with respect to relying extensively on this result.

In the Bamble LU, bimodal magmatism ended in a high-grade metamorphic event at 1140 Ma, traditionally taken as the onset of the Sveconorwegian orogeny, whereas in the Telemark LU extension and sedimentation continued until at least 1110 Ma (Spencer et al., 2014; Fig. 13D), i.e., at least 30 Myr longer than in the presently adjacent Bamble LU. Here, we postulate that by 1150 Ma the long-lived extension had led to separation of the Bamble and Telemark LUs, and that this separation is the reason for the contrasting tectonic evolution of the two domains between 1140 and 1100 Ma. We return to this below.

Detrital zircon data from the Bamble LU metasediments and the Gyadalen metapelites in Rogaland, in the western part of the Telemark LU, document sedimentation from the Bamble through the Telemark LUs at or after ca. 1450 Ma, derived from a dominant 1500 Ma source (Figs. 13A and 14C, D; Nijland et al., 2014). Minimum ages of deposition are not well constrained, but in Rogaland, in the western part

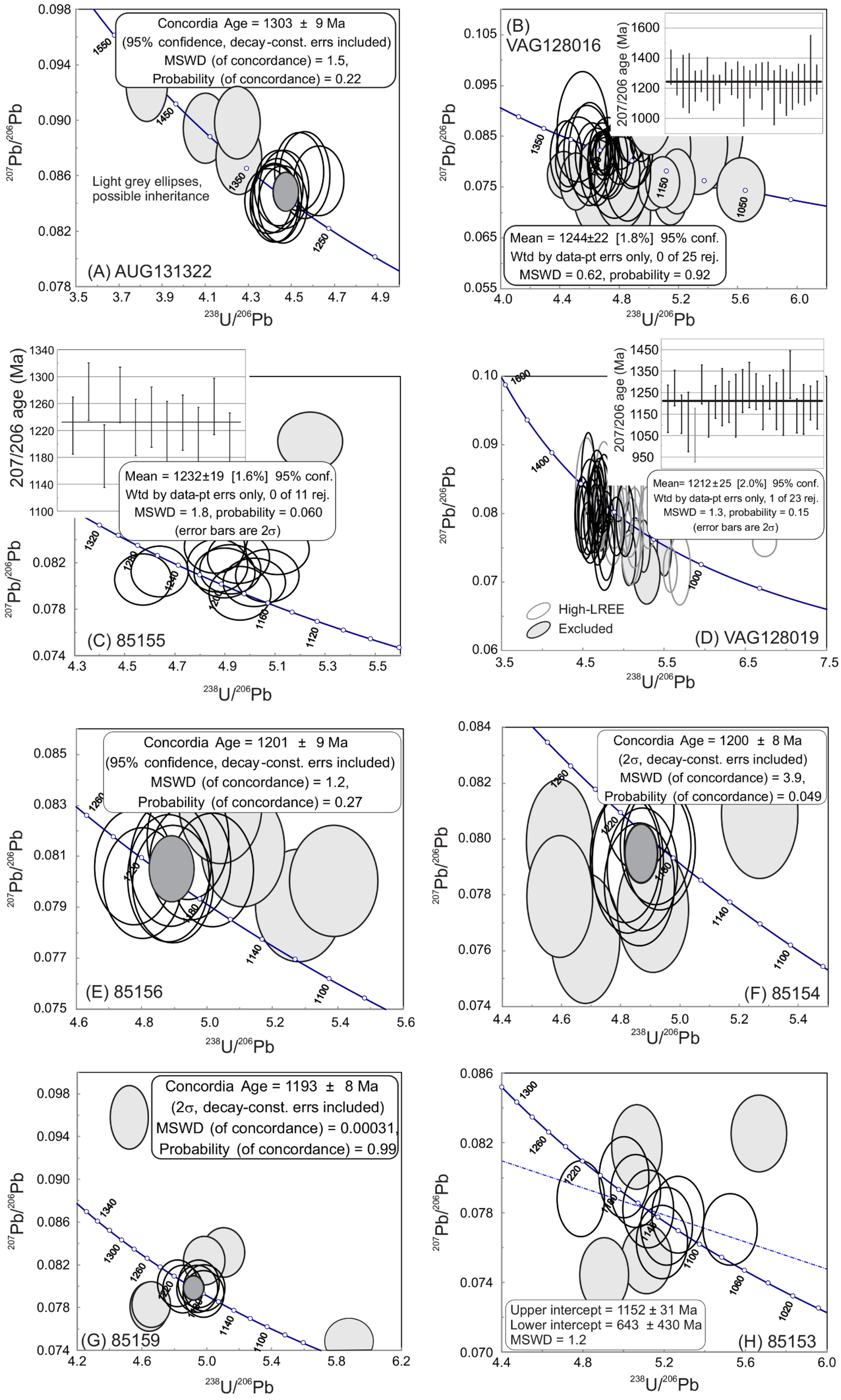

Fig. 7. U–Pb zircon geochronology of pre- to early-Sveconorwegian (ca. 1300–1140 Ma) magmatic events in the Telemark lithotectonic unit. Light-grey ellipses not included in age determinations.

of the orogen, the Gyadalen metapelites are intruded by 1200 Ma granitoids and in the Bamble LU deposition must have taken place before the 1140 Ma high-grade metamorphic event. Considering the scarcity of zircons younger than ca. 1400 Ma, despite well-documented voluminous magmatism until 1150 Ma, it seems likely that deposition took place closer to the maximum than the minimum age of deposition.

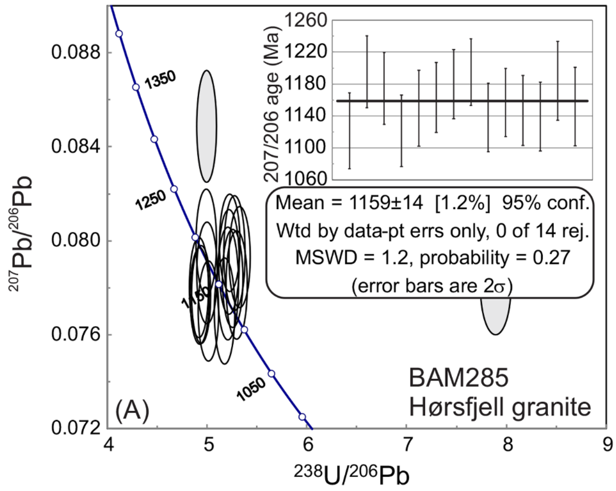

Fig. 8. U–Pb zircon geochronology of pre- to early-Sveconorwegian Hørsfjell granite in the Bamble lithotectonic unit. The granite shows clear cross-cutting relationships with migmatitic gneisses, supporting earlier interpretations that the Bamble lithotectonic unit rocks underwent pre-Sveconorwegian (i.e., ca. 1650–1500 Ma Gothian or Telemarkian high-grade metamorphism). Light-grey ellipses not included in age determinations.

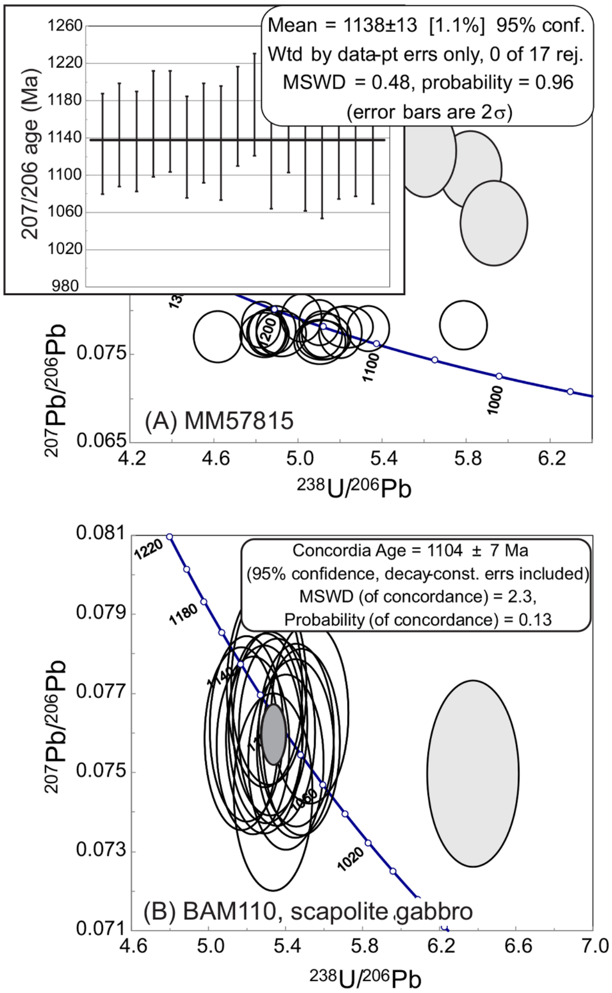

Fig. 9. U–Pb zircon geochronology of early-Sveconorwegian high-grade metamorphism (A) and hydrothermal alteration and scapolitisation (B) in the Bamble lithotectonic unit. Light-grey ellipses not included in age determinations.

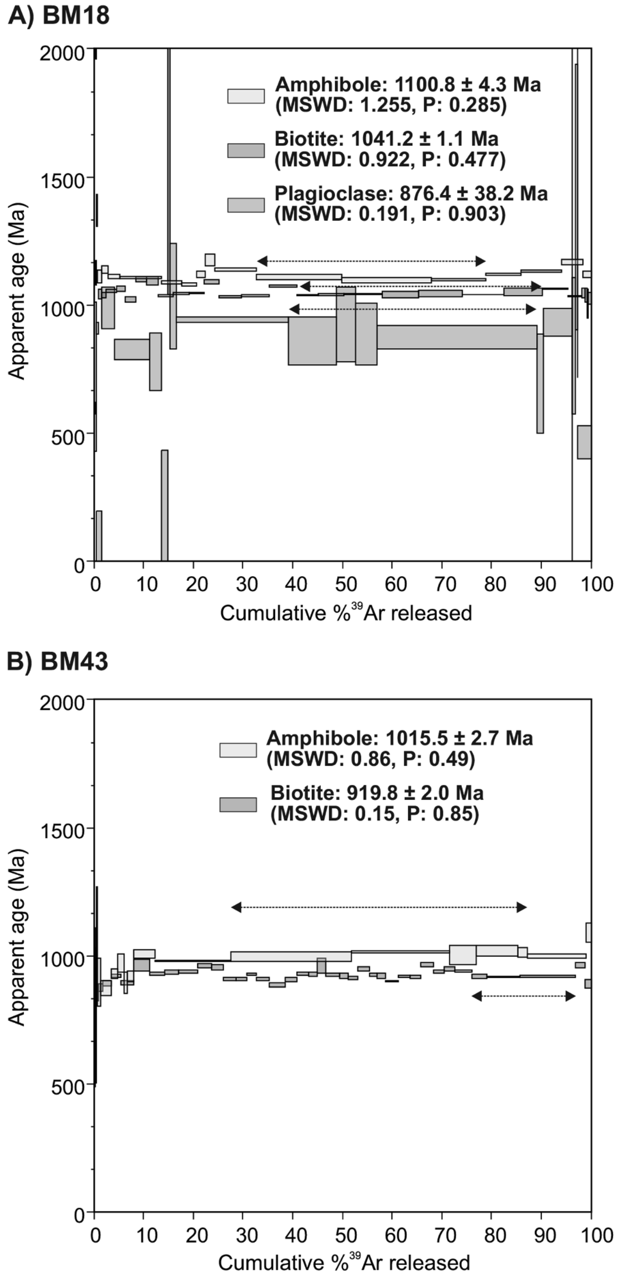

Fig. 10. Stacked ^{40}Ar–^{39}Ar degassing spectra from a sample in the interior of the Bamble lithotectonic unit, structurally above the internal shear zone (BM18, hornblende, biotite and plagioclase) and another from the high-strain zone separating the Bamble and underlying Telemark lithotectonic units (BM43, hornblende and biotite).

The metasedimentary protoliths received detritus from sources as old as 1900 Ma (or alternatively recycled from older sedimentary sequences), consistent with proximity to the Eastern Segment at the time of deposition and in line with the magmatic record suggesting proximity of all Sveconorwegian domains until 1480 Ma. A more far-travelled, Laurentian origin for some of the Sveconorwegian units was proposed by Lamminen and Köykkä (2010), citing a paucity of 1650–1500 Ma detrital zircons corresponding to the 'Laurentian gap' (Gower and Krogh, 2002). However, more recent data presented by Spencer et al. (Spencer et al., 2014) and herein (Fig. 14), do not support the existence of such a gap in the detrital zircon record in the Sveconorwegian Province. On a more general basis, Slagstad and Kirkland (2017) showed that detrital zircons are non-diagnostic with respect to Fennoscandian and Laurentian sources. The available detrital zircon data from the

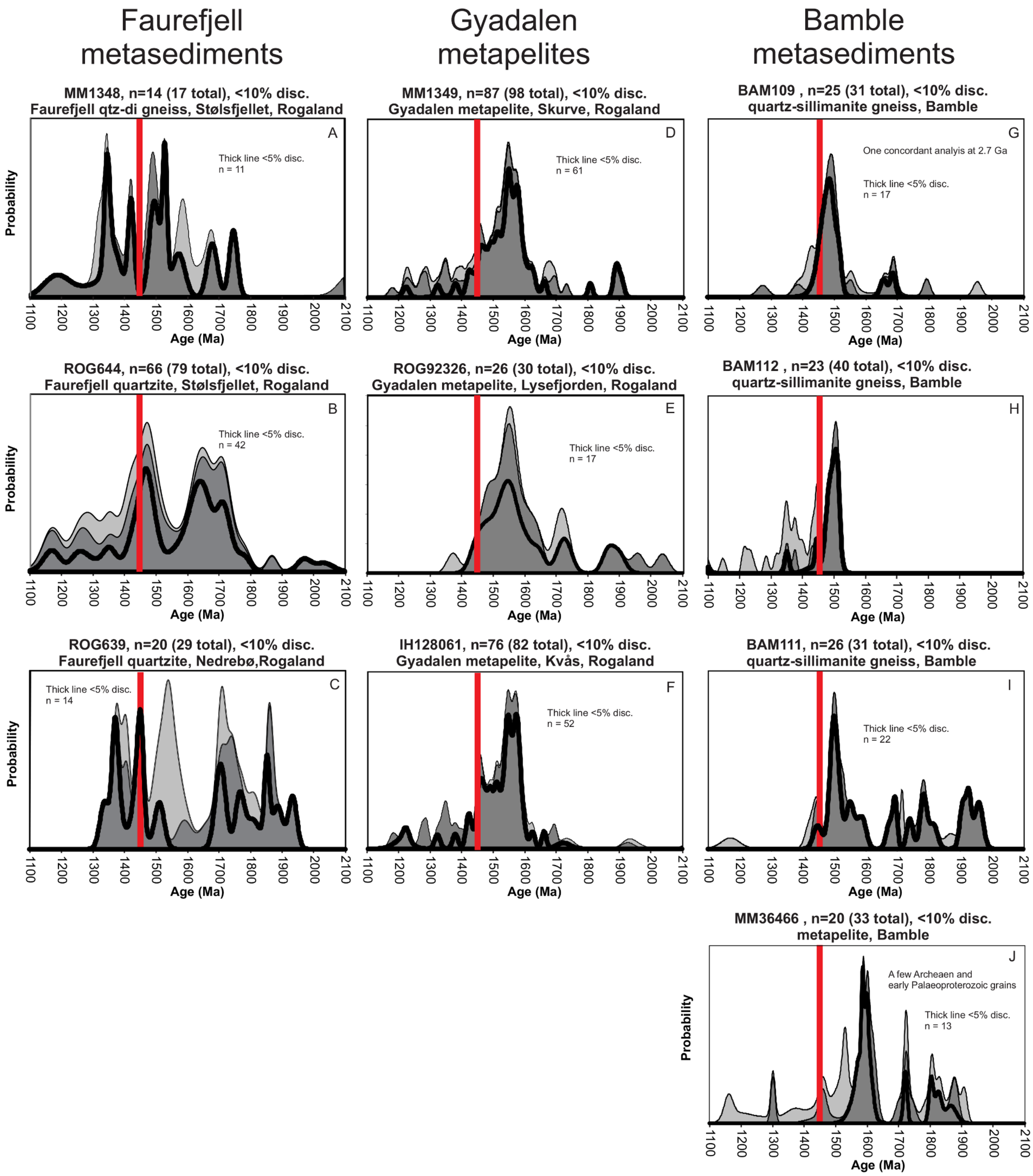

Fig. 11. Probability plots showing detrital zircon age data from metasedimentary units in the Bamble and western part of the Telemark lithotectonic units. Thick, red line at 1450 Ma for reference.

Sveconorwegian Province permit derivation from Fennoscandian sources, consistent with the magmatic record suggesting continuous growth at the Fennoscandian margin, with no need for invoking an exotic origin for any of the units in the Province.

Widespread deposition continued in the Telemark LU until at least 1160–1100 Ma, the age of the youngest detrital zircons in the central and western Telemark LU (Spencer et al., 2014; Fig. 14A, B) and bimodal magmatism in the Telemark LU continued until 1130 Ma. These observations underscore the difference in tectonic environment observed in the Telemark and Bamble LUs; the former in extension until at least 1110 Ma, whereas the latter underwent high-grade metamorphism, generally attributed to the onset of compressional tectonics in the Sveconorwegian Province, from ca. 1140 Ma.

In summary, the available data on the magmatic, metamorphic and depositional history of the different units comprising the Sveconorwegian Province suggest that, although they all formed along the pre-orogenic Fennoscandian margin, their evolution leading up to orogenesis was different. These observations support the notion of a dismembered margin prior to the onset of Sveconorwegian orogenesis (Slagstad et al., 2017), which has major implications for how the ensuing orogeny may have unfolded. Similar continental fragmentation behind active margins has been demonstrated both geologically and numerically, with the SW Pacific presenting the perhaps best examples (Crawford et al., 2003; Rey and Müller, 2010; Sdrolias et al., 2003).

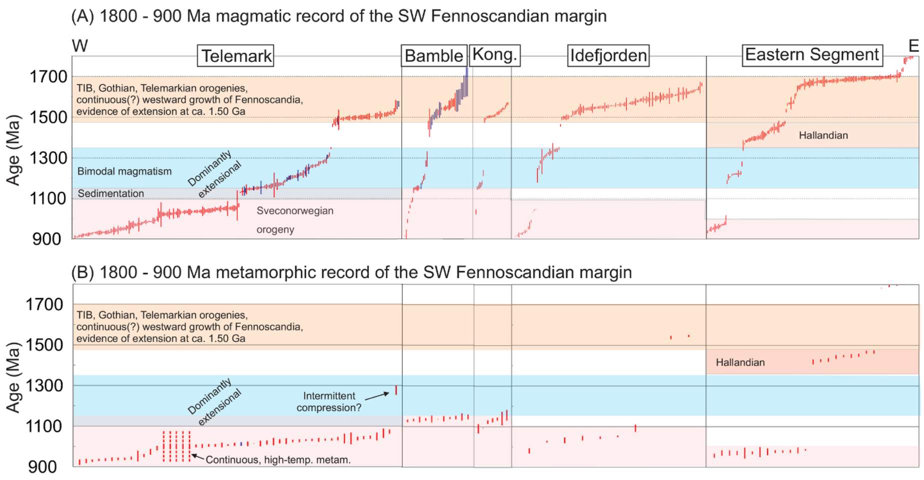

Fig. 12. Compilation of approximately 600 U–Pb zircon crystallisation ages and 80 U–Pb zircon and monazite metamorphic ages previously published from across the Sveconorwegian orogen. Dark blue error bars show new age data presented here. The compilation demonstrates temporal contrasts in magmatism between the different domains constituting the Sveconorwegian orogen, and a first-order interpretation is that these contrasts define periods of separation. This interpretation is consistent with the inference that much of the generally bimodal magmatic activity between ca. 1480 and 1100 Ma apparently represents an extensional (i.e., rifting) tectonic setting. Black, vertical bars represent periods of separation, whereas the grey bar represents possibly separation. The compilations, including references, can be obtained from the corresponding author on request.

5.2. Sveconorwegian reamalgamation of the dismembered SW Fennoscandian margin

An important premise in models of Sveconorwegian orogenesis favouring continent–continent collision is that the Fennoscandian margin was a single crustal entity that collided with another continent-scale crustal entity, resulting in crustal thickening and, in turn, high-grade metamorphism and associated magmatism. The recent model proposed by Bingen and Viola (2018) and Bingen et al. (2018) is also construed as a collisional model, although it is unclear what exactly the role of the collision is in this gravitationally driven orogeny. As discussed above, there was ample time for crustal thinning and rifting during the almost 250 Myr of dominant extensional tectonics between ca. 1340 and 1100 Ma. Thus, geographical separation between different lithotectonic units prior to Sveconorwegian orogenesis is a plausible hypothesis to account for the contrasting magmatic, metamorphic and depositional histories observed in the different units. With this in mind, and in part based on the new data presented herein, we discuss how the reamalgamation may have taken place, focusing on the first ca. 100 Myr of orogenic evolution, between ca. 1150 and 1050 Ma, in the central parts of the orogen.

5.3. 1140 Ma high-grade metamorphism in the Bamble lithotectonic unit; the onset of reamalgamation

High-grade metamorphism in the Bamble LU at ca. 1140–1120 Ma at maximum pressure–temperature conditions of 11.5 kbar and temperatures of > 850 °C (Engvik et al., 2016) is traditionally taken to represent the onset of Sveconorwegian convergence. As discussed above, this high-grade event is coeval with ongoing sedimentation in the Telemark LU, suggesting a distinctly different, extensional tectonic regime there.

The cause of 1140 Ma high-grade metamorphism in the Bamble LU is inherently difficult to ascertain. It is well established that deformation in the Bamble LU was associated with progressive NW-directed thrusting (Henderson and Ihlen, 2004); however, widespread sedimentation in the Telemark LU footwall suggest it was in extension for at least another 30 Myr. It is, therefore, difficult to see how the high-grade event at 1140 Ma could reflect thrusting of the Bamble LU onto the Telemark LU. An alternative interpretation, discussed below, is that deformation and high-grade metamorphism initiated while the Bamble LU was separated from the Telemark LU, and that progressive deformation brought the two units into contact later. The hanging wall of the Bamble LU is concealed by the Skagerrak Sea and the Permian Oslo Graben, which hampers insight into the nature of the rocks that most likely were thrust onto the Bamble LU and caused the 1140 Ma high-grade event.

The Idefjorden LU is a major constituent of the east-central Sveconorwegian Province and is typically interpreted to have undergone two major high-pressure events at 1046 ± 6 and 1026 ± 5 Ma (Söderlund et al., 2008), prior to eastward thrusting over the Eastern Segment at ca. 990 Ma. Pressure–temperature conditions for these metamorphic events in the Idefjorden LU are estimated at 15 kbar and 740 °C and 10 kbar and 700 °C, respectively, and interpreted to reflect imbrication in the Idefjorden LU (Fig. 13F; Söderlund et al., 2008). Apart from these dates, geochronological evidence of high-grade Sveconorwegian metamorphism is comparatively sparse; however, two older metamorphic ages of 1117 ± 23 and 1091 ± 19 Ma are recorded closer to the tectonic contact with the underlying Kongsberg LU, (Bingen et al., 2008a; Bingen and Viola, 2018), and could be interpreted to reflect thrusting of the western part of the Idefjorden LU onto the Konsgberg and Bamble LUs (Fig. 13D). Albeit with large age uncertainties, these ages resemble those for high-grade metamorphism of the Bamble LU. Thus, at present, we interpret the onset of Sveconorwegian convergence to be related to W-directed thrusting of parts of the Idefjorden LU or another, unknown unit onto the Bamble LU, at a time when the latter was geographically separated from the underlying Telemark LU. The tectonic setting and processes responsible for the onset of orogeny in the eastern parts of the Sveconorwegian orogen are poorly understood; however, the possible existence of hyperextended lithospheric domains, formed during long-lived extension, and the apparent lack of arc-like magmatism may indicate a processes akin to Ampferer-type subduction, as suggested for parts of the Alps (e.g., McCarthy et al., 2018, 2020). Further work in the units making up the central and eastern parts of the orogen is needed to resolve the tectonic processes responsible for the onset of orogenic activity.

5.4. 1100 Ma differential cooling of the northeastern Bamble lithotectonic unit constrains Bamble–Telemark lithotectonic unit juxtaposition; implications for metasomatic processes and fluid sources

New mapping from the northeastern part of the Bamble LU suggests that the unit may represent a thrust stack, assembled over a period of > 199 Myr, with different parts of the stack undergoing and recording different events at different times, as indicated by evidence discussed next. More work is needed to confirm and provide details with respect to the evolution of the Bamble LU, and in the ensuing discussion the following, testable assumptions are made (see also Fig. 2A): 1) the northeastern Bamble LU is separated into two domains by the internal and boundary shear zones; 2) hydrothermal activity is restricted to rocks structurally above (southeast of) the internal shear zone; 3) ca. 1200–1140 Ma granitic rocks from the Telemark LU extend across and are deformed by the boundary shear zone, but are absent above the internal shear zone; 4) early-Sveconorwegian, ca. 1140 Ma

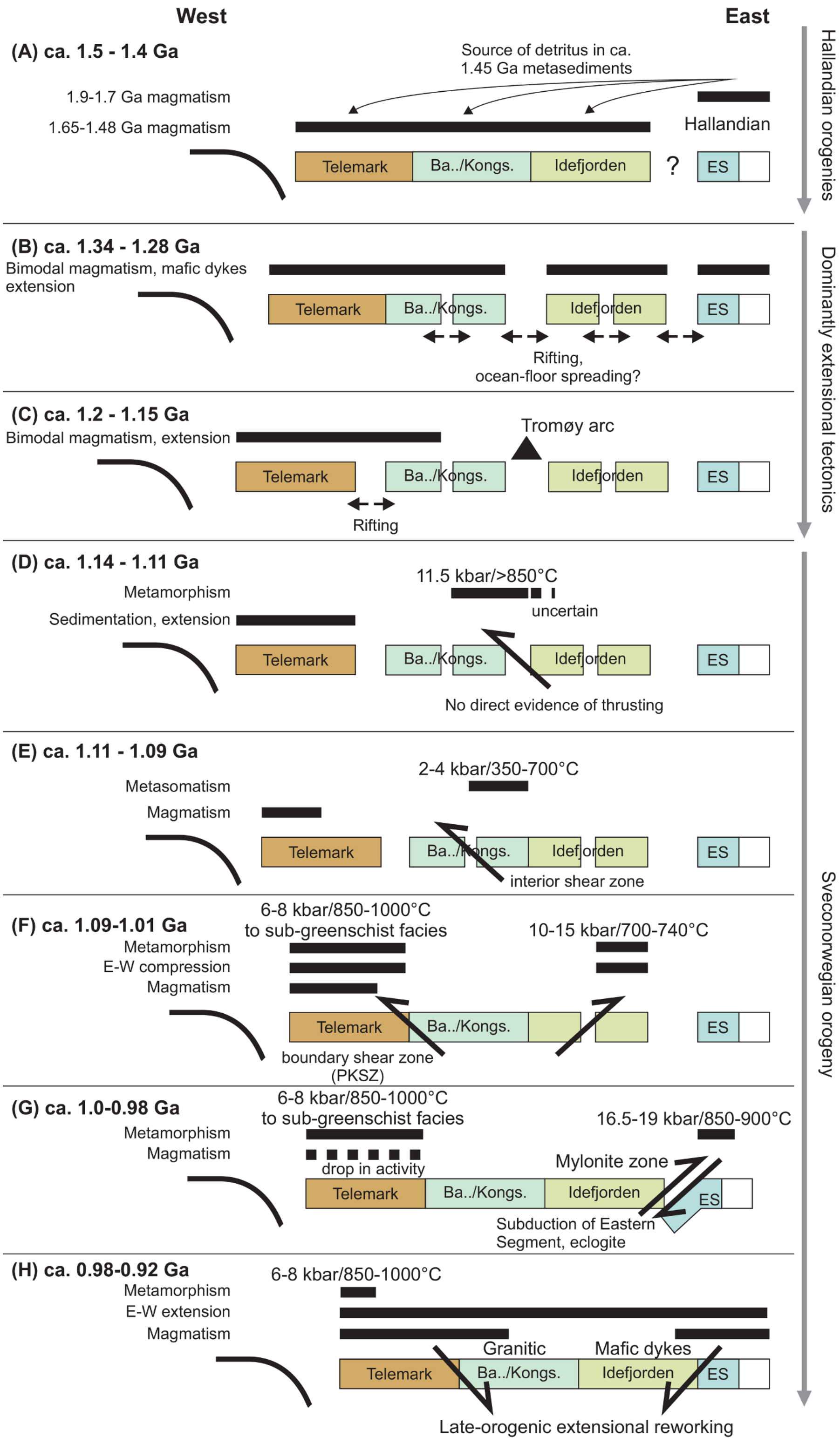

Fig. 13. Cartoon of the pre- through Sveconorwegian tectonic evolution of SW Fennoscandia. The thick, black lines illustrate when processes such as magmatism, metamorphism, deformation and sedimentation were operative in different parts of the orogen. The thick, stippled lines indicate poor control, e.g., due to few data or uncertainties regarding regional significance. The main points of the tectonic model are (see text for details and references): (A) Accretionary tectonics, including the Telemarkian and Hallandian orogenies, appear to have dominated the southwestern Fennoscandian margin until ca. 1400 Ma. Although opinions diverge regarding the Fennoscandian affinity of the various units, the continuous magmatic record and detrital input support an origin close to Fennoscandia. (B) and (C) Between ca. 1340 and 1150 Ma, widespread bimodal magmatism suggests an extensional tectonic regime, with several authors arguing for a continental back-arc setting. Oceanic crust may have formed locally, as indicated by the 1200 Ma Tromøy arc in the Bamble lithotectonic unit. (D) At 1140 Ma, parts of the Bamble lithotectonic unit underwent high-grade metamorphism, whereas other parts of the orogen were either still in extension (Telemark lithotectonic unit) or quiescent (Idefjorden lithotectonic unit, Eastern Segment). (E) and (F) Between ca. 1100 and 1000 Ma, all parts of the orogen, apart from the Eastern Segment, were in compression, with local high-pressure metamorphism in the east (Idefjorden lithotectonic unit) and voluminous magmatism and high-temperature metamorphism in the west (Telemark lithotectonic unit). The Bamble lithotectonic unit appears to have been located at shallow crustal levels at this time, with little or no evidence of the processes taking place elsewhere in the orogen. (G) The last, major compressional event to take place in the Sveconorwegian orogen involved deep subduction of the Eastern Segment under the Idefjorden lithotectonic unit, resulting in eclogite-facies metamorphism in the former. (H) From ca. 990 Ma, the entire orogen, including the foreland east of the Eastern Segment was in extension until ca. 920 Ma. This phase might have resembled the long extensional period prior to the onset of the Sveconorwegian orogeny, and the duration and lack of preceding, widespread crustal thickening suggests an external extensional force, for example a retreating active margin. The evolution can be accounted for by modern-style plate tectonics dominated by horizontal plate movements and does not require more esoteric vertical tectonics, akin to those sometimes inferred for some Archaean regions. The Bamble–Kongsberg lithotectonic unit is split in two to accommodate the tectonometamorphic evolution observed in the Bamble lithotectonic unit; it is not clear how or if this translates to the Kongsberg lithotectonic unit. Similarly, the Idefjorden lithotectonic unit is split in two to accommodate high-pressure metamorphism at 1046 and 1025 Ma, interpreted to reflect assembly of the eastern and western parts of the unit.

high-grade metamorphism is confined to rocks above the internal shear zone.

The new U–Pb zircon data presented here support previously published age data arguing for high-grade metamorphism in the hanging wall to the internal shear zone around 1145–1135 Ma. In addition, a younger metamorphic event is recorded at 1104 ± 7 Ma (U–Pb zircon) in a scapolite gabbro near Ødegården verk, well known for ubiquitous fluid activity and metasomatism before ca. 1090 Ma (U-Pb rutile, Engvik et al., 2011). Table 2 and Fig. 15A show a compilation and plot, respectively, of post-peak metamorphic age data from the northeastern part of the Bamble LU, structurally above the internal shear zone, including U–Pb zircon, titanite and rutile, Ar–Ar hornblende, biotite,

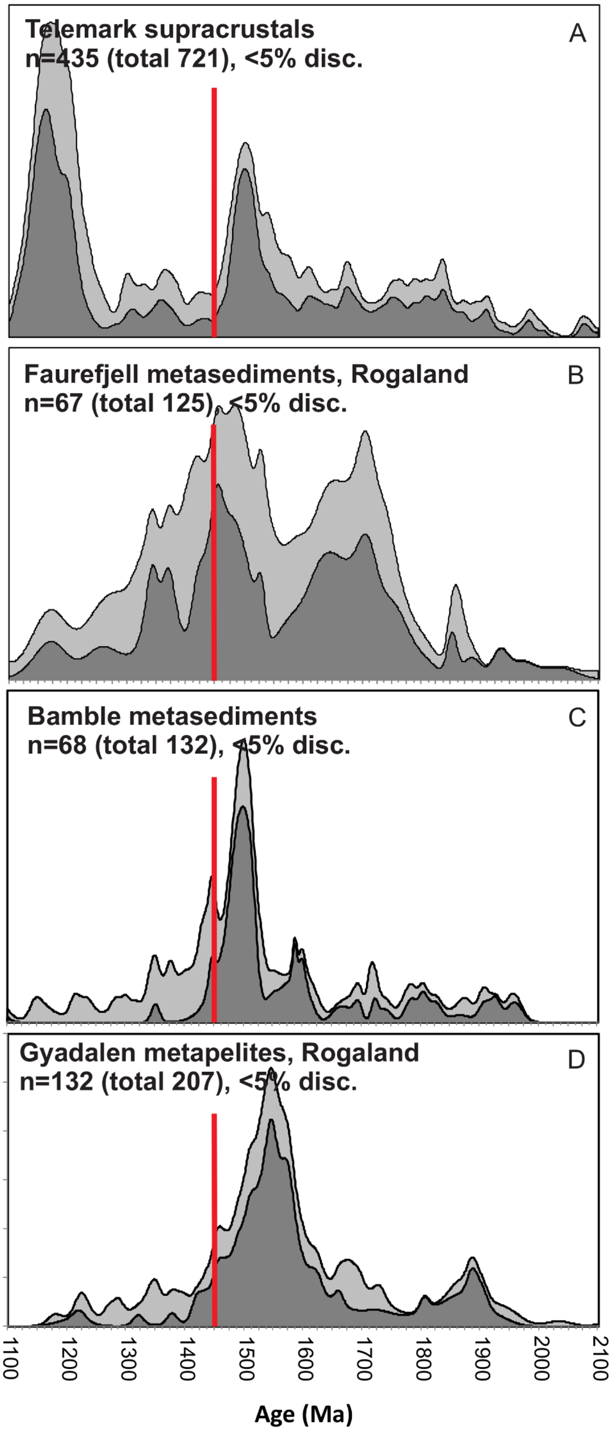

Fig. 14. Compilation of detrital zircon data from the Bamble and Telemark lithotectonic units. The data from the Telemark supracrustals are from Spencer et al. Thick, red line at 1450 Ma for reference.

plagioclase, and Rb–Sr phlogopite data presented here, in Engvik et al. (2011), Cosca and O'Nions (1994) and Cosca et al. (1998). The plot assumes onset of metasomatism (zircon from scapolite gabbro) at 1104 ± 7 Ma at a temperature of 650 ± 50 °C, which is in the higher end of the estimated temperature range for the metasomatic activity (350–700 °C, 2–4 kbar; Engvik et al., 2011; Nijland and Touret, 2001). The rest of the plot assumes temperatures similar to more or less well-established closure temperatures for U–Pb in rutile, Ar–Ar in hornblende, biotite and plagioclase, and Rb–Sr in phlogopite (Table 2). The resulting time–temperature plot (Fig. 15A) shows cooling from ca. 600–700 °C at 1104 Ma to 400–450 °C at ca. 1090–1070 Ma, which translates to a cooling rate of up to 10–12 °C/Myr (Fig. 15A). No thermochronological data exist for areas structurally below the internal shear zone, but extrapolation from low structural levels farther southwest along the PKSZ (Bingen et al., 1998; this study) suggests a much younger cooling history there (ca. 1030–1015 Ma) than above the internal shear zone, indicating a major break in tectonothermal history across this shear zone

Structural observations farther southwest in the Bamble LU, close to the Telemark LU and probably at a structural level similar to our boundary shear zone, show that NW-directed thrusting of the Bamble LU onto the Telemark LU may have been a protracted, possibly episodic process (Henderson and Ihlen, 2004). Unfortunately, this shearing has not been dated directly; pegmatites at higher structural levels yield ages of 1094 and 1060 Ma (Baadsgaard et al., 1984; Scherer et al., 2001), but their relationship to pegmatites at lower structural levels is unclear. Ar–Ar hornblende ages between 1030 and 1015 Ma close to the PKSZ (Bingen et al., 1998; this study) are consistent with cooling following the peak of high-grade metamorphism in the Telemark LU footwall, and are probably the presently best lower age estimate for final thrusting of the Bamble LU onto the Telemark LU (Figs. 13F, 15D). There is currently no evidence for high-grade metamorphism in the Telemark LU until after ca. 1035 Ma (Bingen et al., 2008b), consistent with the 1012 ± 9 Ma age of migmatisation recorded in the central Telemark LU and with the thermochronological data from the Bamble LU indicating final thrusting onto the Telemark LU significantly after high-grade metamorphism in the former.

The shear zone system in the Bamble LU thus appears to be one of the longest-lived structural elements in the Sveconorwegian Province, and further work is clearly warranted in this region, which separates the high-temperature-dominated Telemark LU (800–1000 °C and 6–8 kbar), with voluminous magmatism, from the high-pressure-dominated Idefjorden LU (700–740 °C and10–15 bar) and Eastern Segment (850–900 °C and 16.5–19 kbar).

A characteristic feature of the Bamble LU is widespread metasomatism, including albitisation and scapolitisation (Engvik and Austrheim, 2010; Engvik et al., 2011), possibly related to an influx of saline (seawater) fluids (Engvik et al., 2018; Nijland and Touret, 2001; Nijland et al., 1998). Recent mapping shows that the metasomatic activity is confined to units above the internal shear zone (Marker et al., 2019). Detailed discussions of these metasomatic processes is well beyond the scope of this paper, but the interpretations made regarding the pre-metasomatic (i.e., > 1100–1090 Ma) history of the Bamble and adjacent Telemark LUs may shed some light on potential sources. As discussed above, the metasomatic activity coincided with relatively rapid cooling of the Bamble LU at 1100 Ma, most likely related to thrusting onto Telemark LU-like rocks. Sedimentation in the Telemark LU may have continued right up to the moment of thrusting or even longer, with a maximum age of deposition for the Eidsborg Formation of 1118 ± 17 Ma (Spencer et al., 2014). Thus the lower Telemark LU block could have been replete with fluid-rich sediments that would have dehydrated during overthrusting of the Bamble LU (Fig. 15C). This process could explain the widespread metasomatism and extensive mineralisation in both the Bamble and its correlative Kongsberg LU to the north. A critical point here is the need for improved understanding of the thrust history in the Bamble LU.

5.5. *990 Ma eclogite-facies metamorphism, 1.07–0.93 high-temperature metamorphism and magmatism*

Although yet to be proved conclusively, the different Sveconorwegian units west of the Mylonite Zone (i.e., the Idefjorden and more westerly lithotectonic units) were probably assembled before 1000 Ma, forming a segment of contiguous continental lithosphere on the order of 500 km wide. As discussed in earlier publications (e.g., Slagstad et al., 2017), this entity probably collided with what is now the

Table 2
Age data and closure temperatures used to construct time–temperature evolution of units in the Bamble lithotectonic unit (structurally above the internal shear zone).

Mineral	Isotope system	Significance	Age	Uncert. age	Source Age	Tc	Uncert. Tc	Source Tc
Zircon	U–Pb	Metasomatism	1104	7	2	650	50	6
Titanite	U–Pb	Cooling	1105	3	3	630	30	7
Hornblende	Ar–Ar	Cooling	1113–1083[1]		2, 3, 4	550	50	8
Rutile	U–Pb	Cooling	1087	3	5	420	50	9
Phlogopite	Rb–Sr	Cooling	1055	15	5	435	50	10
Biotite	Ar–Ar	Cooling	1041	1	2	350	35	8
Plagioclase	Ar–Ar	Cooling	876	38	2	260	40	11

[1] Eleven ages. Data sources: [2] This work, [3] Cosca et al. (1998), [4] Cosca and O'Nions (1994), Engvik et al. (2011), [6] Lieftink et al. (1994), [7] Walsh et al. (2013), [8] Quentin de Gromard et al. (2019), [9] Mezger et al. (1989), [10] Willigers et al. (2004), [11] Cassata et al. (2009).
Abbreviations: Uncert. – Uncertainty, Tc – closure temperature.

Eastern Segment, representing the SW margin of Fennoscandia at ca. 990 Ma, causing eclogite-facies metamorphism in the downgoing Fennoscandian margin (Möller, 1998; Fig. 13G), which attained pressure–temperatures conditions of 16.5–19 kbar and 850–900 °C (Tual et al., 2017). This interpretation is compatible with earlier studies proposing that the Mylonite Zone constitutes a major, Sveconorwegian tectonic zone (e.g., Andersson et al., 2002). Following this last compressional event, the entire orogen and its foreland appears to have changed to a largely extensional tectonic regime, as indicated by emplacement of mafic dykes and formation of extensional structures in the eastern and central parts of the orogen (Piñán Llamas et al., 2015; Söderlund et al., 2005), and increased input of mantle melts in the west (Granseth et al., 2020). Thermochronological data from eastern and central parts suggest extension may have lasted until ca. 880–860 Ma (Mulch et al., 2005; Viola et al., 2011), and ca. 850 Ma mafic dykes in the western Telemark LU (Walderhaug et al., 1999) indicate that extension may have lasted on the order of 100 Myr (Fig. 13H).

By ca. 1070 Ma, and quite possibly well before (Roberts and Slagstad, 2015; Spencer et al., 2014), an active margin had been established in the western Sveconorwegian orogen; this margin remained active until at least 930 Ma, as shown in Fig. 13 and discussed in several recent publications (Bybee et al., 2014; Coint et al., 2015; Granseth et al., 2020, in review; Slagstad et al., 2018a,b; Slagstad et al., 2018b; Slagstad et al., 2017; Spencer et al., 2019). This interpretation is borne out from geochemical, geochronological, isotopic and metamorphic evidence (see the references above), and supported by the lack of structural and metamorphic evidence for crustal thickening that is capable of explaining the observed > 150 Myr of continuous magmatic and metamorphic processes in the southwest Telemark LU. Instead, these processes appear to have been sustained by ponding of mafic magma under crust of normal thickness, resulting in lower-crustal melting and high- to ultrahigh-temperature metamorphism at pressures around 6–8 kbar. The geochemical evolution of the felsic magmatism is consistent with a lower-crustal source that became increasingly more depleted, dehydrated and hotter. There are no chemical or isotopic data indicative of introduction of a new crustal source, as might be expected during a collisional event; instead, the isotopic data reveal isotopic pull-downs and pull-ups, matching known compressional and extensional periods, respectively, most easily explained by a cyclic, advancing and retreating active margin, akin to the North American Cordillera (DeCelles et al., 2009). Any model of the Sveconorwegian tectonic evolution needs to account for these significant differences between the eastern and western parts of the orogen. Although long-lived ponding of mafic magma may result in the formation of large bodies of eclogite that in turn can delaminate (e.g., Thybo and Artemieva, 2013), we stress that this process is a result of, not a driving force for, orogenesis.

The evolution following cessation of Sveconorwegian orogenic activity is poorly constrained and understood; however, some recent publications have suggested that active-margin processes continued in a more distal, outboard location (Slagstad et al., 2019, 2017).

5.6. Is Sveconorwegian orogenesis by gravity-driven vertical tectonics (sagduction) a viable tectonic model?

The sagduction modelBriefly, the sagduction model (Bingen and Viola, 2018; Bingen et al., 2018) posits growth of a thick subcontinental lithospheric mantle (SCLM) under the southwestern passive(?) margin of Fennoscandia, following the ca. 1500 Ma Telemarkian orogeny. At 1280 Ma, an active margin west of the present-day coast of Norway was re-established and the thick SCLM under the Telemark LU started to delaminate and founder, resulting in bimodal magmatism. The delamination–foundering process continued until about 1150 Ma and involved formation of an orogenic plateau, with intramontane basins in which the Telemark supracrustals were deposited. The extension was caused by an east-directed roll back of the delaminating SCLM. At 1140 Ma, the delaminating SCLM had migrated to beneath the Bamble LU where, in contrast to the Telemark LU, it drew down the overlying Bamble LU crust with it, thereby causing high temperature/moderate pressure granulite-facies metamorphism until ca. 1100 Ma. At that time, the SCLM dripped off and sank causing the Bamble LU crustal rocks to be exhumed and thrust onto the Telemark LU. These process were all inferred to have taken place behind an active margin. The ensuing orogenic evolution was characterised by an eastward progression of high-pressure metamorphism, from ca. 1150 Ma in the Bamble LU, through 1050–1020 Ma in the Idefjorden LU, culminating at 990 Ma in the Eastern Segment. The driving force of this orogenic evolution was inferred to be continued delamination and foundering of the SCLM. Farther west in the orogen, high- to ultrahigh-temperature metamorphism between 1030 and 925 Ma was inferred to have been caused by upwelling asthenosphere under the orogenic plateau.

5.7. Was there a Sveconorwegian orogenic plateau?

Orogenic plateaus can form where compression, crustal thickening and accompanying radiogenic and minor shear heating result in widespread lower- and/or middle-crustal melting (Chen et al., 2019), which in turn permit flow that eliminates lateral pressure gradients and thus reduces variations in internal topography and crustal thickness (e.g., Clark and Royden, 2000; Culshaw et al., 2006; Jamieson and Beaumont, 2013). Orogenic plateaus can form in different tectonic settings, and the best-known plateaus – the Tibetan plateau and the Altiplano–Puna plateau – formed in continent–continent collisional and accretionary settings, respectively (McQuarrie et al., 2005; Royden et al., 1997). Some of the best studied examples of ancient orogenic plateaus, allowing direct access to the lower, middle and upper orogenic crust, are in the late Mesoproterozoic Grenville Province (Rivers, 2012; Rivers and Schwerdtner, 2015) and the late Archaean Superior Province (Culshaw et al., 2006). In both cases, wide orogenic plateaus are interpreted to have formed as a result of compression leading to crustal thickening through imbrication and formation of nappe-like structures, followed by thermal relaxation and widespread middle-/

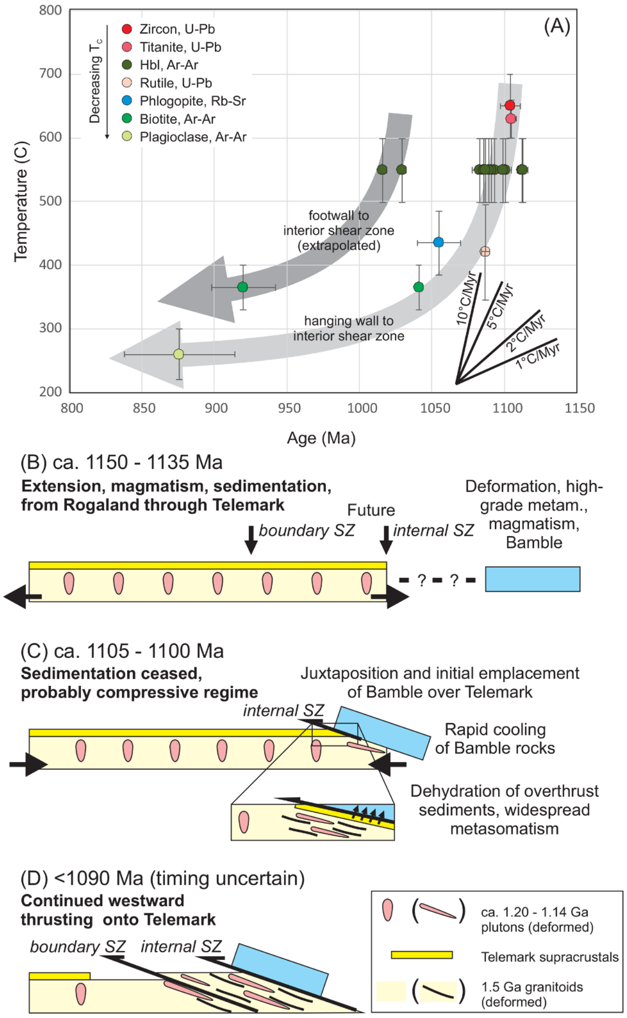

Fig. 15. (A) Time vs. temperature plot for the Bamble lithotectonic unit, based on new and previously published age data from different minerals and isotopic systems, including U–Pb in zircon and rutile, Rb–Sr in phlogopite, and Ar–Ar in hornblende, biotite and plagioclase (Table 2). (B–D) Cartoon showing the possible tectonic evolution of the Telemark and Bamble lithotectonic units, based on the metamorphic/hydrothermal history of the Bamble lithotectonic unit and the sedimentation history of the Telemark lithotectonic unit. (B) At ca. 1150–1135 Ma, parts of the Bamble lithotectonic unit located in the hanging wall of the internal shear zone underwent high-grade metamorphism. At the same time, the Telemark lithotectonic unit underwent geographically widespread extension resulting in bimodal magmatism and sedimentation in. (C) Sedimentation in the Telemark lithotectonic unit, in the footwall to the internal shear zone, ended at ca. 1100 Ma, coeval with rapid cooling of units in the hanging wall, interpreted to reflect thrusting of the Bamble onto the Telemark lithotectonic unit with shearing along the internal shear zone. Although the timing of thrusting along the boundary shear zone is uncertain, we speculate that it may have accommodated further shortening in the central Sveconorwegian orogen. (D) Structural evidence and sparse thermochronological data from the Bamble hanging wall and metamorphic evidence from the Telemark lithotectonic unit footwall suggest that thrusting of the Bamble onto the Telemark lithotectonic unit may have been a protracted process after ca. 1090 Ma. An important implication of this interpretation is that the internal shear zone constitutes the zone along which the Bamble and Telemark lithotectonic units were assembled, whereas the boundary shear zone may reflect continued thrusting and propagation into the Telemark lithotectonic unit footwall.

lower-crustal melting which significantly decreased the viscosity of the deep crust, allowing it to flow laterally. In both recent and ancient examples, plateaus several hundred kilometres across form within a few tens of millions of years of the onset of compression. The Altiplano–Puna plateau, for example, is 350–400 km across and formed over the last 25 Myr (Allmendinger et al., 1997), whereas the 1000 km-wide Tibetan plateau formed over the last 50 Myr (Royden et al., 2008).

In several recent contributions, a Tibetan-like orogenic plateau is argued to have been an important feature in the Sveconorwegian orogen (Bingen and Viola, 2018; Bingen et al., 2018; Möller et al., 2015); however, no direct evidence for such a plateau has been advanced and it is not clear how the interpretation is supported by currently available evidence.. Möller et al. (2015) argued that there are four conceivable models for the formation of ca. 990 Ma eclogites in the Eastern Segment, all involving a Tibetan-style orogenic plateau. However, eclogite-facies metamorphism does not demand the presence of an orogenic plateau; eclogites are a common feature in convergent plate margins and are found in orogens of all sizes and settings, with and without orogenic plateaus (Agard et al., 2009).

The sagduction model postulates the existence of an orogenic plateau spanning the Telemark LUs between 1280 and 925 Ma, and cites the sedimentology and facies associations of the Telemark supracrustals as well as bimodal magmatism as evidence. However, as discussed above, this largely extensional setting does not appear conducive to plateau formation, which is generally restricted to compressional regimes (e.g., Culshaw et al., 2006). In addition, the duration of this orogenic plateau is an order of magnitude longer than the Altiplano–-Puna, Tibetan and Grenvillian plateaus. This does not preclude the possibility that significant portions of the Sveconorwegian orogenic lower crust were molten and highly ductile – one can argue that they probably were – but the important point made by Slagstad et al. (2018b) is that the melting and potential crustal flow did not take place because of crustal thickening and concomitant radiogenic heating, but rather were related to long-lived mafic underplating.

5.8. *Feasibility of vertical tectonics in the Sveconorwegian orogen*

In summary, we consider that there are numerous inconsistencies in the proposed sagduction model. For example, Bingen and Viola (2018) argue that their delamination–foundering model is a “*robust example of intraplate vertical tectonics*” with reference to the work of Raimondo et al. (2014) on intraplate orogens in Australia. However, Raimondo et al. (2014), and more recently Quentin de Gromard et al. (2019), documented horizontal, far-field forces from the plate margins as a main orogenic driving force and argued that predicted intraplate stress generation and deformation resulting from mantle downwelling do not match the observed temporal and spatial scales of the orogens. The work of these authors thus negates, not supports, the proposed mechanism for early Sveconorwegian orogenesis.

Secondly, the sagduction model was invoked to explain the coeval extension and sedimentation in the Telemark LU vs. compression and high-grade metamorphism in the adjacent Bamble and Kongsberg LUs, with steep structures in Kongsberg LU being taken as support for the model. On the other hand, we have argued that the contrasting tectonic evolution can be explained by inferring separation of the Kongsberg–Bamble and Telemark LUs at the time, and have presented evidence that juxtaposition took place significantly later.

Thirdly, from a structural point of view, a drip would be expected to result in circular or ovoid foliation trajectories defining dome-and-basin structures, a radial pattern of increasingly steeper stretching lineations and fold axes towards the point of delamination, as well as extensional shear zones (Chardon et al., 1996; Collins et al., 1998). Geological maps of the Kongsberg and Bamble LUs, in contrast, clearly show predominantly linear foliation trajectories. Specifically, the map of the Bamble LU displays SW–NE-trending elongate lithological units parallel to strike of the border zone, and the stereoplot of bulk foliation and lineation data in this unit (Fig. 3) show well-defined maxima of steep to moderately NW- and SE-dipping foliation planes and S- to SW-plunging lineations; and similarly, the foliation traces in the Kongsberg LU trend approximately N–S with some extent of sigmoidal bending due to shear deformation, but likewise display no circular pattern. Neither can dome and basin patterns be recognised on a larger scale in the Sveconorwegian orogen.

Sagduction settings are characterised by specific distribution patterns of both strain and degree of deformation. Specifically, narrow downwelling areas forming keels with complex steep structure are separated from broader upwelling areas forming gently dipping domes (Bouhallier et al., 1995). However, such structural variability is not observed in the Sveconorwegian orogen. Strain measurements require suitable markers and both total strain and individual strain increments are difficult to evaluate in poly-deformed areas, but measures of total strain such as S > L and L > S fabrics, could at least qualitatively define the shape of the finite strain ellipsoid. Flattening, for example, would be typical along dome flanks while L-tectonites recording vertical constriction would be expected in synformal closures of the troughs (Bouhallier et al., 1995; Collins et al., 1998). Although systematic mapping of rock fabric is not presented for the Kongsberg LU, the small amount of structural information provided by Bingen and Viola (2018) principally emphasises the predominance of roughly N–S-trending steep foliation planes and more variably plunging lineations, as well as locally (Væleren area) dominant subvertical foliations and lineations (including fold axes). Several major, ductile, Sveconorwegian deformation events affected the Kongsberg LU (e.g. Starmer, 1993; Scheiber et al., 2015), and although Bingen and Viola (2018) acknowledged this multiple-stage structural evolution, in the absence of differentiation of the fabric elements according to their origin and age, the implied structural/tectonic evolution is unclear.

Scheiber et al. (2015), on the other hand, describe three to four ductile events of post-1150 Ma age along the Kongsberg–Telemark Boundary Zone beginning with E–W shortening associated with top-to-the-W shearing and creation of a penetrative foliation. Subsequent N–S sinistral transpression with top-to-the-NW reverse kinematics reactivated and overprinted this foliation and formed distinctive mylonitic shear zones. Later folding was followed by formation of extensional shear zones and normal faults with easterly dips inferred to be related to late-Sveconorwegian exhumation of the Telemark LU. The precise ages of the deformation events are not constrained so that the timing remains speculative and is mainly based on regional correlation.

Notably, Scheiber et al. (2015) emphasised the complexity of reactivation and overprinting relationships due to multiple deformation events, which is in contrast to single E–W shortening inferred by Bingen and Viola (2018). For example, Scheiber et al. (2015) showed that a new shear foliation was only formed in the sinistral mylonite zones whereas in adjacent gneisses, the pre-existing foliation was reactivated, overprinted and reoriented associated with development of a new stretching lineation. Similarly, the argued that sinistral transpression is not only recognised in the Kongsberg LU, but is also recorded along the margins of the adjacent tectonic blocks, where deformation was clearly weaker.

According to Bingen and Viola (2018), the fabric mainly formed during the E–W shortening event and was not substantially modified by later deformation stages; the subvertical foliation and steep lineation would indicate a component of near-vertical stretching or near-vertical transport in the crust. However, steep foliations and lineations can be explained by other mechanisms than vertical tectonics. Even if one agrees that the lineation largely formed during a single tectonic event, these structural observations could alternatively be explained by transpression, which involves a component of shortening perpendicular to the shear zone causing rotation of planes towards parallelism to the shear zone and, thus, to a (sub)vertical orientation. The associated stretching lineation in transpressional zones does not necessarily record the transport direction (Tikoff and Greene, 1997) and, in ideal models, either is vertical or switches to vertical after some finite amount of strain (Fossen and Tikoff, 1993). The variably plunging stretching lineations occurring in the Bamble and Kongsberg unit, largely ignored by Bingen and Viola (2018), could be explained by triclinic oblique simple shear (Czeck and Hudleston, 2003; Jones and Holdsworth, 1998; Lin et al., 1998) caused by heterogeneous strain and variations in the direction of extrusion due to, for example, regional variations in lithology and rheology, strain localisation or changing width of the deformation zone (Czeck and Hudleston, 2003). Hence, the observed structures are not convincingly conclusive for vertical transport but can easily be explained by other mechanisms that are in good accordance with the structural evolution observed by other authors (polyphase deformation, including transpression; Scheiber et al., 2015; Starmer, 1993).

5.9. Comparison with intraplate and Archaean orogens, variably driven by horizontal or vertical tectonics

Although many orogens may have formed distally to their respective plate margins, true intraplate orogens are rarely identified in the geological record. The best studied examples include the late Neoproterozoic–Cambrian Petermann orogen and the Palaeozoic Alice Springs orogen (Raimondo et al., 2014), both in central Australia. The tectonic style (i.e., duration, size, metamorphism) of these orogens is comparable to 'normal' plate-margin orogens, and it is therefore conceivable that their intraplate nature would go undetected in ancient orogens where their intraplate location has been obscured by later tectonic processes. Although it has been suggested that these orogenic events may have been driven by vertical tectonics (upper mantle delamination and/or drips; e.g., Gorczyk et al., 2013), Raimondo et al. (2014) made a compelling case that these orogenic events are driven by far-field horizontal stresses transmitted from the plate margins through strong lithosphere. Other, presently active intraplate orogens in Asia, are clearly the result of Himalayan–Tibetan orogenesis (see references in Raimondo et al.), and not of vertical tectonics. Thus, even if the Sveconorwegian orogeny did take place well inboard of the Fennoscandian margin and could be classified as intraplate, there is no reason *per se* that this was related to vertical tectonics; in fact, from what we know of other intraplate orogens, it seems rather unlikely.

Tectonic processes on early Earth were most likely very different from those operating today. Since early Earth was much hotter, with little rigid crust, it seems implausible that processes related to horizontal movement of rigid tectonic plates – the dominant tectonic process today – could have been operative. In its place, many workers have suggested some form of vertical tectonics. The Mesoarchaean Pilbara Craton in NW Australia, the Barberton greenstone belt in the Kaapvaal Craton, South Africa, the Dharwar Craton in India and the Superior Province in Canada are among the best studied examples of possible vertical tectonics, forming dome-and-keel structures characterised by complexly deformed gneissic domes separated by upward-younging, synformal greenstone belts (Chardon et al., 1996; Collins et al., 1998; Lin, 2005; Parmenter et al., 2006; Van Kranendonk, 2011; Van Kranendonk et al., 2004). These dome-and-keel terrains have been interpreted to reflect convective overturn (sagduction) of the crust following voluminous mafic magmatism emplacing dense mafic rocks over less dense, felsic TTG rocks; the mantle does not, however, play an active part in this process. The characteristic structural pattern of the gneiss domes has not been recognised in the Sveconorwegian Province, and there is no voluminous mafic magmatism that could have triggered convective overturn. The time scale of these processes may be on the order of 10 Myr (Robin and Bailey, 2009), not > 100 Myr as suggested for the Sveconorwegian orogen (Bingen and Viola, 2018). Moreover, these process are distinctly different from the sagduction model proposed for the Sveconorwegian.

The "*catalytic delamination-driven model*" for formation of Archaean crust and sub-continental lithospheric mantle by Bédard (2006) may be closer to that proposed by Bingen and Viola (2018) in that the mantle has a more active role. Briefly, in this model, a mantle plume first forms a thick basaltic crust, and continued plume activity leads to partial melting of the lower crust forming tonalite-trondhjemite-granodiorite

(TTG) melts and a dense lower-crustal ultramafic residue. The light TTGs overlain by dense mafic crust triggers convective crustal overturn accounting for the dome-and-keel granite-greenstone terrains, and the dense residues delaminate and sink back into the mantle where they trigger more magmatism. It is, however, very difficult to see any similarities with the anticipated result of this process and that suggested for sagduction in the Sveconorwegian orogen.

We therefore conclude that modern-style plate tectonics, characterised by largely horizontal movements of more or less rigid lithospheric plates, not only explains very well the observed > 200 Myr of Sveconorwegian orogenesis, but that poorly documented processes related to some form of vertical tectonics are highly unlikely to have had a major impact on the orogenic evolution. In many ways, the Sveconorwegian orogeny does not appear to mark a significant break with preceding active-margin processes characterising the southwestern Fennoscandian margin from > 1800 Ma. It seems clear, however, that these processes ended with the Sveconorwegian orogeny at ca. 920 Ma. Earlier work has suggested that termination may have been related to slab roll-back and migration of the margin to a more outboard position (Slagstad et al., 2019, 2018a, 2017), with active-margin processes continuing through much of the Neoproterozoic. This hypothesis is presently confounded by the fact that the evidence, if it exists, is likely to be found in autochthonous basement rocks, strongly overprinted by later tectonic processes, and in allochthonous units of uncertain heritage. Importantly, however, the hypothesis is testable, albeit with difficulty.

6. Conclusions

The new data presented here show that much of the crust constituting the Sveconorwegian Province formed before ca. 1480 Ma. A compilation of magmatic ages and locations from throughout the province shows that there was a gradual younging in magmatism from east to west, rendering previous suggestions of important breaks within the province implausible. Moreover, it provides support for tectonic models that propose the formation of a long-lived active margin in the late Paleoproterozoic or early Mesoproterozoic. Specifically, we argue that starting at 1340 Ma, bimodal, extension-related magmatism and widespread sedimentation was conspicuous in lithotectonic units west of and structurally above the Mylonite Zone, with some units recording extension until ca. 1100 Ma. We infer that this long-lived extension may have caused rifting and separation, but not necessarily drifting, of these units, and that the Sveconorwegian orogeny is essentially a reamalgamation of these units. The Bamble LU is particularly interesting in this regard, and further work may show that this unit preserves a record of long-lived tectonic activity in a critical region of the orogen – at the boundary between a long-lived, high-temperature, magmatically active western part and a more episodic, lower-temperature, high-pressure evolution characterising the eastern part of the orogen, close to the foreland.

Although we have a way to go before fully understanding Sveconorwegian orogenesis, there is at present no pressing need to infer the operation of rather exotic mechanisms involving some form of vertical tectonics. While Sveconorwegian orogenesis almost certainly incorporated tectonic processes involving a vertical component of displacement (e.g., related to subduction, thrusting and delamination) we reject the proposition that vertical tectonics in the form of gravity-driven delamination and foundering of SCLM, operating continuously on time-scales of several hundred million years and across the entire orogen, was the main orogenic driving force. Tectonic processes that in principle are similar to those operating on Earth today, involving movement of relatively rigid tectonic plates, appear sufficient to account for the observed Sveconorwegian tectonic evolution.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

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