Striding – Athabasca mylonite zone: implications for the Archean and Early Proterozoic tectonics of the western Canadian Shield[1]

Simon Hanmer, Michael Williams, and Chris Kopf

Abstract: Study of the northern Saskatchewan – District of Mackenzie segment of the Snowbird tectonic zone suggests that fragments of relatively stiff mid-Archean crust, possibly arc related, have controlled the localization, shape, and complex kinematics of the multistage Striding–Athabasca mylonite zone during the Archean, as well as the geometry of the Early Proterozoic rifted margin of the western Churchill continent. By the late Archean, the Striding–Athabasca mylonite zone was located in the interior of the western Churchill continent, well removed from the contemporaneous plate margins. Except for the Alberta segment, the Snowbird tectonic zone was not the site of an Early Proterozoic plate margin. We suggest that the geometry of the Archean – Early Proterozoic boundary in the western Canadian Shield represents a jagged continental margin, composed of a pair of reentrants defined by rifted and transform segments. These segments were inherited from Early Proterozoic breakup and controlled by the Archean structure of the interior of the western Churchill continent. The geometry of this margin appears to have strongly influenced the Early Proterozoic tectono-magmatic evolution of the western Canadian Shield.

Résumé : D'après l'étude du segment Saskatchewan–Mackenzie de la zone tectonique de Snowbird, des fragments de croûte, peut-être d'arc magmatique, auraient controlé la localisation, la forme et la cinématique de la zone mylonitique complexe de Striding–Athabasca au cours de l'Archéen, ainsi que la géométrie distentionelle d'âge Protérozoïque inférieur de la marge du continent Churchill occidental. Lors de l'Archéen supérieur, la zone mylonitique de Striding–Athabasca se trouvait à l'intérieur du continent Churchill occidental, bien éloignée des marges tectoniques actives. À l'exception du segment Albertain, la zone tectonique de Snowbird n'était pas le site d'une marge tectonique lors du Protérozoïque inférieure. Nous proposons que la géométrie de la limite Archéen – Protérozoïque inférieure dans le Bouclier canadien occidental comprenait une paire de « reentrants », définis par des tronçons rifts et transformes, hérités de l'extension continentale pendant le Protérozoïque inférieur et controlés par la structure Archéene interne du continent Churchill occidental. La géométrie de cette marge aurait étroitement controlé l'évolution tectonomagmatique du bord du Bouclier canadien occidentale.

Introduction

The Snowbird tectonic zone (Hoffman 1988) is a northeast–southwest-trending linear geophysical anomaly, which can be traced in the horizontal gravity gradient map of Canada for up to 3000 km, from the Canadian Rocky Mountains to, possibly, northern Quebec (Goodacre et al. 1987; M.D. Thomas et al. 1988; see Fig. 2 in Hanmer et al. 1994). The geophysical features that mark the Snowbird tectonic zone (see Fig. 2 in Hanmer et al. 1994), historically referred to as the Fond du Lac low, the Kasba Lake – Edmonton low, or the Athabasca axis, have long been familiar to regional geophysicists (e.g., Walcott and Boyd 1971; Walcott 1968; Wallis 1970; Gibb and Halliday 1974; Teskey and Hood 1991) and geochemists (e.g., Burwash and Culbert 1976; Darnley 1981). Apparent truncations within the regional magnetic pattern (Hoffman 1988), or paired anomalies in the regional gravity field (M.D. Thomas and Gibb 1985), have been used to suggest that the Snowbird tectonic zone marks an Early Proterozoic collisional suture between two Archean continental platforms: the Rae and Hearne provinces (Fig. 1) (Hoffman 1988).

The first *geologically based* tectonic model for the Snowbird tectonic zone was proposed by workers in central Saskatchewan, who interpreted it as the northwestern intracontinental limit of the penetrative tectono-thermal influence of the Trans-Hudson orogeny (e.g., Lewry and Sibbald

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S. Hanmer Continental Geoscience Division, Geological Survey of Canada, 601 Booth Street, Ottawa, ON K1A 0E8, Canada.
M. Williams and **C. Kopf** Department of Geology and Geography, University of Massachussetts, Amherst, MA 01003, U.S.A.

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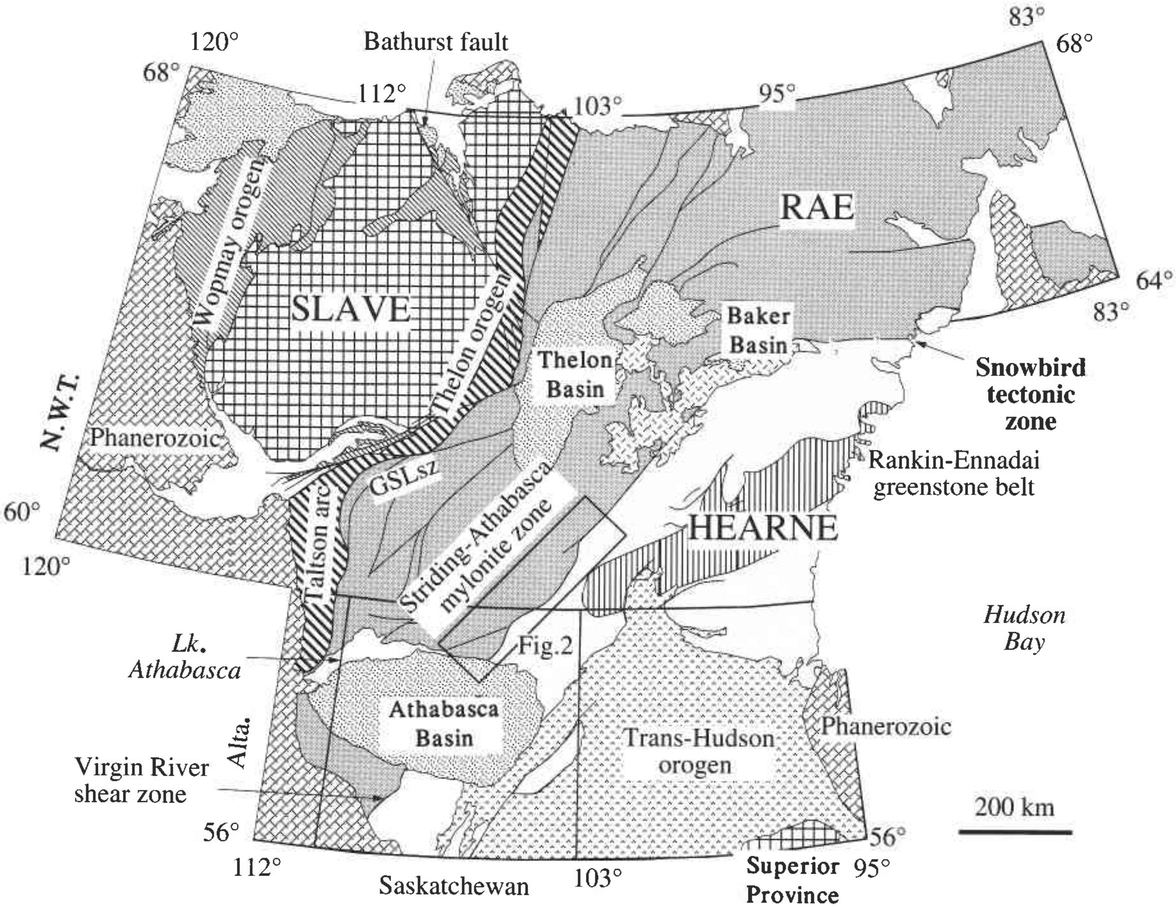
Fig. 1. Generalized geology of the western Canadian Shield. The outline of the Rankin–Ennadai greenstone belt outliers has been grossly simplified. Modified from Hoffman (1988). Discussed in text. GSLsz, Great Slake Lake shear zone.

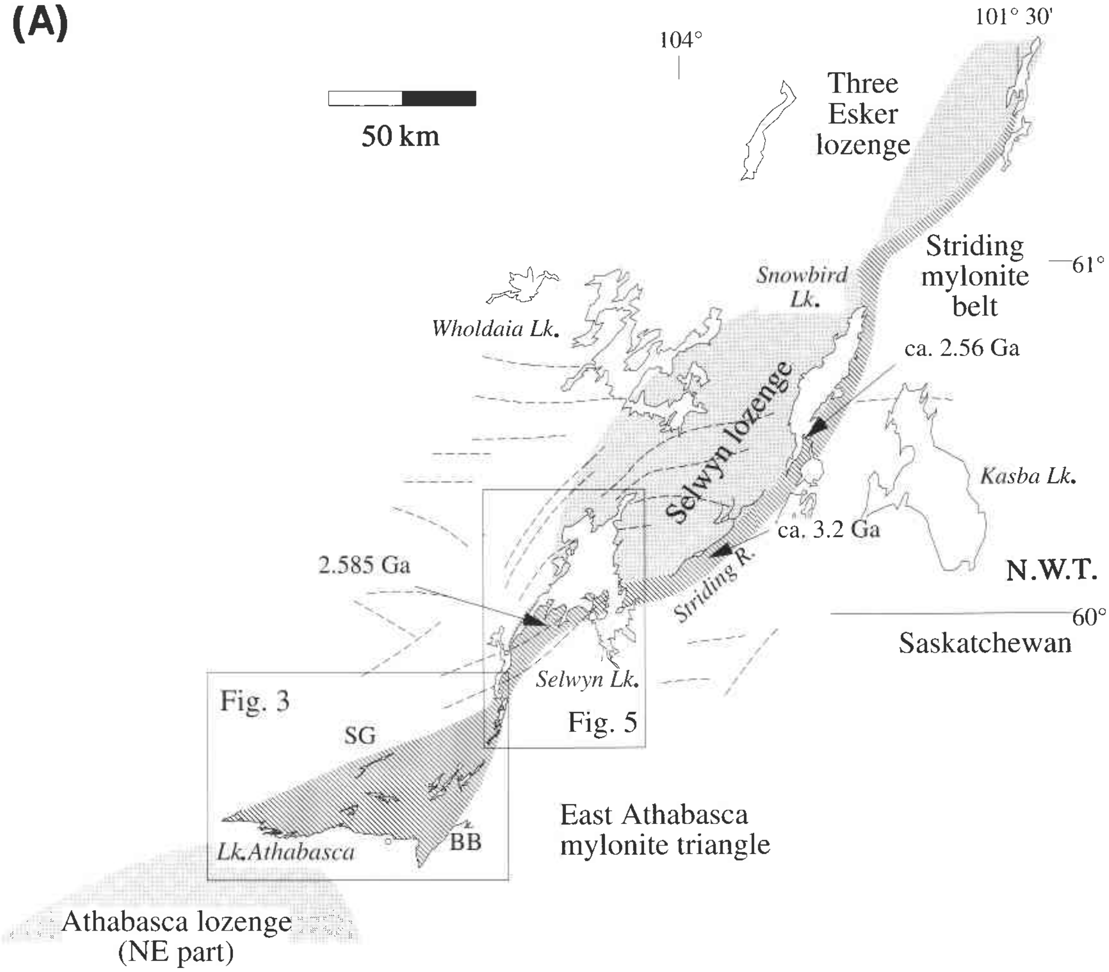


1977, 1980; Lewry et al. 1985). According to this hypothesis, the Snowbird tectonic zone is a shear zone whose location was determined by lateral rheological variation, due to the existence of a thermal dome within the Trans-Hudson hinterland. The soft thermal dome was equated with the Cree Lake (Hearne) zone, flanked to the northwest by the stiff rocks of the Amer Lake (Rae) "cratonic" zone (Lewry et al. 1985), also referred to as the Western Granulite Domain (Lewry and Sibbald 1977; Gibb 1983).

Structural and geochronological examination of the exposed segments of the Snowbird tectonic zone should provide a critical test of these various hypotheses. We have recently completed such a study of a segment of the Snowbird tectonic zone in northern Saskatchewan and southeastern District of Mackenzie, N.W.T. (Fig. 2A) (Hanmer 1987, 1994; Hanmer et al. 1991, 1992*a*; Hanmer and Kopf 1993). Although we shall question the applicability of an Early Proterozoic suturing model, the terms Rae and Hearne are now in general usage and will be retained. The Rae and Hearne crusts will be collectively referred to as the western Churchill continent. A train of three-100-km scale elliptical features occurs in the magnetic field along the trend of the gravity anomaly in the vicinity of the territorial border (Figs. 2B, 2C) (Geological Survey of Canada 1987; Pilkington 1989). We refer to them as the Athabasca, Selwyn, and Three Esker "lozenges" (Fig. 2A) (Hanmer and Kopf 1993; Hanmer et al. 1994). A linked system of granulite facies mylonite belts traces a sinuous course along the train of lozenges: the *Striding–Athabasca mylonite zone*. The mylonite zone consists of two parts: the *East Athabasca mylonite triangle*, at the northeastern apex of the Athabasca lozenge, and its northeastern extension, the *Striding mylonite belt* (Fig. 2A). We have systematically mapped the East Athabasca mylonite zone (Hanmer 1994; Hanmer et al. 1994). A full summer field season was spent in the Striding mylonite belt, the subject of this contribution, undertaking detailed structural mapping of the most accessible and best exposed parts of the mylonite belt itself, combined with short fly camps and air-supported reconnaissance within the Selwyn lozenge and the Rae and Hearne wall rocks.

This paper has three principal goals. The first is to describe structural and geochronological observations of the Striding mylonite belt and to link them to the East Athabasca mylonite triangle, documented in detail in our companion paper (Hanmer et al. 1994). One of our key conclusions is that granulite facies mylonitization in the Striding–Athabasca mylonite zone occurred at ca. 3.2 and ca. 2.6 Ga (all ages cited are U–Pb on zircon, unless otherwise specified). The second aim is to examine the significance of the distinctive lozenge geometry defined by the geophysical and map patterns,

Fig. 2. (A) Striding–Athabasca mylonite zone. Generalized representation of the magnetically defined lozenges (grey shading, Geological Survey of Canada 1987; see also B); the bounding East Athabasca mylonite triangle and Striding mylonite belt (diagonal lines) are illustrated. U–Pb zircon magmatic crystallization ages (Ga) are shown (R. Parrish, unpublished data). Thin lines are foliation trajectories (after Taylor 1963, 1970). Locations of Figs. 3 and 5 indicated. Discussed in text. BB, Black–Bompas fault; SG, Straight–Grease fault. (B) Colour shaded relief map of total magnetic field, National Areomagnetic Data Base (Geophysical Data Centre, Geological Survey of Canada). Sun illumination inclination is 60° above the horizon at a declination of 135°N. Scale 1 : 5 000 000. (C) Principal geographical coordinates, locations, and magnetic lineaments discussed in text, keyed to B. (Fig. 2 concluded on facing page.)



and the possible causal relationship between this geometry and the mylonites, within the context of the Archean history of the western Churchill continent. We consider that the principal deformation occurred in a late Archean intracontinental setting and was localized against a major strength heterogeneity represented by the train of lozenge shapes. Finally, despite an extensive geochronological programme (see also Hanmer et al. 1994), we have not identified significant Early Proterozoic ductile deformation in the Striding–Athabasca segment of the Snowbird tectonic zone. Accordingly, our third aim here is to propose a hypothesis that attempts to reconcile the apparent paradox of an Early Proterozoic history along the Alberta segment of the Snowbird tectonic zone (e.g., Hoffman 1988; Ross et al. 1991) with a mid- to late Archean history along the Striding–Athabasca segment. We suggest that features that have been attributed to Early Proterozoic suturing within the *interior* of the western Churchill continent actually formed in response to subduction beneath the *western margin* of the western Churchill continent, and that the location of these features was controlled by preexisting mid- to late Archean structures of the Snowbird tectonic zone.

Snowbird tectonic zone

East Athabasca mylonite triangle

In our recent work in northern Saskatchewan, we have identified and defined the East Athabasca mylonite triangle, at the east end of Lake Athabasca (Fig. 3) (Hanmer 1987, 1994; Hanmer et al. 1991, 1992*a*). Only a brief review is given here; the reader is referred to Hanmer et al. (1994) for detailed documentation. The East Athabasca mylonite triangle can be divided into an upper and a lower deck, based on the orientation of the dominant structural fabrics (Fig. 3). The lower deck comprises three kinematic sectors: two conjugate, penetratively mylonitic, granulite facies strike-slip

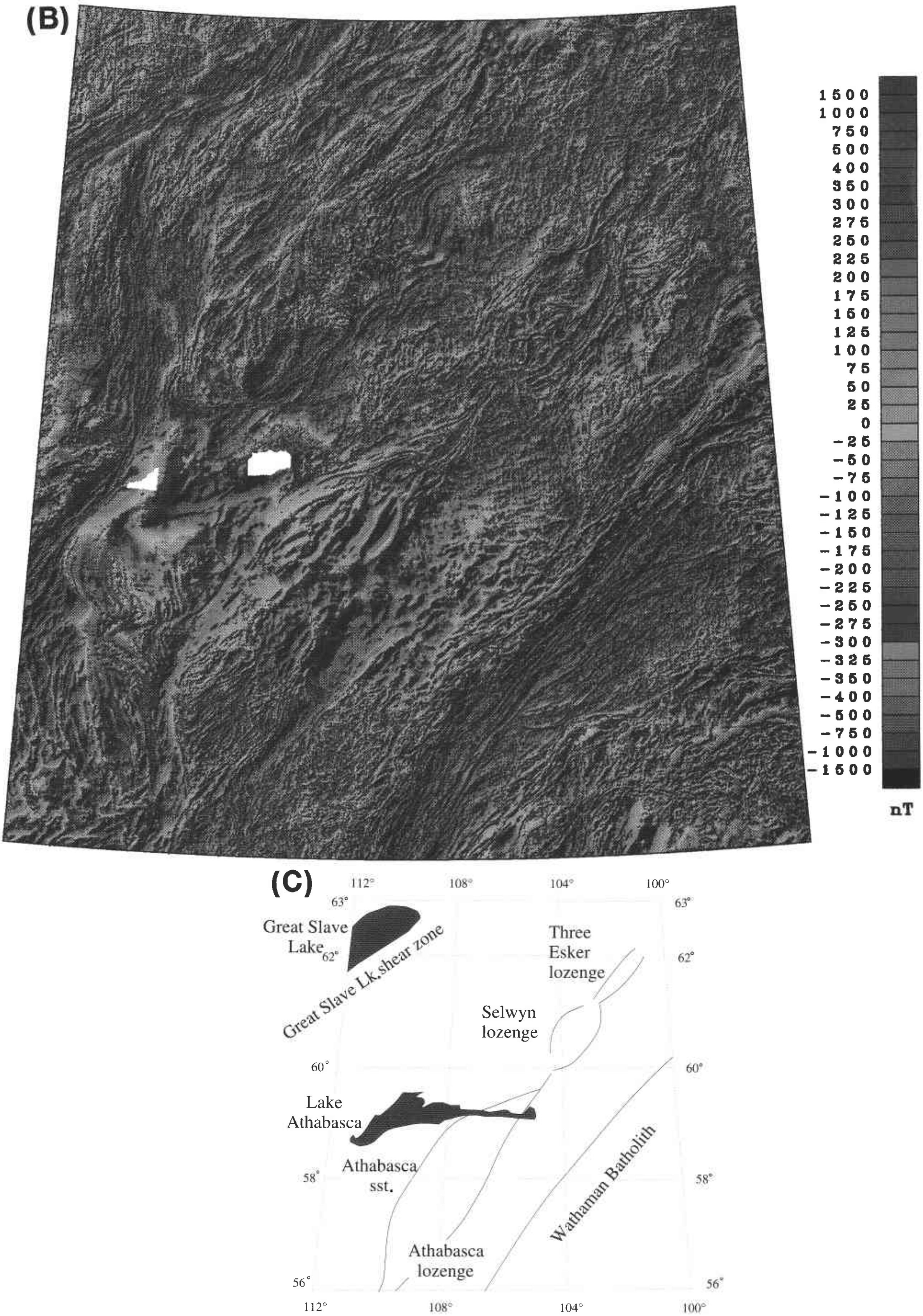
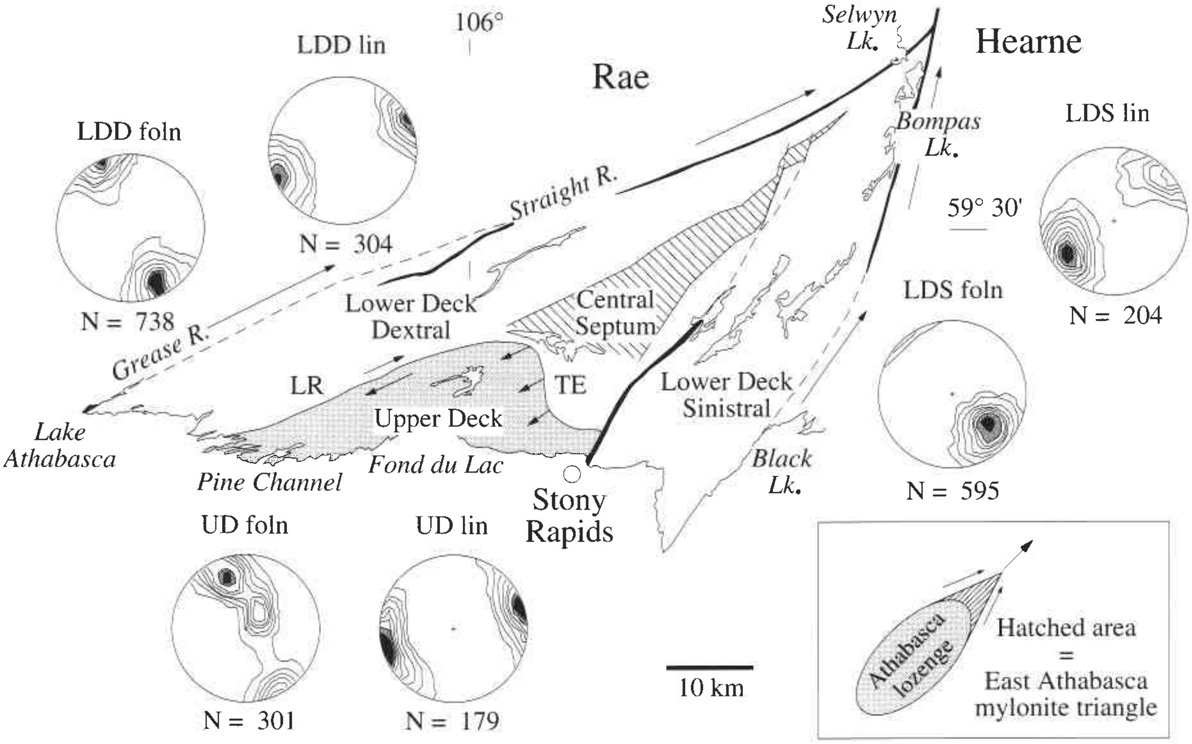

(B)
1500
1000
750
500
400
350
300
275
250
225
200
175
150
125
100
75
50
25
0
-25
-50
-75
-100
-125
-150
-175
-200
-225
-250
-275
-300
-325
-350
-400
-500
-750
-1000
-1500
nT
(C)
112°
108°
104°
100°
63°
62°
60°
58°
56°
Great Slave Lake
Great Slave Lk. shear zone
Three Esker lozenge
Selwyn lozenge
Lake Athabasca
Athabasca sst.
Athabasca lozenge
Wathaman Batholith

Fig. 3. Principal tectonic elements within the East Athabasca mylonite triangle (see Fig. 2 for location). Arrows indicate directions of relative tectonic displacement. The lateral ramp (LR) and trailing edge (TE) of the upper deck are indicated. Stereonets are of poles to foliation and extension lineations for the upper deck (UD), lower deck dextral (LDD), and lower deck sinistral (LDS) kinematic sectors. Inset is a schematic representation of the location of the East Athabasca mylonite triangle at the northeastern apex of the crustal-scale Athabasca lozenge and the general pattern of flow resolved adjacent to the apex (see Hanmer et al. 1994 for details). foln, foliation; ln, lineation.



shear zones, each ca. 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. It is now entirely occupied by a penetratively mylonitic, 10 km thick dip-slip granulite facies shear zone. The upper deck can be described as an extensional shear zone with a lateral ramp, at least with regard to its later granulite and subgranulite facies development. Although the hanging wall to the upper deck shear zone is not exposed, the shape, orientation, and location of the shear zone strongly suggest that it constitutes the sole to the northeastern end of the bulk of the Athabasca lozenge itself (Fig. 2A).

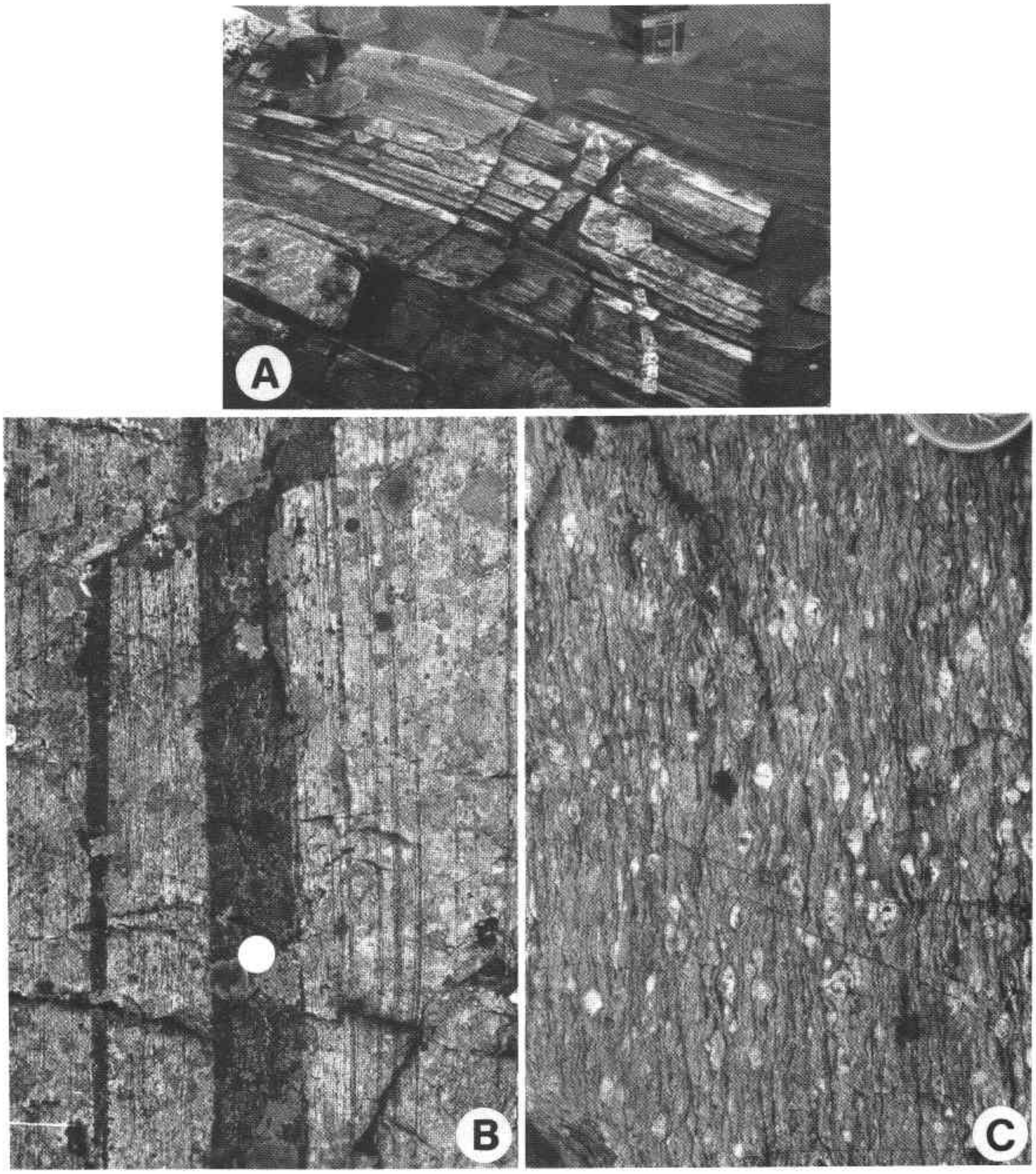
Two granulite facies penetrative mylonitization events are recorded. The first (about 850C, 1.0 GPa) occurred at ca. 3.13 Ga, and is represented by annealed relict fabrics preserved in the eastern part of the triangle. The second (850–1000°C, 1.0–1.5+ GPa), associated with widespread, penetrative ribbon fabrics, occurred ca. 2.62–2.60 Ga, and was accompanied by voluminous mafic and granitic plutonism. During the late Archean event, cooling through amphibolite to greenschist facies led to strain localization and a narrowing of the active shear zones. Early Proterozoic activity within the East Athabasca mylonite triangle is limited to pluton emplacement within the lower deck, and folding of granitic veins during minor reactivation of greenschist mylonites at ca. 1.8 Ga (Hanmer et al. 1994).

Striding mylonite belt

The Striding mylonite belt is a welt of throughgoing granulite to upper amphibolite (2 pyroxene – garnet ± sillimanite) ribbon mylonites, 5–10 km thick, along the southeastern side of the 125 × 50 km Selwyn lozenge, from Selwyn Lake, via Striding River, to the north end of Snowbird Lake (Fig. 2A) (Hanmer and Kopf 1993). It is associated with an upright S ≫ L mylonitic foliation, a subhorizontal extension lineation, and dextral shear-sense indicators (rotated winged inclusions and asymmetrical extensional shears, e.g., Hanmer and Passchier 1991). The fabric symmetry indicates a transpressive regime, an interpretation supported by the occurrence of extensional shear bands and in-plane δ porphyroclasts (Hanmer and Passchier 1991). The mylonites are principally derived from granitoids, anorthosite, mafic rocks, and diatexite (Fig. 4). The trace of the mylonite belt is remarkably sinuous, describing two nearly right-angle bends as it closely follows the southeastern margin of the Selwyn lozenge (Fig. 2A). A large isolated outcrop of these mylonites at Three Esker Lake, in combination with the regional geological trends (Wright 1967), suggests that the highly sinuous character of the trace of the mylonite belt persists to, at least, 61°30′N (Fig. 2A).

At three locations distributed along the length of the Striding mylonite belt, U–Pb analysis of magmatic zircon in syn-

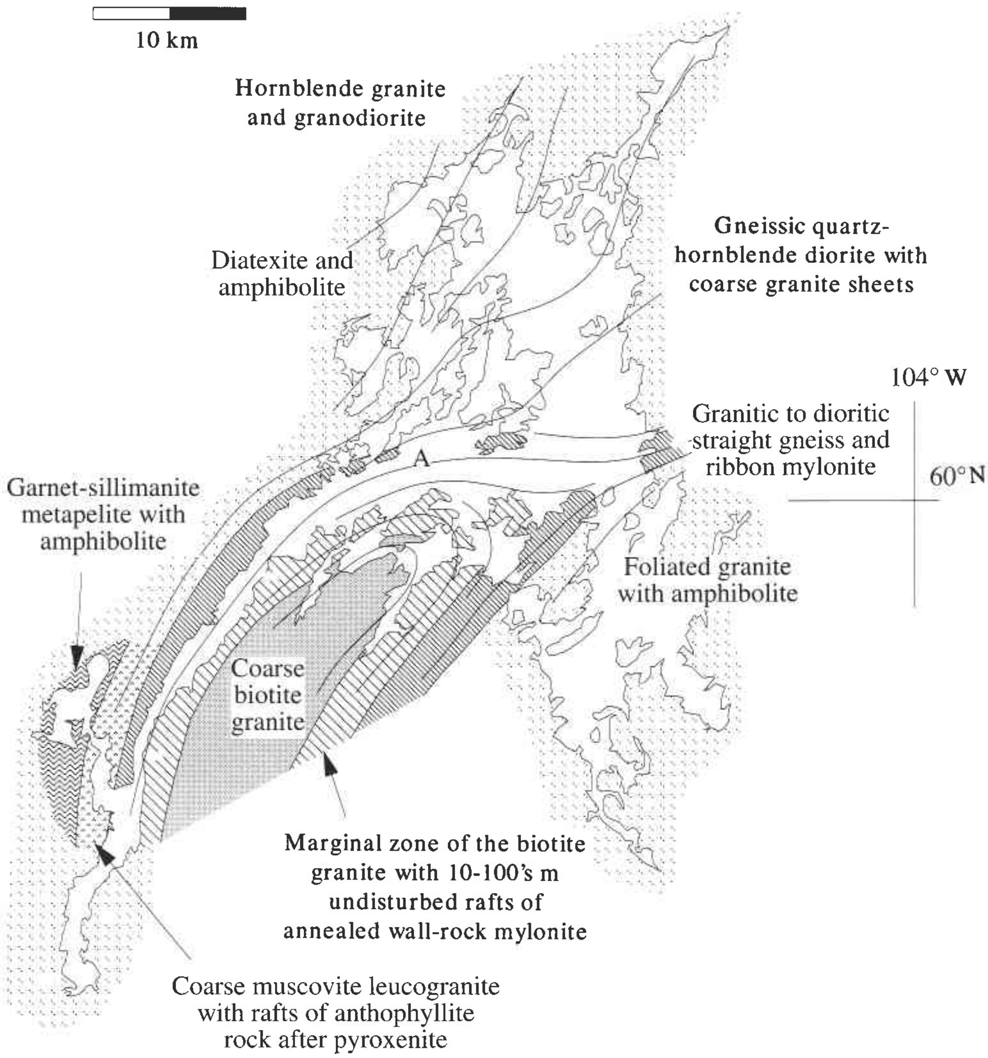
Fig. 4. Granulite facies ribbon mylonites of the Striding mylonite belt. (A) Slabby ribbon ultramylonite in garnet diatexite, central Selwyn Lake. (B) Garnet anorthosite and garnet hornblendite (ex-pyroxenite) ribbon ultramylonite, east of Snowbird Lake. (C) Granitic ribbon mylonite, east Snowbird Lake. All are plan views. Mylonite fabrics strike east–west in A, and approximately north–south in B and C.



tectonic granites indicates that the mylonites are Archean in age (R. Parrish, unpublished data; Fig. 2A). First, at the southern end of Snowbird Lake (Fig. 2A), ribbon mylonites of uncertain, perhaps diatexitic, protolith are cut by coarse-grained white leucogranite. An earlier phase of leucogranite is a ribbon mylonite cut by a later, more porphyroclast-rich mylonitic phase of similar granite. We interpret the leucogranites to be related, and thereby contemporaneous with mylonitization. Accordingly, the ca. 2558 ± 25 Ma age obtained from the porphyroclastic granite mylonite suggests that the mylonitization is late Archean in age.

The Striding mylonite belt bifurcates in the southern half of Selwyn Lake (Fig. 5). To the south, the two strands coalesce just north of their confluence with the apex of the East Athabasca mylonite triangle. In the northwestern strand, north-trending mylonites run up the south arm of Selwyn Lake and swing progressively to strike east–west across the centre of the lake, while the southeastern strand strikes about 045°. Most of the mylonite is present as annealed, slabby quartz leucodioritic-amphibolite straight gneiss (Hanmer 1988*a*), locally with excellent preservation of feldspar porphyroclasts and quartz or feldspar ribbons. However, the

Fig. 5. Geological sketch of bifurcation and 90° bend in mylonites of the Striding mylonite belt at Selwyn Lake. Thin lines are foliation trajectories. See Fig. 2 for location. Discussed in text.



east–west-striking mylonites in the central part of the lake are principally ribbon ultramylonites, derived from garnet–sillimanite diatexite and garnet anorthosite protoliths. The annealed mylonites are cut by an array of 1–10 m thick sheets of generally concordant pink biotite granite in various states of deformation, even within a given outcrop. The most strongly deformed granite sheets are garnetiferous ribbon ultramylonites, which are themselves locally cut by less deformed sheets of similar granite. The mylonites in both strands dip moderately to the west and north, carry a subhorizontal extension lineation, and were formed by dextral transcurrent shear.

The two strands of mylonite are separated by a body of coarse-grained, isotropic biotite granite (Fig. 5). The marginal zone between the granite and the mylonites is composed of the isotropic to poorly foliated granite with large (hundreds of metres long by tens of metres thick) rafts of annealed leucodioritic-amphibolite straight gneiss, somewhat coarser grained than the adjacent material outside of the granite, but still recognizable as the same rocks. The rafts are oriented parallel to the external mylonites, except at the northeastern termination of the granite, where they have been reoriented into parallelism with the pluton contact. The location of the granite just south of a right-handed, near-right-angle bend in the dextral mylonites is surely no coincidence. Such a position would correspond to a releasing bend, or dilational jog (e.g., Sibson 1986*a*, 1986*b*), an excellent site for magma emplacement (e.g., Guinberteau et al. 1987; Hutton 1988; Hutton et al. 1990; Morand 1992). Therefore, we suggest that emplacement of the granite was late syntectonic with respect to the high-grade mylonitization. As noted above, outside of the main body of granite, the straight gneiss wall rocks are cut by veins of similar granite, which vary from isotropic and crosscutting to concordant ribbon mylonites. If these granite sheets were related to the main granite, it would imply that the final emplacement of the pluton was preceded by its own vein cortege, and that the granite is, in fact, syntectonic with respect to the Striding mylonite belt. Accordingly, the 2585 ± 2 Ma age obtained from the granite would indicate the approximate time of mylonitization.

Five kilometres east of the Striding River (Fig. 2A), pink leucogranite ribbon mylonite with locally boudined garnet amphibolite and mafic granulite bands (2 pyroxene–garnet) contains coarse-grained, foliated pink leucogranite in the interboudin necks. The interboudin material, which we regard as syntectonic, has been dated at ca. 3.1–3.3 Ga. Allowing for the imprecision of the determined ages (see Hanmer et al. 1994), the three syntectonic age determinations cited here, when taken together, indicate that the high-grade mid- and late Archean deformation events in the materially contiguous East Athabasca mylonite triangle (Hanmer et al. 1994) are broadly correlative with those recorded by the Striding mylonite belt. The important point being made is that the mylonites are not Early Proterozoic.

Lozenges

In contrast to northernmost Saskatchewan, where the interior of the Athabasca lozenge is mostly obscured by the ca. 1.7 Ga Athabasca Basin, bedrock exposure in the Northwest Territories allows ready access to the interior of the Selwyn lozenge (Figs. 1, 2A) (Hanmer and Kopf 1993). However, the quality of exposure decreases rapidly northeast of Snowbird Lake. Accordingly, the Selwyn example offers the best opportunity to study the geological nature of the geophysically defined lozenges, and the adjacent wall rocks in our study area (see, however, Lewry and Sibbald 1977, 1980 and references therein).

The Rae wall rocks to the northwest of the Selwyn lozenge are a diverse variety of folded granitoid gneisses, intruded by crosscutting granitic and pegmatitic veins, associated with poorly foliated granitoid plutons of unknown age. Diagnostic metamorphic mineral assemblages are not present in these compositions, but the rocks are part of the regionally extensive retrogressed granulite terrane mapped in northern Saskatchewan (e.g., Slimmon 1989 and references therein). The foliation trends are highly variable, but are predominantly east–west (Fig. 2A) (Taylor 1970). To the southeast of the lozenge, the very poorly exposed Hearne wall rock is dominated by a voluminous mass of isotropic biotite granite around Kasba Lake (Taylor 1963). Less than 15 km due east of the lozenge lies a 10 km wide belt of amphibolites and finely laminated pelitic and volcaniclastic metasediments. They are typical greenstone belt rocks, which appear to constitute the southwestern extension of the 600 km long Rankin–Ennadai greenstone belt (Fig. 1) (e.g., Wright 1967). A flat-lying, penetratively developed, bedding-parallel, amphibole–chlorite cleavage is locally deformed about northeast-trending horizontal folds with wavelengths between 1 cm and 10 m, whose axial planes dip shallowly to the southeast or the northwest (Hanmer and Kopf 1993). Similar structural relations are described from elsewhere within the Rankin–Ennadai greenstone belt (e.g., Park and Ralser 1992). Contact relationships with the biotite granite are not exposed, and their time relationship remains unknown.

Selwyn lozenge

The Selwyn lozenge (Fig. 2) comprises an assemblage of weakly deformed rocks of mafic and intermediate composition (Hanmer and Kopf 1993). It is principally composed of homogeneous to banded amphibolite with abundant hornblendite and diopsidite, all intruded by a vein network of leucodiorite to tonalite and metagabbro. Locally it ranges from fine-grained, annealed mafic granulite of uncertain origin to variably deformed hornblende diorites and granodiorites. The extent of these mafic to intermediate lithologies has been underestimated in previous mapping, where they have generally been represented as undifferentiated paragneiss (Taylor 1963, 1970). Parts of the lozenge contain irregularly folded migmatitic granitoid orthogneiss and garnet–sillimanite diatexite. Foliation trends within the lozenge are curvilinear with a significant east–west component, clearly discordant to the magnetically determined lozenge boundaries, but similar to the structural trend pattern in the wall rocks (Fig. 2) (Taylor 1963, 1970).

Although granulite facies mylonites occur on both sides of the Selwyn lozenge, those on the northwest side only comprise minor discontinuous belts, 500 m thick, in a corridor 20 km long located at Wholdaia Lake (Fig. 2A) (Hanmer and Kopf 1993). The throughgoing Striding mylonite belt along the southeastern margin of the lozenge is derived from a lithological assemblage that includes anorthositic, pyroxenitic, and gabbroic/noritic rocks (Fig. 2A). The trace of these mylonites is highlighted on the horizontal gravity gradient map by a very strong positive anomaly along the southeast margin of the lozenge (Goodacre et al. 1987). We suggest that this indicates that the mafic lithological association may be more voluminous at depth (Hanmer and Kopf 1993). The point being made is that the anorthosite–pyroxenite–gabbro/norite association is volumetrically important, and is spatially restricted to the eastern margin of the lozenge.

Athabasca and Three Esker lozenges

Although much of the interior of the Athabasca lozenge is obscured by younger rocks and Quaternary deposits, the East Athabasca mylonite triangle at the northern apex, and the gneisses immediately west of the Virgin River shear zone, near the southwestern apex, are exposed (Fig. 1). Mid-Archean rocks in these two windows into the marginal parts of the Athabasca lozenge are lithologically very similar (Lewry and Sibbald 1977 and references therein; Crocker et al. 1993; Hanmer et al. 1994), and are compositionally comparable to the interior of the Selwyn lozenge. Both lozenges are mafic to intermediate in composition and bounded by granitic to granodioritic Rae and Hearne wall rocks (e.g., Taylor 1963, 1970; Gilboy 1980; Slimmon 1989). Moreover, it appears that the occurrence of anorthosite is associated with the eastern margin of the Athabasca lozenge. Within the East Athabasca mylonite triangle, anorthosite is confined to the sinistral shear zone on the eastern side of the lower deck (Hanmer et al. 1994), and the southern window is located adjacent to the eastern (Virgin River) margin of the lozenge (Crocker et al. 1993). Recalling the distribution of anorthosite along the southeastern margin of the Selwyn lozenge, we suggest that the available data, while not definitive, allow us to draw parallels between it and the Athabasca lozenge. The occurrence of anorthosite–garnet pyroxenite–diatexite ribbon mylonite on the southeastern margin of the Three Esker lozenge suggests that all three lozenges may share a similar internal architecture.

Preliminary Nd data support the proposed lithological correlation of the lozenges. With one exception, Nd model

Table 1. Preliminary Sm–Nd data from Striding–Athabasca mylonite zone.

Sample	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	$\varepsilon_{Nd(3.3)}$	$\varepsilon_{Nd(2.6)}$	$\varepsilon_{Nd(0)}$	T_{dm} age (ma)
Chipman tonalite (ca. 3.2–3.4 Ga)								
CK90 SB	0.88	5.87	0.0906	0.510397	1.42		−43.72	3326
CK90 M076	1.32	13.19	0.0605	0.509628	−0.76		−58.72	3444
CK90 M063	10.20	64.31	0.0959	0.510420	−0.38		−43.27	3452
Pyroxenite (ca. 3.2–3.4 Ga): inclusions in Chipman tonalite								
CK90 J099	0.86	3.32	0.1566	0.511807	0.85		−16.21	3402
Anorthosite (ca. 3.2–3.4 Ga): inclusions in Chipman tonalite								
CK90 S137	0.46	1.85	0.1503	0.511687	1.18		−18.55	3349
Three Esker anorthosite (ca. 3.2–3.4 Ga?)								
CK92 S197	0.51	2.02	0.1527	0.511754	1.50		−17.25	3304
Bohica mafic complex (ca. 2.6 Ga)								
CK91 C531	0.26	1.25	0.1258	0.511373		−0.95	−24.68	2931
CK91 S527	0.79	4.43	0.1078	0.511114		0.00	−29.73	2803
CK91 C534	0.43	1.98	0.1313	0.511502		−0.28	−22.16	2886
CK91 S574	3.66	14.32	0.1545	0.511955		0.78	−13.33	2847
CK91 S576	3.18	11.97	0.1606	0.512016		−0.07	−12.13	3012
CK91 S525	2.54	12.48	0.1231	0.511415		0.78	−23.86	2764
Granulite facies metagranites (ca. 2.6 Ga)								
CK91 C664a	4.44	25.39	0.1057	0.510984		−1.84	−32.26	2938
CK91 C728	10.25	55.78	0.1111	0.511249		1.55	−27.09	2686
CK90 S177	8.23	37.98	0.1310	0.511429		−1.62	−23.58	3019
CK91 S762	8.00	43.51	0.1112	0.511220		0.96	−27.66	2733
Mafic granulite (ca. 2.6 Ga): upper deck								
CK91 C780b	1.37	5.60	0.1479	0.511809		0.15	−16.16	2902
CK90 M205	1.02	4.58	0.1347	0.511589		0.30	−20.46	2837
CK90 S075	1.58	6.96	0.1373	0.511638		0.37	−19.52	2837
CK90 S210	0.91	4.37	0.1259	0.511446		0.43	−23.25	2802
CK90 S251	3.66	20.02	0.1105	0.511097		−1.24	−30.06	2906

NOTES: Samples come from the lower deck of the East Athabasca. mylonite zone (Fig. 3), except for the mafic granulite (upper deck; Fig. 3) and the Three Esker anorthosite. Values for $\varepsilon_{Nd}(3.3)$ and $\varepsilon_{Nd}(2.6)$ were calculated for times of magmatic crystallization based upon the ca. 3.4–3.2 Ga U–Pb zircon ages determined for the Chipman tonalite, and the ca. 2.6 Ga ages for the granulite metagranites and mafic rocks (see Hanmer et al. 1994). Attribution of a mid-Archean age to the Three Esker anorthosite is an extrapolation based upon geological and model age similarity to anorthosite inclusion within the Chipman tonalite. All measurements are by thermal ionization mass spectrometry at the University of California at Santa Cruz. Sm and Nd concentrations were determined by isotope dilution (C.F. Kopf, unpublished data). $T_{(DM)}$ ages are calculated after the method of DePaolo (1981). Discussed in text.

ages for the mid-Archean tonalites and associated anorthosite–pyroxenite assemblage in the East Athabasca mylonite triangle, and the margin of the Three Esker lozenge, fall in the range 3.45–3.30 Ga (Table 1). Crocker et al. (1993) report similar Nd model ages of ca. 3.5–3.1 Ga from mid-Archean rocks in the southern window in the Athabasca lozenge (however, see Bickford et al. 1994).

In the discussion that follows, we will suggest that, according to the combined isotopic, lithological, geophysical, and structural data, the Athabasca, Selwyn, and Three Esker lozenges can be interpreted as a discrete mid-Archean tectonic entity, rheologically distinct from the surrounding Rae and Hearne wall rocks. Furthermore, we will suggest that as a strength heterogeneity, the train of lozenges has strongly influenced the structural behaviour and tectonic evolution of the western Churchill continent during the late Archean and the Early Proterozoic.

Discussion

Our work on the northern Saskatchewan – District of Mackenzie segment of the Snowbird tectonic zone has established the existence and the nature of the Striding–Athabasca mylonite zone (see also Hanmer et al. 1994, in press). Briefly, it is a geometrically and kinematically complex, Archean, right-lateral transpressive structure, ca. 500 km long, formed at granulite facies in the deep crust. Early Proterozoic tectonic activity is recorded as extremely weakly developed and localized plastic deformation, and relatively minor granitoid intrusion. The granulite facies mineral assemblages in these mylonites contrast sharply with the lower metamorphic grade of the immediately adjacent Hearne crust (see also Hanmer et al. 1994). As we have argued elsewhere (Hanmer et al., in press), the sinuous and branching geometry of the Striding–Athabasca mylonite zone is not the result of modification of an initially straight shear zone. Therefore, because of its form, it could not have accommodated significant transcurrent displacement of its wall rocks, despite the associated spectacular mylonite fabrics (Hanmer et al., in press).

Stiff mid-Archean crust

The mylonites are but one component of this segment of the Snowbird tectonic zone. The other component is the train of three lozenges (Fig. 2), lithologically, and at least in part isotopically, distinct from their wall rocks (Table 1) (Crocker et al. 1993). Taking the Selwyn lozenge as our type example, the observation that it is not defined on both sides by throughgoing mylonites suggests that the lozenges have acted to localize the Striding–Athabasca mylonites, rather than the other way around. The control exerted by the lozenges over the complex trace of the mylonites indicates that they have been instrumental in provoking a partitioning of the flow. In other words, they behaved as relatively strong rheological heterogeneities (e.g., Bell 1981, 1985; Bell et al. 1989). This interpretation is supported by their lithological composition relative to that of the wall rocks. However, it raises the question as to why the mylonites should be derived at the expense of relatively stiff lozenge material, for example, anorthosite, rather than the relatively soft wall rocks. In the East Athabasca mylonite triangle, we have shown that emplacement of the voluminous late Archean mafic and granitoid protoliths was contemporaneous with mylonitization (see Hanmer et al. 1994). We suggest that their Nd model ages (ca. 3.0–2.7 Ga; Table 1) indicate juvenile magmas that may have interacted to varying degrees with older, more evolved crust. This suggestion is supported by the ε_{Nd} values calculated for 2.6 Ga (Table 1). We envisage that deformation, localized at the lozenge boundaries, may have focussed the emplacement of mantle and crustally derived magmas, thereby creating a hot, soft zone favourable to the continued localization of plastic deformation (e.g., Simpson 1986; Segall and Simpson 1986). In such a scenario, the evolved continental crust would be represented by the lozenges, with which isotopic interaction would be favoured by the chanelled emplacement process.

Nd model ages (Table 1) (Crocker et al. 1993) suggest that mid-Archean mafic and intermediate composition rocks along the southeast margin of the lozenges were derived from a relatively uniform mantle source, without significant interaction with older crust. We suggest that the lozenges may be fragments of mid-Archean, or older, crust within the western Churchill continent. If they ever formed a single body, dismemberment would have occurred prior to the ca. 2.6 Ga ribbon fabrics of the Striding–Athabasca mylonite zone. Mid-Archean (ca. 3.13 Ga) sinistral mylonites are preserved in the Chipman tonalite batholith in the eastern part of the East Athabasca mylonite triangle (Fig. 3) (Hanmer et al. 1994). Dextral mylonites of similar vintage (ca. 3.1–3.3 Ga) occur on the same side of the Selwyn lozenge, across the interlozenge gap (Fig. 2A). Therefore, it is kinematically possible that the Athabasca and Selwyn lozenges were pulled apart along the mid-Archean mylonites at ca. 3.13 Ga.

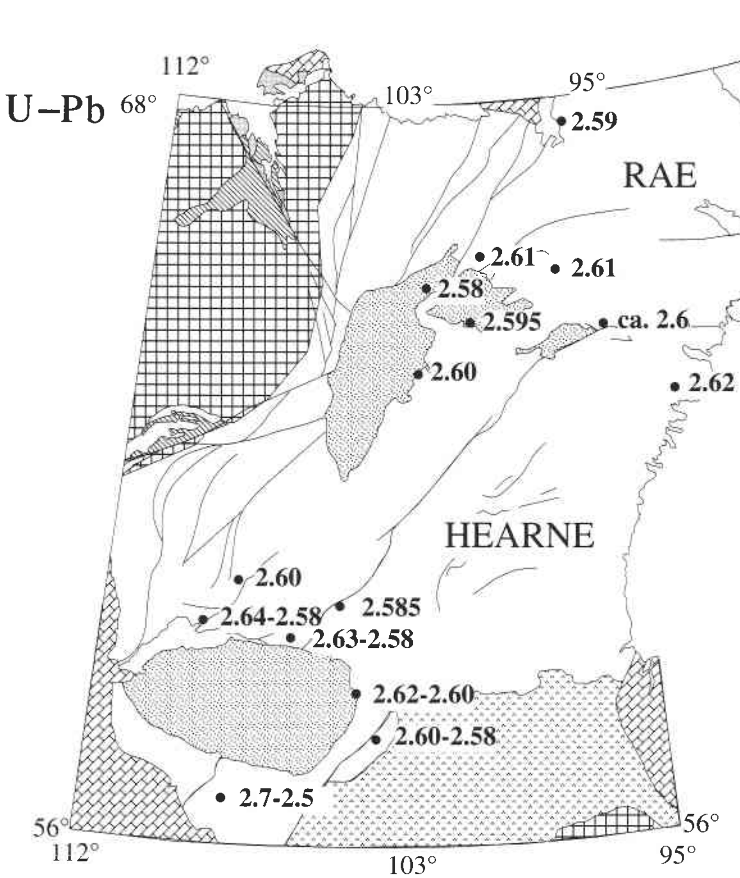
Late Archean intracontinental shear zone

Three independent data sets point to an intracontinental setting for the Striding–Athabasca mylonite zone in the Late Archean. First, the Rankin–Ennadai greenstone belt, the largest Archean tectonic entity identified to date in this part of the Canadian Shield, extends as a zone of outliers across the Hearne crust, over 600 km from Hudson Bay toward the Selwyn lozenge (Fig. 1). U–Pb zircon data from both its northeastern and southwestern parts show that the volcanic rocks of the greenstone belt are ca. 2.7 Ga old (Chiarenzelli and Macdonald 1986; Mortensen and Thorpe 1987; Tella et al. 1992). Very little is currently known about the tectonic context of the formation and deformation of the Rankin–Ennadai greenstone belt. Preliminary work suggests that, at least at its northeastern end, the belt may have originated as an island arc, which was tectonically stacked as a pre-2.65 Ga thrust-nappe pile (Park and Ralser 1992). Syntectonic to late tectonic granitoids were intruded into the arc, prior to its juxtaposition with its present continental wall rocks. However, it is important to note that the tectonic, stratigraphic, or intrusive nature of the gneiss–greenstone belt boundary as a whole remains unresolved. Dating of the granites indicates that thrust stacking occurred at 2.677 Ga, followed by upright shearing and folding just prior to 2.666 Ga (Park and Ralser 1992; see also Cavell et al. 1991). The salient point here is that, according to Park and Ralser (1992), the basin in which the greenstone belt formed had closed prior to ca. 2.67–2.66 Ga. Because there is no evidence for any other major late Archean basin in either the Hearne or the Rae crusts, we suggest that the western Churchill continent was a discrete tectonic entity by ca. 2.67–2.66 Ga.

Second, although the present data set is far from complete, late Archean granites, ca. 2.63–2.58 Ga, broadly contemporaneous with high-grade mylonitization in the Striding–Athabasca mylonite belt, do not appear to be arranged in the belt-like configuration typical of magmatic arcs associated with plate margins (Fig. 6) (e.g., Windley 1984). Third, the same spatial pattern of Archean Nd model age variation, ca. 4.0–2.4 Ga in the southwest compared with ca. 2.9–2.5 Ga in the northeast, is present in both the Rae and Hearne crusts (Fig. 7). This observation would be further supported if the basement rocks beneath the westernmost parts of the Trans-Hudson orogen are indeed part of the Hearne crust (see Bickford et al. 1992). This suggests that the southwestern parts of both the Rae and Hearne crusts are fundamentally different from their northeastern parts. For such a distribution pattern of Nd model ages to accommodate a late Archean suture along the Snowbird tectonic zone some very special boundary conditions would be required, for example, a narrow, short-lived basin, and similar drift and convergence vectors.

Taken together, these arguments suggest that, by ca. 2.62–2.60 Ga, the site of the Striding–Athabasca mylonite zone was not located at a suture, but lay well within the interior of the late Archean western Churchill continent (Fig. 1). We speculate that, had this segment of the Snowbird tectonic zone ever represented a suture, it would most probably have been mid-Archean in age. Combining our observations (see also Hanmer et al. 1994) with those of Crocker et al. (1993), one can construct a schematic sequence for the mid-Archean rocks of the Snowbird tectonic zone. Early semipelitic to pelitic rocks were intruded by a layered mafic complex composed of gabbro/norite, pyroxenite, and anorthosite. The mafic rocks were dismembered by tonalite batholiths. Two

Fig. 6. Distribution of available U−Pb zircon dates (Ga) for magmatic crystallization of granites in the range 2.63−2.58 Ga, approximately contemporaneous with granulite facies mylonitization in the East Athabasca mylonite triangle. Note the absence of an obvious belt-like disposition. Discussed in text. Data from Bickford et al. (1986), Van Schmus et al. (1986), Stevenson et al. (1989), Dudas et al. (1991), LeCheminant and Roddick (1991), Frisch and Parrish (1992), Roddick et al. (1992), Bickford et al. (1992), Annesley et al. (1992), and Tella et al. (1992).

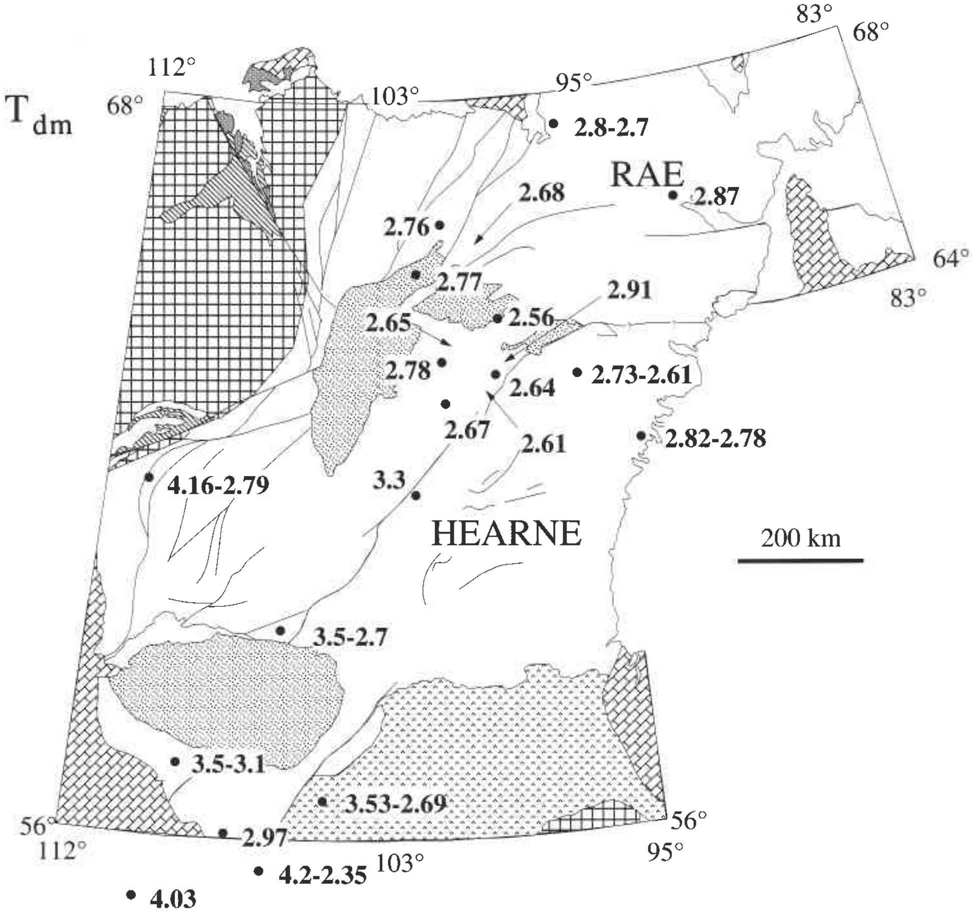


attempts to date the tonalite batholith in the East Athabasca mylonite triangle have yielded poorly constrained mid-Archean ages: 3149 ±100 Ma and 3466^{+327}_{-80} Ma (Hanmer et al. 1994). To obtain a preliminary indication of the magmatic history of these mid-Archean rocks, we have calculated ε_{Nd} values for 3.3 Ga, the average of the two ages (Table 1). The results are suggestive of juvenile magmas with little assimilation of evolved continental crustal material. However, we also note that the younger age of the two U−Pb determinations is similar to the ca. 3.13 Ga age of emplacement of a spatially associated syntectonic, syngranulite facies dyke swarm and associated granite sheets (Hanmer et al. 1994). It is therefore possible that the layered mafic rocks and anorthosite, tonalite, mafic dykes, and granite were emplaced in a sequence of events that culminated in the formation of the mid-Archean mylonites at ca. 3.13 Ga. Allowing that the age constraints remain fairly loose, this synthetic picture is reminiscent of published descriptions of more recently constructed, metamorphosed, and deformed magmatic arcs, e.g., Kohistan (Jan and Howie 1981; Jan 1988; Khan et al. 1989) and Wrangellia (Beard and Barker 1989), as well as Archean examples in the Limpopo Belt and Southwest Greenland (e.g., Windley et al. 1981). In Kohistan, the anorthosites represent relatively small masses that were emplaced into the lower part of the arc, which has subsequently been rotated through 90° about an arc-parallel horizontal axis (Coward et al. 1982). By analogy, the anorthosites in the present study area could represent the lower parts of an arc-related magmatic pile that has been tilted to expose its base along the southeastern margins of the lozenges. According to this scenario, if the Striding−Athabasca segment of the Snowbird tectonic zone was ever a suture, it would have been associated with the incorporation of the arc within the growing western Churchill continent.

Uplift and exhumation

In the Hearne wall rocks of Saskatchewan, metamorphism in the sillimanite−cordierite-bearing metapelites has not exceeded the stage of incipient partial melting (Hanmer et al. 1991, 1994), indicating that the rocks never reached granulite facies (see, however, Lewry and Sibbald 1977, 1980). To the northeast, regional metamorphism in the late Archean Rankin−Ennadai greenstone belt, outliers of which come within 15 km of the Striding mylonite belt, is generally at greenschist facies throughout its 600 km strike length (Fig. 1)

Fig. 7. Distribution of available T_{dm} Nd model ages (Ga) for the western Churchill Province. Note the broader range of ages to the SW and the younger ages to the northeast in both Rae and Hearne crusts. Discussed in text. Data from Bickford et al. (1987), Collerson et al. (1989), Dudas et al. (1991), Frisch and Parrish (1992), Bickford et al. (1992), Crocker et al. (1993), Villeneuve et al. (1993), and S. Tella (personal communication, 1993). See also Frost and Burwash (1986).

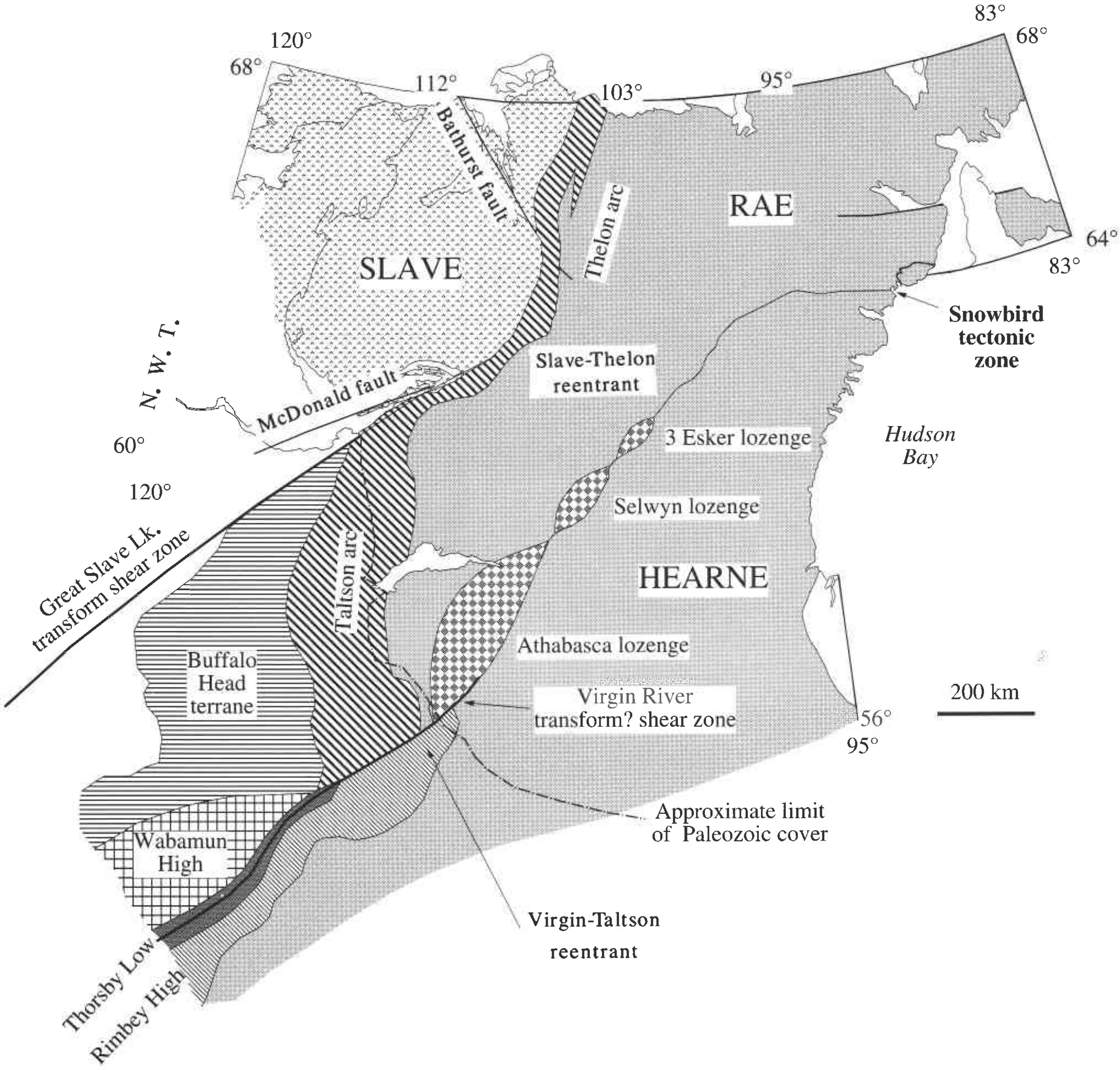


(e.g., Taylor 1963; Park and Ralser 1992; Hanmer and Kopf 1993). Clearly, unless the greenstone belt rocks are allochthonous as a result of major post-2.6 Ga deformation, there is a stark metamorphic contrast between the wall-rock assemblages and the late Archean metamorphism in the Striding–Athabasca mylonite zone. On the other hand, although the timing of metamorphism in the Rae crust is as yet unknown, the presence of extensive granulite facies assemblages in Saskatchewan (e.g., Fraser et al. 1978; Lewry et al. 1978; Slimmon 1989) suggests that there is less of a contrast between the Striding–Athabasca mylonite zone and its wall rocks to the northwest.

At present, there is insufficient data available to adequately account for the juxtaposition of the granulite facies mylonites and the lower grade rocks of the Hearne crust. There may be a number of different components contributing to the overall uplift history of the mylonites, perhaps at different times. Within the East Athabasca mylonite triangle, it is possible that some part of the uplift of the lower deck is related to normal fault displacements at the interface with the upper deck (Fig. 3) (see Hanmer et al. 1994). However, this does not explain the uplift of the upper deck granulites. In the Striding mylonite belt, the generally shallow to subhorizontal plunge of the transport direction makes it extremely improbable that the granulites were uplifted by the deformation that produced the mylonites. Moreover, Hanmer et al. (in press) have argued that the geometry of the Striding–Athabasca mylonite zone precludes tectonically significant displacements along the lineation direction.

If the uplift of the mylonites was not effected by plastic flow, then we are obliged to suggest that the present close juxtaposition of greenschist and granulite facies rocks requires important brittle fault movements. Regional mapping indicates that such faults, if they exist, are cryptic. They are not marked by dip-slip mylonites, cataclasites, breccias, gouge or quartz veining (Hanmer et al. 1991, 1992*a*; Hanmer and Kopf 1993). They do not offset or disrupt the Striding–Athabasca mylonite belt, nor are they visible in the immediately adjacent Hearne wall rocks (Taylor 1963). Even assuming that such faults are present, there are two constraints on their timing. First, the localization of deformation with cooling from granulite to greenschist facies in the East Athabasca mylonite zone occurred during the late Archean (ca. 2.6 Ga; Hanmer et al. 1994). Accordingly, uplift of the

Fig. 8. Sketch of the jagged boundary between Archean western Churchill continent (Rae and Hearne crusts) and the Early Proterozoic Thelon arc, Taltson arc, Rimbey High, Buffalo Head, Wabamun, and Thorsby terranes and domains, Athabasca, Selwyn and Three Esker lozenges, Virgin–Taltson and Slave–Thelon reentrants, and Great Slave Lake shear zone and Virgin River shear zone transforms. Modified from Ross et al. (1991). Discussed in text.



granulite facies mylonites could have occurred at that time, although there is no supporting evidence for this along the Striding mylonite belt. Second, the minimum age of uplift is given by the ca. 1.85 Ga volcanic and sedimentary rocks of the Baker Basin (Fig. 1), which straddle the northeastward extension of these putative faults with relatively minor offset (Hoffman 1988).

Implications for Proterozoic tectonics

Differences between the Archean geological features of the Rae and Hearne crusts, and the apparent truncation of the pronounced magnetic signature of the 1.99–1.90 Ga Taltson magmatic arc beneath the Phanerozoic cover, led Hoffman (1988) to interpret the Snowbird tectonic zone as an Early Proterozoic suture (Figs. 1, 8) (Hoffman 1988; see also Ross et al. 1991; Theriault 1992). This interpretation derives considerable support from the combined geophysical and geochronological evidence for important Early Proterozoic plate tectonic activity in the vicinity of the Snowbird tectonic zone, as it is defined in the subsurface of Alberta (Ross et al. 1991). However, our geological study of the Striding–Athabasca mylonite zone indicates that the tectonometamorphic history of a ca. 500 km long segment of the Snowbird tectonic zone is essentially Archean in age. Accordingly, differences between the Rae and Hearne crusts could reflect several factors, including possible juxtaposition of continental blocks during the mid-Archean, and difference in exposed crustal level. Nevertheless, the fact remains that Early Proterozoic tectonic activity to the west, and along strike to the southwest of the Striding–Athabasca mylonite zone in the time range 2.0–1.80 Ga, and possibly younger, is in apparent contradiction with the absence of significant Early Proterozoic ductile deformation in the Striding–Athabasca mylonite zone itself (see also Bickford et al. 1994). To address this question, it is necessary to briefly review the evidence for tectonic activity along the Alberta segment of the Snowbird tectonic zone.

Ross et al. (1991) have identified a fan-like bundle of alternating wedge-shaped magnetic anomalies to the southwest of the Athabasca lozenge (Fig. 8). They suggest that the Wabamun High is bounded on the southeast by a broad shear zone, the Thorsby Low, and by a more discrete fault on its northwest margin. Sheared gneiss from the Thorsby Low has yielded a U–Pb zircon age of 2.29 Ga, and a deformed pegmatite has given 1.91 Ga. On the southeastern side of the Thorsby Low, the Rimbey High, composed of biotite granites, has yielded U–Pb magmatic crystallization ages between 1.85 and 1.78 Ga. At least some of these granites were derived from old, highly evolved continental crust (Villeneuve et al. 1993; Bickford et al. 1994). The northeastern termination of the Rimbey High (Fig. 8) corresponds to the Junction granite, itself dated at 1.82 Ga (Bickford et al. 1986, 1994), and late-syntectonic with respect to greenschist facies mylonites of the Virgin River shear zone (Lewry and Sibbald 1977, 1980; Carolan and Collerson 1988, 1989). Using geometry to infer a kinematic analogy with the East Athabasca mylonite triangle, Ross et al. (1991) suggest that the Wabamun High is an escape wedge of Rae crust, expelled to the southwest during Early Proterozoic southeastward subduction and continental collision at ca. 1.85–1.82 Ga. In this scenario the Rimbey High would represent a continental arc constructed at the leading edge of the Hearne crust.

There are at least two possible solutions to the apparent temporal contradiction between the Alberta and Striding–Athabasca segments of the Snowbird tectonic zone: (*i*) southeastward subduction associated with the closure of a basin initially formed in the dilational quadrant of a strike-slip fault (Ross 1992), and (*ii*) eastward subduction beneath a jagged continental margin (see below).

Dilational basin

Ross (1992) suggested that the southeastward subduction beneath the Rimbey High was related to closure of a basin floored by thin continental to oceanic crust. In this model, the opening of the basin is related to strike-slip displacements along north–south-oriented faults within the Taltson arc, analogous to recent basins related to major Himalayan faults, such as the South China Sea (Red River fault) and the Andaman Sea (Sagaing Transform). However, to open a basin that extended to the southwest of the Taltson magmatic zone, this model requires the existence of a north–south-trending, crustal-scale, dextral fault, of appropriate age (pre-1.85 Ga). The basin must be large enough to sustain southeastward subduction and generation of the Rimbey magmas for a period of at least 70 Ma (1.85–1.78 Ga). Accordingly, the fault displacements must be correspondingly important. Even for a simple orthogonal geometry (maximum basin opening – fault slip ratio), and a conservative rate of plate motion (1 cm/a), the basin width and corresponding fault displacement should both be of the order of 700 km, at least. To close the basin, similar-magnitude sinistral displacements are required, perhaps along the same fault, during the period 1.85–1.78 Ga. However, the faults within the Taltson magmatic zone form an anastomosing system of discontinuous segments, with both dextral and sinistral elements, for which no throughgoing master fault has yet been proposed (Bostock 1987, 1988; McDonough et al. 1993*a*, 1993*b*). The cited authors also show that, where it can be determined, dextral shearing postdates sinistral shearing (see also Hanmer et al. 1992*b*), the converse of Ross' (1992) model. Moreover, the shear zones appear to be contemporaneous with plutonism in the 1.99–1.90 Ga Taltson magmatic arc (McDonough et al. 1993*a*; McNicoll et al. 1993), and are thereby too old to have accommodated the 1.85–1.78 Ga closure of the proposed basin.

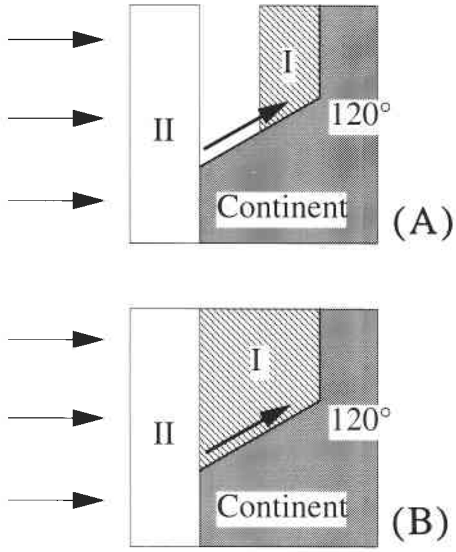
Reentrants, jagged margins, and transform faults

We suggest an alternative hypothesis. The Virgin River shear zone, exposed just north of the Paleozoic cover (Fig. 1), is a narrow greenschist to lower amphibolite facies mylonite zone (Lewry and Sibbald 1977, 1980). It is located on the same side of the Athabasca lozenge as the mylonitic Black–Bompas fault in the lower deck of the East Athabasca mylonite triangle (Figs. 1, 3, 8) (Hanmer et al. 1994). If the mylonites along this margin of the lozenge were part of a throughgoing master fault, they should show the same sense of shear. The Virgin River mylonites are strike lineated and associated with strike-slip displacements, but in contrast to the sinistral Black–Bompas fault, they have a dextral shear sense (Carolan and Collerson 1988, 1989; Hanmer et al. 1994). This implies that the Virgin River shear zone does not extend along the boundary of the Athabasca lozenge: *it must terminate south of Lake Athabasca*. Even if it was suggested that the Virgin River shear zone is more properly related to the mylonitic Straight–Grease fault on the northwest (dextral) side of the East Athabasca mylonite triangle (Figs. 1 and 3) (Hanmer et al. 1994), the kinematic consequences remain unchanged. Such a relationship would imply that the shear zone was composed of discontinuous, left-stepping segments, and the Virgin River segment would still terminate south of Lake Athabasca. Regardless, our initial geochronological data indicate that the Straight–Grease fault has only suffered minor reactivation during the Early Proterozoic (see Hanmer et al. 1994).

The orientations of the Early Proterozoic rocks, and their collective boundary with the Archean western Churchill continent in the Alberta subsurface, describe a 120° angle (Figs. 8, 9) (Ross et al. 1991). Jagged continental margins, composed of promontories and reentrants, are common features (e.g., Burke and Dewey 1973; Dewey and Burke 1974; Rankin 1976; W.A. Thomas 1977). In some cases it has been shown that primary offsets in the rifted continental margin directly reflect the intersection of the average plane of rifting and preexisting planes of weakness, such as old faults within the initially intact continent (e.g., Howarth 1977, 1980).

We hypothesize that the generally meridional (present-day coordinates) initial Early Proterozoic margin of the western

Fig. 9. Terrane docked at a jagged continental margin (I), and arrival of a second outboard terrane (II). Our hypothesis predicts that the Virgin River shear zone was a long-lived, initially transform fault, which accommodated relative displacements both at the active continental edge and within the accreted terranes. Compare with Fig. 8. A fault can remain active during subduction, convergence, and accretion if (*i*) terrane I does not fill the reentrant (A), or (*ii*) it is softer than the continental material (grey) and the rigid reentrant–promontory pair continues to act as a strength heterogeneity within the continental margin during subsequent compression (B). The initial transform fault may evolve in B to become an general intraplate strike-slip fault.



Churchill continent was jagged, with a sharp reentrant in what is now Alberta, possibly related to the southwest continuation of the train of mid-Archean lozenges (Fig. 8). We suggest that the Virgin River shear zone acted as a dextral transform fault along the southern arm of the reentrant (Fig. 9). The restriction of the Rimbey High to the Hearne edge of the Alberta reentrant (Fig. 8) suggests that the Virgin River shear zone acted to localize intrusion of the Rimbey granites, as has been shown in fault zones elsewhere (e.g., Hanmer et al. 1982; Strong and Hanmer 1981; Guineberteau et al. 1987; Hutton 1988; Hutton et al. 1990; Tikoff and Teyssier 1992). This scenario predicts that the Virgin River shear zone was active over much of its length throughout the period of plutonism (ca. 1.85–1.78 Ga). If the Virgin River shear zone was inherited from the geometry of the initial rifted western Churchill continental margin, one can predict that it would have influenced the accretion of pre-Rimbey crustal fragments. Accordingly, the Buffalo Head terrane, which docked as early as 2.08–2.05 Ga (Bostock and van Breemen 1994), should be offset or truncated by the proposed transform fault. Ross et al. (1991) present a clear geometrical interpretation of the aeromagnetic data, which indicates that this could well be the case. Therefore, we would predict that further geochronological work should show that parts of the Virgin River shear zone were active for 300 Ma (2.08–1.78 Ga; see also Bickford et al. 1994).

The abrupt termination of the magnetic expression of the buried southern segment of the ca. 1.99–1.90 Ga Taltson magmatic arc (Fig. 8) has been taken as prima facie evidence for truncation by important Early Proterozoic displacements across the trace of the Snowbird tectonic zone (e.g., Hoffman 1988; Ross et al. 1991). However, our hypothesis allows for an alternative explanation. Dewey and Lamb (1992) describe a relationship between gaps in the lateral continuity of subduction-related magmatic arcs in the Andes that reflects variations in the dip of the subducting plate. One might envisage segmentation and variable dip of the subducting plate beneath the leading edge of the western Churchill continent, on either side of a Virgin River transform fault, resulting in primary termination of the Taltson arc at the geophysical trace of the Snowbird tectonic zone.

An internal test

Our hypothesis makes two further predictions, which we can compare with the available regional geological data. First, a collisional event in the time range ca. 1.85–1.78 Ga must have occurred to the west of the accreted terranes (Fig. 8), in order to drive the displacements along the Virgin River shear zone after accretion of the Buffalo Head terrane. Second, a jagged continental margin would be expected to contain more than one reentrant–promontory pair. Regarding the first prediction, the ca. 1.85–1.78 Ga Rimbey plutonism was broadly contemporaneous with latitudinal (present-day coordinates) shortening and low-temperature faulting adjacent to the Slave Craton (Fig. 1) (Hoffman 1988; Henderson et al. 1990). The dextral McDonald fault and its sinistral conjugate, the Bathurst fault, are genetically related and coeval with a network of faults generated by collision of the Nahanni terrane with the Wopmay orogen, west of the Slave Craton, at ca. 1845 Ma (Figs. 1, 8) (Villeneuve et al. 1991; see also Hoffman 1988; Henderson et al. 1990). We suggest that the docking of the Nahanni terrane is a component of the western collisional event required by our hypothesis, although we must point out that the same prediction was made by Ross (1992).

Regarding the second prediction, we call attention to the deflections of the western margin of the Rae crust, highlighted by the 2.0–1.9 Ga Taltson–Thelon magmatic arc in the vicinity of Great Slave Lake (Figs. 1, 8). They have been attributed entirely to 1.97–1.90 Ga postcollisional indentation by the Slave Craton indentor (Hoffman 1987; Hanmer et al. 1992*b*). However, the northeast-trending segment of the Rae–Slave boundary, marked by the ca. 2.0–1.9 Ga Great Slave Lake shear zone, contains an apparently anomalous belt of concordant pre-2.56 Ga mylonites (Hanmer et al. 1992*b*). In the context of the present discussion, these late Archean mylonites may have localized a second reentrant in the initial Early Proterozoic margin of the western Churchill continent, which controlled the orientation of Great Slave Lake shear zone, albeit modified by collision and indentation. Note that both the Great Slave Lake shear zone and the younger McDonald fault have been interpreted as transform faults (Hoffman 1987; Henderson et al. 1990). Significantly, the proposed Slave–Thelon and Virgin–Taltson reentrants are of similar scale. Finally, if the jagged margin hypothesis holds, the remarkable spatial coincidence between Great Slave Lake shear zone and an associated syntectonic granite batholith (Hanmer and Connelly 1986; Hanmer 1988*b*; Hanmer et al. 1992*b*) is perhaps a well-exposed analogue of the relationship we propose between the Virgin River shear zone and emplacement of the Rimbey plutons.

Conclusions

Our study of the northern Saskatchewan – District of Mackenzie segment of the Snowbird tectonic zone suggests that fragments of relatively stiff mid-Archean crust have controlled the localization, shape, and complex kinematics of the multistage Striding–Athabasca mylonite zone during the Archean, as well as the geometry of Early Proterozoic rifting at the margin of the western Churchill continent in what is now Alberta. Further petrological investigation is required to test the possibility that the mid-Archean crust may have formed as a magmatic arc, now lying at the site of an Archean suture within the western Churchill continent. By the Late Archean, the Striding–Athabasca mylonite zone was located in the interior of the western Churchill continent, well removed from the plate margins. Except for the Alberta segment, the Snowbird tectonic zone was not the site of an Early Proterozoic plate margin. We suggest that the geometry of the Archean – Early Proterozoic boundary in the western Canadian Shield represents a jagged continental margin, composed of a pair of reentrants defined by rifted and transform segments, inherited from Early Proterozoic breakup and controlled by the Archean structure of the interior of the western Churchill continent. The geometry of this jagged margin appears to have strongly influenced Early Proterozoic tectonics and magmatism at the edge of the western Canadian Shield.

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