Pergamon

Journal of Structural Geology, Vol. 17, No. 4, pp. 493 to 507, 1995

0191-8141/95 $9.50 + 0.00

0191-8141(94)00070-0

Modest movements, spectacular fabrics in an intracontinental deep-crustal strike-slip fault: Striding-Athabasca mylonite zone, NW Canadian Shield

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(Received 27 September 1993; accepted in revised form 28 May 1994)

Abstract—Geometry and strain partitioning within lower-crustal *intraplate* strike-slip shear zones can be extremely complex, compared with analogous structural levels of *interplate* strike-slip shear zones sited at plate margins. Striding-Athabasca mylonite zone, Canadian Shield, is a spectacular *ca.* 500 km long granulite facies continental intraplate shear zone. The shear zone is composed of Middle Archean granulite facies annealed mylonites (*ca.* 3.13 Ga) and Late Archean (*ca.* 2.62–2.60 Ga) granulite facies ribbon mylonite belts, which thread a sinuous course along a chain of crustal-scale 'lozenges' cored by relatively stiff rocks of mafic to intermediate composition. To the northeast, the mylonites form a N–S-trending, 5–10 km thick, dextral strike-slip belt. To the southwest, this bifurcates into a pair of conjugate strike-slip shear zones, overlain by a contemporaneous dip-slip shear zone.

Striding-Athabasca mylonite zone was kinematically inefficient as a strike-slip fault and cannot have accommodated large wallrock displacements. Nevertheless, spectacular granulite facies ribbon mylonites were formed throughout the shear zone, reflecting the very high temperatures (*ca.* 850–1000°C), high recrystallization rate/strain rate ratios, and the transpressive nature of the deformation ($W_k < 1$), possibly accommodated by significant volume loss by magma migration.

INTRODUCTION

Large-scale, continental strike-slip faults, both fossil and modern, have long been the subject of geological research, particularly regarding their relationships to plate tectonics. Most studies have focused upon *interplate* faults formed at active plate boundaries. The surface traces of interplate faults are characteristically straight, even corresponding to terrestrial great circles (e.g. Hoffman 1987), with strike lengths on the order of thousands of kilometres. Kinematically, the deep-crustal parts of interplate faults tend to be relatively simple, closely reflecting the generalized pattern of plate interaction (e.g. Lacassin 1989, Hanmer *et al.* 1992a). Compared to interplate faults, little work has explicitly focused on the geology of large-scale *intraplate* strike-slip faults formed within the continental crust, far removed from active plate boundaries. Our work on high-temperature shear zones suggests that, in contrast to interplate faults, the geometry and strain partitioning within intraplate faults can be extremely complex and are significantly influenced by local, crustal-scale rheological variations (e.g. Hanmer & Kopf 1993, Hanmer *et al.* in press a,b)

In this contribution, we examine the structural and metamorphic history of an intraplate, crustal-scale, high grade, Precambrian fault from the western Canadian Shield (Fig. 1); the Striding-Athabasca mylonite zone (Hanmer *et al.* in press b). Our principal conclusion is that, despite the visually spectacular nature of its associated ribbon fabrics, the Striding-Athabasca mylonite zone cannot have accommodated significant displacement of its wallrocks. The development of ribbon fabrics in such a kinematically limited shear zone implies that tectonic flow deviated strongly from simple shear. This, combined with very high temperatures, and high recrystallization rate/deformation rate ratios, allowed the formation of ribbon ultramylonites at relatively low shear strains.

STRIDING-ATHABASCA MYLONITE ZONE

The Striding-Athabasca mylonite zone (Hanmer *et al.* in press b) underlies the central portion of the geophysically defined Snowbird tectonic zone (Fig. 1). The Snowbird tectonic zone (Hoffman 1988) is a high amplitude linear anomaly in the horizontal gravity gradient map of the Canadian Shield which extends almost 3000 km from the Canadian Rocky Mountains to the eastern side of Hudson Bay (Goodacre *et al.* 1987, Thomas *et al.* 1988). In northern Saskatchewan and the adjacent parts of the Northwest Territories (Fig. 2), low amplitude anastomosing linear elements in the magnetic field occur along the trend of the gravity anomaly and define a train of three crustal-scale elliptical features (Geological Survey of Canada 1987, Pilkington 1989), referred to as the Athabasca, Selwyn and Three Esker 'lozenges' (Fig. 2;

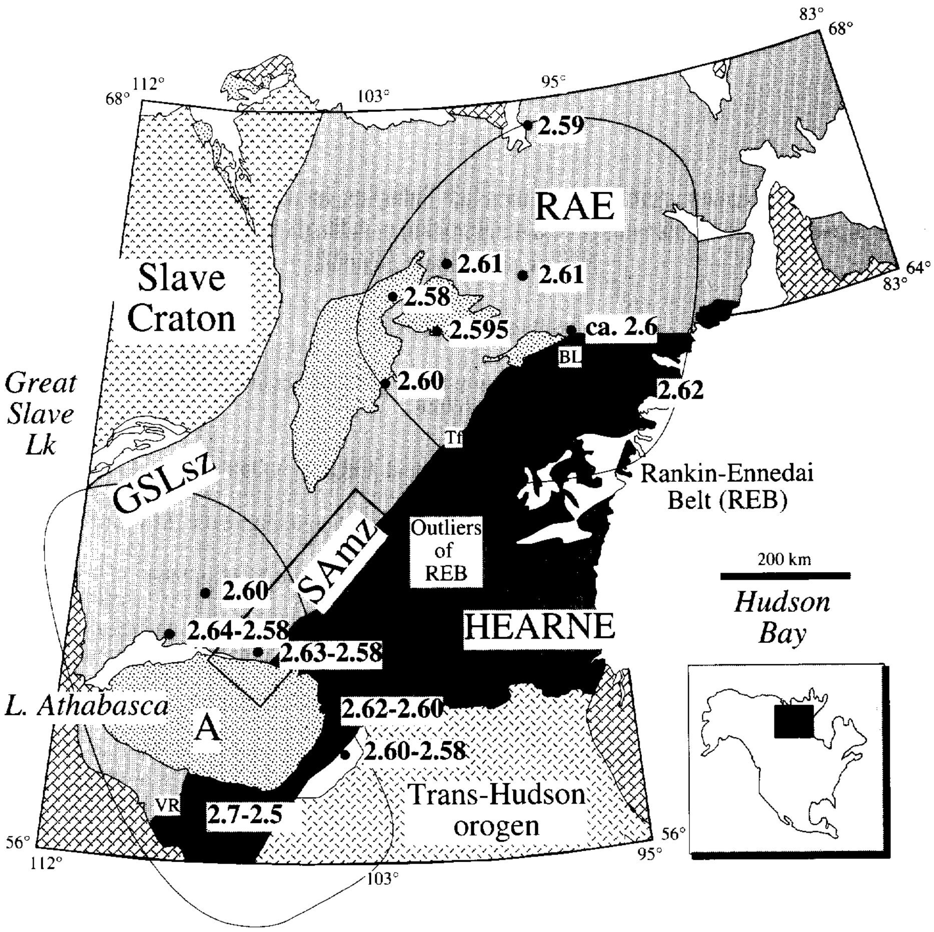

Fig. 1. Location of the intraplate Striding-Athabasca mylonite zone (SAmz) along the contact (Snowbird tectonic zone) between the Rae and Hearne crusts, collectively referred to as the western Churchill continent, the interplate Great Slave Lake shear zone (GSLsz) at the contact between the western Churchill continent and the Slave Craton, and the approximate extent of the Rankin-Ennedai Belt (REB) and its outliers. Numbers are available U–Pb zircon ages of granites in the range 2.63–2.58 Ga, approximately coeval with the Late Archean deformation in Striding-Athabasca mylonite zone. Note the absence of a belt-like distribution. The enclosed areas are a schematic representation of the distribution of available $T_{(DM)}$ Nd model ages (Ga) for the western Churchill continent, with a broader range of model ages (*ca.* 4.0–2.4 Ga) in the southwest enclosed area vs more clustered, younger ages (*ca.* 2.9–2.5 Ga) in the northeast enclosed area, in both Rae and Hearne crusts. See Hanmer *et al.* (in press b) for data sources. Discussed in text. The location of Fig. 2 is shown. The 'brick-wall' shading is Phanerozoic cover. A—Athabasca Basin; BL—Baker Lake; Tf—Tulemalu fault; VR—Virgin River shear zone.

Hanmer & Kopf 1993, Hanmer *et al.* in press b). Until recently, little modern structural work had been undertaken along the Snowbird tectonic zone. Nevertheless, it has been interpreted as the Early Proterozoic suture between two Archean continents [Rae and Hearne (Hoffman 1988), herein referred to collectively as the western Churchill continent]. However, other geological and geophysical interpretations represent the Snowbird tectonic zone as an Early Proterozoic intraplate fault (e.g. Lewry & Sibbald 1977, Gibb 1978, Lewry & Sibbald, 1980, Gibb *et al.* 1983, Lewry *et al.* 1985, Thomas & Gibb 1985, Symons 1991), and geochronological data from our study indicates that it is an Archean structure (Hanmer *et al.* in press a).

We have examined the geology of a *ca.* 400 km long segment of the Snowbird tectonic zone where it crosses the Saskatchewan-NWT border (Fig. 2; Hanmer 1987a, Hanmer *et al.* 1991, 1992b, Hanmer & Kopf 1993, Hanmer 1994, Hanmer *et al.* in press a,b). The geophysical expression of the Snowbird tectonic zone corresponds to two principal geological components: mafic to intermediate composition gneisses which underlie the magnetically defined crustal-scale lozenges, and the granulite facies Striding-Athabasca mylonite zone adjacent to them (Fig. 2). In contrast to the above mentioned Early Proterozoic tectonic scenarios, new geochronological results indicate that the Striding-Athabasca mylonite zone is a multistage, granulite facies, Middle to Late Archean structure in which mylonites were generated both at *ca.* 3.13 Ga, and *ca.* 2.62–2.60 Ga (Fig. 3; Hanmer *et al.* in press a; all dates quoted herein are U–Pb zircon magmatic crystallization ages, R. Parrish, unpublished data, unless otherwise noted). Here, we shall focus principally on the Late Archean granulite facies history of the Striding-Athabasca mylonite zone. The mylonites are considered in two parts; the East Athabasca mylonite triangle and the Striding mylonite belt (Fig. 2).

East Athabasca mylonite triangle

At the east end of Lake Athabasca (Fig. 1), the Snowbird tectonic zone is geologically underlain by the East Athabasca mylonite triangle (Hanmer 1987a,

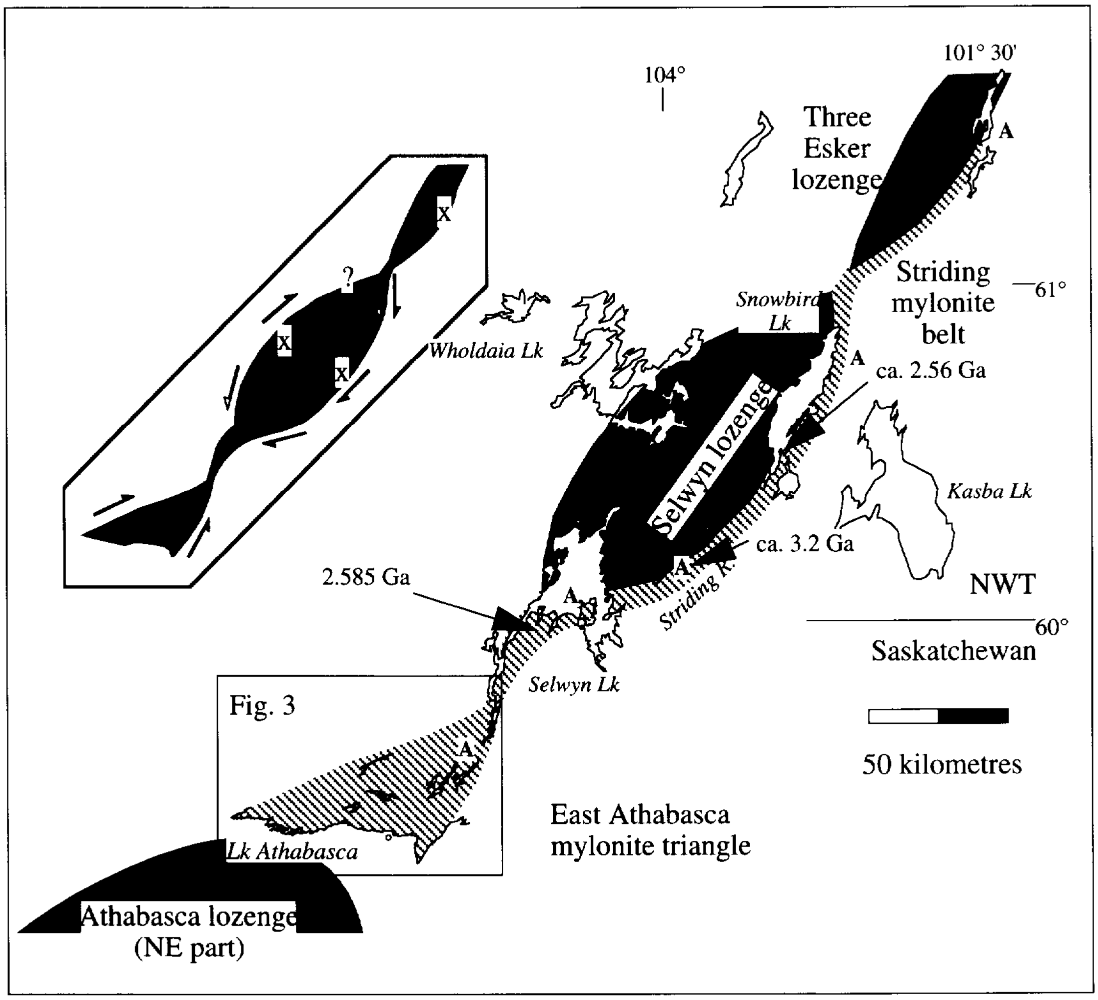

Fig. 2 Location of the East Athabasca mylonite triangle and the sinuous Striding mylonite belt (cross-hatch) with respect to the magnetically defined (Geological Survey of Canada 1987), crustal-scale Three Esker, Selwyn and Athabasca lozenges (grey). The locations of U–Pb zircon magmatic crystallization ages (Ga) are indicated. A = anorthositic mylonites. Inset: schematic representation of observed flow distribution in the East Athabasca mylonite triangle and around the margins of the lozenges (black arrows). The '?' indicates lack of exposure. White arrows indicate the additional shear couples, and 'X' represents gaps in mylonite development, predicted by models which would interpret the nonlinear trace of the mylonite belts in terms of late disruption or distortion of an initially straight shear zone. Discussed in text.

Hanmer *et al.* 1991, 1992b, Hanmer 1994, Hanmer *et al.* in press a), a 75 by 80 by 125 km triangle of penetratively developed, granulite facies mylonites (Table 1; 850–1000°C, 1.0 GPa, Williams unpublished data). This structure is located at the northeastern end of the 300 by 100 km Athabasca lozenge (Fig. 3), itself generally obscured by the overlying *ca.* 1.7 Ga Athabasca Basin (Fig. 1; Cumming *et al.* 1987). Shear-sense indicators are present throughout the mylonites, e.g. rotated winged inclusions, asymmetrical extensional shear bands, asymmetrical pull-aparts, strain insensitive foliations, pressure shadows and oblique boudin trains (see Hanmer & Passchier 1991 and references therein). Many examples of these shear-sense indicators are diagnostic of transpressive general noncoaxial flow (kinematical vorticity number $W_k < 1$; e.g. Means *et al.* 1980, Hanmer & Passchier 1991) at the scale of observation, e.g. in-plane δ-porphyroclasts and asymmetrical extensional shear bands (see detailed discussions in Hanmer 1990, Hanmer & Passchier 1991). The East Athabasca mylonite triangle is structurally divided into an upper and a lower deck (Fig. 3a).

Upper deck. The upper deck is principally composed of granulite facies diatexite and mafic mylonite (Fig. 3; Table 1; Hanmer *et al.* in press a). The diatexite (Mehnert 1971, Brown 1973) predominates in the lower structural level of the upper deck. It is a leucocratic, straight banded rock, with extremely well developed ribbons of quartz and feldspar (0.5 mm by up to 100 mm) which wrap around abundant 2–20 mm garnets (Fig. 4a). The banding is a complex alternation of 0.1–1.0 m thick bands of ultramylonitic garnet granitoid leucosome, mesosome and mafic granulite, with concordant 10–20 cm thick sheets of isotropic to strongly foliated garnet-bearing pegmatite. Compositionally, the pegmatites are identical to the leucosome. Accordingly, the contrast in the degree of their fabric development, combined with their intimate spatial association, suggests that migmatization was syntectonic with respect to the mylonitization. Locally, map-scale lenses of non-mylonitic migmatite preserve agmatitic, raft, ptygmatic and schlieren structures (Fig. 4b; Mehnert 1971, pp. 10–11). The mafic granulite occupies much of the structurally higher parts of the upper deck (Fig. 3b). It is a dark grey

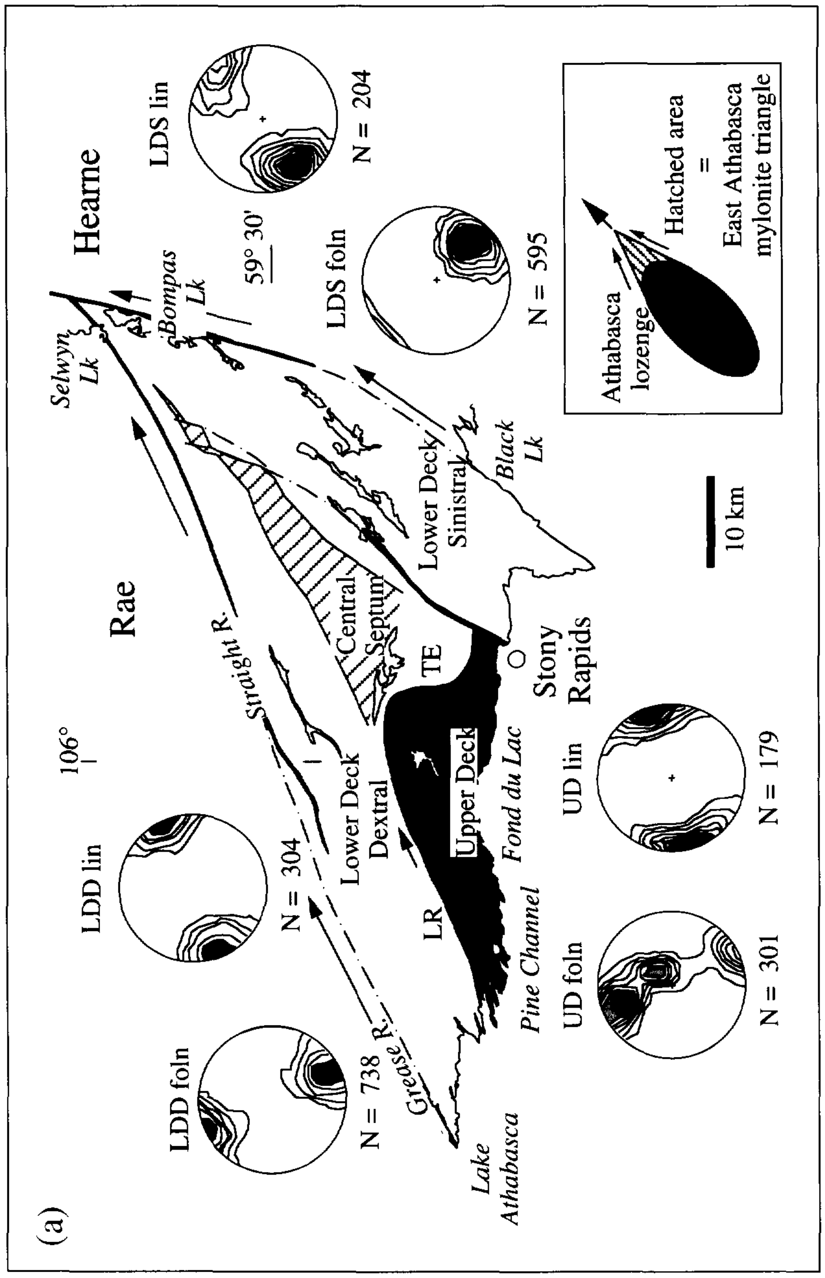
(a)
Rae
Hearne
LDD foln
N = 738
LDD lin
N = 304
106°
UD foln
N = 301
UD lin
N = 179
LDS foln
N = 595
LDS lin
N = 204
59° 30'
Lake Athabasca
Grease R.
Straight R.
Pine Channel
LR
Lower Deck Dextral
Upper Deck
Fond du Lac
TE
Central Septum
Stony Rapids
Lower Deck Sinistral
Black Lk
Selwyn Lk
Bompas Lk
10 km
Athabasca lozenge
Hatched area
=
East Athabasca mylonite triangle

Fig. 3(a).

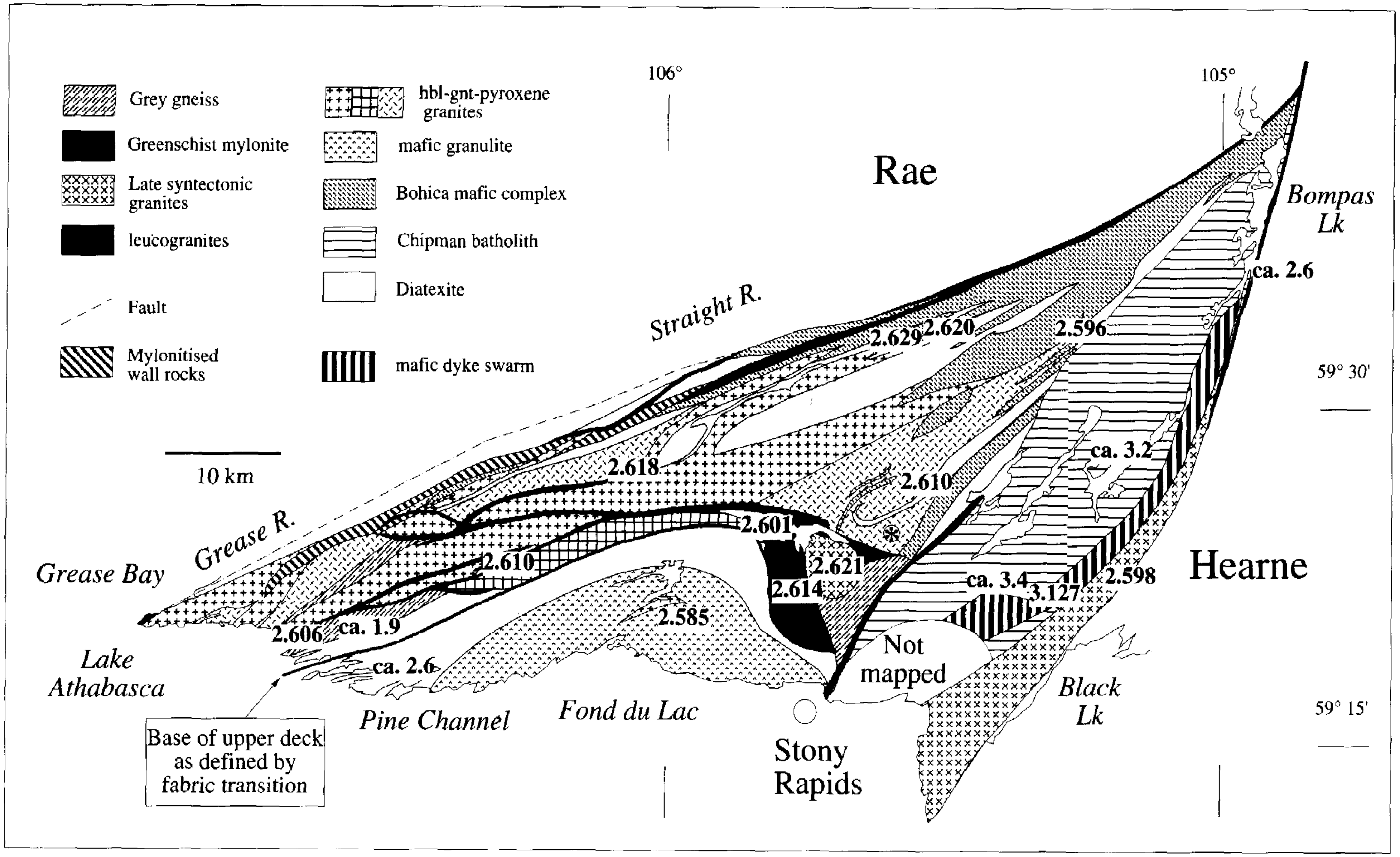

Fig. 3. (a) Principal kinematic elements of the East Athabasca mylonite triangle and relationship of the mylonites to the Athabasca lozenge. Arrows indicate directions of relative tectonic displacement. The lateral ramp (LR) and trailing edge (TE) of the upper deck are indicated. Steeonets of mylonitic foliation and lineation for the dextral (LDD) and sinistral (LDS) kinematic sectors of the lower deck and for the upper deck (UD) are shown. Inset: schematic representation of the location and kinematic setting of the East Athabasca mylonite triangle with respect to the Athabasca lozenge. (b) Geological map of the East Athabasca mylonite triangle. Note that most of the map units are mylonites, except where noted in the text. The approximate locations of U–Pb zircon magmatic crystallization ages (Ga) are indicated. * lies just southeast of a folded panel of diatexite of kinematic significance. Discussed in text. See Fig. 2 for location.

Table 1. Granulite facies mineral assemblages in principal lithologies discussed in text, East Athabasca mylonite triangle

Upper deck:	
Diatexite:	garnite–sillimanite–orthopyroxene
Mafic granulite:	orthopyroxene–clinopyroxene–plagioclase ± garnet
Lower deck:	
Diatexite:	garnet–sillimanite ± orthopyroxene
Bohica mafic complex:	hornblende–orthopyroxene–plagioclase ± garnet and hornblende–clinopyroxene–garnet–plagioclase
Chipman tonalite:	hornblende ± garnet and clinopyroxene–garnet
Metadykes:	hornblende–clinopyroxene–garnet
Syntectonic granulite granitoids:	hornblende–garnet–clinopyroxene–orthopyroxene ± biotite

to brown, fine-grained, sugary textured rock, with 2 by 30 mm polycrystalline streaks of orthopyroxene or plagioclase. Coarse gabbroic to microgabbroic ophitic textures are locally preserved in the mafic granulites, indicative of a plutonic origin for the protolith, dated at *ca.* 2.6 Ga. Magmatic zircons in gabbroic and granitic protoliths, and metamorphic monazite in diatexite mylonite, yield ages of *ca.* 2.62–2.60 Ga, demonstrating that the ribbon mylonites were formed during that time window.

Geometrically, the upper deck (Fig. 3a) is composed of two parts. In the east it is essentially a gently W–SW-dipping homocline. The foliation progressively steepens and changes azimuth westwards to form an upright, SE-dipping panel. In the east, the extension lineations are approximately down-dip, whereas in the west they are subparallel to the strike (Fig. 3a). Layer-parallel, $S > L$ ribbon mylonite fabrics, remarkably homogeneous at all scales, are penetratively developed throughout the upper deck, except for the local lenses of migmatite. Feldspar porphyroclasts are extremely rare. Minor folds, boudins and asymmetrical extensional shear bands are scarce, despite the abundance of potentially stiff layers, such as garnet–quartz and garnet–clinopyroxene bands, or stress raisers, such as large garnet crystals. This implies that the spatial distribution of strain rate was so homogeneous that perturbations in the flow failed to amplify. Consequently visible shear-sense indicators are uncommon. Nevertheless the few that we have found, mostly asymmetrical extensional shear bands and rotated winged inclusions (Fig. 4c), consistently indicate a top-down to the southwest sense of shear, i.e. normal faulting in present coordinates.

As with any shear zone, the lower boundary of the upper deck must be defined by a strain gradient, and/or a discontinuity. In the west, we have mapped the base of the upper deck at a fabric transition from penetratively, but relatively heterogeneously, mylonitized, folded and veined diatexite in the lower deck, into the remarkably homogeneous diatexite mylonite described above (see Fig. 3b). Passing eastward, the basal fabric transition coincides with the lithological contact between the diatexite and subjacent hornblende–garnet–pyroxene granite, then transgresses the lithological contact to lie within the granite. Still further to the east, the fabric transition has been intruded by younger, underlying hornblende–biotite leucogranite, itself (proto)mylonitized at lower amphibolite facies (Fig. 3b). In brief, the upper deck is a penetratively mylonitic segment of a curved, 10 km thick, granulite facies Late Archean shear zone, whose hangingwall has not yet been identified. It can be described as an extensional shear zone with a lateral ramp.

Lower deck. The lower deck is principally composed of three large granulite facies metaplutonic complexes or batholiths, and their mylonitic equivalents (Fig. 3b; Table 1; Hanmer *et al.* in press a). Its preserved structural history is more complex than that of the upper deck, comprising Middle Archean granulite facies mylonitization and extensive Late Archean granulite facies mylonitization, immediately followed by the localization of amphibolite to greenschist facies shearing (Hanmer *et al.* in press a). The plutonic rocks were all originally intrusive with respect to diatexite host rocks, or their protoliths (Fig. 3b), and range from the Middle Archean (*ca.* 3.4–3.2 Ga) Chipman tonalitic batholith, to the Late Archean (*ca.* 2.62–2.60 Ga) Bohica mafic complex and granitic plutons. The diatexite is lithologically and structurally very similar to that of the upper deck. In the east, it is intruded by the tonalitic Chipman batholith (Fig. 3b). The batholith is cut by a syntectonic swarm of mafic dykes and 3.13 Ga granitic veins which were emplaced during the development of now annealed Middle Archean granulite facies mylonites (Fig. 4d) preserved within the eastern part of the tonalite (see below).

In the north and west, the diatexite is intruded by the *ca.* 2.62–2.60 Ga Bohica mafic complex (Fig. 3b). In the main body of the mafic complex, locally preserved well foliated granulite facies meta-norite, gabbro and diorite, with relict igneous textures (Fig. 5a), are enclosed within hornblende-bearing granulite ribbon mylonite (Fig. 5b; Table 1). The western and central parts of the lower deck were intruded *ca.* 2.62–2.60 Ga by a voluminous suite of syntectonic granitic, granodioritic and monzonitic plutons, locally preserved within the mass of their granulite facies ribbon mylonite equivalents (Fig. 5c; Table 1). The later members of this granitoid suite form a linked network of amphibolite facies hornblende–biotite–garnet leucogranite sheets, and their ribbon mylonite equivalents, which cut the high-grade ribbon mylonites (Fig. 3b). Because less thoroughly mylonitized plutons cut earlier, strong mylonitized ones (Fig. 3b), the granitoid suite is syntectonic. This establishes that the ribbon mylonites were formed during the Late

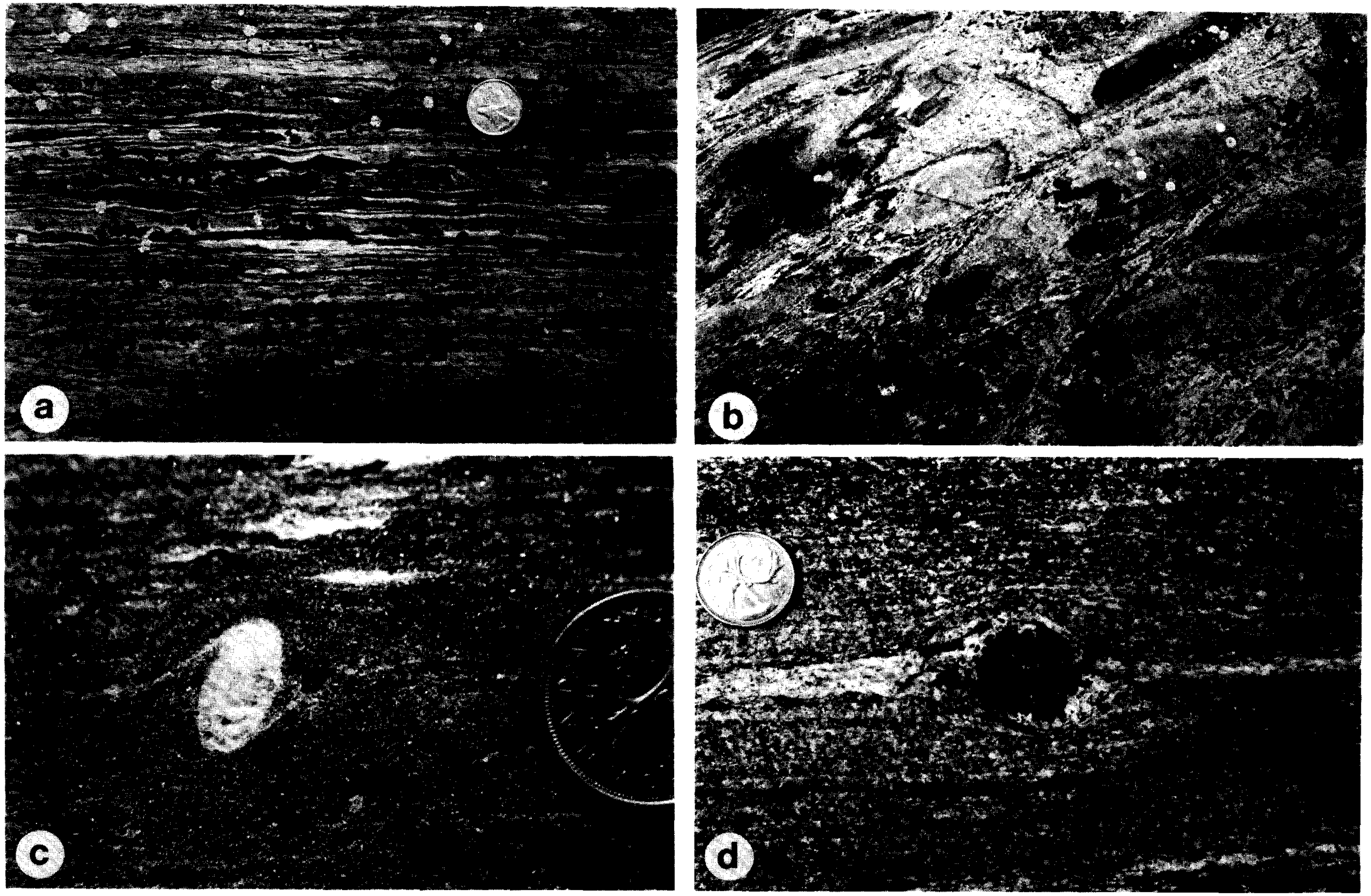

Fig. 4. (a) Diatexite ribbon mylonite equivalent, derived from (b) moderately deformed migmatite of irregular aspect; both are from the upper deck garnet–sillimanite–orthopyroxene diatexite. (c) A rotated winged inclusion (feldspar), in a fine grained annealed mylonite after a coarse granitoid protolith, close to the lower deck/upper deck interface, Lake Athabasca. Dextral shear sense. (d) Large monocrystalline plagioclase porphyroclast with polycrystalline tails in coarsely annealed straight gneiss, Chipman tonalite batholith, eastern lower deck. Note the absence of shape fabric.

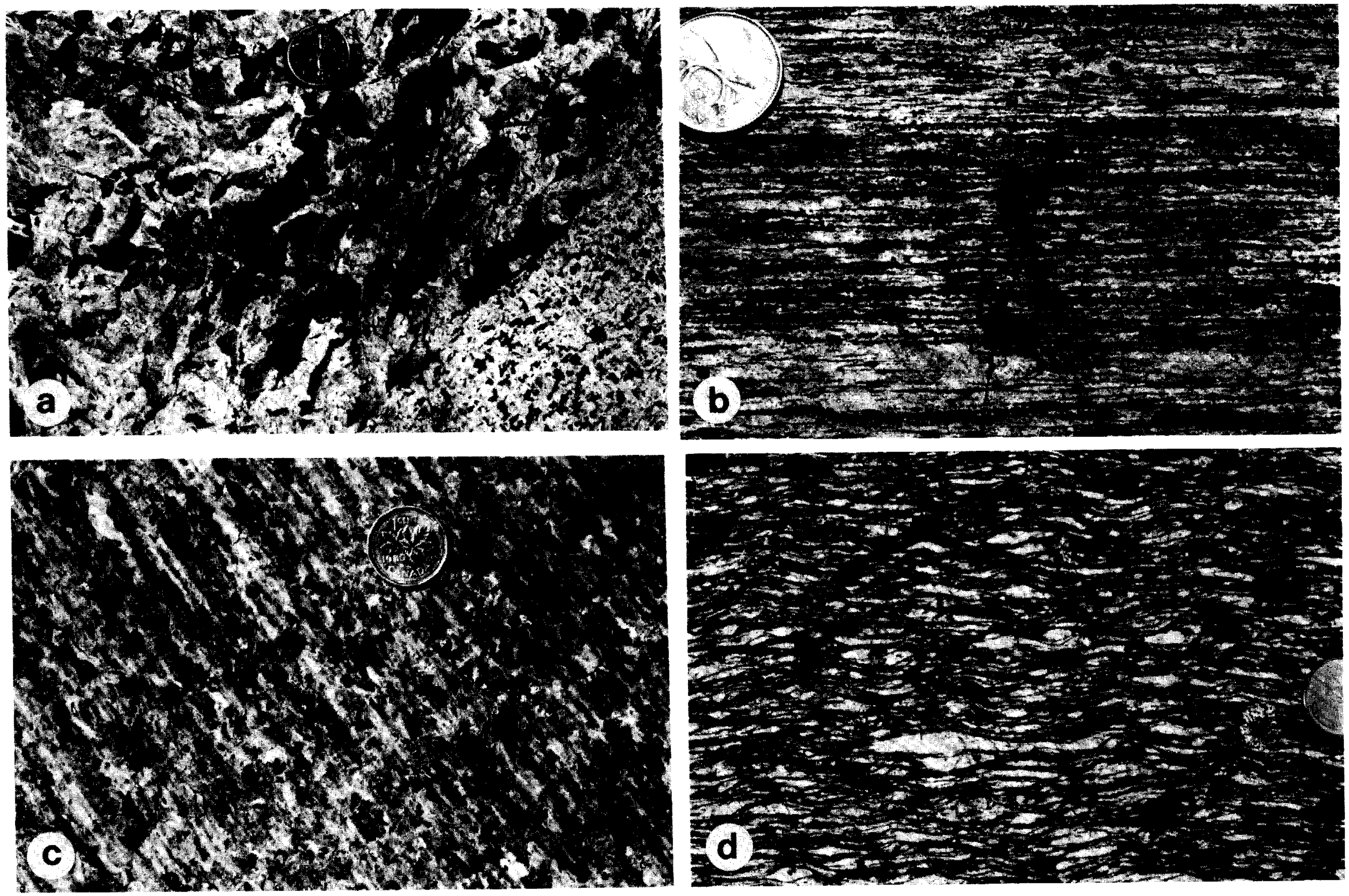

Fig. 5. (a) Coarse igneous textures and (b) orthopyroxene–plagioclase mylonite, Bohica mafic complex, western lower deck. (c) Streaky granite mylonite, studded with garnet porphyroclasts, from the lower deck. The light grey streaks are polycrystalline feldspar set in a homogeneous, grey quartzo-feldspathic matrix. (d) A thoroughly recrystallized tectonite derived by the deformation of a very coarse granitoid protolith in the lower deck. Note the light grey polycrystalline feldspar fields derived from single feldspar parent grains and the absence of preserved porphyroclasts. This fabric is interpreted in terms of a high recrystallization rate/strain rate ratio. See text.

Archean at *ca.* 2.62–2.60 Ga (see Hanmer *et al.* in press a for detailed discussion). The important point to retain here is that the extensive ribbon mylonites are younger and quite distinct from the locally preserved annealed mylonites (*ca.* 3.13 Ga).

The lower deck comprises three kinematic sectors; a central septum which has experienced progressive bulk pure shear, flanked by penetratively mylonitized sinistral and dextral shear zones to the east and west, respectively (Fig. 3a). The widespread, though patchy, preservation of relict igneous textures in the plutonic rocks of the central septum indicates that it accumulated relatively low finite strain.

The dextral shear zone is a *ca.* 15 km wide mylonite belt. Upright, concordant mylonitic foliations and lithological contacts strike *ca.* 060–070° and carry subhorizontal extension lineations (Fig. 3a). The sinistral shear zone is a 20 km wide composite belt of *ca.* 020–050° striking, steeply NW-dipping mylonitic fabrics and extension lineations which plunge to the southwest (Fig. 3a). Within this zone, the fabric elements of the *ca.* 3.13 Ga annealed mylonites, confined to the eastern part of the Chipman batholith (Fig. 3b), are parallel and kinematically identical to those of the *ca.* 2.62–2.60 Ga ribbon mylonites. Nevertheless, it is clear that the latter rework the former (see Hanmer *et al.* in press a). A diverse assemblage of shear-sense indicators occurs throughout the lower deck (see above). At the northern apex of the lower deck, where the dextral and sinistral shear zones meet, exposure is very poor and it is not possible to observe their relative time relations directly. However, within each shear zone, the sense of shear is consistent, regardless of deformation intensity. In other words, there is no evidence of shear-sense reversal during the histories of either shear zone. Taken together, these field observations indicate that the ribbon mylonites of the dextral and sinistral shear zones of the lower deck are approximately coeval and appear to form a conjugate pair.

Central septum. Within the lower deck, the central septum is a wedge-shaped area of nonpenetrative anastomosing mylonite belts, *ca.* 15 km across the base (Fig. 3a). It is principally composed of poorly foliated to isotropic garnet–pyroxene–hornblende metagranite, associated with panels and rafts of moderately deformed Bohica mafic complex. The granite also contains a large curved panel of annealed straight-banded diatexite (* in Fig. 3b), whose hinge line plunges gently to the southwest. The panels of both Bohica mafic complex and diatexite appear to have acted as guides for the flow patterns in their host granites, according to whether they are oriented parallel to the large-scale dextral or sinistral shear zones bounding the central septum. A steeply dipping panel of Bohica mafic complex, striking about 020°, is associated with sinistral strike-slip shearing in the adjacent mylonitic garnet–pyroxene–hornblende granite. The same mylonitic granite, adjacent to *ca.* 070° striking upright panels of Bohica mafic complex, contains dextral shear-sense indicators. In the hinge zone of the folded diatexite panel within the granite, the diatexite and the immediately overlying granite are transformed to ribbon mylonites associated with a top-down to the southwest sense of shear.

At the scale of the central septum as a whole, mylonitic fabrics of the northwest side tend to reflect dextral non-coaxial flow and those of the southeast side tend to be sinistral. However, clearly contradictory shear-sense indicators also occur locally. All of these observations indicate that the flow pattern within the central septum faithfully reflects the orientation distribution of flow and vorticity in both the upper and lower decks of the East Athabasca mylonite triangle. Moreover, if the three dimensional pattern of shear planes reflects a coherent, contemporaneous set of slip systems, as suggested by the geochronological (Fig. 3b) and metamorphic data (Table 1), then bulk flow within the central septum was essentially coaxial, with a generally prolate symmetry. The subhorizontal, northeast–southwest direction of maximum extension would have been associated with two principal directions of shortening, the one steeply plunging and the other subhorizontal and trending-NW–SE. The horizontal principal shortening direction is kinematically compatible with the conjugate shear zones of the lower deck, whereas the steeply plunging one is compatible with the extensional shearing in the upper deck. In other words, the apparent kinematic complexity of the ribbon fabrics in the East Athabasca mylonite triangle can be simply rationalized as the result of convergent flow of material away from the northeastern apex of the stiff Athabasca lozenge during northeast–southwest extension (Fig. 3a inset).

Contact relationships. During the *ca.* 2.62–2.60 Ga time window, flow in the East Athabasca mylonite triangle evolved with cooling from the pervasive high temperature granulite facies regime described here, to a more localized middle to lower amphibolite facies regime (e.g. Hanmer 1988; see detailed discussion in Hanmer *et al.* in press a,b). This is manifested by the focusing of deformation within the leucogranites at the contact between the upper and lower decks (Fig. 3b), as well as amphibolite facies reworking of the granulite 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, and a mylonitic fault within the lower deck trending northeast from Stony rapids (Fig. 3a; see Hanmer *et al.* in press a for details).

Selwyn lozenge and Striding mylonite belt

The interior of the Selwyn lozenge (125 × 50 km), is readily accessible and fairly well exposed, in contrast to the Athabasca lozenge and the Quaternary-covered Three Esker lozenge (Fig. 2; Hanmer & Kopf 1993). It is principally composed of homogeneous to banded amphibolite and hornblendite intruded by a vein network of leucodiorite–tonalite. Other major components include metagabbro, monotonous fine grained, annealed mafic granulite of uncertain origin, and variably deformed hornblende diorite to granodiorite. For the most part, the rocks are not strongly deformed. They preserve cross-cutting relationships, open fold profiles and igneous textures in foliated plutonic rocks. Foliation trends are curvilinear with a significant east–west component, clearly discordant to the magnetically determined lozenge boundaries, and similar to the structural trends in the wallrocks (Taylor 1963, 1970).

The Striding mylonite belt, made of through-going granulite facies (two pyroxene–garnet ± sillimanite) ribbon mylonites, occurs along the southeastern side of the Selwyn lozenge, from Selwyn Lake, via Striding River, to the north end of Snowbird Lake (Fig. 2). The trace of the belt is remarkably curvilinear and closely follows the magnetically defined trace of the Selwyn and Three Esker lozenges. The mylonite belt is 5–10 km thick, except at Selwyn Lake where it bifurcates into two narrow strands (Hanmer & Kopf 1993). It is everywhere associated with an upright mylonitic foliation, parallel to a straight $S \gg L$ compositional layering, a subhorizontal extension lineation and dextral shear-sense indicators (rotated winged inclusions, asymmetrical extensional shear bands; e.g. Hanmer & Passchier 1991). The mylonites are principally derived from granites, anorthosite, mafic rocks and diatexite (Hanmer & Kopf 1993). Syntectonic granites from Selwyn and Snowbird Lakes indicate that the granulite facies mylonitization occurred at *ca.* 2.60 Ga (Hanmer *et al.* in press b).

As discussed above, combined flow along three coeval shear zones in the East Athabasca mylonite triangle is compatible with a bulk finite strain ellipsoid that lies in the prolate symmetry field. In comparison, the relatively simple $S \gg L$ fabrics in the Striding mylonite belt indicate more oblate finite strains (Hanmer & Kopf 1993). However, the apparent simplicity is complicated by the existence of the two near right-angle bends in the map trace of the mylonite zone along the southeastern margin of the Selwyn lozenge (Fig. 2). If these bends are primary, rather than the result of modification of a once straight shear zone, they must have imposed severe constraints on the ability of the mylonites to accommodate significant wallrock displacements. Alternatively, there are three principal deformation scenarios which could lead to the geometrical modification of a hypothetically initially straight Striding mylonite belt: (i) heterogeneous shortening (folding), (ii) heterogeneous extension (boudinage), and (iii) brittle disruption.

Crustal-scale folding of the mylonite belt could be accomplished by flexural slip or by internal tangential longitudinal strain (e.g. Ramsay 1967). The simplest argument against modification of an initially straight mylonite zone by heterogeneous shortening is provided by the colinearity of the long axes of the relatively stiff Athabasca, Selwyn and Three Esker lozenges. If the lozenges had been subjected to bulk shortening along their line of centres, they would have rotated toward the corresponding plane of flattening; clearly, they did not. Moreover, in the flexural slip model, clockwise rotation of the mylonite fabrics from their initial NE–SW trend would have engendered sinistral shear along the rotating foliation planes. Accordingly, the E–W-trending mylonites in the mid-section of Selwyn Lake should show indications of sinistral shear imposed upon earlier dextral shear fabrics. This is not the case. Alternatively, in the internal tangential longitudinal strain model, the buckling of the initially straight mylonite belt would have been accommodated by layer-parallel and layer-normal extensional strains on the outer and inner arcs of the folds, respectively. No such accommodation structures are present (Hanmer & Kopf 1993; see also Taylor 1963, 1970).

Formation of the discrete lozenges by crustal-scale boudinage at a relatively late stage in the mylonitization history would lead to a predictable spatial distribution of shear fabrics and shear-sense (Fig. 2). First, the lozenges should be bounded by continuous mylonite belts, with minimum boudinage-related deformation in their mid-sections (X's in Fig. 2). Although high-grade mylonite belts, up to several hundred metres wide, occur along the northwestern side of the Selwyn lozenge at Wholdaia Lake (Fig. 2; see Hanmer & Kopf 1993), they do not extend more than 20 km along the lozenge boundary. Secondly, indications of sinistral shear-sense should be present along the lozenge boundaries in the southwest and northeast quadrants (white arrows in Fig. 2). However, there are no mylonites in the southwest quadrant and there is no sign of sinistral shear-sense in the northeastern quadrant, i.e. at Snowbird Lake (Fig. 2).

It is possible to produce a strong deflection (bending) of an initially straight deformation belt by the action of superposed brittle faulting (e.g. Cunningham 1993). Such a mechanism would obviate the requirement for sinistral shear-sense indicators and inner arc/outer arc accommodation strains. The faults could either be long and discrete, or short components of a closely spaced array. In the former case, they would clearly dislocate the initially straight mylonite belt. No such offsets are present (Hanmer & Kopf 1993; see also Taylor 1963, 1970). In the latter case, rotations of the order of 90° would lead to severe disruption and dilation of the mylonite belt and its wallrocks. This is not observed.

To summarize, in the absence of brittle dislocation and dilation, ductile models for deforming a once straight Striding mylonite belt either predict mylonites where there are none (southwest quadrant), poor mylonite development where excellent mylonites are present (mid-points on either side of the lozenge), or sinistral shear where the vorticity is consistently dextral (northeast quadrant). Furthermore, the absence of through-going, continuous mylonites along the northwestern margin of the Selwyn lozenge (Fig. 2) demonstrates that the lozenge is older than, and has served to localize, the Striding mylonite belt (Hanmer & Kopf 1993). We would suggest that the same conclusion applies to the Three Esker lozenge. Accordingly, the bends in the Striding mylonite belt are indeed primary and represent the initial structural configuration, dictated by the geometry and distribution of the lozenges.

DISCUSSION

Intraplate or interplate?

In the first geologically-based tectonic model for the Snowbird tectonic zone, it was interpreted as the northwestern intracontinental limit of the penetrative tectonothermal influence of the Early Proterozoic (1.9–1.8 Ga) Trans-Hudson orogeny (Fig. 1; Lewry & Sibbald 1977, 1980, Lewry *et al.* 1985). We have already shown that the new geochronological data does not support the proposed Early Proterozoic age. Three data sets point to a Late Archean intracontinental setting. First, the *ca.* 2.7 Ga Rankin-Ennadai greenstone belt (Chiarenzelli & Macdonald 1986, Mortensen & Thorpe 1987, Tella *et al.* 1992) extends over 600 km as a zone of outliers across the Hearne crust, from Hudson Bay toward the Selwyn lozenge (Fig. 1). The basin within which it was deposited was closed by continental collision prior to 2.65 Ga (Park & Ralser 1992). Secondly, Late Archean granites, *ca.* 2.63–2.58 Ga, broadly contemporaneous with high grade mylonitization in the Striding-Athabasca mylonite belt, are not arranged in the belt-like configuration typical of magmatic arcs associated with plate margins (Fig. 1; e.g. Windley 1984). Thirdly, the same spatial pattern of Archean Nd model age variation, *ca.* 4.0–2.4 Ga in the southwest vs *ca.* 2.9–2.5 Ga in the northeast, is present in both the Rae and Hearne crusts (Fig. 1). Such a distribution pattern might be expected if the Rae and Hearne crusts had been united for much of the Archean, but would require rather specific circumstances (e.g. narrow, short-lived basin, very similar drift and convergence vectors) in order to accommodate a Late Archean suture along the Snowbird tectonic zone. Taken together, these data sets indicate that by *ca.* 2.62–2.60 Ga the site of the Striding-Athabasca mylonite zone was not a suture, but sat well within the interior of the Late Archean western Churchill continent.

Intraplate faults

The rheological structure of the continental crust (and lithosphere) is complex (e.g. Ranalli & Murphy 1987). In different circumstances its interior parts, well removed from plate boundaries, may double in thickness by penetrative deformation (e.g. Dewey & Burke 1973), fold (e.g. Stephenson & Cloetingh 1991), or merely flex (e.g. Beaumont *et al.* 1988). In addition, intracontinental faulting, well removed from lateral plate boundaries, is known to be associated with both strike-slip and dip-slip faults.

Classical examples of *strike-slip intracontinental faults* occur in China and Tibet, north of the Himalayan orogenic belt. Many workers consider them to constitute a fault array related to the stress field induced by the Indian indentor (e.g. Tapponier & Molnar 1976, Molnar *et al.* 1987). Nevertheless, one of the most important members of this array, the Red River fault (e.g. Tapponier *et al.* 1990), apparently owes its exceptional development to its intersection of a lateral free boundary, the continental margin with the China Sea (Tapponier *et al.* 1982). This does not appear to apply to the case of the Striding-Athabasca mylonite zone.

Among the most spectacular examples of *dip-slip intracontinental faults* are contractional structures, such as the Wind River thrust and associated Laramide uplifts in the southwest United States (Brewer *et al.* 1980, Brewer *et al.* 1982), the Kapuskasing uplift in the eastern Canadian Shield (Percival *et al.* 1989), and the Redbank deformed zone, central Australia (Goleby *et al.* 1989, Wright *et al.* 1990). These faults appear to have been kinematically efficient, in the sense that they have accommodated displacements of sufficient magnitude to expose thick crustal cross-sections. However, the dip-slip nature of these intracontinental faults makes them unsuitable as analogues for the Striding-Athabasca mylonite zone.

Fault localization

The Striding-Athabasca mylonite zone does not appear to extend much beyond the strike length we have examined. Granulite facies mylonites, indeed thick belts of mylonite at any metamorphic grade, are absent along the trace of the Snowbird tectonic zone to the south of the Athabasca Basin (e.g. Crocker *et al.* 1993), as well as between Three Esker lozenge and Baker Lake (Fig. 1; Eade 1985, Tella & Eade 1985). Either the lateral extensions of the Striding-Athabasca mylonite zone were cut out by later Early Proterozoic faulting along the Virgin River shear zone (e.g. Crocker *et al.* 1993) and the Tulemalu fault (Eade 1985, Tella & Eade 1985), or the mylonite zone never extended much beyond its present length (*ca.* 500 km). In light of the absence of relics of high-grade mylonites, we do not favour the former hypothesis. Accordingly, we shall now examine the latter possibility.

The Selwyn and Three Esker lozenges have clearly controlled the sinuous trace of the Striding mylonite belt and its oblate fabrics, and the Athabasca lozenge has played a pivotal role in the location and complex kinematic development of the East Athabasca mylonite triangle and its prolate bulk flow (Figs. 2 and 3; Hanmer *et al.* in press a,b). The lozenges can only have exerted this influence as the result of a rheological contrast between themselves and the softer wallrocks, reflecting systematic compositional differences. The Athabasca and Selwyn lozenges are more mafic to intermediate in composition, compared with the granodioritic to granitic composition of the wallrocks (Hanmer *et al.* in press a,b; see also Crocker & Collerson 1988, Crocker *et al.* 1993, Hanmer & Kopf 1993).

Theoretical treatments of plane strain (Vilotte *et al.* 1984) and three-dimensional flow (England & Houseman 1985) have examined the role of lithospheric variation in rheology in the localization of intracontinental strike-slip faults. Both studies sought to model the formation of the Altyn Tagh fault on the southeastern margin of the relatively stiff Tarim Basin, north of the

Himalayas. The study by Vilotte *et al.* (1984) specifically addressed the question of the size of the rheological heterogeneity and determined that even small strength perturbations are significant. We suggest that the Athabasca–Selwyn–Three Esker lozenges, considered as a unit, are a small-scale rheological *analogue* of the Tarim Basin in the cited theoretical studies.

Modest movements, spectacular fabrics

The geometry of the Striding-Athabasca mylonite zone imposed an important constraint on its potential to accommodate significant strike-slip displacements between the Rae and Hearne crusts. The sinuous course of the Striding mylonite belt, and the conjugate character of the lower deck in the East Athabasca mylonite triangle combine to represent an extremely inefficient geometry for transcurrent movements (e.g. Lamouroux *et al.* 1991, Saucier *et al.* 1992). The bends in the fault trace represent a high-amplitude relief which acts to impede easy strike-slip motion along the mylonite zone and prevent the accumulation of significant lateral displacements of the wallrocks. The apparent kinematic inefficiency of the Striding-Athabasca mylonite zone, predicated upon direct observation of the geometry of the structure, raises several important questions: (i) *Why didn't a favourably oriented segment of the mylonite zone propagate to form a straight, through-going shear zone?* and (ii) *How were spectacular ribbon fabrics generated in what is essentially a failed fault?*

We have noted that the deformation fabrics and shear-sense indicators of the Striding-Athabasca mylonite zone as a whole are indicative of shortening across a bulk dextral shear plane. In the Striding mylonite belt, this is principally expressed as oblate finite strains, with some supporting evidence from shear criteria. Most of the shear-sense indicators in the East Athabasca mylonite triangle occur in the lower deck. At the local scale they show that deformation occurred by progressive, general non-coaxial flow, rather than simple shear. However, the coeval operation of the large scale conjugate shear zones, which make up much of the lower deck, combined to accommodate a component of shortening approximately normal to the general trend of the Striding-Athabasca mylonite zone. Together, these observations indicate that, whereas the local fabrics can reflect the immediate influence of the stiff lozenges (e.g. prolate flow at the tip of the Athabasca lozenge), the bulk of the Striding-Athabasca mylonite zone accommodated a transpressive deformation (e.g. Sanderson & Marchini 1984). Displacement on a fault which terminates within continental crust must, by definition, attenuate to zero at the fault tips (Freund 1974). Fault tips propagate in response to the accumulation of increments of displacement on the main part of the fault (e.g. Swanson 1992, Cowie & Scholz 1992a). However, in a strongly transpressive deformation regime, coupled with a kinematically inefficient geometry, the simple shear/pure shear ratio will be low. Accordingly, in a wide shear zone, displacements may be modest and the fault is unlikely to propagate very far from the rheological heterogeneities upon which it initially localized.

The evolution of mylonite fabrics is a reflection of the partitioning of flow within the rock (e.g. Hanmer 1987b). The number of attempts to define 'mylonite' is partly a reflection of the historical ambivalence regarding a genetic designation (e.g. Higgins 1971, Zeck 1974, Bell & Etheridge 1976, White *et al.* 1980, White 1982, Tullis *et al.* 1982, Wise *et al.* 1984, Mawer 1986, Hanmer 1987b). However, it is probably also a reflection of the diversity of mylonitic fabrics and microstructures developed in different lithologies, deformed at different conditions of strain rate, flow regime and metamorphic environment (PTX). Many, though by no means all, descriptions of mylonites include a reference to the presence of monomineralic ribbons, often of quartz or feldspar which, at least partially, define the mylonitic foliation (e.g. Boullier & Bouchez 1978). They are commonly interpreted as the highly attenuated, deformed equivalents of rock-forming minerals of the initially coarse parent material. The most spectacular ribbons are developed in ultramylonites where the scarcity of porphyroclasts allows the ribbons to adopt a homogeneous planar aspect. The development of relatively fine grained ultramylonite from an initially coarse protolith is classically described in terms of a fabric path whereby porphyroclastic phases are progressively refined and incorporated into the microstructural matrix by dynamic recrystallization (e.g. Sibson 1977, White 1982, Wise *et al.* 1984). The magnitude of the finite strain required for the elimination of the porphyroclast population is therefore a partial function of the rate of dynamic recrystallization of the porphyroclastic materials within the aggregate (Hanmer 1987b). However, this perspective is a reflection of the historical bias toward the study of mylonitization in relatively low grade metamorphic environments.

Compared with low-temperature mylonites, few studies have specifically focused upon the microstructural evolution and fabric paths of high-temperature mylonite fabrics (e.g. Etheridge 1975, Bell & Etheridge 1976, Brodie 1981, Brodie & Rutter 1985, 1987, Biermann & van Roermund 1983, Hanmer 1982, 1984, White & Mawer 1986, Hanmer 1987b, White & Mawer 1988). It is clear that the size, or even the existence, of porphyroclasts in any mylonitic rock is a function of the scale of deformation partitioning within the grain-scale aggregate (Hanmer 1987b, Bell & Johnson 1989). Extrinsic factors which influence the deformation partitioning include those governing the strength of minerals, i.e. P, T, P_{H_2O} and strain rate. The persistence of stiff feldspar porphyroclasts in low-temperature mylonites and ultramylonites is well known, but at appropriately high temperatures and slow strain rates, feldspar can become as soft as quartz (Tullis 1983, Tullis & Yund 1980, 1985, Tullis *et al.* 1990). Under appropriate conditions, a fabric path may lead directly from the protolith to an ultramylonite without the intermediate development of a matrix/porphyroclast bimodality (see Figs. 4a, 5b,c & d). In other words, if high *recrystallization rate/*

strain rate ratios pertain throughout the grain-scale aggregate, a straight, planar ribbon fabric can develop at relatively low finite strain magnitudes (see Ramsay & Graham 1970, Fig. 5d; fig. 3a in Hanmer 1987b; fig. 13 in Hanmer *et al.* 1992b). If, in addition, the deformation is the product of a progressive general non-coaxial flow (kinematical vorticity number $W_k < 1$; e.g. Means *et al.* 1980, Hanmer & Passchier 1991), with a significant component of shortening across the flow plane ($W_k \ll 1$), *longitudinal strain* along the flow plane can accumulate very quickly (Pfiffner & Ramsay 1982). Accordingly, the development of a ribbon mylonite fabric could represent even lower magnitudes of *shear strain* than in the case of progressive simple shear. Our field observations indicate that feldspar porphyroclasts are poorly preserved throughout most of the voluminous mylonites of the Striding-Athabasca mylonite zone. Thermobarometric determinations of metamorphic conditions in the East Athabasca mylonite triangle indicate very high temperatures (850–1000°C, Williams unpublished data) and our structural observations indicate significant deviation from ideal simple shear. Therefore, in a transpressive regime, under high-temperature granulite facies metamorphic conditions, a thick mylonite zone such as the Striding-Athabasca example need not represent major wallrock displacements.

Our interpretation of the Striding-Athabasca mylonite zone as a deep-crustal, intraplate, transpressive fault during the Late Archean begs the question of how the coaxial component of the flow was accommodated, given the shallow to moderate plunges of the associated extension lineations. Under conditions of constant volume deformation and significant extension of the flow plane along the transport direction, a subhorizontal principal extension direction would generate an important room problem at the fault terminations. Moreover, a strongly transpressive, volume-constant, progressive deformation would tend to generate steeply plunging extension lineations (Sanderson & Marchini 1984). However, in the Striding-Athabasca mylonite zone, melts were moving through the presently exposed structural level. Crustal (granites) and probable mantle melts (gabbro/norites) were emplaced from below, but potentially voluminous granitic melts from the diatexites would have migrated upward to higher structural levels. It is unlikely that we will ever be able to evaluate the bulk volume change in the Striding-Athabasca mylonite zone. However, we suggest that significant volume loss by magma migration may be a fundamental process in deep-crustal, high-temperature shear zones.

CONCLUSIONS

Geometry, strain symmetry and strain partitioning within lower-crustal intraplate shear zones can be extremely complex, and may be significantly influenced by local, crustal-scale rheological variation. Striding-Athabasca mylonite zone is a Late Archean, granulite facies, intraplate, transpressive, dextral shear zone formed in the lower continental crust. Its highly sinuous trace and locally conjugate configuration rendered it kinematically inefficient as a strike-slip fault. It cannot have accommodated large displacements. Movement on the incipient shear zone was geometrically inhibited by the high amplitude relief on the fault plane, and was kinematically discouraged by the low simple shear/pure shear strain rate ratio of the general noncoaxial (transpressive) flow regime. Spectacular ribbon mylonites developed throughout the shear zone are not diagnostic of significant tectonic displacements of the bounding wallrocks. Rather, they reflect the very high temperatures, high recrystallization rate/strain rate ratios and the transpressive nature of the deformation, potentially coupled with significant volume loss by magma migration. We suggest that the ribbon fabrics of the Striding-Athabasca mylonite zone were generated in the western Churchill continental interior in response to Late Archean far-field plate convergence, continental collision or post-collisional shortening at the, as yet indeterminate, plate margins. Rather than a transcontinental fault, the mylonite zone represents a crustal-scale adjustment structure, i.e. an incipient shear zone.

Acknowledgements—This paper was presented orally at the International Conference on *Structures and Tectonics at Different Lithospheric Levels*, held at Graz University, Austria, 6–8 September 1993. We thank Dick Brown, Shaocheng Ji, Steve Lucas, Sally McEachern, Randy Parrish, Mary Sanborn-Barrie, Subhas Tella and Cees van Staal for stimulating discussion and critical reading of an early version of this manuscript, and Mark Darrach and Shaocheng Ji for their participation in the fieldwork. We also thank Randy Parrish for permission to quote the unpublished geochronological data from the Geological Survey of Canada Geochronology Laboratory. Journal reviews by Tim Bell and Art Snoke were much appreciated. MW acknowledges receipt of NSF grant EAR-9106001 and CK was funded by NSF grant EAR-9105284 to K. Collerson. This is Geological Survey of Canada Contribution 27993.

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