Striding-Athabasca mylonite zone: Complex Archean deep-crustal deformation in the East Athabasca mylonite triangle, northern Saskatchewan[1]

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The geophysically defined Snowbird tectonic zone is manifested in northernmost Saskatchewan as a deep-crustal, multistage mylonitic structure, the East Athabasca mylonite triangle. The triangle, located at the northeastern apex of a stiff, crustal-scale "lozenge," is composed of mid-Archean annealed mylonites and late Archean ribbon mylonites, formed during two granulite facies events (850–1000°C, 1.0 GPa). The flow pattern in the mylonites is geometrically and kinematically complex, and corresponds to that expected adjacent to the apex of a stiff elliptical volume subjected to subhorizontal regional extension parallel to its principal axis. The late Archean mylonites are divided into an upper structural deck, entirely occupied by a dip-slip shear zone, and an underlying lower deck. The latter is divided into two upright conjugate strike-slip shear zones, separated by a low-strain septum, which deformed by progressive coaxial flow. The flow pattern in the mid-Archean mylonites is compatible with that of the late Archean mylonites, and suggests that the crustal-scale lozenge influenced deformation since the mid-Archean. In the interval ca. 2.62–2.60 Ga, deformation in the upper and lower decks evolved from a granulite facies pervasive regime to a more localized amphibolite facies regime. With further cooling, deformation was localized within very narrow greenschist mylonitic faults at the lateral limits of the lower deck. By the late Archean, the East Athabasca mylonite triangle was part of a deep-crustal, intracontinental shear zone. This segment of the Snowbird tectonic zone was not the site of an Early Proterozoic suture or orogen.

Dans la Saskatchewan septentrionale, la zone tectonique de Snowbird définie selon des critères géophysiques comprend une structure mylonitique complexe, le triangle mylonitique de l'Athabasca oriental, formée en plusieurs étapes dans la partie inférieure de la croûte continentale. Le triangle, localisé à la terminaison nord-orientale d'un grand corps crustal rigide en forme d'ellipse, le boudin d'Athabasca, comprend des mylonites recuites d'âge archéen moyen et des mylonites à rubans d'âge archéen tardif, formées au cours de deux évènements tectonométamorphiques au faciès granulite (850–1000°C, 1,0 GPa). Le fluage tectonique dans les mylonites à rubans était complexe et correspond à celui attendu auprès de la terminaison d'un corps elliptique rigide dans un régime d'extension orientée selon son axe majeur. Les mylonites tardi-archéennes se présentent en deux parties : un compartiment structural supérieur entièrement constitué par une puissante zone de cisaillement extensif, et un compartiment structural inférieur qui comprend deux zones coulissantes conjugées, séparées par un zone à faible déformation coaxiale progressive. Le fluage dans les mylonites d'âge archéen moyen est comparable avec celui des mylonites tardi-archéennes et pourrait indiquer que le boudin d'Athabasca existe depuis l'archéen moyen. Pendant la période ca. 2,62–2,60 Ga, la déformation dans les deux compartiments structuraux a évolué d'un régime pénétratif sous faciès granulite à un régime localisé sous faciès amphibolite moyen ou même inférieur. Avec le refroidissement progressif, la déformation s'est encore localisée dans des failles mylonitiques sous faciès des schistes verts, aux limites latérales du compartiment structural inférieur. À la fin de l'Archéen, le triangle mylonitique de l'Athabasca oriental faisait partie d'une zone de cisaillement intra-continentale à grande profondeur. Cette partie de la zone tectonique de Snowbird ne correspond pas au site d'une suture ou d'un orogène d'âge protérozoïque précoce.

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Introduction

A high-amplitude linear anomaly in the horizontal gravity gradient map of the Canadian Shield, recently referred to as the Snowbird tectonic zone (Hoffrnan 1988), extends almost 3000 km from the Canadian Rocky Mountains to the coast of Hudson Bay, possibly as far as northern Quebec (Goodacre et al. 1987; Thomas et al. 1988). It has been proposed as the boundary between major crustal blocks within the western Canadian Shield (Figs. 1, 2), variously referred to as the Rae and Hearne provinces (Hoffman 1988), or the Amer Lake (cratonic) and Cree Lake (ensialic mobile) zones (Lewry et al. 1985). Low-amplitude anastomosing linear elements in the magnetic field occur along the trend of the gravity anomaly in the vicinity of the Saskatchewan – Northwest Territories border. They define a train of three 100 km scale elliptical features (Geological Survey of Canada 1987; Pilkington 1989), referred to by us as the Athabasca, Selwyn, and Three Esker "lozenges" (Fig. 3) (Hanmer and Kopf 1993).

Field-based structural geological studies of an –400 km long segment of the Snowbird tectonic zone have recently been undertaken where its geophysical trace crosses the territorial border (Figs. 1–3) (Hanmer 1987, 1994; Hanmer et al. 1991, 1992; Hanmer and Kopf 1993). Along this segment, the geophysical elements correspond to a geometrically and kinematically complex, dextral transpressive structure of Archean age, the Striding-Athabasca mylonite zone (Hanmer and Kopf 1993) (Figs. 1, 3), with two principal geological components: (i) a chain of relatively stiff fragments of mafic to intermediate middle Archean crust, corresponding to the Athabasca, Selwyn, and Three Esker lozenges and (ii) sinuous belts of Archean high-grade ribbon mylonites bounding them. Understanding the origin of these two components and the genetic relationship

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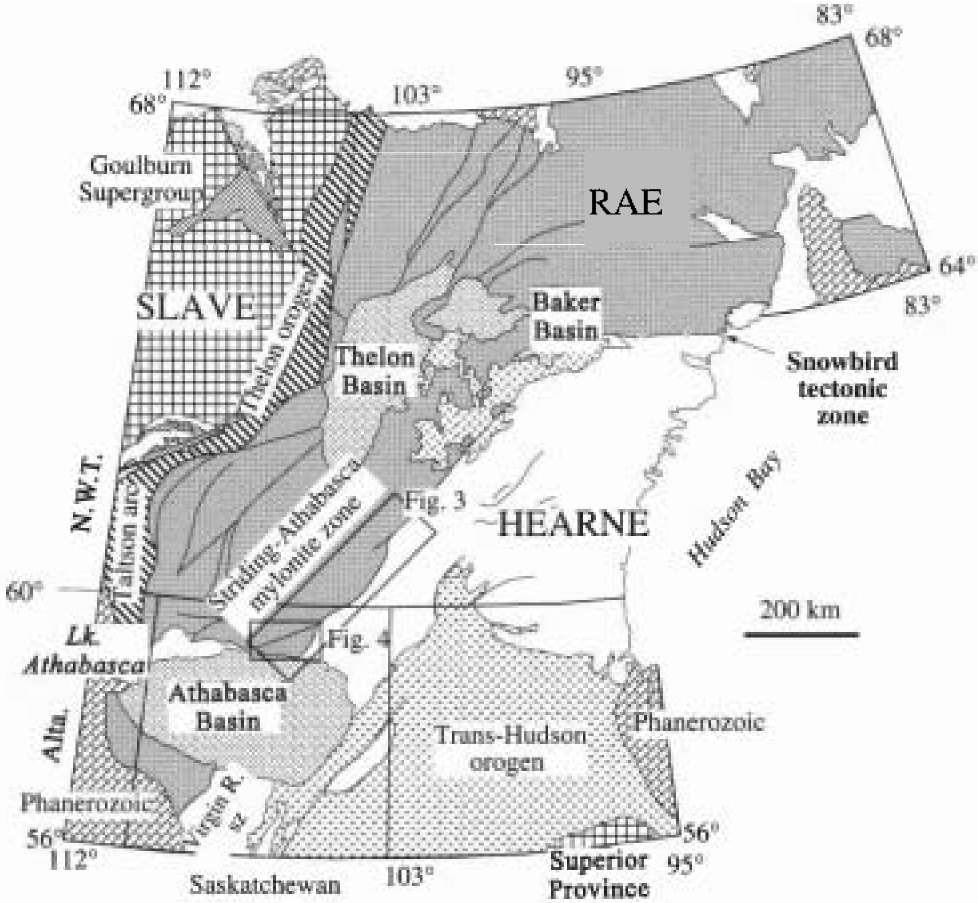

FIG. 1. Generalized representation of the distribution of the principal geotectonic features in the western Canadian Shield. The Rae and Hearne crusts, collectively referred to as the western Churchill continent, are separated by the Snowbird tectonic zone, shown here running from the Virgin River shear zone (sz), via the Striding-Athabasca mylonite zone, to Hudson Bay. Adapted from Hoffman (1988). Locations of Figs. 3 and 4 are given.

between them is the keystone to our current view of this segment of the Snowbird tectonic zone. The original definitions of the Rae and Hearne provinces were predicated upon the recognition of the Snowbird tectonic zone as an Early Proterozoic suture (Hoffman 1988). However, as will be shown, the mylonite zone we have studied is intracontinental and Archean. Since the terms Rae and Hearne are now well entrenched in general usage, we shall retain the Rae and Hearne nomenclature. All magmatic crystallization ages cited herein are U–Pb determinations from zircon (Table 1) (R. Parrish, unpublished data).

We have systematically mapped the geology of the Snowbird tectonic zone and its immediate wall rocks in the Stony Rapids area, northern Saskatchewan (Fig. 4) (Hanmer 1994). Our work from Selwyn Lake to north of Snowbird Lake (Fig. 3) (S. Hanmer et al.[2]) represents a structural, as well as a more general, upgrading of the preexisting geological database (Taylor 1963, 1970). Details of our structural, petrological, and geochronological studies will be presented elsewhere (e.g., Hanmer et al. 1994). In this contribution, we present the geology of the Saskatchewan part of our study (East Athabasca mylonite triangle, Figs. 3–5). Our observations in the Northwest Territories (Striding mylonite belt, Fig. 3), as well as a regional discussion and tectonic synthesis, are the subject of a companion paper (S. Hanmer et al.[2]).

East Athabasca mylonite triangle

In the Stony Rapids area, at the east end of Lake Athabasca, the geophysical expression of the Snowbird tectonic zone is geologically underlain by the East Athabasca mylonite triangle (Hanmer 1987, 1994; Hanmer et al. 1991, 1992). This 75 km × 80 km × 125 km triangle of high-grade mylonites, separated from the Rae and Hearne wall rocks by narrow greenschist mylonite belts, is located at the northeastern end of the magnetically defined, 300 km × 100 km Athabasca lozenge (Figs. 3–5). The triangular area, well known to earlier workers (Alcock 1936; Furnival 1940, 1941*a*, 19416; Mawdsley 1949, 1957; Kranck 1955; Colborne 1960, 1961, 1962; Colborne and Rosenberger 1963; Johnston 1960, 1961, 1962, 1963, 1964; Baer 1969; Gilboy 1980; Slimrnon and Macdonald 1987; Slimmon 1989), corresponds to the Tantato domain of Gilboy (1980). The simple lines of the foliation trajectories and the trace of layering within the East Athabasca mylonite triangle contrast strongly with the curvilinear trajectories of foliations and lithological traces in the wall rocks of the Rae and Hearne provinces (e.g., Gilboy 1980; Slimrnon 1989; Hanmer 1994).

[2]S. Hanmer, M. Williams, and C. Kopf. Striding-Athabasca mylonite zone: Implications for the Archean and Early Proterozoic tectonics of the western Canadian Shield. Submitted for publication.

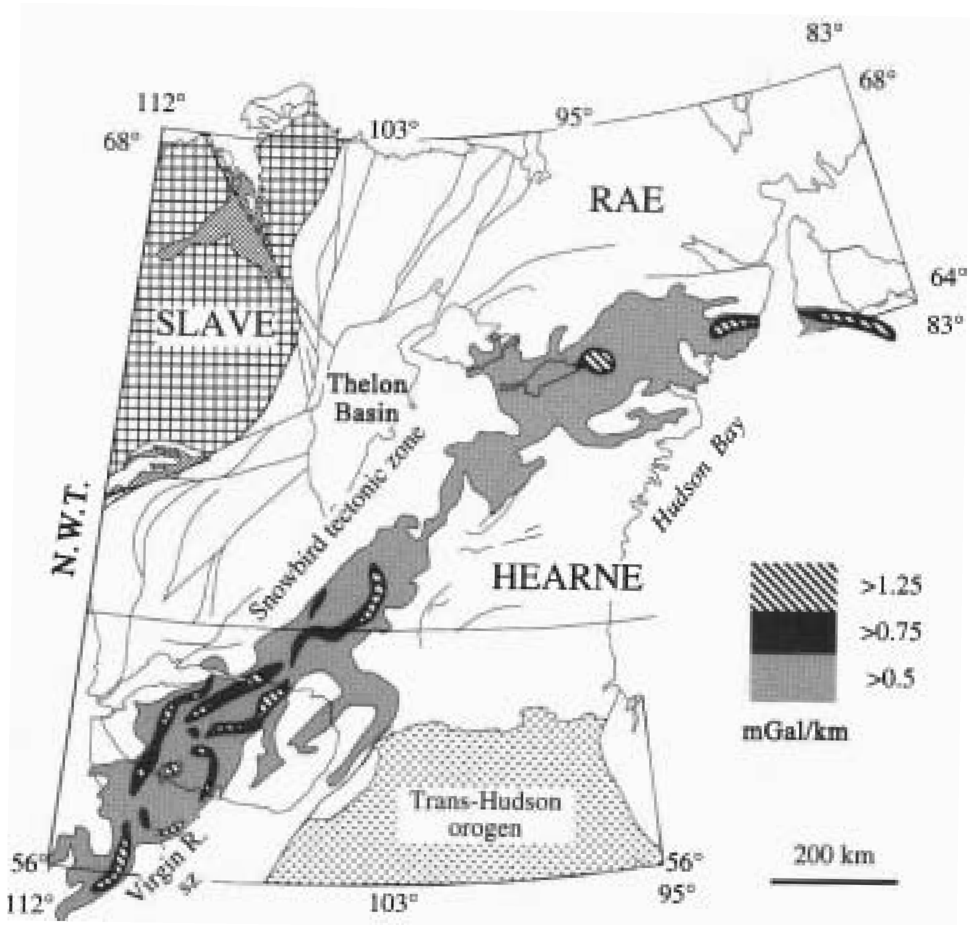

FIG. 2. Expression of the Snowbird tectonic zone in the horizontal gravity gradient of the western Churchill Province (grey tones and diagonal bars). Drawn from Goodacre et al. (1987).

To the west, the immediately adjacent Rae wall rocks are predominantly amphibolite and sillimanite–garnet–andalusite metapelitic migmatite. Farther from the East Athabasca mylonite triangle, they are composed of variably retrogressed, apparently complexly folded, granulite facies gneisses, intruded by voluminous granitic plutons, of unknown age (e.g., Baer 1969; Slimmon 1989). To the east, the Hearne wall rocks are composed of garnet–sillimanite–cordierite schists with incipient development of partial melt and concordant leucogranite sheets. However, within a few kilometres of the East Athabasca mylonite triangle, they are represented by dome-like nonfoliated granite plutons of Early Proterozoic age (ca. 1.84 Ga; Table 1), with an anastomosing carapace of foliated, very coarsely annealed pelitic, carbonate, and amphibolitic supracrustal rocks (Mawdsley 1957).

The East Athabasca mylonite triangle is structurally divided into an upper and a lower deck,[3] both of which were extensively and penetratively mylonitized at granulite facies (Figs. 4, 5) (Hanmer et al. 1994). As we shall explain, the upper deck was apparently initially emplaced as a hanging wall along a shallowly dipping, discrete thrust plane at its base. Subsequently, a dip-slip mylonite zone, at least 10 km thick, developed in the lower part of the hanging wall, parallel to the shallowly dipping basal thrust. The hanging wall to the dip-slip mylonites (the Athabasca lozenge?) is still buried beneath the overlying Athabasca Basin (ca. 1.7 Ga, Cumming et al. 1987; Cumming and Krstic 1992). Finally, the mylonites are displaced as a rigid hanging wall to later, localized dip-slip mylonites. Accordingly, in order to avoid repetitious clarification of which hanging wall we are referring to throughout the text, we couch our descriptions in terms of upper and lower decks. Layering and foliations in the lower deck are steeply dipping, with subhorizontal to shallowly southwest plunging extension lineations. The lower deck comprises three kinematic sectors: a central septum[4] which has experienced bulk progressive pure shear and relatively low finite strain, flanked by sinistral and dextral shear zones to the east and west, respectively (Fig. 5). The upper deck is entirely composed of a shallowly southwest dipping, dip-slip shear zone in the east, which progressively becomes a steeply dipping dextral shear zone in the west. Layering and foliations are parallel to the shear zone boundaries. Extension lineations plunge downdip in the east and porpoise moderately about the strike direction in the west (Fig. 5). The dip-lineated mylonites were associated with top-side-down displacements, at least during the later stages of their high-grade deformation history. Accord-

[3]We refer to upper and lower decks to convey the image of two structural compartments with geometrically distinct internal fabrics, separated by a sharp discontinuity.

[4]We use the term septum in the sense of a relatively narrow panel seperating two volumes, specifically the two shear zones within the lower deck.

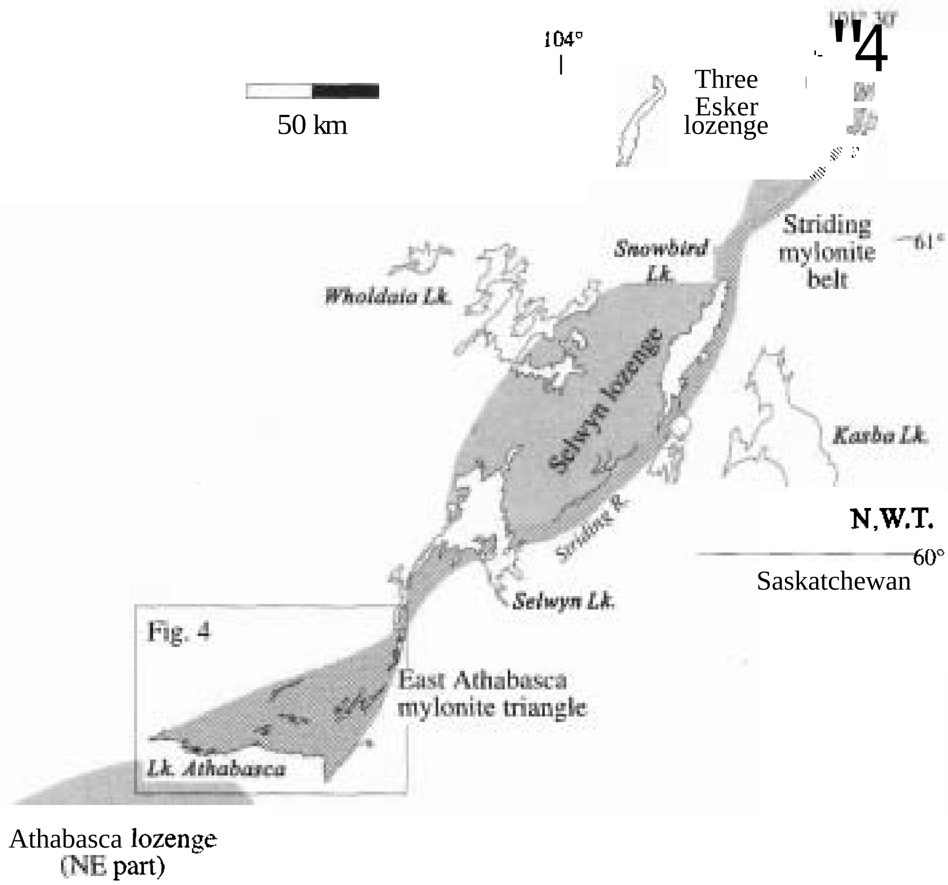

FIG. 3. Sketch of the spatial distribution of laterally continuous, extensive, high-grade mylonites (diagonal line rule) with respect to the magnetically defined crustal-scale Athabasca, Selwyn, and Three Esker lozenges (drawn from Geological Survey of Canada 1987). The locations of Fig. 4, the East Athabasca mylonite triangle, and the Striding mylonite belt are indicated.

ingly, in its present configuration, the upper deck is entirely occupied by an extensional shear zone with a lateral ramp.

Upper deck

The upper deck is composed of two principal lithologies: garnet – sillimanite ± orthopyroxene ± kyanite diatexite mylonite, and an orthopyroxene – clinopyroxene – plagioclase ± garnet mafic granulite mylonite (Figs. 4, 5). In the structurally higher level, discrete, kilometre-thick sheets of mafic granulite predominate, underlain principally by diatexite in the lower level. The mafic granulites coincide spatially with an important positive gravity anomaly that extends under the northern margin of the overlying Athabasca Basin (Fig. 6).

Diatexite

The diatexite (Fig. 4) is a leucocratic, straight-layered rock, with extremely well developed 0.5 mm by up to 100 mm ribbons of quartz and feldspar, which deflect about abundant 2–20 mm lilac garnets (Fig. 7). The term diatexite should be understood sensu lato, because a variable proportion of mesosome (modified parent rock) is locally present in sufficient proportion for the rock to be a metatexite (see Mehnert 1971; Brown 1973). The layering is a complex alternation of 0.1–1.0 m thick layers of ultramylonitic garnet granitoid leucosome, mesosome, garnet – quartz ± clinopyroxene, garnet – clinopyroxene ± quartz and mafic granulite, with concordant 10–20 cm thick sheets of isotropic to strongly foliated garnet pegmatite. Compositionally, the pegmatites are identical to the mylonitic leucosome. Accordingly, the contrast in the degree of their fabric development, combined with their intimate spatial association, suggests that migmatization was, at least in part, syntectonic with respect to the mylonitization. A U–Pb monazite age of ca. 2.62 Ga from the diatexite indicates that the high-temperature synmetamorphic mylonitization is no younger than late Archean in age (Table 1).

Although sillimanite is widespread in diatexite throughout the East Athabasca mylonite triangle, kyanite is confined to the lower structural levels of the upper deck. It occurs in intimate association with quartz, plagioclase, and K-feldspar as inclusions within garnets in the diatexite, as well as within the matrix. At one outcrop in the kyanite-bearing diatexites, we have found an occurrence of mafic bands of omphacitic clinopyroxene–garnet reacting to orthopyroxene–albite symplectites. We interpret these rocks as eclogites, partially retrogressed to granulite facies, which may have experienced pressures of at least 1.5 GPa (M. Williams, unpublished data).

Axis mafic granulite

The mafic granulite (Fig. 4) is a dark grey to brown, fine-grained, sugary-textured rock, with 30 mm × 2 mm polycrystalline streaks of orthopyroxene or plagioclase (Fig. 8). Delicate, postkinematic coronitic microstructures involving orthopyroxene cores with concentric clinopyroxene, garnet, and vermicular orthopyroxene rims are common in the Axis Lake area. In the structurally lower level of the upper deck, volumetrically subordinate bands of clinopyroxene–garnet–

TABLE 1. Generalized table of geological events in the East Athabasca mylonite triangle

Event		Age (Ma)
Minor reactivation of Straight–Grease fault		1788^{+28}_{-15}
Intrusion of isotropic granites into Hearne wall rock		ca. 1840
Robillard–Patterson gneissic plutons		ca. 1900
Minor dextral shearing in Rae wall rock		2181 ± 5
Localized greenschist facies mylonitization	D	ca. 2600
McGillivray granite		2621 ± 3
Fehr dykes		
Fehr granite		2598 ± 3
Clut and Melby–Turnbull granites	D	2614^{+9}_{-7}, 2611 ± 2
Felsic Lake leucogranite sheets		ca. 2629^{+26}_{-20}
Localization of flow		
End of pervasive granulite facies mylonitization (~700°C[a])		2619^{+9}_{-6} (l), 2629 ± 2 (u)
Hawkes granite*		ca. 2610, 2604 ± 1
Rea granite*		2584^{+40}_{-15}
Bradley and Brykea granites*	D	
Godfrey granite*		2601 ± 4
Mary granite (multiphase)*		2618 ± 4, 2606^{+13}_{-11}
Beginning of syn-granulite facies pervasive mylonitization		
Axis gabbro–norite (mafic granulite)		ca. 2600
Bohica mafic complex		2596 ± 12, 2620^{+34}_{-21}
Syn-granulite facies mylonitization, Chipman dykes and granites	D	3126^{+6}_{-5}
Chipman batholith		ca. 3150, >3400
Layered anorthositic mafic complex		
Pelitic–semipelitic sediments		

NOTES: All ages are U–Pb on zircon, except for the diatexites (l and u), which are U–Pb on monazite. Note that Rea granite is from the upper dcck. There is a significant degree of latitude in the geologically determined order of emplacement of the granulite facies metagranites (indicated by *) because they are usually only seen in direct contact with the Mary granite, which is itself a multiphase body composed of several plutonic units of different age. Note that within the sequence of high-grade events there are apparent contradictions between the proposed geological history and some of the age determinations. Discussed in the text. D. major mylonitization event (beiween broken lines); l, lower deck; u, upper deck.

[a]Monazite (see Pamsh 1990).

plagioclase, possibly unrelated to the main structurally overlying volume of mafic granulite, contain the above-mentioned eclogite assemblage. Coarse gabbroic to rnicrogabbroic ophitic textures are locally preserved in the mafic granulites, indicative of a plutonic origin for the protolith. Subtle compositional banding is only locally visible, suggesting that the protolith was not a layered mafic complex. The relationship of the mafic granulite protolith to the diatexites has been the subject of disagreement between partisans of an intrusive relationship (e.g., Alcock 1936; Furnival 1940; Mawdsley 1949) and those who interpret the mafic rocks as supracrustal in origin and stratigraphically intercalated with the diatexite protolith (e.g., Kranck 1955). We agree with Baer (1969) that, due to the scarcity of crosscutting contacts, field relationships are equivocal. However, our geochronological data suggest that the plutonic precursor to the mafic granulites is no older than ca. 2.6 Ga (Table 1).

Lower deck

The lower deck is principally composed of three large granulite facies metaplutonic complexes, or batholiths, and their mylonitic equivalents, all of which contain map-scale panels and rafts of diatexite (Figs. 4, 5). Except for the absence of kyanite and of relict eclogitic assemblages, garnet – sillimanite ± orthopyroxene diatexite is lithologically and structurally very similar to that described above for the upper deck, even including ca. 2.62 Ga monazite (Table 1). The three plutonic suites were intruded in the following sequence (Table 1): (i) Chipman tonalite (older than ca. 3.2 Ga), cut by syntectonic granite sheets (3.13 Ga) and mafic dykes, (ii) Bohica mafic complex (ca. 2.6 Ga), (iii) early high grade, and later lower grade, syntectonic metagranites (2.62–2.60 Ga).

Chiprnan *tonalite* batholith

In the eastern part of the lower deck, diatexite is in concordant contact with the tonalitic ca. 3.4–3.2 Ga Chipman batholith (Fig. 4; Table 1). Included map-scale panels and rafts of the former within the latter suggest an intrusive relationship. Deformation fabrics within the tonalite are composite, comprising an older annealed mylonite preserved in the eastern part of the batholith, reworked by younger ribbon mylonites principally to the west. The Chipman batholith is

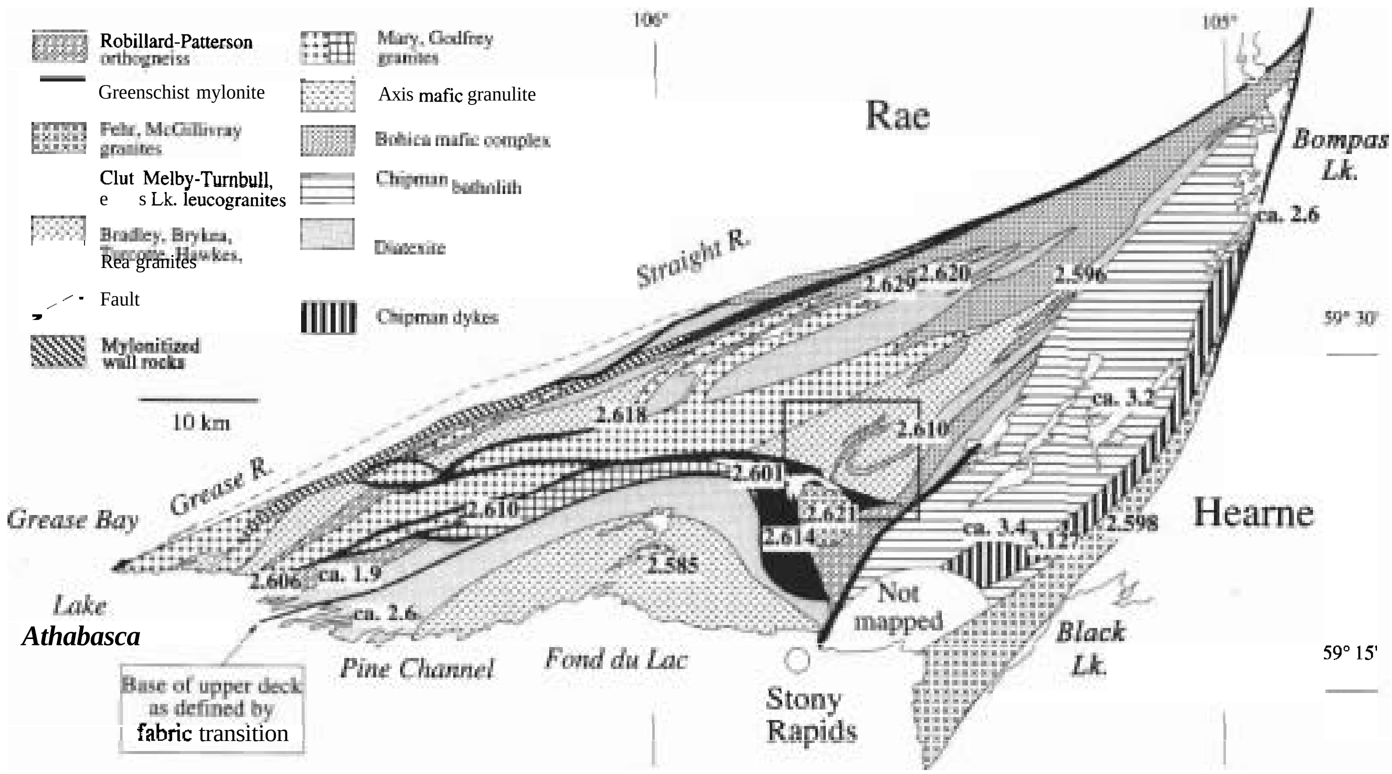

FIG. 4. Generalized geology of the East Athabasca mylonite triangle, Stony Rapids area, northernmost Saskatchewan. U–Pb zircon ages given in Ma. Location of Fig. 13 indicated by box. Discussed in text.

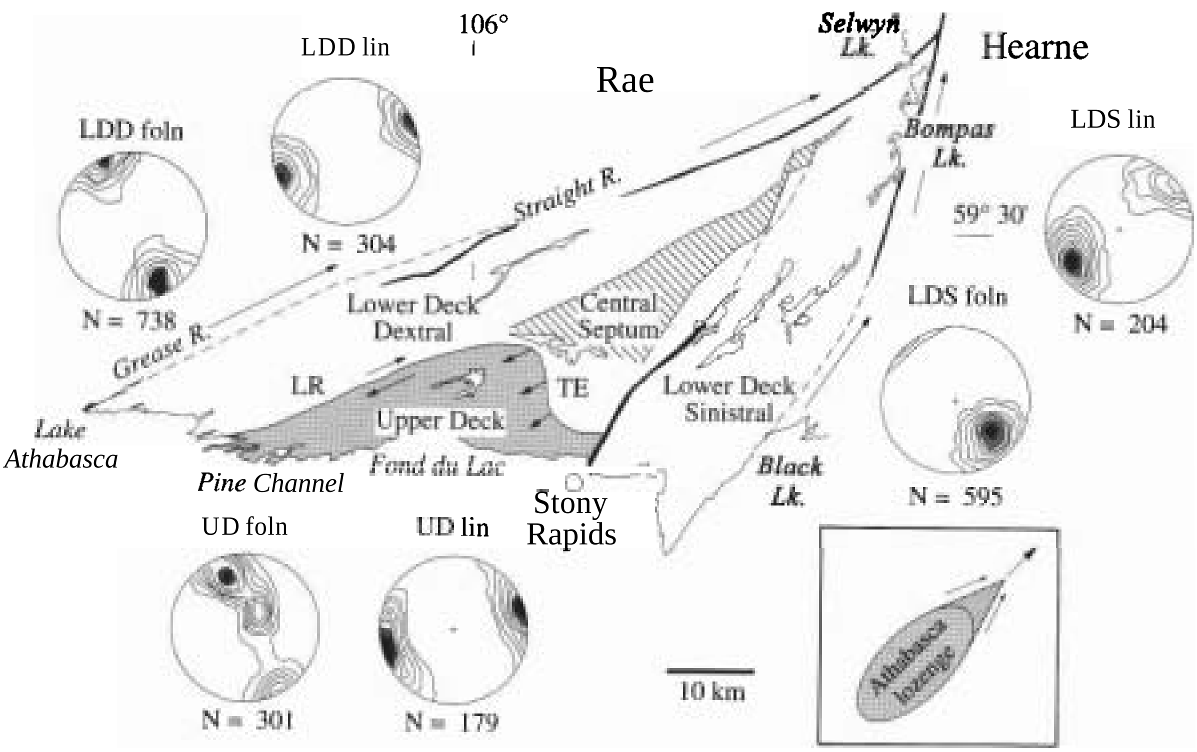

FIG. 5. Principal tectonic elements within the East Athabasca mylonite triangle (see Fig. 4 for geology). Arrows indicate directions of relative tectonic displacement. The lateral ramp (LR) and trailing edge ('SE) of the upper deck are indicated. Stereoplots are of poles to foliation and extension lineations 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 Fast Athabasca mylonite triangle (diagonal lines) at the northeastern apex of the crustal-scale Athabasca lozenge and the general pattern of flow resolved adjacent to the apex. Discussed in text. foln, foliation; lin, lineation.

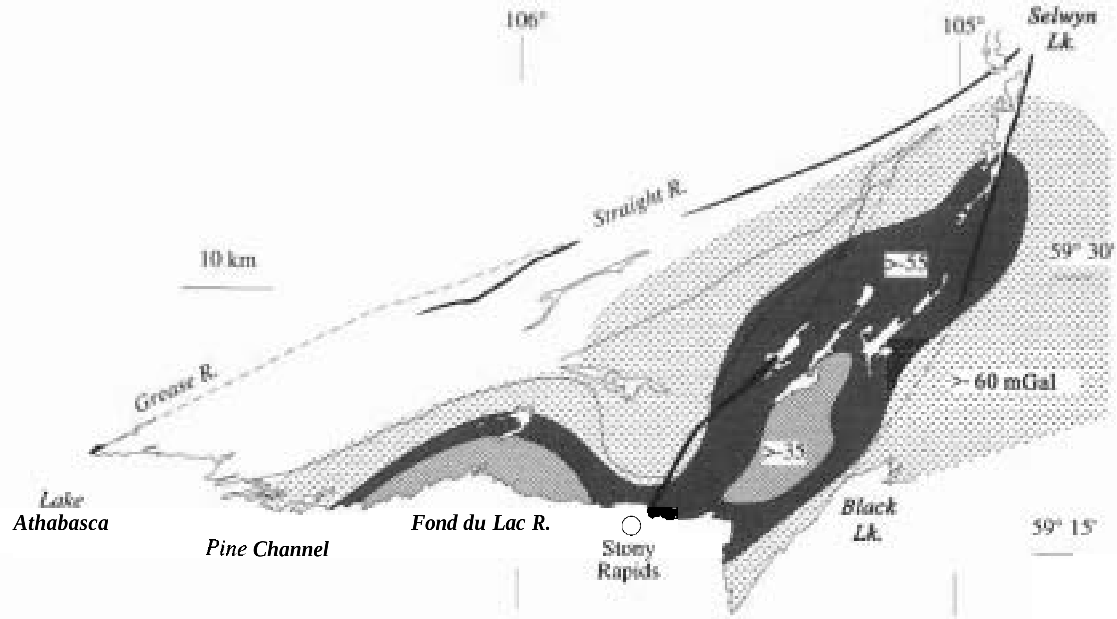

FIG. 6. Generalized representation of the Bouguer gravity field in the vicinity of the East Athabasca mylonite triangle. Unpublished 1 : 50 000 scale Bouguer anomaly compilation, Geophysics Division, Geological Survey of Canada. Discussed in text.

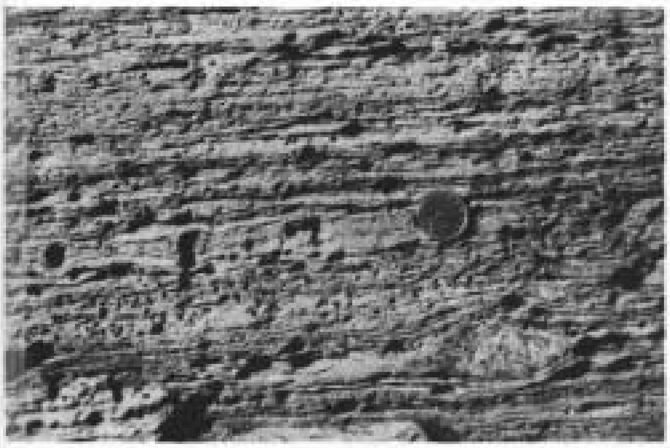

FIG. 7. Granulite facies diatexite ribbon mylonite, western upper deck. The pock marks are well-presewed garnet sites. Note the coarse feldspar just below the coin.

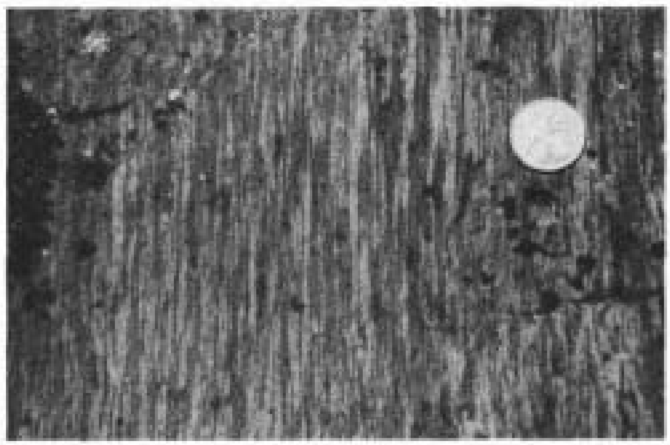

FIG. 8. Streaky aspect of mafic granulite mylonite, western upper deck. The light streaks are plagioclase, the dark streaks are orthopyroxene.

apparently a single magmatic body, divided into several structurally and metamorphically defined map units. In its central part, hornblende tonalite, with coarse igneous textures and a nebulous banding of dispersed amphibolite and clinopyroxenite inclusions, passes progressively westward, into clinopyroxene–garnet tonalitic ribbon mylonite. Although similar hornblende-bearing ribbon mylonites also form a 500 m wide band along the eastern margin of the batholith, most of the eastern part of the tonalite is transformed into an annealed, medium- to fine-grained hornblende ± garnet straight gneiss (Fig. 9; e.g., Hanmer 1988*a*), devoid of shape fabrics. Relict isoclinal folds and 2+ cm feldspar porphyroclasts indicate that the straight gneiss is an annealed mylonite, the product of folding, transposition, and significant grain-size refinement (e.g., Hanmer 1987). All units of the Chipman batholith contain abundant, dispersed, fist-size to map-scale inclusions of variably deformed anorthosite, layered amphibolite, pyroxenite, anthophyllite rock, and fresh peridotite, which appear to represent fragments of a dismembered layered mafic cornplex, dispersed throughout the tonalite. Clinopyroxene–garnet anorthosite mylonite, with 1–2 m thick garnet–pyroxenite bands, forms two panels, up to 2 km thick, along the northwest margin of the Chipman tonalite batholith. Initial geochemistry suggests that these panels are affiliated with the inclusions in the Chipman tonalite (S. Hanmer et al.[2]).

Syntectonic dykes

The batholith is intruded by a swarm of subconcordant, 1–100 m thick hornblende–plagioclase, hornblende–garnet–plagioclase, clinopyroxene–garnet mafic metadykes (Fig. 4), referred to here as the Chipman dykes (Chipman sills of Macdonald 1980). In the east, the dykes may be so closely

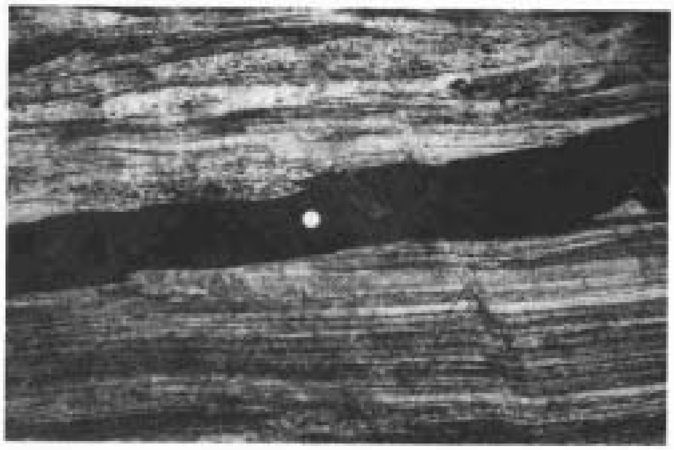

FIG. 9. Late Chipman dyke cutting across granulite facies annealed straight gneiss banding in eastern part of Chipman tonalite, eastern lower deck. Note the apophyses on either side of the dyke, which indicate emplacement by the dilation of an array of left-stepping sinistral fractures (e.g., Nicholson and Pollard 1985).

FIG. 10. Granulite facies Bohica mafic complex ribbon mylonite, northwestern lower deck. Bands reflect variation in plagioclase – orthopyroxene ± clinopyroxene ratio. Lens cap for scale.

spaced that they occupy 60–100% of the outcrop. The main part of the dyke swarm is spatially coincident with the tonalitic straight gneisses. Without implying any correlation with the upper deck units, we note that it also coincides with an extension of the positive gravity anomaly, centred on the mafic granulites of the upper deck (Fig. 6). This suggests that the volume of dyke material may increase with depth. Because they preserve all deformation states from isoclinally folded, transposed, and dismembered to crosscutting joint-controlled contacts with apophyses, even within a single outcrop, the dykes were emplaced during the deformation that produced the Chipman batholith straight gneiss (Fig. 9). The syntectonic features of the dyke swarm are totally obliterated in the ribbon mylonites of the western part of the Chipman batholith, providing the observational evidence for reworking of the annealed mylonites during the formation of the younger ribbon fabrics.

The eastern part of the Chipman batholith is also cut by an extensive array of leucogranite sheets, generally 10–100 cm thick, referred to here as the Chipman granite. The deformation state of the granite sheets varies from unfoliated and crosscutting with respect to the annealed straight gneiss fabric of the host tonalite to complete transposition into the straight gneiss banding. The granite sheets crosscut strongly deformed members of the Chipman dyke swarm, but are systematically crosscut by later, less-deformed to undeformed Chipman dykes. Accordingly, the Chipman batholith was intruded by broadly coeval, syntectonic granite and mafic sheets, although the period of emplacement of the latter outlasted that of the former. A crosscutting member of the granite sheet array has been dated at ca. 3.13 Ga (Table 1), indicating that the annealed mylonites and the syntectonic mafic dykes are mid-Archean in age.

Bohica mafic complex

In the northern and western parts of the lower deck, diatexite interfingers with, and is included within the Bohica mafic complex (Fig. 5), an association of orthopyroxene–clinopyroxene–garnet–plagioclase metanorite, gabbro, and diorite. The main body of the Bohica mafic complex is divided into well-foliated garnet–pyroxene mafic rocks with relict igneous textures (ca. 2.62 Ga; Table 1), flanked to the northwest by mafic mylonite (Fig. 10). Although much of the mylonite is principally composed of hornblende-bearing assemblages, anhydrous garnet–clinopyroxene and orthopyroxene–clinopyroxene-bearing assemblages are also well preserved.

Syntectonic metagranites

In the western part of the lower deck, the diatexite and the Bohica mafic complex are intruded by a voluminous suite of hornblende–garnet–biotite, hornblende – garnet – clinopyroxene ± orthopyroxene, and garnet–clinopyroxene granitoids, and their mylonitized equivalents, which were emplaced and metamorphosed at granulite facies (Fig. 4). Similar mylonitized garnet–clinopyroxene granite is also present in the upper deck. We have been able to distinguish a number of discrete map units, apparently corresponding to deformed plutons. Other variably deformed hornblende – biotite ± garnet granites are spatially associated with the interface between the upper and lower decks, and also occur adjacent to the outer margins of the lower deck. These were emplaced and deformed at amphibolite facies. Because the granitoids were emplaced syntectonically, they provide the opportunity to determine the timing of deformation within the East Athabasca mylonite triangle. Our geochronological data (Table 1) indicate that the ribbon mylonite fabrics throughout the upper and lower decks are late Archean in age, ca. 2.62–2.60 Ga (Table 1). It is therefore appropriate that we now establish the syntectonic nature of the granites.

Granulite facies metagranites

The 2.62–2.605 Ga Mary granite (sensu lato), the largest compositionally coherent granitic body, occupies much of the western lower deck (Fig. 4). It is penetratively mylonitized, but locally preserves small volumes of coarse (5 cm) plutonic protolith. Although referred to as a granite, much of it is composed of granodiorite and diorite. When fresh, it exhibits white streaks and ribbons of polycrystalline feldspar, separated by a fine "steel" grey matrix of feldspar–quartz–hornblende–garnet–clinopyroxene–orthopyroxene, studded with 2–5 mm hornblende, blood-red garnet, and locally clinopyroxene porphyroclasts (Fig. 11). Outcrop-scale observations, in windows of relatively low strain within the Mary granite, show that more strongly mylonitized parts of the granite are crosscut by less strongly mylonitized components.

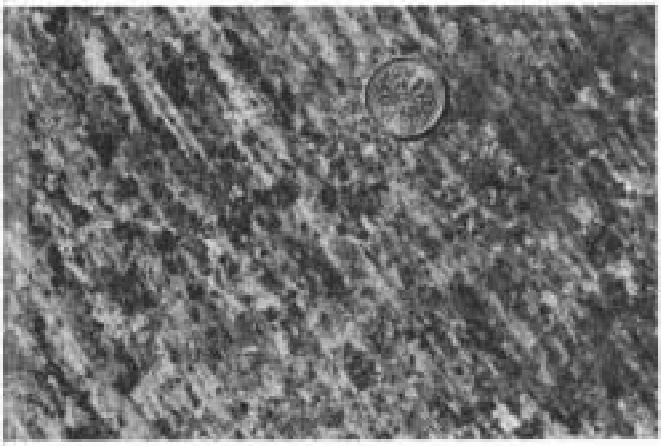

FIG. 11. Granulite facies streaky Mary granite mylonite, central lower deck. The dark "studs" are garnet, clinopyroxene, and orthopyroxene porphyroclasts.

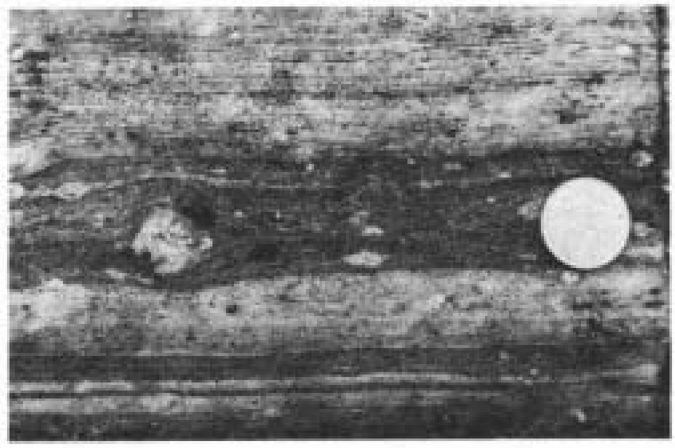

FIG. 12. Amphibolite facies porphyroclastic Mary granite mylonite, western lower deck. The light grey porphyroclasts are feldspar, the small dark ones are hornblende and garnet.

From this, we deduce that the Mary granite is syntectonic with respect to the granulite facies mylonitization that affects it.

The Mary granite is flanked by several garnetiferous metagranites. To the east, it apparently grades into, yet is included within, the ca. 2.61 Ga Hawkes leucogranite (Fig. 4). We remain uncertain about the relationship between the two because of the compositional variability within the Mary granite itself. The Hawkes granite varies structurally from an isotropic, equigranular, medium-grained (1–2 cm) plutonic rock to a ribbon ultramylonite. However, all the structural variants carry the same metamorphic mineral assemblages and metamorphic textures as those described for the Mary granite. Whether the Hawkes granite is part of the Mary granite, or an independent pluton of similar age, it is syntectonic with respect to the granulite facies mylonitization.

To the south, the Mary granite is partly flanked by the coarse (5 cm) garnet – hornblende ± clinopyroxene mylonitized Godfrey granite (2.60 Ga; Fig. 4). To the west, the Bradley and Brykea leucogranites (Fig. 4) are both garnet–clinopyroxene bearing, and penetratively mylonitized. In the absence of direct evidence of syntectonic emplacement, the Godfrey granite only yields a maximum age of mylonitization.

Amphibolite facies metagranites

Hornblende – biotite ± garnet metagranites, syntectonically emplaced into the granulite facies mylonites and metagranites just described, have been variably deformed at amphibolite facies. Because they contain no trace of relict pyroxene, they were emplaced after granulite facies metamorphism. Immediately beneath the eastern part of the upper deck, the (proto)mylonitic Clut granite (2.614 Ga) forms a shallowly south and west dipping kilometric-scale sheet (Fig. 4), with a concordant mylonitic foliation and a dip-parallel extension lineation. Abundant asymmetrical extensional shear bands indicate that mylonitization was associated with top down to the south and west displacements. It extends into the western part of the lower deck as the anastomosing vertical array of Melby–Turnbull metagranite sheets (2.61 Ga). The latter comprise two principal branches (Fig. 4). The more northerly branch extends due west from the Clut granite. The more southerly one lies subparallel to the lateral ramp of the upper deck, but within the lower deck. The granite sheets are concordantly foliated, strike lineated, and variably mylonitized. They were exclusively emplaced into a network of 1 km wide, particularly fine grained and porphyroclastic (Fig. 12), dextral strike-slip hornblende–garnet (pyroxene-absent) mylonite belts within the Mary granite. From the foregoing, we deduce that the Melby–Turnbull granite sheets were syntectonically emplaced toward the end of dextral shearing along a network of localized porphyroclastic mylonite belts within the lower deck under amphibolite facies conditions.

In the western part of the lower deck, thin intrusive sheets of dextral, thoroughly mylonitized biotite leucogranite form a concordant swarm. The Felsic Lake leucogranite ultramylonite, located along the western boundary of the lower deck (Fig. 4) and identical to the granite sheets, probably represents a map-scale component of the swarm. Locally, mylonitized later members of the swarm (ca. 2.63 Ga) crosscut earlier, concordant, mylonitized leucogranite sheets at a low angle. Accordingly, the leucogranites are syntectonic with respect to mylonitization.

Along the southeast margin of the lower deck, the hornblende – biotite ± garnet Fehr granite (2.598 Ga) cuts the Chipman batholith and its associated syntectonic mafic dykes (Fig. 4), but is itself cut by a later set of subconcordant hornblende–plagioclase–garnet mafic metadykes (Fehr dykes). The granite is very coarse and isotropic in its southwestern part, but grades into sinistral hornblende–garnet-bearing amphibolite facies mylonite to the northeast. At the southwestern end of the central septum, the biotite–hornblende McGillivray megacrystic granite (2.62 Ga) cuts the Clut granite (Fig. 4). It is very coarse grained, with very little internal tectonic fabric development and locally preserves a magmatic alignment of 5–10 cm euhedral feldspar megacrysts. Although its circular outcrop pattern and internal deformation state are suggestive of postkinematic emplacement, the pluton is in fact syntectonic. It contains misoriented rafts of mylonitic Clut granite, but is itself cut by a narrow (<100 m) west-side-down hornblende-bearing ribbon mylonite zone, along its western contact.

Early Proterozoic plutons

Hanmer et al. (1992) described several upright panels of Robillard–Patterson orthogneiss in the lower deck (Fig. 5). Although they have the field appearance of coarsely recrystallized, vertically layered, horizontally rodded and folded orthogneisses, they lack any indication that they were ever subjected to granulite facies metamorphism and show no sign of being annealed mylonites. Hanmer et al. (1992) suggested that these panels were relics of older material, but were uncertain of their significance. Our geochronological data indicate that these panels were emplaced at ca. 1.9 Ga (Table 1) and suggest that they are of magmatic origin. The poorly exposed boundaries with the adjacent mylonites are broad, ill-defined zones and could be progressive intrusive contacts. In the absence of Early Proterozoic regional deformation in the adjacent high-grade mylonites (Table 1), it would appear that the deformation of the orthogneisses is local and related to their plutonic emplacement, perhaps in a manner analogous to that described for the Donegal Granite in Ireland by Pitcher and Berger (1972).

Greenschist mylonites

The geometrical and kinematic pattern of high-temperature strike-slip shear zones in the lower deck is reflected in a set of low-temperature mylonites (Figs. 4, 5) (Hanmer et al. 1991). Biotite–chlorite and chlorite–sericite mylonites occur discontinuously along the northwestern side of the dextral shear zone and along the eastern side of the sinistral shear zone. The mylonites are spatially associated with very discrete brittle faults. Pseudotachylite is practically absent, cataclasite is very rare, and quartz-carbonate veining is infrequent.

In the sinistral shear zone of the lower deck, the transition from the hornblende-bearing ribbon mylonites is abrupt, albeit progressive. The main belt of greenschist facies chlorite–sericite mylonites, the sinistral Black–Bompas fault, is 100–200 m thick and derived, in part, at the expense of the metatexitic paragneisses of the eastern wall rock. At Bompas Lake (Fig. 4), a 50 m wide, ca. 2.6 Ga leucogranite sheet (Table 1) cuts two-mica mylonitic metatexite and biotitized amphibolite, which form the eastem margin of the Black–Bompas fault. The leucogranite then acted as a soft medium within which the subsequent development of the chlorite–sericite mylonites of the Black–Bompas fault was localized. We suggest that the leucogranite was anomalously soft because it was in a warm, subsolidus state imrnediately following magmatic crystallization. Accordingly, we consider the Black–Bompas fault to be late Archean in age.

The Straight–Grease fault in the dextral shear zone is an array of mylonites and anastomosing fault strands that preserves a progressive, retrograde relationship with the high-temperature ribbon mylonites of the lower deck. Mylonitization and retrogression of the northwestern wall rocks during dextral shearing occur over a distance of less than 100 m. However, within the lower deck the transition from granulite facies ribbon mylonites to chiorite-bearing mylonites occurs progressively over a zone up to 1.5–2.0 km wide. Garnet–hornblende–clinopyroxene–orthopyroxene-bearing ribbon mylonites derived from diatexite, Mary granite, and Bohica mafic complex protoliths become progressively finer grained towards the chlorite-bearing mylonites, such that their ribbon structure is invisible to the naked eye and the rocks are distinctively nonfissile. The syntectonic granite sheets of the ca. 2.63 Ga Felsic Lake leucogranite swarm occur within the amphibolite facies part of the transition zone. We interpret these observations in terms of the reworking of higher temperature mylonites by younger, colder mylonites towards the close of the late Archean shearing events described herein. Accordingly, the 15 km wide belt of granulite facies mylonites has been reworked to produce a 2 km wide amphibolite facies belt, which was itself reworked to form a belt of lower greenschist facies mylonites, several hundreds of metres wide. This reflects the progressive narrowing of the active zone of shearing with cooling through time (localization, e.g. Hanmer 19886; Hanmer et al. 1992), and contrasts with the mylonites of the Black–Bompas fault, which cut across and truncate the higher temperature mylonites in an abrupt manner (Fig. 4). Thin (1 m), ca. 1.8 Ga (Table 1) leucogranite veins, which crosscut the greenschist mylonite fabric of the Straight–Grease fault at a high angle, are themselves moderately folded and mildly foliated. They demonstrate that minor tectonic reworking of the greenschist mylonites occurred during the Early Proterozoic.

Complex *flow*

The distribution pattern of tectonic fabrics in the East Athabasca mylonite triangle is complex. The penetrative development of ribbon mylonites (and the significant finite strain that such fabrics represent) allows us to equate the local mylonitic foliation with the local flow plane (see Hanmer and Passchier 1991). Recall that, geometrically, the East Athabasca mylonite triangle is divided into an upper and a lower deck (Fig. 5). This is most clearly seen in the eastern half of the interface between the two decks, where the shallowly south and west dipping foliation of the upper deck is highly discordant to the upright, northeast-trending foliation and lithological contacts within the lower deck.

Dip-slip shear zone (upper deck)

Geometrically, the upper deck can be considered in two parts (Fig. 5). In the east it is essentially a shallowly west to southwest dipping homocline. The foliation and layering progressively steepen and change azimuth westwards to form an upright, southeastward-dipping panel. In the east, the extension lineations are approximately downdip, whereas in the west they are subparallel to the strike (Fig. 5). Layer-parallel ribbon mylonite fabrics are penetratively developed throughout the upper deck. The striking feature of the mylonite fabrics is their homogeneity at all scales. Visible shear-sense indicators (e.g., Hanmer and Passchier 1991) are rare. Nevertheless, the few that we have found, mostly asyrnmetrical extensional shear bands, consistently indicate a top down to the southwest sense of shear.

From a purely structural perspective, the upper deck is a penetratively mylonitic segment of a curved, 10 km thick, granulite facies shear zone, whose upper boundary has not been observed. 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 remarkably homogeneous diatexite mylonite, which was just described (see Figs. 4, 5). Passing eastward, the basal fabric transition initially coincides with the lithological contact between the diatexite and the underlying Godfrey granite, then transgresses the lithological contact to lie within the Godfrey granite. Still farther to the east, the fabric transition has been intruded by the younger, mylonitic Clut granite (Figs. 4, 5). Initially, the continuity of lithological units across the fabric transition in the west might suggest that the upper deck is parautochthonous with respect to the lower deck. However, from a metamorphic perspective, the contrast in the high-grade assemblages across the boundary between the structural decks strongly suggests that the upper deck is allochthonous and relatively far traveled. Therefore, we suggest that the original base of the upper deck lay at the lower limit of the diatexite, including that which now occurs immediately on the lower deck side of the fabric transition. The original contact has been subsequently obscured by the emplacement of the Godfrey granite and the Robillard–Patterson gneissic pluton. This would imply that the upper deck was initially emplaced as a thrusted hanging wall, whose lower level evolved into the 10 km thick shear zone, within which the marginal fabric transition is preserved. Although the early kinematic evolution of the shear zone is masked by the homogeneity of the ribbon fabrics, it acted as an extensional shear zone, at least during the latter part of its history.

Strike-slip shear zones (lower deck)

The tripartite geometrical subdivision of the lower deck represents three kinematic sectors (Fig. 5). The northwestern side of the lower deck is an ~15 km wide belt of upright mylonitic foliations and lithological contacts, which strike ca. 060–070°. The eastern side is a 20 km wide belt of about 020-050°-trending, moderately to very steeply northwest dipping mylonites. The two belts are separated from each other by a wedge-shaped area of anastomosing mylonite belts, ca. 15 km across the base, referred to here as the central septum (Fig. 5).

Outside of the central septum, ca. 2.6 Ga ribbon mylonite and ca. 3.13 Ga annealed mylonite fabrics are extensively developed (Table 1). Diverse assemblages of shear-sense indicators, such as winged porphyroclasts, asymmetrical extensional shear bands, asymmetrical pull-aparts, pressure shadows, oblique boudin trains, and oblique internal fabrics in shear-parallel veins (see Hanmer and Passchier 1991 for review and references) occur throughout the lower deck. The northwestern and eastern mylonite belts are strike-slip dextral and sinistral shear zones, respectively, with gently to moderately southwestward plunging extension lineations (Fig. 5). 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 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.

Central septum (lower deck)

The central septum is principally composed of mylonitic to isotropic Hawkes leucogranite, associated with panels and rafts of variably deformed Bohica mafic complex and Mary granite (Figs. 4, 5). In contrast to the flanking, kinematically homogeneous, dextral and sinistral shear zones, the spatial distribution of shear-sense within the central septum is heterogeneous, as illustrated by the following observations:

(i) A large panel of Mary granite occupies the middle of the central septum (Figs. 4, 5, and 13). It encloses steeply dipping, map-scale rafts of moderately deformed Bohica mafic complex, yet is itself mylonitized. Where the rafts of Bohica mafic complex strike ~020°, strike-slip shear sense in the adjacent mylonitic Mary granite is sinistral. Where the rafts of Bohica mafic complex strike ~070°, the Mary granite contains dextral shear-sense indicators (Fig. 13).

(ii) In the eastern part of the central septum, the Hawkes granite contains a large curved panel of annealed, straight-banded diatexite, whose fold axis plunges gently to the southwest (Fig. 13). In the fold hinge zone, the diatexite and the immediately overlying granite are transformed to ribbon mylonites associated with a top down to the southwest sense of shear.

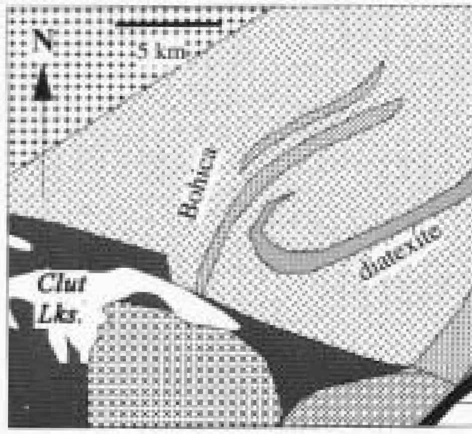

FIG. 13. A detail from the geological map of the East Athabasca mylonite triangle (Hanmer 1994) to illustrate the kinematic pattern in part of the central septum (see Fig. 4 for location and legend). Panels of Bohica mafic complex and of diatexite are labelled. Discussed in text.

(iii) At the scale of the central septum as a whole, although the mylonitic fabrics of the northwest side tend to reflect dextral non-coaxial flow and the southeast side tends to be sinistral, unequivocal sinistral and dextral shear-sense indicators (see Hanmer and Passchier 1991) also occur locally in close proximity.

Taken together, these observations indicate that, in terms of the relationship between the orientation of the flow plane and the sense of shear, the flow pattern within the central septum faithfully reflects the distribution of flow and vorticity in the East Athabasca mylonite triangle as a whole (Fig. 5). Moreover, if the three-dimensional pattern of ribbon fabrics reflects a coherent, contemporaneous set of slip systems, then we suggest that bulk flow within the central septum was essentially coaxial, with a generally prolate symmetry. The northeast–southwest direction of maximum extension would have been associated with significant shortening along two principal vectors, the one subhorizontal and the other steeply plunging.

Timing of deformation, metamorphism, and plutonism

Geochronological results (Table 1) demonstrate that the granulite to greenschist facies ribbon fabrics were all generated during the late Archean at ca. 2.62–2.60 Ga. However, it is also clear that a mid-Archean granulite facies, sinistral shearing event, which had entered its waning phase by 3.13 Ga, is preserved in the annealed mylonites and syntectonic intrusions in the eastern part of the Chipman batholith. To justify extending the interpretation of the dated samples (Table 1) to the rest of the East Athabasca mylonite triangle, it is appropriate here to explicitly examine the geological evidence for structural correlation of late Archean ribbon fabrics.

The principal geological observations in favour of the contemporaneity of late Archean granulite facies ribbon fabrics in the dextral and sinistral shear zones of the lower deck are (i) the similarity of metamorphic grade of the mylonitic fabrics in both shear zones, (ii) the conjugate disposition and spatial exclusivity of the dextral and sinistral shear fabrics, and (*iii*) the absence of kinematic reworking, i.e., reversal of shear-sense, or crosscucting. relations. We emphasize that these observations also apply within the central septum, where they pertain to the relationship between the strike-slip shearing and the southwestdirected extensionai shearing. This strongly suggests that granulite facies ribbon fabrics in the upper and lower decks are contemporaneous. The geological interpretation is well supported by our U–Pb data on magmatic zircon, as well as metamorphic monazite, from mylonites in both decks (Table 1).

The following observations suggest that amphibolite facies *displacement* of the penetratively mylonitized upper deck was contemporaneous with localization of amphibolite facies *deformation* within the lower deck. (*i*) The eastern part of the upper deck – lower deck contact is occupied by the Clut granite (Figs. 4, 5). The granite was mylonitized at amphibolite facies, during extensional sheanng, which accommodated southwest-directed displacement of the upper deck. (ii) The Clut granite extends westward into the lower deck as the Melby–Turnbull granites (Fig. 4). The latter were syntectonically emplaced exclusively within narrow dextral shear belts and variably mylonitized at amphibolite facies. (iii) Because the dextral shear plane in the Melby–Tumbull granites is subparallel to the dextral lateral ramp of the upper deck, the deformation within the lower deck is kinematically compatible, as well as materially continuous. with the Clut granite deformation fabric immediately beneath the upper deck (Figs. 4, 5). Furthemore, we recall that the southern branch of the Melby–Turnbull granites lies parallel to the upper deck – lower deck contact, but within the lower deck (Figs. 4, 5). In effect, during the amphibolite facies displacement. part of the lower deck was parautochthonously accreted to the upper deck. Finally, we note that this transition from penetrative granulite facies mylonitization to localized amphibolite facies deformation is analogous to, and contemporaneous (Table 1) with, similar localization of shearing along the northwestern side of the dextral shear zone, and within the Fehr granite in the sinistral shear zone, in the lower deck.

Syntectonic granites revisited

Although we have already considered the geological evidence in favour of a syntectonic interpretation of the Clut and McGillivray granites, we return to the topic to present another perspective on this question, which we were unable to address before considering the kinematic history of the upper deck. Because the Clut granite contains no trace of relict pyroxenes, it was emplaced under subgranulite metamorphic conditions, after granulite facies mylonitization. Therefore, during the amphibolite facies mylonitization of the Clut granite, the upper deck was displaced as a coherent, relatively stiff hanging wall above an extensional shear zone. We suggest that the spatial relationship between the thickest part of the Clut granite and the trailing edge of the upper deck (Figs. 4, 5) is more than a coincidence. The granite is located at the site of dilation, which would result from the top down to the southwest displacement of the stiff upper deck. In other words, the emplacement of the Clut granite was syntectonic with respect to the amphibolite facies displacement of the upper deck.

The McGillivray granite is also located adjacent to the trailing edge of the upper deck (Figs. 4, 5). It cuts and includes xenoliths and rafts of already mylonitic Clut granite. However, the mylonitic foliation within the Clut granite bifurcates around the McGillivray pluton (Fig. 4). This indicates that the McGillivray granite exerted a mechanical influence on the pattern of tectonic flow within its wall rocks prior to its final emplacement, presumably during its ascent. We suggest that the McGillivray granite was emplaced into a steeply oriented dilational zone, during the amphibolite facies displacement of the upper deck, while the Clut granite was actively deforming along a shallowly dipping shear plane. The present post-kinematic outcrop pattern simply reflects the final stages of its emplacement history.

Discussion and summary

Our study of the Snowbird tectonic zone in northernmost Saskatchewan has identified a kinematically complex, deep-crustal, multistage mylonitic structure, the East Athabasca mylonite triangle. We are well aware of the internal contradictions in the details of the emplacement sequence determined from geological observation and our late Archean syntectonic geochronological data. In view of the high metamorphic grade of the associated deformation, this is perhaps not surprising. Therefore, we shall confine our interpretation of the geochronological data to indicating that mylonitization occurred during the mid- (ca. 3.13 Ga) and late Archean (ca. 2.62–2.60 Ga).

The geological history of the East Athabasca mylonite triangle began with the deposition of semipelitic to pelitic sediments. The sediments were intruded at ca. 3.4–3.2 Ga by a tonalite batholith (Chipman), which dismembered an anorthosite–pyroxenite–gabbro–peridotite mafic complex during the early stages of its ascent. Although we favour the mafic complex hypothesis, we are as yet unable to eliminate the possibility that the mafic rocks might represent cumulate rocks, contemporaneous with, and perhaps related to, the tonalite. At ca. 3.13 Ga, the tonalite batholith was mylonitized under granulite facies metamorphic conditions (approximately 850°C, 1.0 GPa; M. Williams, unpublished data) and invaded by a syntectonic mafic dyke swarm during sinistral strike-slip shearing.

No geological record is preserved for the interval ca. 3.13–2.62 Ga. At ca. 2.62–2.60 Ga, the metasediments were invaded by two large mafic plutonic bodies and by a voluminous suite of syntectonic granitoids. The granitoids were emplaced during granulite facies mylonitization (850–1000°C, 1.0 GPa; M. Williams, unpublished data). During at least the later stages of mylonitization, the spatial distribution of deformation resulted in the localization of the emplacement of the younger granite plutons. The complex pattern of flow during mylonitization is that which one would predict adjacent to the apex of a stiff, elliptical, crustal-scale rheological heterogeneity (Athabasca lozenge) subjected to northeast–southwest extension, i.e., sinistral and dextral conjugate shear zones in the lower deck and an extensional shear occupying the upper deck. However, during the initial stages of late Archean deformation, flow was probably even more complex because the distribution of metamorphic mineral assemblages indicates that the upper deck was initially emplaced as an allochthonous thrust sheet. Although the hanging wall to the upper deck mylonites is not exposed, their shape, orientation, and location strongly suggest that they constitute the sole to the northeastern end of the Athabasca lozenge itself.

The total tectonic flow pattern is clearly multistage. The oldest preserved component of the flow is represented by the mid-Archean sinistral mylonites of the Chipman batholith. These could be the remains of a simple sinistral shear zone, or they could indicate the existence of a mid-Archean flow pattern very similar to that preserved by the younger ribbon mylonites, implying that the Athabasca lozenge was already a discrete entity by ca. 3.13 Ga. During the ca. 2.62–2.60 Ga time window, flow in the East Athabasca mylonite triangle evolved with cooling, from a pervasive high-temperature granulite facies regime to a more localized middle to lower amphibolite facies regime. This is manifested by the focusing of deformation within the Clut and Melby–Turnbull granites, as well as the 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. 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 the Straight–Grease fault at ca. 1.8 Ga.

The sequence of geological events that we have established for the East Athabasca mylonite triangle (Table 1) is of more than local significance. It is very similar to that described by Crocker et al. (1993) for the southeastern margin of the Athabasca lozenge, adjacent to the Virgin River shear zone, south of the Athabasca Basin (Fig. 1). They describe an early, layered mafic complex, associated with metasediments, engulfed by a mafic–tonalitic complex, which was subsequently deformed and intruded by mafic dykes and metamorphosed at granulite facies. Although they note a moderately northeasterly plunge to associated extension lineations, compatible with a tectonic setting on the underside of the Athabasca lozenge, they do not report high-temperature mylonites. Later lower amphibolite to greenschist facies mylonites (Virgin River shear zone) were cut by late discordant granitic dykes. However, in the absence of any syntectonic U–Pb zircon data, the dates that Crocker et al. (1993) attribute to specific geological events are speculative.

The absence of important Early Proterozoic regional deformation and metamorphism from the East Athabasca mylonite triangle indicates that this segment of the Snowbird tectonic zonewas not the site of an Early Proterozoic suture, or orogen (cf. Hoffman 1988). As we show elsewhere (Hanrner et al. 1994, S. Hanmer et al.[2]), the late Archean ribbon mylonites formed as vart of an intracontinental structure. far removed from the contemporaneous active plate margins. However, it is possible that the Chipman batholith might be part of a plutonic arc, perhaps emplaced into a sedimentary accretionary wedge, above a mid-Archean subduction zone during the early construction of the western Churchill continent.

When considered at the scale of the East Athabasca mvlonite zone and its imrnediately adjacent wall rocks, the geological history we have outlined here fails to address many important questions, e.g., (i) the tectonic context of the mid- and late Archean events and the cause of the granulite facies metamorphism, (ii) an explanation for the absence of significant Early Proterozoic tectonic activity, given the recent data on contemporaneous crustal formation in the Alberta basement (e.g., Ross et al. 1991) and the Trans-Hudson orogen (e.g., Lewry and Stauffer 1990), (iii) the timing and mechanisms of uplift of the high-pressure granulites, (iv) the relationship of plutonic and tectonometamorphic events within the mylonites to those within their wall rocks, both locally and at the scale of the western Churchill Province as a whole. Lacking the space to tackle them here, we have chosen to specifically address some of these questions in a companion contribution (S. Hanmer et al.[2]) and forthcoming papers, to which we refer the interested reader.

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