Journal of the Geological Society, London, Vol. 154, 1997, pp. 111-116, 4 figs. Printed in Great Britain

Shear zone reactivation at granulite facies: the importance of plutons in the localization of viscous flow

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Abstract: Localization of flow in natural quartzo-feldspathic shear zones at granulite-facies temperatures does not occur purely by dislocation creep and dynamic recrystallization. From a review of some natural shear zones, initial faulting focused along a pre-existing shear zone, and/or the boundary of a volume of relatively strong crust or lithosphere, guides emplacement of hot mantle-derived melts into the base of the crust, leading to extensive melting. When crystallized, the sub-solidus plutons represent a focused zone of thermally softened material, capable of deforming readily and developing the full range of crystal-plastic microstructures and fabrics associated with mylonites, contrasting with the intrusion-free wall rocks.

Keywords: Shear zones, plutons, granulites, reactivation, faults.

Localization of flow occurs where the material within a band yields more readily than the wallrocks. When the flow is non-coaxial, the band is geologically referred to as a shear zone. In many mid-crustal shear zones, initial localization of flow is associated with mechanical strain softening, the rheological result of dynamic recrystallization (e.g. Urai et al. 1986; Hirth & Tullis 1992). It has long been axiomatic that localized shear zones remain inherently weak, and that they are susceptible to later reactivation (e.g. Watterson 1975). This is generally predicated on the assumption that dynamic grain size refinement leads to softening by the enhancement of grain-size dependent deformation processes, such as grain-boundary diffusion creep and diffusion-accommodated grain-boundary sliding (e.g. Schmid et al. 1977; Kirby 1985; Rutter & Brodie 1988; Rubie 1990). However, depending on the temperature and strain rate of the deformation, dynamic recrystallization does not necessarily result in softening (e.g. Tullis 1990) and, as will be suggested here, the localization of flow in hot shear zones is not necessarily the product of grain-size refinement.

Shear zone localization

Where faults are stress-controlled, it is likely that the boundary conditions of localization relate to the load-bearing upper to middle crust, more specifically the inflection 'point' (brittle– plastic transition) which figures so prominently in crustal strength profiles (e.g. Meissner & Strehlau 1982). However, large-scale shear zones, whose true width approaches the thickness of the continental crust, are more likely to be kinematically controlled, and localization will be determined by the behaviour of their deeper-seated, thermally softened segments.

Because the continental crust is essentially quartzofeldspathic in composition, recent experimental investigations of the role of dynamically recrystallizing quartz and feldspar in flow localization are of great significance to understanding shear zones. Where deformation occurs entirely by dislocation creep, strain softening does not occur in either quartz or feldspar mosaics in the rotational dynamic recrystallization regime (e.g. Tullis *et al.* 1990; Hirth & Tullis 1992). The density of free dislocations within the rotating and deforming subgrains remains at the same order of magnitude as in the parent material, thereby maintaining the strength of the recrystallizing aggregate. However, strain softening is inherent in the migrational dynamic recrystallization regime which results in the formation of small, initially strain-free new grains. Because of their small size, the volumes within deformed new grains are readily swept clean of dislocations by the next local grain boundary migration event (e.g. White *et al.* 1985).

The different thermal sensitivities of dislocation climb rates in quartz and feldspar results in different strain softening behaviours in the two minerals at sub-granulite temperatures. The sluggishness of dislocation climb in feldspar favours the accommodation of dislocation glide by migrational recrystallization (Tullis & Yund 1992). Since dislocation climb in quartz is efficient at comparable temperatures, strain softening in feldspar will control flow localization within the deforming microstructure in quartzofeldspathic rocks, where the feldspar component is load-bearing (Tullis *et al.* 1990).

However, at granulite facies, feldspar enters the rotational recrystallization regime, and localization of high temperature flow in quartzo-feldspathic rocks is suppressed (Tullis et al. 1990). It therefore follows that successful localization of flow in natural quartzo-feldspathic shear zones at granulite facies temperatures did not occur purely by dislocation creep and dynamic recrystallization (Tullis 1990). One alternative microstructural path which might lead to rheological softening is the enhancement of grain-size sensitive deformation processes, such as grain boundary diffusion creep. However, diffusion in anhydrous rocks is minor, even at elevated temperatures, and its enhancement by fluids remains equivocal (e.g. Tullis 1990; Farver & Yund 1995; Tullis et al. 1996). In any case, water is unlikely to reside in anhydrous rocks at ambient granulitefacies conditions, and its transient presence in hot quartzofeldspathic materials would likely lead to melting (e.g. Tullis & Yund 1980; Dell'Angello et al. 1987). Accordingly, it is appropriate to review the geology of some natural, crustal-scale, granulite facies shear zones in order to identify the common factor which may have lead to successful flow localization.

Great Slave Lake shear zone

In the NW Canadian Shield (Fig. 1), the boundary between the Archaean Slave and Western Churchill continents is marked

RAE SLAVE SLAVE McDomit Unit Athabasca lozenge Virgin River shear zone Approximate limit of Paleozoic cover 200 km

Fig. 1. Distribution and location of principal tectonic elements in the western Canadian Shield. Locations of Figs 2 and 3 are show by boxes, illustrating details of Great Slave Lake shear zone and East Athabasca mylonite triangle (EAMT), respectively. Striding mylonite belt (SMB) runs along the southeast margins of the Selwyn and Three Esker lozenges, or crustal-scale boudins of remnant Middle Archaean arc material. Western Churchill continent is composed of Rae and Hearne crusts, separated by the Snowbird tectonic zone.

by the Great Slave Lake shear zone, a discrete corridor of granulite (850–900°C, 800 MPa) to lower greenschist-facies mylonitic rocks (Hanmer 1988, 1991; Hanmer *et al.* 1992). The shear zone is up to 25 km wide and has been traced in outcrop

for approximately 200 km along strike (Fig. 2). It is a strikeslip structure which formed within part of an active Early Proterozoic magmatic arc at the leading edge of the upper plate during oblique convergence and continental collision at 1.97 Ga (Tirrul & Grotzinger 1990; Hanmer *et al.* 1992). The 2.05–1.92 Ga granitoid protoliths of Great Slave Lake shear zone constitute the Laloche batholith, contiguous with the granitic rocks of the 2.0–1.9 Ga Thelon and Taltson magmatic arcs (Fig. 2; see Hanmer *et al.* 1992 for references). Collectively, the arc plutons underlie one of the most pronounced positive magnetic anomalies in the western Canadian Shield (Geological Survey of Canada 1987).

From field mapping, kinematic study and geothermobarometry, the shear zone records the progressive uplift, cooling and narrowing of an active dextral strike-slip shear zone from granulite to greenschist facies (Hanmer 1988, 1989). The shear zone is subdivided into longitudinal belts of mylonite, according to their metamorphic grade, relative age of mylonitization and kinematic significance (Hanmer 1987, 1988). All the belts and fault zones are upright. For the most part, the mylonitic foliation and layering are vertical, the extension lineation is subhorizontal and shear-sense criteria indicate a dextral sense of strike-slip displacement (see Hanmer 1986, 1990; Hanmer & Passchier 1991, 1995). The Great Slave Lake shear zone mylonites are principally derived from pyroxene-hornblendebiotite \pm garnet granites, clinopyroxene tonalites and included panels of mixed paragneiss. The granites were emplaced during the tectonic activity of the shear zone, coeval with intrusion of dispersed mafic dykes. By dating the granites (U-Pb on zircon; Hanmer et al. 1992), it has been shown that granulite-facies

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Fig. 2. Generalized geology of Great Slave Lake shear zone (after Hanmer *et al.* 1992). 'Mylonites' are derived from predominantly granite protoliths, unless otherwise indicated. See Fig. 1 for location.

and lower amphibolite-facies mylonites are younger than c. 1.98 Ga and 1.924 Ga, respectively. However, upper amphibolite-facies, strike-lineated protomylonites and mylonites of the NE part of the shear zone were formed prior to the emplacement of a 2.562 Ga granite, thereby indicating that the Early Proterozoic component of Great Slave Lake shear zone is in fact a reactivated structure.

There are two important points to retain from the Early Proterozoic shear zone geology. First, the shear zone was 25 km in true width when it was reactivated at granulite facies; second, the shear zone is spatial coincident with the Laloche batholith over its exposed strike length (Hanmer 1988).

Striding-Athabasca mylonite zone

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The Striding-Athabasca mylonite zone is a 400 km long segment of the Snowbird tectonic zone, a NE-SW-trending linear gravity anomaly which can be traced across the western half of the Canadian Shield (Fig. 1; Hanmer et al. 1995a). It is a linked system of granulite-facies mylonite belts (850-1000°C, 1.0-1.5+GPa; Hanmer et al. 1995a, b; Williams et al. 1995; Snoeyenbos et al. 1995) which trace a sinuous course along a train of three 100 km scale, magnetically defined 'lozenges' (Athabasca, Selwyn and Three Esker; Fig. 1). The lozenges appear to be the remnants of an island arc assemblage of mafic to intermediate composition which were dismembered (crustalscale boudinage) during continental suturing at c. 3.2 Ga, an early event in the growth of the Western Churchill continent (Hanmer et al. 1995a). During the Late Archaean, granulitefacies fabrics of the Striding-Athabasca mylonite zone were penetratively developed during dextral strike-slip shearing along the southeastern margins of the crustal-scale boudins at c. 2.6 Ga. The intimate spatial and kinematic relationship between the Late Archaean mylonites and the boudins indicates that the lozenges were relatively stiff and served to focus the deformation (Hanmer et al. 1995b). However, contrary to expectation, the mylonites were not derived at the expense of the relatively softer granitic wallrocks.

The mylonite zone is geometrically complex and consists of two parts: the East Athabasca mylonite triangle (Hanmer 1994: Hanmer et al. 1994) at the northeastern apex of the Athabasca lozenge, and its northeastern extension, the Striding mylonite belt (Fig. 1; Hanmer et al. 1995a). The East Athabasca mylonite triangle is structurally divided into an upper and a lower deck (Fig. 3). The lower deck comprises three kinematic sectors; two upright, conjugate, penetratively mylonitic, granulite facies strike-slip shear zones, each about 15 km thick, separated by a central septum. In the central septum the bulk finite strain is of relatively low magnitude, and the bulk deformation path approximates to progressive pure shear. The upper deck overlies the lower deck and was initially emplaced along a discrete basal thrust, as indicated by the presence of relict eclogites (Snoeyenbos et al. 1995). It is now entirely occupied by a penetratively mylonitic, 10 km thick dip-slip granulite-facies shear zone, whose latest movements were extensional (Fig. 3). The Striding mylonite belt is a zone of through-going granulite to upper amphibolite ribbon mylonites, 5-10 km thick, extending from the East Athabasca mylonite triangle along the southeastern side of the Selwyn and Three Esker lozenges (Fig. 1). It is steeply dipping, strikelineated, and has a uniform dextral shear-sense. The overall kinematic framework of the Striding-Athabasca mylonite zone as a whole is one of strongly transpressive, dextral, strike-slip shearing (Hanmer et al. 1995a).



Fig. 3. Generalized distribution of kinematic elements of the East Athabasca mylonite triangle, corresponding to horizontal extensional flow at the northeastern end of the Athabasca lozenge, or crustal-scale boudin (inset). Stereonets of poles to mylonitic foliation, and extension lineations are shown. LDD, LDS and UD are Lower Deck Dextral, Lower Deck Sinistral and Upper Deck, respectively. Shear-sense sectors correspond to *c*. 2.6 Ga Late Archaean fabrics, except for LDS which includes Middle Archaean *c*. 3.2 Ga fabrics (see Fig. 4a). Arrows indicate latest displacements of the upper deck.

The geological history of the Striding-Athabasca mylonite zone (Hanmer et al. 1994, 1995a; Hanmer 1996), best documented in the East Athabasca mylonite triangle, began with the deposition of semipelitic to pelitic sediments, now represented by diatexites. At c. 3.2 Ga, the sediments were intruded by an arc assemblage, comprising a tonalite batholith (Fig. 4a), a major mafic dyke swarm and pervasive granite sheeting. The dykes, granite sheets and later tonalites were contemporaneous with pervasively developed granulite facies shearing and mylonitization (Williams et al. 1995). The shear-sense in the preserved mylonites (Fig. 3) is compatible with bulk NE-SW subhorizontal extension which Hanmer et al. (1995a, b) relate to the initial dismemberment of the arc during accretion and continental collision. At c. 2.6 Ga, widespread, penetrative ribbon fabrics developed, coeval with the emplacement of two large mafic plutonic bodies and a voluminous suite of synkinematic granulite-facies hornblende-garnet-clinopyroxeneorthopyroxene metagranitoids, associated with granulite-facies diatexite, throughout the upper deck and the western and central parts of the lower deck (Fig. 4b and c). Similar granulite-facies mylonitized diatexite, mafic and granitic mylonites, and syntectonic c. 2.6 Ga granites are present along the length of the Striding mylonite belt. It is these plutonic and migmatitic rocks that constitute the protolith to the ribbon mylonites throughout the Striding-Athabasca mylonite zone. However, it is important to note the total absence of penetrative Late Archaean deformation fabrics in the eastern part of the lower deck, where the 3.2 Ga mylonitic tonalite (Fig. 4a) is free of any significant Late Archaean plutonism.

With cooling from c. 2.6 Ga granulite-facies metamorphism, the high-grade metagranitic rocks in the East Athabasca mylonite triangle were intruded by synkinematic hornblende– biotite granites and leucogranites and their mylonitic equivalents, emplaced just below the upper deck, or adjacent to the lateral limits of the mylonite triangle (Fig. 4c; Hanmer *et al.* 1995*a*). Significantly, the only *c*. 2.6 Ga ribbon mylonites associated with the Middle Archaean tonalite in the eastern part of the lower deck are granulite-facies fabrics developed at its western contact with hornblende–garnet–clinopyroxene–

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Fig. 4. Distribution of c. 3.2 Ga tonalite (a), and c. 2.6 Ga mafic (b) and granite plutons (c) in the East Athabasca mylonite triangle. All other space within the triangle is underlain by diatexite.

orthopyroxene granitoids, and amphibolite-facies fabrics developed at its eastern contact with a hornblende-biotite granite pluton (compare Figs 4a and c). During the development of the ribbon mylonites, tectonic flow in the East Athabasca mylonite triangle evolved with cooling from a pervasive hightemperature granulite-facies regime to a more localized middle to lower amphibolite-facies regime. This is manifested by the focusing of deformation within the later sub-granulite-facies granites, as well as the amphibolite-facies reworking of the granulite-facies mylonites along the northwest margin of the lower deck. With further cooling and localization, deformation was confined within the very narrow Black–Bompas and Straight–Grease greenschist mylonitic faults at the lateral limits of the lower deck (Fig. 3).

There are three important points to retain from the mylonite zone geology. First, both the initial localization (c. 3.2 Ga) and the subsequent reactivation (c. 2.6 Ga) occurred at granulite facies. Second, both initial localization and reactivation were temporally and spatially related to voluminous batholithicscale mafic and granitic plutons and extensively melted paragneiss. Third, not only are c. 2.6 Ga ribbon fabrics developed in the presence of large synkinematic intrusions, but reactivation of the older mylonite fabrics simply does not occur in the absence of voluminous Late Archaean plutons.

Discussion

The foregoing highlights three first order common features in Great Slave Lake shear zone and Striding–Athabasca mylonite zone. The first feature is the evidence for reactivation of older structures, which at Great Slave Lake are represented by an earlier mylonite zone. In the Striding–Athabasca mylonite zone, the older structure comprises a crustal-scale strength anomaly, as well as earlier mylonite fabrics. The second feature is the intimate spatial and temporal relationship between large volumes of synkinematic melt and the localization of deformation. The third feature is the association of mafic intrusions with the more voluminous granitic plutons. The difference in volume between the mafic dykes at Great Slave Lake and the voluminous mafic plutons north of Lake Athabasca reflects a difference in exposed crustal level, as indicated by the geothermobarometric evidence.



These associations suggest that mafic melts of upper mantle origin were emplaced into the lower continental crust where they gave rise to a thermal anomaly and provoked extensive crustal melting to produce granitic magmas and migmatites. The role of faults in focusing and volumetrically accommodating magma emplacement (e.g. Hutton 1988; Petford et al. 1993; Brown 1994; Speer et al. 1994), and the genetic relationship between faulting and crustal-scale strength heterogeneities (Vilotte et al. 1984; England & Houseman 1985; Tomassi and Vauchez 1995), are both well established. It is also readily apparent that a pre-existing anisotropy can act to focus faulting (e.g. Jaeger 1961; Price & Cosgrove 1990, pp. 134-137). Accordingly, one may envisage a scenario whereby initial discrete faulting focused along either a pre-existing shear zone, and/or the boundary of a volume of relatively strong crust or lithosphere, guides the emplacement of hot mantle-derived melts into the base of the continental crust. The resulting crustal thermal anomaly leads to extensive melting and the production of granitoids and diatexite. Magma rising from the melt zone would be similarly guided by the same faulting. When crystallized, the subsolidus plutons would represent a focused zone of thermally softened material, capable of deforming readily and developing the full range of crystal-plastic microstructures and fabrics associated with mylonites (e.g. Gapais & Barbarin 1986; Blumenfeld et al. 1986), compared with the intrusion-free wall rocks. This scenario is quite distinct from the enhancement of localized viscous deformation by the presence of small-scale melts (Hollister & Crawford 1986).

The foregoing raises a number of questions. What is the potential role of strain heating in nominally dry, possibly strong, granulite facies rocks? Strain heating is related to the absolute strength of the deforming material at a given ambient temperature (e.g. Brun & Cobbold 1980), whereas localization is a function of relative strength. At temperatures in the range 850–1000°C, it is unlikely that the strain heating of even anhydrous quartzo-feldspathic rocks would give rise to a significant further strength decrease, compared with already thermally softened material (e.g. Fleitout & Froideveaux 1980). What is the origin of the mantle-derived mafic melt? In the case of Great Slave Lake shear zone, one might speculate about the possibility of fault-related decompression melting

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(Harris & Massey 1994; Bown & White 1995), or peridotitehosted strain heating at mantle depths (Fleitout & Froidevaux 1980). However, as demonstrated by Hanmer *et al.* (1995*b*), the bends in the primary geometry of the Striding–Athabasca mylonite zone preclude the possibility of significant displacements on the strike-slip shear zone. Accordingly, the origin of the mafic melts remains undetermined.

Conclusions

Initiation and reactivation of localized flow in quartzofeldspathic rocks, at elevated temperatures typical of the deep continental crust, do not occur solely by microstructural strain softening. Large-scale rheological variations or heterogeneities appear to focus discrete faulting, plutonism and melting in the lower crust, leading to the formation of a localized belt of thermally softened subsolidus material in which progressive non-coaxial viscous deformation occurs more readily than in the intrusion-free wall rocks, i.e. a shear zone. The spatiotemporal association of on-going viscous deformation and synkinematic plutonism during the active life of such shear zones suggests a positive feedback loop between continued localized flow and the presence of large-scale melts.

This is GSC contribution No. 1996236.

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- Received 24 April 1996; revised typescript accepted 3 September 1996. Scientific editing by Geoff Lloyd and Alex Maltman.