Precambrian Research 265 (2015) 121–149

Contents lists available at ScienceDirect

Precambrian Research

journal homepage: www.elsevier.com/locate/precamres

# Polyphasal foreland-vergent deformation in a deep section of the 1 Ga Sveconorwegian orogen

Aránzazu Piñán Llamas[a,*], Jenny Andersson[b], Charlotte Möller[c], Leif Johansson[c], Edward Hansen[d]

[a] Department of Geosciences, Indiana University – Purdue University Fort Wayne, 2101 East Coliseum Boulevard, Fort Wayne, IN 46805-1499, USA
[b] Geological Survey of Sweden, Box 670, SE-751 28 Uppsala, Sweden
[c] Department of Geology, Lund University, SE-223 62 Lund, Sweden
[d] Geological and Environmental Sciences Department, Schaap Science Center, 2017, 35 East 12th Street, Holland, MI 49423-3605, USA

ARTICLE INFO

*Article history:*
Received 29 September 2014
Received in revised form 30 April 2015
Accepted 6 May 2015
Available online 27 May 2015

*Keywords:*
Sveconorwegian
Synkinematic
Migmatite
Folding
Structural evolution
Zircon

ABSTRACT

Metamorphic belts in Precambrian shields expose deep interiors of orogens and are often challenging to interpret in tectonic terms. The Eastern Segment of the 1.1–0.9 Ga Sveconorwegian orogen is a metamorphic belt, which was metamorphosed at high-pressure granulite and upper amphibolite facies at 35–40 km depth and shows highly complex fold patterns. We use detailed structural analysis in combination with U–Pb SIMS dating of complex zircon to identify the structural and tectonic evolution in a composite migmatitic orthogneiss complex of the Eastern Segment. We link four fold phases to late-orogenic foreland-vergent flow, and date $D_2$ and $D_3$ at 0.97–0.95 Ga. Leucosome and mesosome of felsic metasupracrustal migmatitic gneiss contain igneous zircon that dates the crystallization of the source rock or protolith at 1695 ± 8 Ma, and 1690 ± 8 Ma, respectively, demonstrating a temporal link to unmetamorphosed or little metamorphosed igneous rocks east of the Sveconorwegian orogen (the Transscandinavian Igneous Belt). Early migmatization attributed to Hallandian orogenesis is dated by formation of secondary zircon in two leucosome samples at 1402 ± 12 and 1386 ± 7 Ma. The pre-Sveconorwegian structure ($S_c$), which is strongly overprinted by Sveconorwegian deformation and migmatization, is a composite coarse gneissic layering made up of a primary compositional layering and (variably present) Hallandian leucosome veins. The dominant foliation, a pervasive gneissic banding ($S_1$), is axial planar to intrafolial $F_1$ folds and developed as a result of tectonic overprint of $S_c$; $S_1$ is associated with a strong ESE-trending aggregate stretching lineation ($L_1$). $S_1$ and $L_1$ were folded by asymmetric SE-vergent $F_2$ folds during foreland-vergent flow. Crystallization of Sveconorwegian zircon in syn-$F_2$ leucosome dates this phase at 970 ± 5 Ma. The sequence was subsequently deformed by symmetric and asymmetric $F_3$ folds that are S- to SE-vergent. Syn-$F_3$ leucosome, mineral parageneses and microtextures associated with $D_3$ show that this deformation occurred under still high temperatures. The last ductile phase ($D_4$) also involved the generation of leucosome synkinematic with N-S trending folds that deformed all previous structures under amphibolite facies conditions. K-feldspar-rich, originally coarse-grained and strongly deformed metapegmatite contain two generations of zircon: texturally old 1414 ± 5 Ma cores and fragments, and voluminous Sveconorwegian envelopes, and new grains that demonstrate the presence of melt as late as 958 ± 7 Ma. Ductile structures are similar in metasupracrustal and metaplutonic orthogneiss complexes. Likely, these units were tectonically juxtaposed during $D_1$, while $D_2$–$D_4$ structures reflect a common tectonic evolution after their emplacement. We interpret $D_{1-4}$ structures as recording WNW-ESE convergence ($D_1$) and ESE-vergent flow ($D_2$ and $D_3$), followed by E-W gentle upright folding ($D_4$). Sveconorwegian 0.98–0.96 Ga foreland-vergent deformation, accompanied by migmatization at all four stages, was responsible for formation of the polyphasal deformation pattern in this part of the orogen.

Published by Elsevier B.V.

## 1. Introduction

Old and deeply eroded orogens in Precambrian shields are the roots of mountain ranges similar to modern orogenic systems (e.g. Molnar and Tapponnier, 1975; Davidson et al., 1982; Allegre et al.,

* Corresponding author. Tel.: +1 2604816253.
*E-mail addresses:* pinana@ipfw.edu (A. Piñán Llamas), jenny.andersson@sgu.se (J. Andersson), charlotte.moller@geol.lu.se (C. Möller), leif.johansson@geol.lu.se (L. Johansson), hansen@hope.edu (E. Hansen).

http://dx.doi.org/10.1016/j.precamres.2015.05.009
0301-9268/Published by Elsevier B.V.

1984; Hossack and Cooper, 1986; Molnar, 1986; Treloar and Searle, 1993; Searle et al., 1987; Dallmeyer, 1988; Stephens, 1988; Rivers et al., 1989; Searle, 1991, 2006; Culshaw et al., 1997; St-Onge et al., 2006; Rivers, 2008; Rivers et al., 2012; Möller et al., 2015), whether accretionary (Andean type) or collisional (Himalaya type). While modern orogens preserve topography and upper parts of the tectonostratigraphy (suprastructure; Culshaw et al., 2006, and references therein), old orogens provide unique access to the very deep interior levels (infrastructure). The deep levels are structurally complex and characterized by high-temperature metamorphism and partial melting, rendering identification of lithological and tectonometamorphic differences difficult. Nevertheless, deciphering the tectonic evolution and mechanical properties of the deep roots of orogens is essential to our understanding of mountain-building dynamics in general. In particular, the presence of even small amounts of partial melting can dramatically reduce the strength of continental crust (Kohlstedt and Mackwell, 1995; Rosenberg and Handy, 2005) and induce lateral flow (e.g. Jamieson et al., 2011, and references therein).

The Sveconorwegian orogen (Fig. 1) represents the deep root of a collisional mountain belt situated on the southwest margin of the Baltic Shield. It is a counterpart to the 1.1–0.9 Ga Grenville orogeny in the Canadian Shield (Bingen et al., 2005, 2008; Möller et al., 2015). The Eastern Segment is a 100 km wide belt that includes the orogenic frontal zone in the east (Fig. 1) and consists of Baltica continental crust. The internal section of the Eastern Segment (Fig. 1) was located at 35–40 km depth (10 kbar) during the late stage of a continental collision that resulted in the emplacement of an eclogite-bearing nappe into its present westernmost part (Möller et al., 2015; Tual et al., 2015). The Eastern Segment shows highly ductile east-vergent flow toward the orogenic front; c. 25 km west of the Sveconorwegian Front lower temperature metamorphic conditions prevailed (lower amphibolite to greenschist facies) and strain was non-penetrative.

In this paper we document the style of deformation, kinematics and polyphasal geometric relations in a high-grade, interior, and well-exposed key area of the Eastern Segment. We link deformation to late-orogenic foreland-vergent flow at deep crustal levels (35–40 km) by dating synkinematic melts using U–Pb SIMS analysis of zircon, thereby demonstrating the style of deformation under the deep frontal parts of an orogenic plateau.

## 2. Geological setting

The Sveconorwegian orogen and its tectonic equivalent, the Grenville orogen in North America, were part of a collisional orogenic system that led to the assembly of supercontinent Rodinia. The Sveconorwegian orogen consists of crustal segments that are separated by roughly N-S trending ductile deformation zones (Fig. 1, Bingen et al., 2008). From east to west these crustal blocks are: the Eastern Segment, and the Idefjorden, Bamble, Kongsberg, and Telemarkia terranes. The orogenic frontal zone in the east (frontal wedge, Fig. 1) is a fan structure with a steep central part (in the literature referred to as the Protogine Zone; e.g. Gorbatschev, 1980). Its eastern part, coined Sveconorwegian Frontal Deformation Zone (Wahlgren et al., 1994) is characterized by east-vergent non-penetrative deformation. To the east of the Sveconorwegian Front are metamorphic and igneous rocks of the 2.0–1.8 Ga Svecokarelian orogen (Stephens et al., 2009), partly overlain by supracrustals and intruded by voluminous 1.7 Ga old igneous rocks; this includes 1.86–1.66 Ga plutonic and volcanic rocks of the Transscandinavian Igneous Belt (Andersson and Wikström, 2004). These rocks can be followed westward into the Eastern Segment where they become gradually more penetratively deformed at successively higher temperatures (Wahlgren et al., 1994; Söderlund et al., 1999). Protolith ages and bulk chemical composition of orthogneisses in the Eastern Segment are equivalent to ~1.7 Ga old granitic to quartzmonzodioritic rocks of the Transscandinavian Igneous Belt (Wahlgren et al., 1994; Petersson et al., 2013). The frontal wedge accommodated displacement during the latest stages of the Sveconorwegian orogeny and represents the eastern boundary of the Eastern Segment. The western boundary of the Eastern Segment is an arcuate shear zone, the Mylonite Zone (MZ in Fig. 1; Magnusson, 1937; Stephens et al., 1996; Berglund, 1997; Viola and Henderson, 2010; Viola et al., 2011) that represents a major terrane boundary (Andersson et al., 2002) and that has been characterized as a top-to-the-SE thrust of Sveconorwegian age that juxtaposed the allochtonous Idefjorden terrane in the hanging wall against the parautochtonous Eastern Segment in the footwall (Fig. 1; Viola et al., 2011). Areas of the Mylonite Zone (hereinafter MZ) have been overprinted by a younger down-to-the-west extension (Berglund, 1997; Viola et al., 2011) that has been interpreted as related to gravitational instabilities resulting from crustal overthickening during the shortening phase of the Sveconorwegian orogeny (Viola et al., 2011). Hornblende, biotite, and white mica (Ar–Ar ages; Page et al., 1996; Viola et al., 2011) and titanite (U–Pb ages c. 920 Ma; Johansson and Johansson, 1993) from the Mylonite Zone have given cooling ages in the range of 922–860 Ma.

In the Eastern Segment, metamorphic grade ranges from greenschist to amphibolite facies in the frontal wedge, to high-pressure granulite and upper amphibolite facies in the internal section exposed in the west (Fig. 1; Möller et al., 2015). Geothermobarometry applied to mafic rocks in the internal section yield temperatures between 680 and 795 °C and pressures of 0.8–1.2 GPa (Johansson et al., 1991; Wang and Lindh, 1996; Wang et al., 1998; Möller, 1998). Retrogressed eclogites, occurring in a fold nappe (Möller et al., 2015), indicate metamorphic pressures exceeding 1.5 GPa (Möller, 1998, 1999; Austin Hegardt et al., 2005). U–Pb zircon dating of eclogite metamorphism has yielded ages of 990–980 Ma (Johansson et al., 2001; Möller et al., 2015). Eclogite metamorphism was followed by exhumation and nappe emplacement (Möller et al., 2015). This took place under regional-scale penetrative ductile deformation and partial melting at high-pressure granulite to upper amphibolite-facies conditions at c. 970 Ma (Andersson et al., 1999, 2002; Söderlund et al., 2002; Möller et al., 2007, 2015; Tual et al., 2015).

The Eastern Segment also hosts crosscutting metadolerites that in many places are weakly to moderately strained. In the lower structural levels these metadolerites have been metamorphosed at high-pressure granulite facies. Still younger granitic and undeformed pegmatite dikes have been dated at 0.96–0.94 Ga (Christoffel et al., 1999; Möller et al., 2007; Möller and Söderlund, 1997).

The southern parts of the Eastern Segment record a high-grade polymetamorphic evolution. Pre-Sveconorwegian, regional scale, high-grade metamorphism and deformation (Hallandian orogenesis) are constrained to 1.47–1.42 Ga (Christoffel et al., 1999; Söderlund et al., 2002; Austin Hegardt et al., 2005; Möller et al., 2007; Brander et al., 2012; Ulmius et al., 2015). This event was followed by intrusion of gabbro, anorthosite, syenite, granite and charnockite, and widespread injections of pegmatitic and granitic dikes at 1.41–1.38 Ga (Hubbard, 1989; Söderlund, 1996; Andersson et al., 1999; Christoffel et al., 1999; Möller et al., 2007; Brander and Söderlund, 2009; Brander et al., 2012; Petersson et al., 2013). Simultaneous fluid-assisted high-temperature metamorphism has been recorded as charnockitisation of gneisses in southwestern parts of the Eastern Segment (Halmstad-Varberg; Hubbard, 1989; Harlov et al., 2006; Rimsa et al., 2007). In the internal parts of the Eastern Segment, Hallandian migmatitic structures were variably transposed by Sveconorwegian ductile deformation at high-grade conditions (e.g. Möller et al., 2007).

Download English Version:

https://daneshyari.com/en/article/4722656

Download Persian Version:

https://daneshyari.com/article/4722656

Daneshyari.com