How does orogenic crust deform? Evidence of crustal-scale competent behaviour within the partially molten middle crust during orogenic compression

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Abstract

Granitic partial melts are generally thought to significantly weaken the orogenic crust and, ultimately, lead to the collapse of an orogen. Studies from different orogens have shown that the syn-melt deformation behaviour at the orogenic scale is, however, more complex. In addition, once fully crystalline, granitic material strengthens the crust. Linking the evolution from melt-present to melt-absent deformation at the scale of the orogen is a challenging but necessary task if one wishes to investigate the overall behaviour of the middle crust. In this paper, we make observations of orogen-scale deformation, developed over a period of 30-40 Ma *during* crustal partial melting of the middle and lower crust of a Palaeoproterozoic orogen. The crust shows a globally common pattern where coeval partial melting, compressional deformation, and transtensional structures co-exist. We demonstrate this complex interaction through an integrated approach using multiple datasets. The key observation is that of widespread, regular, orogen-scale shortening by folding of sub-horizontal anatectic granitic sheets and highly migmatized lithologies; this shortening interacted with modest orogen-parallel lateral stretch expressed as local extensional/transtensional structures. The pervasive and dominant deformation style is that of folding of the syn-kinematic granite sheets and voluminous migmatites: this demonstrates that the principal deformation style of deformation *within* the partially molten mid-crust is *competent* rather than weak. The observed evidence of weak behaviour such as strike-slip or transtensional shear zones accommodating lateral escape are localized in comparison and not necessarily

associated directly with the granite sheets or the highly migmatized volumes of the crust. Moreover, the strain intensity (fold wavelengths and amplitudes) seems independent of the overall melt fraction. The implications are that, while widespread volumes of partial melt will weaken the crust overall (compared to the brittle upper crust) and while individual melt bodies can possibly persist for up to some millions of years in the middle crust, i) the *relative* strength between (partially) molten volumes is not primarily controlled by melt fraction; and ii) individual melt volumes may be too short-lived at the time-scale of orogenic deformation (i.e. orogenic strain rates) to significantly influence the overall deformation style.

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1. Introduction

In orogens of all ages, broadly syn-convergent, crustal-scale partial melting (anatexis) of the middle and lower crust is well documented globally: exposed roots of hot orogens invariably show voluminous migmatisation and granitic partial melts (e.g. Brown, 1994; Vanderhaege, 2009; Chardon et al., 2011; King et al., 2011; Clemens and Stevens, 2016). Migmatites and syn-orogenic granites are common in modern orogens as well. In the Himalayan-Tibetan system, migmatites are currently exposed along the Himalayan high-T orogen-frontal zone, where their presence has been controversially explained with the so-called mid-crustal channel extrusion model (e.g. Beaumont et al., 2001; Grujic et al., 2002; Godin et al., 2006). Furthermore, at least one partially molten channel is interpreted in the Tibetan lower crust, suggested to facilitate orogenic collapse and the eastward escape of the upper crust (e.g. Clark and Royden, 2000; Bird, 1991; Royden et al., 1997; England and Houseman, 1989; Klemperer, 2006; Searle, 2013). Migmatites have also been suggested to exist in e.g. the Andes (e.g. Yuan et al., 2000; Schilling and Partzsch, 2001; ANCORP Working Group, 2003). The presence of anatectic granites and migmatites is a feature of many models of how metamorphic core complexes evolve (e.g. Vanderhaeghe and Teyssier, 2001; Zeitler et al. 2001; Rey et al., 2009; Langille et al., 2012).

The timing and dynamics of crustal-scale partial melting, the longevity (residence time) of partial melts in the crust, and the factors controlling melt mobilization, transport, and crystallization is crucial with respect to the stress evolution and strain partitioning from orogen scale to outcrop. The interplay between melt - mobilization - crystallization processes and crustal deformation is, however, still only partially understood: the exact relationships between anatexis and migmatisation, rheological behaviour, and stress and strain distribution either locally or at crustal scale, are still debated. The interaction and relative timing/duration of anatectic vs. deformation processes directly influences the models of the behaviour of orogens in general, and the models relying on 'weak' mid-crustal behaviour in particular.

Initiation of partial melting has been experimentally and numerically shown to significantly weaken the bulk rock volume (e.g. Rosenberg and Handy, 2005; Rutter et al., 2006). In addition, syn-melt shearing has been modelled to enhance melt extraction from country rock into shear zones, so that any deformation should very effectively facilitate melt escape (e.g. Brown, 1994; Holzmann et al., 2003). An overall feedback relationship is postulated for crustal models, where melting weakens bulk rheology, leading to deformation and shearing, which focuses melts into shear zones causing even more weakening (e.g. Brown and Solar, 1998; Holtzman et al., 2003). However, various field observations show that the interaction of syn-orogenic crustal deformation (both compressional and extensional) and crustal melting is not as straight-forward as predicted by models. For example, ubiquitous observations globally show that significant amounts of melts are retained more or less *in situ*, even in the vicinity of shear zones (e.g. up to 20% in Lee et al., 2018).

Melts can of course migrate and escape, forming anatectic granite intrusions at various scales, often as foliation-oblique dykes or foliation-parallel sills or laccoliths. However, these anatectic melt bodies, and other syn-orogenic granitoids and pegmatites, are often internally relatively undeformed, or only weakly deformed, unless they coincide with major shear zones in which case they localise deformation and can show intense deformation fabrics. This can often be explained by rapid crystallization: a 1 km thick granitic sill intrusion, if intruded in a crust of 600°C, will completely crystallize in c. 150,000 years and will have accumulated a negligible amount of strain in that time (using a 'typical' orogenic, crustal strain rate of 10^{-14} s^{-1}; Davidson et al., 1992, 1994). However, if a shear zone develops in a (partially) molten volume strain rates will be higher than this. On the other hand, the initial localization of these granite bodies is often explained either by an extensional phase or at least a "pause" in the far-field compressional stresses; a "protected" lens within an otherwise deforming crust; or a localized zone of extension (e.g. Holdsworth et al., 2001; Druguet and Carreras 2006; Nironen and Kurhila, 2008). Localized zones of extension such as fault jogs do host granite intrusions, such as porphyric granite stocks which can carry important ore mineralization (e.g. Sillitoe, 2000), but applying this or other 'non-compressional' models to any internally undeformed intrusive or migmatitic body is problematic: it is especially problematic if applied to anatectic granite dykes, sills and sheets common in the migmatitic lower orogenic crust. Such anatectic granites occur orogen-wide; they do not necessarily correlate with any major shear zones or extensional jogs; they are often foliation-parallel regionally; and the timing of their formation is usually coeval with the continuing convergence of lithospheric plates.

The key question, therefore, is: does 'melt weakening', observed in experiments and models, significantly affect the deformation style within the middle crust *at an orogenic scale*? In order to answer this question, we need to investigate the geometric responses to deformation: structural and geometric relationships are an excellent tool to infer relative viscosity contrasts of deforming material (Fig. 1). Crucially, these observations need to be made at the orogenic scale, as we are dealing with an orogen-scale problem. In modern orogens, crustal roots are not yet exposed and cannot be directly studied. Exposed old orogens offer a very valuable analogue for lower crustal processes. Careful combination of different techniques and data is needed to study the temporal and spatial relationships of tectonic and anatectic processes at a crustal scale, because the original configuration of the orogen may not be known in detail. In this paper, we use such an integrated approach to discuss the timing and styles of crustal-scale deformation in relation to an episode of syn-convergent, voluminous partial melting and anatectic granite magmatism in the Svecofennian orogen, south-central Fennoscandia. The study area typifies many other migmatite terranes globally: although the orogen is a Precambrian 'hot' orogen, and as such may not be directly comparable with all orogens globally, the observations of general behaviour of partially molten rocks in the middle and lower crust should nevertheless be transferable to other orogens. We combine age and geothermobarometric data, crustal-scale cross sections, and geological maps to illustrate the contemporaneous nature of anatexis and granite magmatism, large-scale folding, shear zone development, and (late) syn-compressional orogen-oblique extension; and discuss the implications of the observations to models of mid-crustal orogenic deformation in general. We show that the overall, crustal-scale patterns of deformation within the highly molten middle crust are perhaps counter-intuitive in that they are dominated by fairly evenly distributed competent features (folds), regardless of the inferred total volumes of granites and partial melts. In other words, unless associated with shear zones, the apparent melt fraction within the migmatitic middle/lower crust does *not* seem to control the deformation style at crustal scale, implying that melts are heterogeneous and transient features, not present in (semi)liquid state for long enough to control the deformation style in the middle crust at orogenic strain rates.

2. Geological setting

In order to better understand modern, non-exposed orogenic roots, exposed ancient analogues offer an excellent opportunity to study lower crustal processes. Our results come from the exhumed deep crust of the Palaeoproterozoic Svecofennian orogen in south-central Fennoscandia (Figs. 2a, b; e.g. Ehlers et al., 1993).

The current erosional level exposes lower amphibolite facies to granulite facies rocks, depending on the location and palaeocrustal level (e.g. Väisänen and Hölttä, 1999). Most of the exposed crust is migmatitic and forms an approximately E-W trending, c. 120 km wide ‘Late-Svecofennian Granite-Migmatite belt’ LSGM (after Ehlers et al., 1993; Fig. 2a). The metamorphic peak conditions in southern Finland have been estimated at c. 750-825°C and c. 4-7 kbar (e.g. Schreurs & Westra, 1986; Väisänen & Hölttä, 1999; Johannes et al., 2003; Mouri et al., 2005; Torvela and Annersten, 2005; Torvela et al., 2010b). Near Sammatti and Turku (Fig. 2b), this peak metamorphism has been dated at c. 1825-1815 Ma from migmatite leucosomes (Väisänen et al., 2002; Mouri et al., 2005). Outside the granulite facies areas, the amphibolite facies LSGM schists and gneisses are typically still migmatitic; peak conditions have been determined at c. 100-150°C and 0.3-2 kbar lower than in the granulite areas. PT conditions as low as c. 600°C and 3 kbar occur in the non-migmatitic Kisko-Orijärvi triangle zone, SW of Sammatti (e.g. Latvalahti, 1979; Schreurs & Westra, 1986; Väisänen & Hölttä, 1999).

The orogen is multi-phased. Between c. 1.92-1.86 Ga large amounts of juvenile crust formed in volcanic arcs, which were subsequently accreted onto the Archaean craton in the present NE Fennoscandia (Fig. 2a; e.g. Lahtinen et al., 2005). This resulted in pervasive crustal deformation and stacking in present Southern Finland, now mostly identified as recumbent/isoclinal folds (F1) with c. E-W striking axial planes, associated thrust sheets, and minor migmatisation (Fig. 3a; e.g. van Staal & Williams, 1983; Bleeker and Westra, 1987; Ehlers et al., 1993; Väisänen & Hölttä, 1999). After this first convergence, a phase of tectonic quiescence seems to have taken place, perhaps with minor crustal extension and rifting (Fig. 3b) although this is poorly constrained: shallow-water and fluvial quartz-rich sandstones, now quartzites, were deposited at c. 1.85 Ga and are now found locally in the vicinity of the presumed palaeomargin (the ‘intraorogenic metasediments’ in Fig. 2b; Bergman et al., 2008; Nironen and Mänttäri 2012). In any case, orogenic convergence resumed by 1.84 Ga, resulting in renewed deformation (e.g. Lahtinen et al., 2005; Torvela et al., 2008). This convergence refolded the earlier recumbent folds and flat-lying crustal fabrics into open to tight, upright folds with E-W striking fold axes (F2; Fig. 3c) synchronously with extensive high-T, low-P metamorphism, voluminous migmatisation and anatectic granite magmatism, and formation of local- to crustal-scale, steep, strike-slip shear zones throughout S Finland and in central Sweden (Fig. 4; e.g. Patchett & Kouvo, 1986; Bleeker and Westra, 1987; Ehlers et al., 1993; Lahtinen & Huhma, 1997; Väisänen & Hölttä, 1999; Väisänen & Mänttäri, 2002; Hermansson et al., 2007; Högdahl et al., 2008; Torvela et al., 2008, 2010a, 2010b; Skyttä and Mänttäri, 2008). This (second) convergence, called 'Fennian orogen' by Lahtinen et al. (2005), is the focus of this paper. The Fennian orogen, or 'Fennian phase', constrained by various authors to c. 1.85-1.81 Ga, is responsible for the majority of the current crustal configuration in the study area, although convergence from the south-southeast continued until at least c. 1.79 Ga (e.g. Levin et al., 2005; Torvela et al., 2008; Torvela and Ehlers, 2010a). Controversial suggestions imply that a distinct extensional phase occurred at c. 1.82-1.81 Ga, before the final convergence, based on e.g. multiple stages of gold mineralization; the presence of the internally undeformed Veikkola granite complex near Vihti; and local extensional structures such as boudinaged pegmatites (Fig. 2b; e.g. Nironen and Kurhila, 2008; Skyttä and Mänttäri, 2008; Saalman et al., 2009).

Partly overlapping the Fennian orogenic phase, convergence of Amazonia from the west at c. 1.81-1.79 Ga has been suggested to weakly modify the crustal structures throughout the Svecofennian domain through ~E-W compression, although evidence for this is debated (e.g. Lahtinen et al., 2005). This late E-W compression has been suggested to be responsible for the current dome-and-basin configuration of the crust (Lahtinen et al., 2005), although alternative explanations have been suggested through subvertical stretching during D2 compression (Cagnard et al., 2007); localised extensional shearing around domes (Torvela et al., 2013); or a combination of 'cross-folding and diapirism' (Bleeker and Westra, 1987).

The present dominating map pattern is that of fold interference that formed mostly due to the interaction of the initial 'F1' folds (pre-1.86 Ga) and the subsequent 'F2' folding episode of the Fennian orogen (Figs. 2b, 3). These composite folds generally have ~E-W striking steep F2 axial planes, and verge either north or south (more typically north). In addition to folding, both transpressional and transtensional shear zones, some of which are reactivated as post-orogenic brittle faults, are observed. The transpressional zones are dominantly dextral, strike mostly W-E to NW-SE, and are interpreted to reflect N-S to NNW-SSE compression (Fig. 2b). A significant amount of evidence suggests that the main deformation along most of these occurred c. 1.83-1.82 Ga, possibly as early as c. 1.85 Ga, i.e. during the Fennian collisional phase (e.g. Väisänen and Skyttä, 2007; Torvela et al., 2008). Large-scale thrusting has been suggested to facilitate the crustal thickening and deformation during the Fennian crustal thickening (e.g. Lahtinen et al., 2005 and references therein; Levin et al., 2005; Skyttä et al., 2006). Few thrusts of Fennian age have, however, been identified in the field; most speculated low-angle thrusts formed early, during D1 in association with the large-scale recumbent F1 folds, and are now folded by the F2 folds and intruded by anatectic granites (van Staal & Williams, 1983).

Few large-scale 'extensional' structures have been identified with confidence. Outcrop-scale boudinaged pegmatite and granite dykes near Kisko-Orijärvi area were interpreted to reflect syn-Fennian spreading and dated at. 1.83 Ga and, therefore, be approximately coeval with convergence and crustal thickening (Fig. 4; Skyttä and Mänttäri, 2008). Some transtensional high-T shear zones have been interpreted in the field and from seismic reflection data within the granulite-facies West Uusimaa Complex (e.g. Torvela et al. 2013). These shear zones have recently been dated and the results show that they too are broadly coeval with the other crustal features (Torvela and Kurhila submitted).

The Fennian deformation was accompanied by widespread crustal-scale anatexis and migmatisation of the pre-Fennian schists and gneisses, the dated leucosomes in southern Finland giving syn-orogenic ages of around 1.82 Ga (Fig. 4). 'Microcline granite' bodies and sheets of various sizes were emplaced throughout the area, interpreted by geochemistry to source from crustal partial melts (i.e. the migmatites; e.g. Stålfors & Ehlers, 2005). The microcline granites consistently show Fennian ages of c. 1.84-1.82 Ga (Fig. 3). It should be noted that the mapping convention of the Geological Survey of Finland does not discriminate between 'true' intrusives and very high migmatitic melt fractions that may not have migrated long distances: all outcrops estimated during mapping to contain >50% granitic melt is mapped as 'microcline granite and migmatite'; the implications of this will be discussed later. There are some identifiable major anatectic granite intrusions: the most important ones for this paper being the Veikkola composite granite and the Perniö granite (Fig 2b). Both show gradational boundaries with the surrounding migmatites, where observable. At a crustal scale, these and other syn-orogenic 'true' granites are often gently folded but internally undeformed or show only weak cleavages at outcrop, unless within or in the vicinity of large shear zones; these shear zones have been interpreted to function as transport channels of at least some of the magmas (e.g. ; Selonen et al, 1996; Stålfors & Ehlers, 2005).

Post-Fennian (i.e. 'post-orogenic' in terms of major crustal thickening) ages of ~1.81-1.79 Ga have been found for individual small intrusions, and these ages are increasingly common towards easternmost Finland (e.g. Suominen, 1991; Johannes et al., 2003; Kurhila et al., 2011). Convergence continued between 1.81-1.79 Ga but during this time deformation is mostly restricted to steep, mylonitic to semi-brittle shear and fault zones under upper greenschist to lower amphibolite facies conditions (e.g. Lindroos et al., 1996; Levin et al., 2005; Torvela et al., 2008). Within the highly migmatitic Sammatti area, for example, the PT conditions had dropped to c. 600°C and 5 kbar by c. 1.80 Ga (Fig. 4; Mouri et al., 2005). There is no evidence that significant crustal thickening through thrusting or large-scale folding occurred at this stage: except for the still active strike-slip shear zones, lower-grade (retrograde) metamorphism and structural features such as axial plane schistosity are largely absent.

3. Methods

Field observations and crustal-scale cross-sections across the LSGM have been combined with published age and geothermobarometric data and geological maps. These are used to investigate the overall architecture and, therefore, the deformation style of the migmatitic crust at the scale of the orogen.

A key role of the cross-sections is to illustrate the overall deformation style of the orogen, and to help put the field data and the interpretations in crustal-scale context. The cross-sections were constructed using 1:100 000 geological maps of the Geological Survey of Finland, along four N-S profiles across the migmatite belt of southern Finland (Figs. 2b, 5). The detail available on the geological maps varies, but most of them show a good density of structural measurements along and in the vicinity of the cross-section profile lines. The existing geological map data and data published in the literature are supported by field data collected by the first author in the West Uusimaa area, the Turku-Hanko archipelago, and Åland archipelago between 2002 and 2013. The structural patterns are summarized in Fig. 6.

In the generalized cross sections the rocks are divided into: i) mica schists ('paragneisses'), with pre-Fennian protoliths, commonly migmatized and internally isoclinally folded (F1) and deformed; ii) mafic and felsic pre-Fennian metavolcanics, migmatized during the Fennian phase; iii) igneous mafic and felsic pre-Fennian rocks; and iv) the younger, Fennian 'microcline granites' originating from anatectic melting of the crust (Fig. 7). The convention of the Geological Survey of Finland dictates that migmatites that are estimated to consist of >50% of anatectic material are marked on the map as 'microcline granite', whereas for melt contents of <50% the lithology of the host rock is used. Therefore, the lithological boundaries especially between the Fennian microcline granites and the pre-Fennian rocks are normally gradational and the contacts indicated on geological maps and cross sections are, therefore, estimations.

The profile orientations approximate the tectonic transport direction during the orogenic compression. Due to the complex tectonic history involving refolding of folds, anatexis, and the dome-and-basin style geometries, the cross sections cannot be extrapolated to great depths and cannot be balanced. However, their purpose here is to demonstrate the overall structural styles and the lithological relationships resulting from the syn-anatectic Fennian deformation, especially the distribution and the deformation styles of the migmatites and the anatectic microcline granite sheets across the migmatite belt. The cross sections are also used to estimate, albeit very roughly, the minimum amount of lithospheric shortening at this crustal level during the syn-migmatitic compression at ~1.84-1.81 Ga. The sections also demonstrate the relationship of the few known transtensional shear zones with the compressional structures, mostly in the West Uusimaa area.

4. Observations

The cross sections show pervasive F2 folding within the LSGM, with some localized deformation expressed as both normal and reverse/thrust faults and strike-slip shear zones. There are a number of steep normal faults seen in the sections and in the field. They cross-cut and displace the high-T rocks and clearly belong to a later, brittle deformation phase which is not discussed further in this paper except where they may affect elongation estimations. The regional-scale F2 folding patterns with approximately E-W to SW-NE striking axial planes reflect the approximately N-S to NW-SE compression related to the Fennian orogen.

In some areas the folds are tight (e.g. north of Gullkrona in the Turku section; Fig. 5a) and the fold tightness often correlates with nearby significant fault zones. The increasing tightness of the folding seems to be the mainly associated with major reverse faults and strike/oblique-slip transpressional shear zones and, apart from Somero-Karkkila Fault

Zone S-KFZ, there seems to be no correlation between significant changes in the dominant lithology and the presence of fault zones. Another example of tighter folds is the Rosala area in the southern part of the Turku section which shows a curious, tightly folded 'flower structure' of mainly felsic metavolcanics squeezed between two openly-folded granite-dominated blocks; the map pattern reveals a granite-poor zone squeezed between a granite-dominated, east-plunging anticline pair (Fig. 6; see also Torvela, 2017 for a detailed map of this area).

The intensity (amplitude and wavelength) of folding does not seem to be controlled by the dominant lithology, including the presence or absence of 'microcline granites' i.e. significant volumes of partial melts (Fig. 8). For example, in Turku section (Fig. 5a), the microcline granite-dominant lithologies (which include the highly migmatitic units) show both gentle to open, and tight folding; and the same is true for the mica schist-dominated parts of the section. The central part of the Salo section (Fig. 5b) is also dominated by granitoids (both intrusives and 'microcline granites' which includes highly migmatitic rocks). Published structural measurements are sparse especially between Salo and the major shear zone at the southern margin of the Perniö granite, compared to the rest of the section or the other sections, but the same observation can be made: the deformation intensity as indicated by the amplitude and, especially, the wavelength of the folding, does not seem to vary depending on the lithology. The presence of anatectic granites does not, in other words, seem to enhance strain accumulation and the granite-dominated bodies are, in fact, often less folded than the supracrustal-dominated areas (Fig. 8). On the other hand, along the Vihti section (Fig. 5d), the composite Veikkola granite is not folded at all. It is also worth mentioning here that the contact of the anatectic microcline granites and the Perniö granite (Fig. 5b) is diffuse and gradational, the intrusion showing little internal deformation; deformation features become fairly abundant close to the shear zone at the S margin of the granite (see also Selonen et al, 1996).

The foliations in all areas are folded around relatively well-defined regional beta directions, and the F2 fold axial planes are normally steep with a slight tendency to verge towards NNW (Fig. 6). The beta directions coincide with fold axes plunging moderately to E or ENE, and the lineation mean principal direction is very close to both the beta direction and to the fold axis maxima (Fig. 6). Many of the stereonets reveal a weak N-S oriented trend (e.g. the N and S plunging fold axes in Rosala and Vihti areas, or the N and S plunging lineations in Sammatti area). Some of them can possibly be explained by the presence of domes at various scales (e.g. the Sammatti and Karkkila areas; Fig. 8a) which would introduce some variability in fabric orientations, including those of the fold axes. Another possible explanation is the putative post-Fennian E-W directed compression of the Nordic orogen. However, in Turku area where domes are less well developed or absent, these N-S patterns do not occur, which would be unexpected for a regional E-W compressional event.

The lineations show a gently to moderately doubly-plunging trend. The observed lineations are mostly mineral/stretching lineations or crenulation lineations. They are typically weak but relatively consistent in most areas, implying that they are mostly related to D2 rather than D1, although some folding of earlier lineations does probably contribute to the data as some scatter along great circles can be observed (e.g. Rosala and Salo areas in Fig. 6).

5. Interpretation

5.1 Deformation styles and intensity

Originally recumbent tight F1 folds, now refolded by F2, are interpreted in many locations along the cross section lines: for example, there are several folds around Sammatti area (Fig. 5c) that seem to refold earlier F1 folds. Other good examples can be seen in most sections, such as the Parainen refolded F1 fold in Turku section (Fig. 5a). There are also some possible D1 palaeothrust planes, now folded by F2; the best example of this is the palaeothrust surface just north of Teijo in Salo section where slightly higher-grade rocks are found on top of lower-grade felsic volcanic rocks (Fig. 5b).

The folding style is expected to be variable across the cross sections. At outcrop, concentric (parallel) F2 folds appear to be much more common than similar folds with sheared and thinned limbs (Fig. 7e, f). Both similar folds and concentric F2 folds are inferred in the cross sections; the folding style at large scale is difficult to determine with certainty, but concentric folds are more commonly interpreted than similar folds, partly based on the field observations of more dominant concentric folding. Many of the 'microcline granite'-rich layers seem to favour a concentric fold interpretation as opposed to especially the paragneiss migmatite-dominated areas.

Regardless of the folding style, the quantification of the fold amplitude-wavelength ratios (a/λ) should give a reasonable first-pass estimation of the intensity of folding (i.e. deformation intensity) across the area (Fig. 8). The amplitude measurements carry more uncertainty than the wavelength measurements, but if all folds are extrapolated in the same way, representative patterns should still emerge in terms of relative a/λ. Fig. 8 confirms the qualitative estimate of the relative deformation intensity: the highly migmatized areas (dominated by the microcline granites) have equal or smaller a/λ than the areas dominated by the supracrustal rocks migmatized to lesser degrees (compare e.g. the S and N parts of Turku section in Fig 5a). In addition, 'true' microcline granite intrusions are clearly less deformed than their surroundings: this is especially true for the 1.85-1.82 Ga Veikkola composite granite laccolith in the Vihti section (Fig 5d). Veikkola granite shows very little folding, and the internal deformation, observed in the field, is very weak to non-existent. The oldest parts of the Veikkola granite are dated at c. 1.85 Ga and, therefore, possibly slightly pre-date the Fennian compression. It is likely that parts of the granite had crystallized by the time most of the F2 folds started to develop, allowing the granite to act as a competent body during the main compression phase and as such protect the later intrusive pulses into the granite complex from the effects of crustal deformation. The northern margin of the granite was possibly slightly thrusted upon the Karkkila granulite dome, the flanks of which have later collapsed along the extensional granulite shear zones that are dated at c. 1.82 Ga (Torvela and Kurhila, submitted). Either way, the margins of the Veikkola granite gradually steepen into the surrounding migmatites. The other 'true' granite that is little affected by F2 folding is the c. 1.83 Ga Perniö granite (Salo section; Fig. 5b.; Selonen et al., 1996). Some thinner microcline granite bodies can, however, be seen to form tight folds which may indicate some localisation of deformation (e.g. around Sammatti area; Fig 5c).

Some reverse faults and thrusts, including a large imbricate stack in the southern part of the Sammatti section, are interpreted in the sections. Most of the reverse faults are quite steep, implying that these may have formed early during D2 as thrusts and subsequently rotated into steeper dips during progressive deformation. Later, more gently dipping thrusts may have formed but they are difficult to recognize in the field data due to lack of topography and distinct marker horizons. One such potential low-angle thrust is interpreted near Parainen in Turku section. The interpreted imbricate stack in the southern part of the Sammatti section is largely un-studied: it is interpreted from the geological map of the area, based on the repetition of units across each inferred thrust.

5.2 Amount of shortening and orogenic strain rate

Where not overly obscured by granite intrusions or effects of folding interference, the cross sections were used to roughly estimate the crustal-scale elongation in different parts of the migmatite belt. Unbalanced cross sections do not in principle allow for accurate quantification of elongation (e.g. Dahlstrom, 1969). However, the field data presented in this paper and in the literature (e.g. Cagnard et al., 2007; Skyttä & Mänttäri, 2008) suggest that the dominant compressional structural pattern, i.e. the folding patterns seen in the cross sections and in the stereoplots, are chiefly a result of the Fennian D2 convergence: if the crustal fabric was largely 'reset' to sub-horizontal before the Fennian phase (Fig. 3), any previously formed fabrics would have only minor effect on the subsequent crustal-scale elongation. In addition, the microcline granite intrusions and the voluminous, approximately layer-parallel partial melt sheets are demonstrably syn-D2 and are, therefore, reliable markers of the orogenic deformation. The large-scale deformation patterns of these granites and migmatites should, consequently, give a reasonable approximation of the syn-orogenic elongation. Some complications are still likely to be caused by the presence of large-scale early fold interference patterns; these interact in some places with the main shortening patterns (possibly suggesting that the fabric was not 'reset' everywhere), causing further interference. In Vihti section (Fig. 5d), the southern part is dominated by the undeformed syn-compressional Veikkola granite, and consequently the cross section and the elongation estimations were not extended as far south as the other sections farther west. Complications are possibly also caused by: i) the likely presence of other, unrecognized syn-Fennian low-angle thrusts and late-Fennian low-angle extensional shear zones; and ii) the intrusion of various non-anatectic igneous rocks prior to, during, and after the orogenic compression. The selected cross-section lines deliberately avoid non-Fennian granitoid intrusions and other complex areas where possible. Nevertheless, various complications and uncertainties are present especially toward the southern parts of the cross sections where geological maps tend to be old and lack sufficient structural data; the southern parts were not included in the elongation calculations due to larger uncertainties. We consider that the elongation calculations give reasonable estimates of the minimum shortening in the northern parts of the sections, but they should be treated with caution.

The late- and post-orogenic extension along low-angle shear zones and brittle normal faults is difficult to quantify without reliable markers, but we approximate it to be in the order of e=0.14 (14% extension) in the Karkkila granulite dome (Fig. 5d). Extension interpreted in the cross sections decreases towards W, being only about 5% along the northern part of the Somero section, and seems to be largely absent in Turku area. However, as noted earlier, unrecognised transtensional/extensional low-angle shear zones may (or may not) be present in the west; on the other hand, the known transtensional shear zones seem to be associated with dome structures in Karkkila area and possibly in Sammatti, but similar domes are absent in Salo and Turku areas. After removing the interpreted ductile and brittle extension (where observed), all four cross sections imply minimum shortening in the order of e=-0.3 (30% shortening) in the N-S direction. The elongation in the southern part of the sections is not quantified, but it seems that it may be more variable: the interpreted large imbricate stack in the southern part of Somero section may imply more shortening in this area compared to Salo and Turku sections; on the other hand, the Veikkola granite is virtually undeformed.

In addition to the strain distribution, the cross sections give an insight into the possible strain rates in the middle crust. Extrapolated to the present width of the migmatite zone (c. 100 km), the elongation calculations imply roughly 30 km of shortening. If the main D2 compression and F2 folding occurred c. 1.84-1.81 Ma i.e. over a 30 Ma period, the strain rate is in the order of 3.2×10^{-14} s^{-1}. For 50 Ma duration of the convergence, the strain rate is approximately 1.9×10^{-14} s^{-1}, and for 15 Ma c. 6.3×10^{-14} s^{-1}. These figures are consistent with those suggested for strain rates for orogenic deformation of the continental crust (e.g. Boutonnet et al., 2013). The corresponding average shortening rate for 30 km is 0.10 cm/a in 30 Ma, again consistent with modern observations of shortening rates in active orogens (e.g. ~0.10-0.14 cm/a for the Tibetan system; Zhang et al., 2004).

Perniö granite in the Salo section (Fig. 5b) may potentially be used to estimate the timing of the F2 folding in more detail. The cross sections suggest that the granite laccolith has experienced c. 20% shortening since the zircons within it reached closure temperatures at c. 1830 Ma, if we assume that the granite folding occurred mostly after full solidification. If this age represents the full solidification age of the granite, then for overall crustal shortening of 30%, 10% of the crustal shortening occurred prior to 1830 Ma, with the remaining 20% post-1830 Ma. Although it is likely that some folding occurred before the granite was fully crystalline, this granite body is a good marker of the timing of the overall crustal deformation.

5.3 Timing of metamorphism with respect to crustal events

Regarding published geothermobarometric data, the spatial coverage of the data is still fairly poor at the scale of the orogen. Along and in the vicinity of the Turku section, data are available in the northern part of the section. Only one area with PT data is available along the Salo section. Along the Sammatti section, the existing PT data are focussed on the south-central part of the section, and along the Vihti section the only available data are from the Veikkola granite.

Along the westernmost, Turku cross section, the peak metamorphic conditions are in the order of 800°C and 6-6.5 kbar (c. 18-21 km palaeodepth). The age of the peak metamorphism is not constrained, but younger 1814 Ma leucosomes show PT conditions of c. 680°C and 4 kbar (Väisänen & Hölttä, 1999; Väisänen et al., 2002). In the Sammatti section, the peak PT conditions of c. 800°C and 5-5.5 kbar (15-17 km palaeodepth) are dated at 1821-1814 Ma (Mouri et al., 2005). Based on these numbers we can estimate the age of the peak metamorphism (c. 19 km palaeodepth) as c. 1825 Ma, and subsequent exhumation to c. 16 km by c. 1815 Ma, giving an average exhumation rate of 0.03±0.02 cm/year. Similar exhumation rates were implied by Torvela and Ehlers (2010b) in the SW archipelago (SE of the SJSZ; Fig. 2b).

The Salo section has no PT data along or in the immediate vicinity of the section. There are some PT data points on the island of Kemiö c. 30 km west of Teijo-Perniö area in the cross-section line. The data from Kemiö imply a local palaeopressure variation of about 1 kbar with the higher-grade, sillimanite-bearing horizon on top of a sillimanite-absent one; this has been interpreted as a D1 palaeothrust later folded by F2 (van Staal and Williams, 1983; Fig. 5b). More recent results have implied that the sillimanite growth is younger, c. 1824 Ma, so that the juxtaposition would have resulted from D2 north-verging thrusts along F2 fold limbs (Levin et al. 2009). Either interpretation is consistent with our overall cross section interpretation. Associated phase assemblage changes are observed at outcrop also along the cross-section line and are indicated in the cross section with a red dotted line.

The south-central parts of Sammatti section (around Koski) show generally slightly lower observed PT conditions than elsewhere in the analyzed area. The general area is close to the Orijärvi triangle zone which is a remnant lower-grade crustal block; there, the PT conditions have been reported to be generally in the region of 600-700°C and 3-5 kbar (e.g. Latvalahti, 1979; Schumacher and Czank, 1987; Schneiderman and Tracy, 1991) with little or no migmatization that can be reliably linked to Fennian deformation (Nironen et al., 2016). Overall, there seems to be a c. E-W striking, westward-narrowing belt within which the metamorphic grade may be slightly lower than in the surrounding areas (Fig. 2b). In the Sammatti section, the largest drop in palaeotemperature occurs across the pyroxene-in transition (going from the Orijärvi domain in the south to Sammatti area to the north). However, this is apparently not accompanied by a significant drop in the palaeopressure according to available geothermobarometric results. The nature and cause of this relationship has been debated (thermal doming/CO_2 influx vs. tectonic juxtaposition; Schreurs & Westra,1986) and a detailed discussion of this is outside the scope of this paper, but it does seem from the structural relationships in the cross-sections that the variations may be principally structurally controlled by our interpreted normal faults in the area.

Along the Vihti section, there are no PT data from within the migmatites and granulites, but some data exist from the microcline granites and migmatite xenoliths within them in the southern part of the section (the composite Veikkola granite; Fig. 5d). The data are highly variable from c. 500°C and 4-5 kbar to c. 700°C and 5-6 kbar. The lower PT conditions seem to have been reached by c. 1825 Ma (Kurhila et al. 2005, 2011; Nironen and Kurhila, 2008), which is much earlier than in the Sammatti or Karkkila areas (Mouri et al., 2005). The implication is that the crustal level in the Veikkola area was higher than elsewhere in the studied cross sections along the orogenic strike. Its present position might be explained partly by the regional folding followed by late-orogenic extensional/transtensional shearing of the fold flanks. bringing the deeper sections such as the Karkkila and Sammatti domes towards the surface.

5.4 Strike-slip deformation

In addition to folds, the role of the steep, transpressional shear zones in orogens is equally if not more important for the overall strain partitioning. There are few crustal-scale strike-slip shear zones interpreted in the sections and in the field within the LSGM. The most significant ones in the study area are the Perniö shear zone following the southern margin of the Perniö granite (Salo section; Fig. 5b), and the Somero-Karkkila fault zone (S-KFZ) present in several sections. In addition, there is another distinctive, crustal-scale, dextral strike-slip zone within this part of the Svecofennian domain: the Sottunga-Jurmo shear zone SJSZ (Fig. 2a; no cross section is made here due to lack of sufficient data to produce a regional cross section: the area is mostly under the Baltic Sea).

The SJSZ has been studied in some more detail than the other crustal-scale strike-slip zones. Although it is not present in the cross sections, we consider it important to discuss its overall evolution in the orogenic context. This c. 1 km wide shear zone has been suggested to accommodate a lateral slip of at least some tens of kilometres, possibly up to ~100 km, but also some SW-side up component (Torvela and Ehlers, 2010a). The strike of the shear zone is NW-SE, i.e. ~45-30° to σ1 of the N-S to NNW-SSE orogenic compression. The shear zone has been active from at least ~1830 Ma, possibly as early as 1850 Ma, in upper amphibolite facies conditions of c. 600-750°C and 5.5-7 kbar; to ~1800-1790 Ma in lower amphibolite facies to upper greenschist facies conditions (Torvela and Annersten, 2005; Torvela et al., 2008). For 50 km lateral slip over 30 Ma deformation, strain rate is 5.3 x 10^{-14} s^{-1} and the displacement rate is 1.7 mm/a. These figures are consistent with other observations and models for crustal-scale shear and fault zones (e.g. Sassier et al., 2009) although Zhang et al., (2004) have observed slip rates from 1.5 mm/a to up to 12 mm/a in strike slip zones in the Himalayan-Tibetan system. Simple geometric relationships reveal that 50 km of lateral slip alone is capable of accommodating at least 35 km of shortening parallel to σ1, more if the σ1 was NNW-SSE orientated rather than N-S. These calculations only account for the shortening accommodated by lateral slip and is in the same order of magnitude as that estimated from folding farther to the east.

Compared to SJSZ, relatively little is known about S-KFZ which is seen in many of the sections, but it too seems to be a long-lived feature; field observations indicate that there is at least one, possibly dextral, phase of shearing at a high-grade (at least upper amphibolite facies), the age of which is unknown. The last (semi-)ductile phase at greenschist facies conditions is dated by Torvela and Kurhila (submitted) at 1.80-1.79 Ga. This age agrees well with other ages and observations across LSGM that the entire area had reached lower amphibolite to greenschist facies conditions by about 1.80 Ga (e.g. Lahtinen et al., 2005; Mouri et al. 2005; Torvela et al., 2008). The S-KFZ does not seem to be as extensive as the SJSZ, and can be traced with confidence for about 80 km. Therefore, the maximum slip (presumably at the centre of the shear zone) is much less than for the SFSZ. The S-KFZ has an overall strike which is at a larger angle to regional σ1, 60-90°. At 75°, e.g. 20 km of maximum slip along this shear zone can accommodate ~5 km of shortening, so it is unlikely to be a major contributor to the overall crustal shortening unless there is a significant dip-slip component, the evidence of which is uncertain. The Vihti cross section does imply that there may be several km of vertical motion along the S-KFZ, although the role of the post-Fennian greenschist facies deformation is unclear. Regardless, the S-KFZ is likely to follow an important crustal discontinuity because the lithology, the metamorphic grade, and the structural style change considerably across the shear zone. The nature of this discontinuity remains unknown, but we can preliminarily postulate either a major, F2-folded F1 thrust, and/or some form of structural inheritance stemming from the configuration of the palaeo-arc and/or –basin (e.g. an underlying basin-bounding fault) as an underlying cause.

The exact characteristics of the Perniö shear zone along the southern margin of the Perniö granite are unknown. It is relatively high grade for most parts and therefore broadly syn-D2. Its kinematics are unknown but are inferred to be dextral-oblique by

Selonen et al. (1996); this would agree with the regional c. N-S orientation of σ1 during the Fennian orogen. It too may have been a major F1 thrust or an early D2 thrust (possibly facilitating the intrusion of Perniö granite sheet), later steepened and reactivated as a dominantly strike-slip shear zone during D2 shortening.

6. Discussion

6.1 Deformation of the partially molten crust in S Finland

The post-1.85 Ga orogenic compression and the associated deformation in the study area was coeval with the large-scale migmatisation and the formation of the microcline anatectic granite sheets (Fig. 4; e.g. Suominen, 1991; Väisänen et al., 2002; Ehlers et al., 2004; Mouri et al., 2005; Kurhila et al., 2005; Nironen and Kurhila, 2008; Skyttä and Mänttäri, 2008; Kurhila et al., 2011). This syn-melt deformation was dominantly accommodated by pervasive, open to tight folding, some thrusting, and discrete strike-slip transpressional shearing (Fig. 5). The orogen-perpendicular folding was locally accompanied by ~ENE-WSW striking transtensional shear zones that formed along the flanks of the growing fold structures. As signs of N-S fold axes are rare or absent in some areas, we suggest that the granulite dome formation in e.g. Karkkila and Sammatti areas is related to this process rather than the putative post-Fennian E-W 'Nordic compression' suggested by Lahtinen et al. (2005).

The overall compressional high-T deformation was at its most intense around 1.83-1.82 Ga as constrained by several authors and by the observations made in this paper (e.g. Väisänen and Hölttä, 1999; Levin et al., 2005; Torvela et al., 2008; Fig. 4). Conversely, Skyttä and Mänttäri (2008) suggest that there was a "break" in compressional deformation and regional crustal extension at c. 1835-1825 Ma to explain the crustal-scale migmatisation and the apparent foliation-parallel boudinage and other "flattening" structures in their study area. We consider that the foliation-parallel extension/flattening structures may form along flanks of large-scale folds due to limbparallel and/or outer arc stretching, or else in other localities where foliations are sub-perpendicular to $\sigma1$ and need not reflect crustal-scale dynamics (Fig. 3c). Furthermore, the suggestion by e.g. Nironen (1997) that the migmatization of the Fennian middle crust was caused by orogenic collapse and extension is not supported by our observations. The structural context of the late-syn-compressional granulite-facies transtensional shear zones implies that transtensional/extensional features can form locally in response to e.g. growth of large-scale fold structures and need not reflect orogen-scale collapse; additionally, the shear zones post-date peak metamorphism (Torvela and Kurhila, submitted).

A genetic relationship between the migmatites and the 'true' microcline granite intrusions has been demonstrated by e.g. Väisänen and Hölttä (1999) and Stålfors and Ehlers (2006). Selonen et al. (1996) and Stålfors and Ehlers (2006) propose a model for how the steep strike-slip shear zones such as the Perniö shear zone acted as channels transporting the partial melts into higher crustal levels, where they are emplaced as sheets and subsequently folded. Once emplaced as horizontal sills or laccoliths however, the granites crystallize rapidly and act as rigid bodies, controlling the wavelength and amplitude of the crustal-scale folding. Apart from the 'true' anatectic microcline granite intrusions, there are ubiquitous granite-rich bodies that may not have migrated long distances, as evidenced by the gradational boundaries with the surrounding migmatites and by the 'ghost structures' of the remnant gneissose fabric present in these bodies. It is important to re-emphasise that these magma-rich migmatite volumes seem to show similar deformation styles and intensities to those lithologies that show lesser degrees of partial melting (Fig. 8).

To summarize, there is no evidence at a regional scale that the migmatisation and voluminous anatectic granite magmatism was associated with an extensional event, a wide-scale crustal 'collapse', or a 'break' in the compressional deformation; nor does the deformation intensity vary with the overall melt fraction within different parts of the middle crust. The crustal-scale folding occurred syn-melt but the deformation patterns are similar across the region regardless of the lithology or the volume of granitic material in the migmatites. Extensional features do occur but they are spatially limited, not necessarily associated with melt bodies, and very local and secondary to the dominant deformation style: folding. The ages of the granites and of the partial melts from which they are derived show a great range and cover the entire compressional phase from c. 1.85-1.81 Ga, implying that the partial melting was not coeval throughout the crust. Melts were, therefore, probably *not* available within entire middle crust simultaneously, meaning that any relative strength differences caused by partial melt fraction variations were transient (unless within shear zones). This has some important implications as to how the partially molten orogenic mid-crust behaves as a whole.

6.2 Implications for the deformation of migmatitic middle crust

The recognition that granitic (partial) melts exist in the orogenic middle-lower crust has had a fundamental role in the formulation of orogenic models. As Brown (2007) and others have shown, the feldspar content and the grain size of fully crystallized granitic bodies are normally larger than those of the surrounding rocks, making crystalline granites more competent than their host rocks. On the other hand, weakening of the deep crust by (assumed) coeval, voluminous, long-lived anatectic melts has been a crucial prerequisite for models evoking weak deformation style of the middle crust, such as the channel extrusion and channel flow models for the Himalayan-Tibetan system (e.g. Royden et al., 1997; Grujic et al., 2002; Searle, 2013). At a smaller scale, some models for metamorphic dome formation suggest melt-enhanced processes such as diapiric emplacement of granites which trigger formation of domes flanked by extensional shear zones (e.g. Ayoa et al., 2005; Langille et al., 2012).

In terms of the coexistence of melts and deformation, it is crucial that space and time scales of both processes are considered. In order to significantly weaken large parts of the crust, any syn-kinematic melt body needs to be long-lived enough at the time scale of the accumulation of strain (crystallization rate versus strain rate). For (granitic) intrusions, it is generally accepted that 'exotic' granitic bodies will start crystallizing very soon upon emplacement, and that the time it takes to completely crystallise a granite body depends on the intrusion volume/thickness and the temperature of the ambient crust. It takes in the order of 5-8 Ma to crystallise a 1-2 km thick granite sheet if it intrudes into rocks that are at c. 700°C, even longer if the country rock is near the melt solidus temperature (e.g. Davidson et al., 1994). On the other hand, the crystallisation time of granite bodies of any significant volume can be very short: a 1 km thick granitic sill intrusion, if intruded in a crust of 600°C, will completely crystallise in c. 150,000 years and will have accumulated a negligible amount of strain in that time (using a geologically 'typical' orogenic strain rate of 10^{-14} s^{-1}; Davidson et al., 1992, 1994). However, controversies still exist as to how much any individual granite bodies can focus deformation before fully crystallised: under a reasonably constant stress field, deformation is likely to continue through the crystallisation process of the melts, leaving deformation fabrics behind, but a significant amount of deformation may have been accumulated within shear zones at early melt stages. These controversies are exemplified by the so-called aneurysm model and other models for metamorphic core complex development that invoke 'diapirism' of melts. Such models commonly hypothesize that synkinematic melts that form by erosion-promoted decompression focus further deformation and exhumation, enabling a formation of a vertical 'channel' into which further melts are emplaced (e.g. numerical models of Rey et al., 2009; the Leo Pargil dome of Langille et al., 2012; or the Nanga Parbat syntaxis controversy; Zeitler et al. 2001; Koons et al., 2002; Crowley et al., 2009; Butler, 2018). Indeed, Butler (2018) argues against this traditional 'melt-weakening' approach, showing that the granite intrusions along the thrusts accommodating the rise of the Nanga Parbat massif show strong inclusion behaviour, not the weak inclusion behaviour expected from the model where the granite melts would focus deformation. As is the case of both Leo Pargil dome and Nanga Parbat, the most intensely deformed shear zones have not developed within the supposedly weaker granite-rich portions of the dome margins.

Interestingly, albeit at a very different scale, a study by Butler and Torvela (2018) of pegmatite dykes intruded in a shear zone suggests that a strong inclusion behaviour can be achieved already before the granite bodies are fully crystallized; this behaviour would be controlled by the preferential initial crystallization along the intrusion walls. The anatectic granite bodies and sheets in southern Finland seem to similarly show behaviours that are more consistent with competent deformation patterns (folding, or not deforming at all at a crustal scale) rather than weak ones (e.g. large-scale shear zone localization into volumes of high melt fraction; flattening of crustal fabrics; see also Fig. 1). In each case, the apparent lack of significant deformation focusing into the granite bodies (i.e. weak behaviour) can be explained with (partial) crystallization vs. strain rate considerations: assuming a duration of e.g. 5 Ma for the crystallization of a granitic sheet, and taking the estimated orogen-perpendicular shortening (strain) of 30-40% over a 20 Ma duration of orogenic compression, the amount of shortening accumulated while the granites were still molten becomes very small (<10 km across the ~100 km long cross sections in the case of Southern Finland). Where the granite sheets are less than ~2 km thick, the crystallization is up to 1-2 orders of magnitude faster and, consequently, the amount of accumulated strain is even lower.

With respect to migmatites, these considerations are more complicated. Some studies suggest that, once formed, partial melts within the crust are fairly stable and can persist for millions of years because the ambient temperature does not change. E.g. Bea et al. (1994) and Cesare et al. (2003) estimate residence times of >3 Ma and 5-10 Ma, respectively, for metapelitic migmatitic melts and related S-type granites in Spain; Rubatto et al. (2013) suggests 5-7 Ma for the metapelitic migmatites in the Higher Himalayan Crystallines. Conversely, Ayres et al. (1997) suggest <50 ka for the residence time of the anatectic metapelite melts in the Himalayas; Villaros et al (2009) conclude that residence time of melts in the metapelitic source for S-type granites is as little as 500 years in South Africa. Whilst ambient temperature is likely to be fairly stable, other physio-chemical conditions may change due to deformation-induced melt migration/segregation over relatively short distances, resulting in chemical disequilibrium and "back-reactions" with phases of rapid crystallisation (e.g. Kriegsman, 2001; White and Powell, 2010). White and Powell (2010) show that once early biotite starts to crystallize from partial melts, a process which is enhanced by partial melt migration/segregation, the H_2O content of the residual melt fraction decreases rapidly, leading to more or less instant crystallisation of the remaining melt. In orogenic systems prevalent in Southern Finland and globally, the various migmatites almost ubiquitously show biotite associated with the leucosomes: once crystallisation starts with segregation of biotite (or other hydrous minerals) it will happen very rapidly, effectively instantly 'freezing' that part of the once-partially molten rock. Further, it is unlikely that all parts of the crust melted at the same time or produced the same amount of melt, due to significant lithological heterogeneity in the continental lithosphere (e.g. White et al., 2017). This means that both melting and 'freezing' is likely to happen at different times in different parts of the crust. Either way, even if individual melt volumes survived for up to 5 Ma, the same consideration applies as for intrusive granite sheets: the amount of strain that can be accumulated within individual melt bodies in this time frame is fairly small in the context of the evolution of the entire orogen. This does not, of course, apply to shear zones which typically have 1-3 orders of magnitude higher strain rates than large-scale orogenic deformation: if anatectic melts formed within or intruded into a high-strain zone they do accumulate significant deformation. While melts were present in such shear zones they would, therefore, be weak; this weakness is likely to be enhanced by the high degree of fabric organization in the shear zones (Lee et al., 2018). A 'side-effect' of this strain localisation, on the other hand, would be to further reduce the strain accumulation in and weak behaviour of the partially molten volumes *outside* the shear zones.

From the above discussion, combined with the observation of overall competent deformation behaviour at a crustal scale in S Finland, the main conclusion is that sheets of granitic magma and volumes of partially melted (migmatitic) crust exert only minor control on the overall deformation style and strain distribution. Our findings do not support models which suggest that wide-scale late-orogenic collapse facilitated by weak middle crust and high gravitational potential energy is an inevitable consequence in hot orogens. The observed extensional/transtensional features are fairly local and rare compared to the overall compressional deformation style. Our findings are more compatible with a scenario for overall thickening with probable lateral escape facilitated mainly by local deformation zones, as a mechanism to accommodate orogenic compression (e.g. Chardon et al., 2011). However, we expand such models to suggest that despite melting and lateral escape, the dominant deformation style at the crustal scale is, at least in cases like the Svecofennian orogen, still relatively competent: the migmatitic middle crust must be strong enough to accommodate pervasive, large-scale folding. The Svecofennian orogen may or may not be directly comparable with e.g. the Tibetan-Himalayan system (we do not, for example, know how high the gravitational potential energy was), and we cannot observe the current mid-crustal deformation style within that system, but it is interesting to note that there are suggestions of potentially strong, not weak, behaviour within the middle crust in the Himalayan-Tibetan system and elsewhere from e.g. numerical models (Copley et al., 2011).

Another implication is related to the proposed positive feedback relationship between shear localization and increasing melt fraction (e.g. Brown and Solar, 1998; Holtzman et al., 2003). While strain localization into melts undoubtedly occurs locally in shear zones (see e.g. Lee et al., 2020), once certain melt fractions are exceeded the melt volume does not seem to be important for strain localization. The syn-orogenic shear zones and thrusts that have been detected in our study do not seem to be directly governed by the degree of migmatisation and anatectic melts. For example, the transtensional shear zones within the Karkkila granulite dome are interpreted to have formed along flanks of pre-existing fold structures and are, therefore, superimposed onto and controlled by the synchronous or slightly older compressional structural grain, not by melt fraction. The relative un-importance of melt fraction within the partially molten crust to material strength can be at least partly explained by considering the strength-melt fraction relationships over a large range of melt fractions (Fig. 9): once the melt connectivity threshold (MCT) is superseded, the absolute aggregate strength variation is fairly small with changing melt fraction. In other words, increasing the melt volume above c. 10% melt fraction introduces only minor effects to the aggregate strength, compared to 1-10% melt, as the strain becomes increasingly more distributed in the entire volume. Recent detailed studies have yielded similar results: Lee et al. (2018) found that the development of a intensely deformed shear zone in a highly migmatised crust in the Western Gneiss Region, Norway, was not controlled by the melt fraction (which reached 20% in some parts) but by the degree of the fabric organization. They show that once a shear zone is established, the rest of the volume undergoes relatively little strain (and little melt escape), despite the high melt fraction. The implication is that the presence of significant volumes of anatectic melts within the middle crust is only one factor that may control deformation localization and bulk strength, but that other factors such as inherited or developing structures, overall crustal fabric (an)isotropy and organisation, or lithological changes, are likely to be equally or more significant.

7. Conclusion

We have analysed the deformation style within the partially molten (migmatitic) orogenic mid-crust of Southern Finland, as an analogue to modern, unexposed orogenic roots. Our analysis covers both the time and spatial scales of the entire orogenic compression. We show that the overall melt (~leucosome/granite) volume percentage does *not* influence the deformation style within the migmatitic part of the crust. The dominant deformation style is competent shortening accommodated mostly by folding, not weak extensional collapse and/or flattening. We explain this by observing melt crystallization rate vs. orogenic strain rates: for both migmatitic partial melts and more voluminous granitic sheets, the amount of strain that can be accumulated while the melts are still (partly) liquid is fairly small in the context of the evolution of the entire orogen.

This combined with the fact that the melts are unlikely to be synchronous within the orogen, nor are they likely to be homogeneously distributed, leads us to conclude that the melt fraction (i.e. the volume of the melt) exerts only minor control on the *overall* deformation style and strain distribution at an orogenic scale. While melts probably locally and transiently accommodated significant strain (shear zones), the overall deformation style within the migmatitic mid-crust is competent, not weak.

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Figure captions

Fig 1. Basic geometric styles in response to deformation of rocks under general shear (shortening and shearing), with varying rheological contrasts, based on a number of field observations, analogue experiments and numerical modelling studies. a) Closely spaced alternating layers of weak and strong material (e.g. Cobbold et al., 1971; Carreras et al., 2013). The resulting deformation style is that of relatively regular, penetrative folding/crenulation of the fabric; b) Weak subhorizontal layers within a homogeneous stronger matrix (e.g. Gardner et al., 2017). The weak layers will focus strain, forming shear zones, whereas the more competent matrix may accommodate shortening by forming a sub-vertical fabric such as cleavage; c) Dispersed weak inclusions within a homogeneous stronger matrix (e.g. . Kohlstedt et al., 2010; Gardner et al., 2017). Strain will localise into the weak inclusions and in some cases link up to form continuous shear zones whilst matrix can develop deformation fabric; d) Strong subhorizontal layers within a homogeneous weaker matrix (e.g. Ramsay, 1967; Lister & Williams, 1983). The shortening is accommodated by folding whilst weak matrix can develop deformation fabric; deformation 'intensity' i.e. fold amplitude and wavelength is controlled by the thickness of the competent layer(s) and the rheological contrast; e) Strong inclusions within a homogeneous weak matrix (e.g. Griera et al., 2013). The strong inclusions behave passively, possibly rotating, whilst the weaker matrix develops a deformation fabric deflecting around the inclusions.

Fig. 2. a) Simplified geological map of the Fennoscandian shield. Key: **1** Archaean 3.2-2.5 Ga; **2-3** Early Palaeoproterozoic metasupracrustal and metavolcanic rocks 2.5-1.9 Ga; **4** Early Svecofennian supracrustal gneisses and migmatites 2.0-1.85 Ga; **5** Early Svecofennian pre- and synorogenic magmatic rocks 1.95-1.85 Ga; **6** Late Svecofennian (i.e. ‘Fennian’) granites and migmatites 1.85-1.77 Ga; **7** Sandstones ~1.2 Ga; **8**

Anorogenic rapakivi granites 1.65-1.4 Ga; **9** Sveconorwegian rocks 1.25-0.9 Ga; **10** Caledonian rocks 0.6-0.4 Ga; **11** Phanerozoic sedimentary cover <0.57 Ga; **CFGC** = Central Finland Granitoid Complex; **LSGM** = Late Svecofennian Granite Migmatite belt. See text for references and more detailed descriptions. Location of Fig. 2b is shown.

b) Simplified geological map of the Late Svecofennian Granite Migmatite belt and surrounding areas. Main cities mentioned in the text and the cross section line locations (Fig. 5) are shown, along with major fault zones. Note that in this map, the metavolcanic rocks in the cross sections (Fig. 5) are grouped together with the 'felsic intrusive' rocks for clarity. Note that the 'microcline granites' include both 'true' intrusives and highly migmatitic rocks with various protoliths (see text). The normal faults post-date the orogenic deformation, and many of the strike-slip zones show prolonged and complex kinematic histories (see text). Key geological locations identified in the text: **O** = Kisko-Orijärvi triangle zone; **V** = Veikkola granite; **P** = Perniö granite; **K** = Karkkila granulite dome **SKFZ** = Somero-Karkkila Fault Zone; **SJSZ** = Sottunga-Jurmo Shear Zone. Modified from Korsman et al. (1997), various 1:50 000 geological maps published by the Geological Survey of Finland, and from own field observations.

Fig. 3. Schematic structural evolution of southern Finland during the composite Svecofennian orogen. a) Pre-Fennian phase (D1) with crustal stacking and recumbent folding, and minor migmatization and granite magmatism; b) a period of tectonic quiescence and relaxation enhancing the recumbent folds and mainly flat-lying crustal fabrics (D1 thrusts omitted for clarity); c) Fennian phase (D2) with upright folding, voluminous anatectic magmatism and migmatization, and formation of large strike-slip shear zones (not shown). Note the suggested deformation styles of both pre-Fennian and syn-Fennian anatectic granite sheets. See text for references.

Fig. 4. Compiled published age data from southern Finland, with error bars where available. Note especially the synchronicity of gneiss metamorphic ages (proxies for orogenic deformation), and the migmatite leucosome and anatectic granite magmatic

ages between c. 1830-1805 Ma. Younger (post-1810-1805 Ma) ages are commonly related to cooling and lower amphibolite to greenschist facies deformation along discrete shear zones. Age data from 1) Suominen (1991); 2) Väisänen et al. (2002); 3) Ehlers et al. (2004); 4) Eklund and Shebanov (2005); 5) Jurvanen et al. (2005); 6) Kurhila et al. (2005); 7) Levin et al. (2005); 8) Mouri et al. (2005); 9) Skyttä et al. (2006); 10) Nironen and Kurhila (2008); 11) Pajunen et al. (2008); 12) Skyttä and Mänttäri (2008); 13) Torvela et al. (2008); 14) Kurhila et al. (2011); and 15) Torvela and Kurhila (submitted).

Fig. 5. Simplified cross sections along the lines shown in Fig. 2. a) Turku section; Age and PT data from Suominen (1991), Väisänen & Hölttä (1999), Väisänen et al. (2000, 2002), Johannes et al. (2003); b) Salo section; Age and PT data from Kurhila et al. (2005), Levin et al. (2005); c) Sammatti section; Age and PT data from Schumacher & Czank (1987), Schneiderman & Tracy (1991), Suominen (1991), Jurvanen et al. (2005), Mouri et al. (2005), Kurhila et al. (2006), Skyttä et al. (2006), Skyttä & Mänttäri (2008); d) Vihti section; Age and PT data from Kurhila et al. (2005, 2011), Nironen and Kurhila (2008), and Torvela and Kurhila (submitted). Note that the category 'microcline granite' encompasses 'true; intrusives i.e. syn-compressional anatectic granite intrusions but also highly migmatized (>50% melt estimate at outcrop) paragneisses and metavolcanic rocks, as per the convention used by the 1:100 000 bedrock maps of the Geological Survey of Finland (see also Fig. 6). The most notable feature of the crustal-scale deformation patterns for the purposes of this study is that the apparent strain intensity (e.g. wavelength/tightness of the folding) does not correlate with the presence or absence of highly anatectic or migmatized parts of the crust. Neither are the extensional shear zones within Karkkila granulite dome in D) associated with very high anatectic magma volumes, suggesting that extensional features were local and controlled by other factors than melt volume. See text for further descriptions and discussion.

Fig. 6. Summary of structural patterns illustrating the overall orientations of the crustal fabrics in representative parts of the study area. Equal angle lower hemisphere projections plotted using GEOrient v9.4.5. See Fig. 2b for geological map legend.

Fig. 7. Examples of typical rock types and fold structures within each rock type grouping used in the cross sections. Where a photo is annotated as 'highly migmatized' would in the cross sections and geological maps normally be included in the 'microcline granite' category. Therefore, the boundaries between the 'true' granites and the other 'microcline granites' is often diffuse and transitional.

Fig. 8. Plot showing the approximate fold wavelength-amplitude (a/λ) relationships for different dominant lithologies. The intensity of the deformation (folding) is inferred from the a/λ: higher ratio implies more shortening, assuming that all folds are of approximately the same type (see text). The 'microcline granite' lithologies i.e. those that consist of either anatectic granite intrusions or of migmatites with >50% leucosome at outcrop are highlighted with the shaded area. The plot shows that granite- and leucosome-rich lithologies are not more intensely deformed than the other lithologies: on the contrary, these lithologies tend to show slightly lower a/λ ratios than especially the paragneiss-rich parts of the crust.

Fig. 9. Strength vs. melt fraction plot, modified from Rosenberg and Handy (2005). After the melt connectivity threshold (MCT) is reached during partial melting, the aggregate strength weakening rate with increasing melt fraction slows down. There is a minor drop in strength when liquid-solid transition is reached, but overall there is little relative difference in aggregate strength above a melt fraction of ~7-8%. This implies that in partially molten (>~7%) crustal volumes, melt fraction may be less important for deformation style and deformation localization than other factors; a conclusion also reached in this paper.