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Constriction structures related to viscous collision, southern Prince Charles Mountains, Antarctica

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ABSTRACT

Macroscopic structures are investigated in a zone of highly contorted migmatites from the southern Prince Charles Mountains, Antarctica. Here, L-tectonite fabrics, rods, mullions, boudin pods, elongate enclaves, and fold hinges, are persistent linear features all plunging gently to the northeast. In contrast, amoeboid folds, ptygmatic folds and folded boudins with different orientations are the characteristic structures in transverse sections (perpendicular to the lineation). No consistent shear sense is recognised in any dimension. Together with strain and shape analysis, these observations strongly suggest that the deformation pattern is one of folding and stretching by constriction. Previous timing constraints indicate that this deformation overlapped with the waning stages of anatexis during decompression at approximately 510 Ma, up to 30 million years after initial orogeny at 540 Ma. The zone affected by constriction is several kilometres wide and has a contorted flower-like shape confined between two broad domal antiforms. In this context, the constricted zone is interpreted as a relatively late tectonic feature that could have formed via deep-seated viscous collision in response to orogenic collapse and doming.

1. Introduction

Constriction is a deformation process that involves the elongation of matter in a preferred axial direction accompanied by orthoradial shortening (Fig. 1). In the viscous regime, large-scale constriction will limit the ability of metamorphic rocks to flow in the transverse directions, and they may repeatedly fold, shear and flatten, to fill a narrowing region of space. The flow therein will not be steady or simple; it will locally vary with ongoing changes in geometry, melt content and strain-rate between the deforming heterogeneous materials (e.g., McLellan, 1984; Jiang and White, 1995; Burg and Vigneresse, 2002; Misra et al., 2014).

Where constrictive folding of layers occurs, intricate patterns of interfering folds are produced that resemble those of several superposed folding events. Examples of this phenomenon are shown in constricted model clay layers (Ramberg, 1959; Ghosh et al., 1995; Zulauf et al., 2016), salt flows (Talbot and Jackson, 1987), and deeply eroded migmatites (Chetty and Bhaskar Rao, 2006). If present, obstacles including large boudins will perturb the flow, making the fold patterns considerably more complicated. Yet, although constriction produces complex fold structures in the transverse sections, it can also produce a relatively simple and constant stretching lineation marking the average maximum elongation in the longitudinal direction (e.g., Kobberger and Zulauf, 1995; Zulauf and Zulauf, 2005; Ziv et al., 2010). In metamorphic rocks the lineation will likely include L-tectonite fabrics, which can be equated with prolate ellipsoids indicative of constrictional strain (i.e., $X \gg Y \geq Z$; Flinn, 1965; Sullivan, 2013).

Given the similarities between stretching lineations in constriction and in simple shear flow, their tectonic interpretation is not always clear (e.g., Passchier et al., 1990; Flinn, 1994). The presence or absence of a persistent sense of rotational shear parallel or perpendicular to the lineation is a key distinction. However, it is widely apparent that stretching lineations in constrictional strain regimes of the lower crust are produced through various combinations of pure and simple shearing. For example, they are frequently found in transpression or transtension settings (e.g., Dias and Ribeiro, 1994; Tikoff and Greene, 1997; Krabbendam and Dewey, 1998; Solar and Brown, 2001; Vassallo and Wilson, 2002; Dewey et al., 2006). They also appear extensively where

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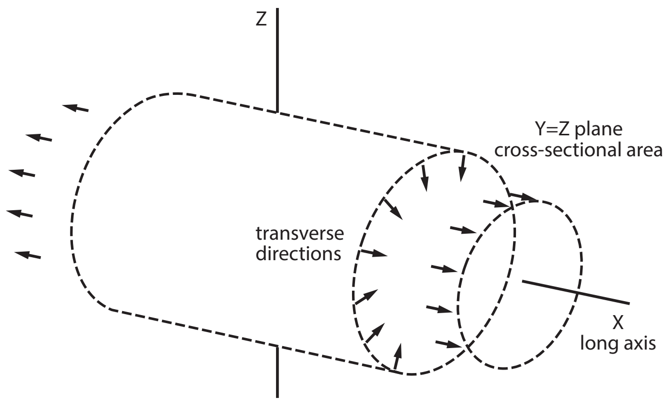

Fig. 1. Idealised constriction deformation with uniform radial shortening in *YZ* planes and elongation parallel to *X* (after Bayly, 1992).

penetrative ductile flows, of either non-coaxial or purely elongational nature, interfere with vertical extrusion flows linked to diapiric and domal uplifts (e.g., Brun et al., 1981; Poli and Oliver, 2001; Cagnard et al., 2006; Chardon et al., 2011; Duclaux et al., 2007; Sullivan, 2013).

With the above in mind, this paper is concerned with macroscopic structures inferred to have resulted from the constriction of migmatites in the Turk Glacier area of the Mawson Escarpment, southern Prince Charles Mountains (Fig. 2). Previous regional mapping places these rocks within a steeply dipping shear zone dissecting two domal antiforms that modify an older flat-lying transposition foliation (Fig. 2e; Boger and Wilson, 2005; Boger et al., 2008). New structural observations, orientation and strain measurements, show that the steeply dipping zone has a contorted flower-like shape and internally complex fold patterns that can be attributed to constriction. The viscous collision model of Rey et al. (2011) describes how a steep median zone showing belated constrictional deformation can occur between two domes in orogenic collapse and is a hypothesis considered below.

2. Geological setting

The Mawson Escarpment is the largest cross sectional exposure (130 km long) of some of the oldest, and most tectonically evolved, Precambrian rocks in Antarctica (Fig. 2). It has two major tectonic subdivisions; the Ruker and Lambert complexes (e.g., Kamenev et al., 1993; Mikhalsky et al., 2001; Boger et al., 2001).

The Ruker Complex, in the southern part of the escarpment, is dominated by trondhjemitic and granitic gneisses whose granitoid protoliths were emplaced at 3.39–3.37 Ga and 3.17 Ga, respectively (Boger et al., 2006; Mikhalsky et al., 2006). Supracrustal rocks were deposited after 3.15 Ga (Phillips et al., 2006), and deformed together with the gneisses under amphibolite facies conditions at 2.79–2.77 Ga prior to the intrusion of pegmatitic granites at 2.65 Ga (Boger et al., 2008). Thereafter the Ruker Complex was largely stabilised, although several crosscutting mafic dyke swarms intruded the gneisses during extensional events in the Proterozoic (Mikhalsky et al., 2013). Subsequently, these mafic dykes were offset by late-stage mylonitic shear zones thought to be active in the latest Neoproterozoic or Cambrian (Boger et al., 2006).

To the north, the Lambert Complex is characterised by large volumes of 2.49–2.42 Ga granodioritic gneiss and 2.18–2.08 Ga granitic gneiss (Mikhalsky et al., 2006; Boger et al., 2008; Corvino et al., 2008). An older core to these rocks may be represented by 3.52 Ga trondhjemitic gneiss that is exposed for several kilometres in the middle of the escarpment (Boger et al., 2008). Interleaved metasupracrustal rocks, including thick marble sequences, were possibly deposited in a marine shelf environment following one or several rifting events sometime after 2.45 Ga, but before 960 Ma. These rocks, and the Palaeoproterozoic gneisses, were then repeatedly reworked during orogenies at approximately 960–900 Ma and 540–490 Ma. The degree of overlap of reworking, anatexis and related granitoid magmatism for these intervals, and the microstructural relationships between the two metamorphic peaks, are difficult to delineate. In general, rocks in the north were metamorphosed to granulite facies during the 960–900 Ma event, whereas those in the south record peak upper amphibolite facies assemblages related to the 540–490 Ma event (Boger and Wilson, 2005; Phillips et al., 2009; Corvino et al., 2011).

The study area lies in the southern part of the Lambert Complex. Here, the regional structure is dominated by a series of large-scale buckle antiforms each in the order of ten kilometres or more across (Boger and Wilson, 2005). However, the plunge variations of these antiforms suggest a periclinal or radial curvature typical of dome structures. Internally they show a gently dipping foliation that is macroscopically defined by gneissic compositional layering and, on smaller scales, by similarly oriented grains or aggregates of biotite and sillimanite. Attenuation and boudinage of layers is common and can be clearly seen along the limbs of isoclinal folds up to a kilometre in the cliff faces (Fig. 3). Hence, this foliation is interpreted to be a product of large-scale transposition (in the sense of Williams, 1983; Williams and Jiang, 2005); whereby older structural elements have been rotated and flattened into strong parallelism via isoclinal folding in a non-coaxial flow regime (Boger et al., 2005; Corvino et al., 2011).

Separating the domes are steeply dipping zones, up to several kilometres wide, in which the transposition foliation is rotated and further flattened by upright folds and shear zones striking E–W or NE–SW. Development of both the domes and steep zones is related to collision of the Lambert and Ruker complexes at 540–490 Ma during the assembly of Gondwana (Boger and Wilson, 2005; Boger, 2011). A steeply dipping zone 15 km south of the study area is plausibly interpreted as the suture (Boger et al., 2001). These deformations are superimposed on the transposition foliation, which may belong partly or fully to the same orogeny, or to the 960–900 Ma event that is predominant north of the study area.

3. Study area description: a contorted steep zone

Complexly folded and contorted rocks of the study area are well exposed along a southwest facing cliff, over 3 km long and 500 m tall, on the north side of the Turk Glacier (Fig. 2d). The cliff is differentiated into three parts based on changes in rock type and dominant dip direction; these are referred to as the Western, Central and Eastern domains. Orientations of lineations, foliations, and leucogranitic rocks of anatectic and intrusive origin (emplaced before, during and after folding), are shown on stereograms in Fig. 4.

The Western Domain consists predominantly of granitic to granodioritic biotite gneiss. Characteristically, this gneiss includes large amphibolite boudins, up to several tens of metres across, and is probably correlative with rocks in the dome at McIntryre Bluff to the north (Fig. 2e). Whereas the boudins are widely separated in the dome, here they are strongly folded and irregularly stacked giving a gross indication of their subsequent shortening and distortion (compare Figs. 2d and 3). Some of these boudins appear subcircular in transverse sections containing the fold profiles, although they are elongate parallel to the fold axes. Outside of the boudins, complex series of narrow leucogranite veins are the most conspicuous and persistent feature, though they, too, are almost everywhere shortened by ptygmatic folding (see below). In most