Journal of Structural Geology, Vol. 14, No. 7, pp. 757 to 773, 1992
Printed in Great Britain

0191-8141/92 $05.00+0.00

Great Slave Lake shear zone, NW Canada : mylonitic record of Early Proterozoic continental convergence, collision and indentation

SIMON HANMER

Continental Geoscience Division, Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario, Canada K1A 0E8

SAMUEL BOWRING

Department of Earth, Atmospheric and Planetary Science, Massachusetts Institute of Technology, Cambridge, MA 02139, U.S.A.

and

OTTO VAN BREEMEN and RANDALL PARRISH

Continental Geoscience Division, Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario, Canada K1A 0E8

(*Received* 12 *March* 1991; *accepted in revised form* 3 *February* 1992)

Abstract—Great Slave Lake shear zone, in the NW Canadian Shield, is an excellent example of the kinematic, magmatic and thermal evolution of a crustal-scale shear zone associated with a collisional continental boundary. It is a 25 km wide corridor of granulite to lower greenschist facies mylonites and cataclastic fault rocks developed in the deep-seated parts of an Early Proterozoic (2.0–1.9 Ga) magmatic arc, constructed on the upper (Rae) plate at the contact between the Archean Slave and Rae continents. The rocks of Great Slave Lake shear zone, the Thelon magmatic arc and the Taltson magmatic zone are all components of the same magmatic arc, but their geological histories reflect different aspects of the continental interaction.

INTRODUCTION

IN THE NW Canadian Shield (Figs. 1 and 2), the boundary between the southeastern margin of the Archean Slave Province and the western edge of the Archean Rae Province is marked by a discrete corridor of granulite to lower greenschist facies mylonitic rocks (Reinhardt 1969, Hanmer 1988a,b), recently identified as Great Slave Lake shear zone (Hanmer & Lucas 1985, Hanmer & Connelly 1986, Hanmer & Needham 1988). The shear zone is up to 25 km wide and has been traced in outcrop for over 200 km along strike. It is a vertical, strike-slip structure which formed within part of an active Early Proterozoic magmatic arc of magnetite-series granitoids at the leading edge of the Rae continent during oblique convergence and collision with the Slave continent (Fig. 2) (Hoffman 1987, 1988a). The 2.051–1.92 Ga granitoid protoliths of Great Slave Lake shear zone (van Breemen *et al.* 1990, Bowring & Hanmer in preparation) are continuous with the granitoids of the Thelon magmatic arc (2.0–1.92 Ga, Henderson *et al.* 1982, Thompson & Ashton 1984, Henderson & Macfie 1985, Hoffman 1987) and those of the Taltson magmatic zone (1.986–1.906 Ga, Bostock *et al.* 1987). Collectively they are associated with one of the most pronounced positive magnetic anomalies in the western Canadian Shield (Geological Survey of Canada 1987). According to preliminary U–Pb zircon dating in the foredeep developed in front of the Bear Creek Hills foreland thrust–fold belt on the east margin of the Slave continent (Fig. 2), collision began *ca* 1.97 Ga (Tirrul 1985, Bowring & Grotzinger 1989, Tirrul & Grotzinger 1990).

Removal of the 70–125 km of brittle right-lateral strike-slip movement on the post 1.87–1.86 Ga McDonald fault (Fig. 1) (Hoffman *et al.* 1977, Bowring *et al.* 1984, Henderson & Macfie 1985) brings the northeastern part of Great Slave Lake shear zone into close proximity with the southeastern corner of the Slave Province. In previously elaborated tectonic models, the Slave Province has been viewed as a continental indentor. The indentor models group naturally into two genetically unrelated types: (i) those which only consider the phase of indentation that directly followed collision between the Rae and Slave continents and was accommodated by high-temperature plastic deformation (Hoffman 1987), and (ii) those which only examine later movements associated with the late, brittle McDonald–Bathurst fault system (Fig. 1) (Gibb & Thomas 1977, Gibb 1978, Gibb *et al.* 1983, Henderson *et al.* in press). In the present contribution, we are only concerned with the first type of tectonic model, according to which the southeastern corner of the Slave continent separates the N–S-trending frontal face of the indentor (Thelon tectonic zone, Thompson & Henderson 1983) from the NE–SW-trending transform margin (Great Slave Lake shear zone). Accordingly, given the kinematic contrast between the two zones predicted by the model, it should be possible to document the history of convergence,

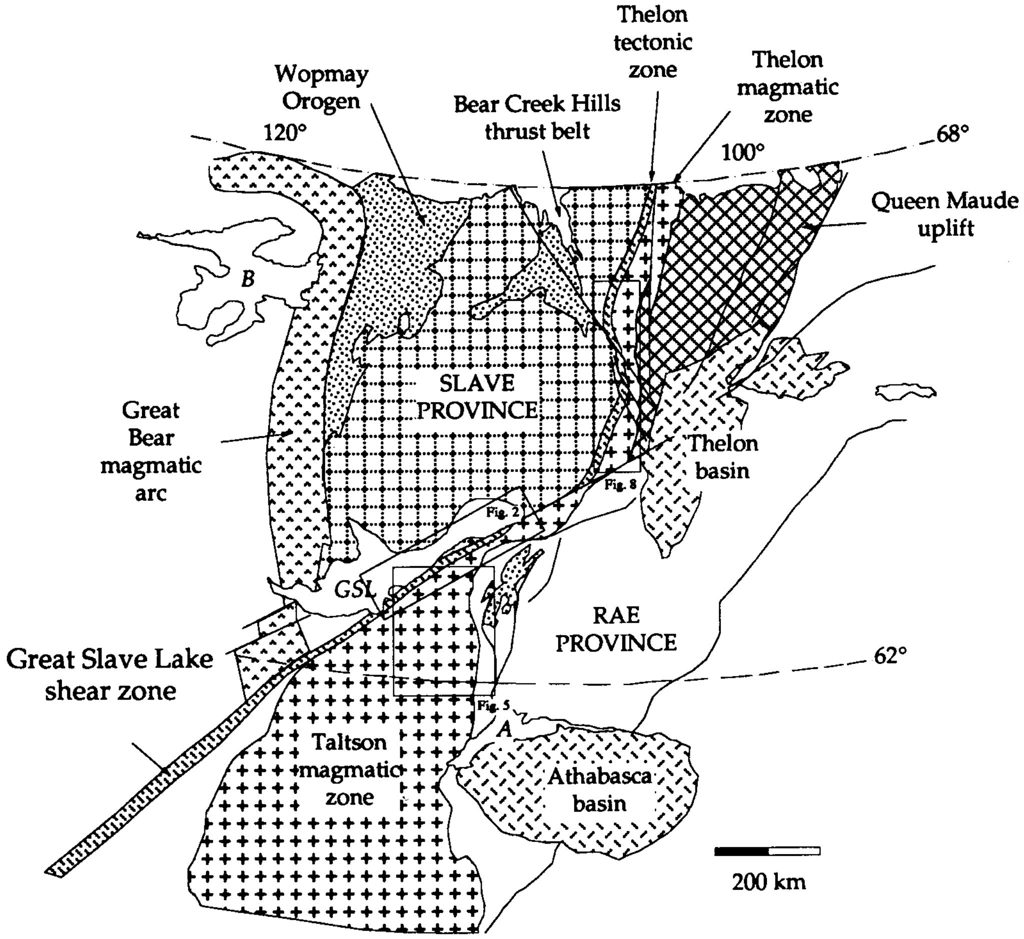

Fig. 1. Selected tectonic elements of the NW Canadian Shield and structures referred to in the text. The locations of other text figures are indicated by boxes. GSL, Great Slave Lake; B, Great Bear Lake; A, Lake Athabasca (see Hoffman 1989).

collision and indentation in the mylonites developed within the deeply excavated roots of the magmatic arc and associated, post-collisional granites. The progressive evolution of the transcurrent–transform part of Great Slave Lake shear zone, furthest removed from the frontal face of the indentor, has been described elsewhere (Hanmer 1988a). From field mapping, kinematic study and geothermobarometry, this Southwest segment records the progressive uplift, cooling and narrowing of an active dextral strike-slip shear zone with time (Hanmer 1987, 1988a,b) and its progressive passage through the plastic–brittle transition (Hanmer 1989). More recent mapping, closer to the eastern margin (frontal face) of the Slave indentor (Hanmer & Needham 1988) and U–Pb zircon dating (van Breemen *et al.* 1990, Bowring and Hanmer in preparation), now allows us to document the tectonic evolution of Great Slave Lake shear zone as an Early Proterozoic intra-arc transform fault and to compare it with the tectonothermal history of other parts of the Slave–Rae boundary and the adjacent Rae Province.

GREAT SLAVE LAKE SHEAR ZONE

In what follows, it is sufficient to briefly summarize the detailed descriptions of Great Slave Lake shear zone which have been presented elsewhere (Hanmer & Lucas 1985, Hanmer & Connelly 1986, Hanmer & Needham 1988). The mylonites of Great Slave Lake shear zone were derived by intense deformation of a mixed protolith of hornblende–biotite, megacrystic to equigranular, magnetite-bearing granite and granodiorite (referred to for simplicity as granite; Fig. 2). The metamorphic grade at which the mylonitization took place can be estimated from the syntectonic mineral assemblages in paragneisses which are intimately associated with the granitic mylonites. They occur as small xenoliths, rafts and laterally continuous screens up to 50 km long in the hornblende–biotite granites.

On the northwest side of Great Slave Lake shear zone, the wall rocks are predominantly poorly foliated, two-mica, equigranular Archean leucogranites (2.55 Ga), with rafts of open folded supracrustal rocks. All of the components of the wall rock are cut across by a swarm of closely spaced, upright, mafic dykes, oriented parallel to Great Slave Lake shear zone (Fig. 4). Similar, but more widely separated dykes (post-2.2–2.1 Ga) are known throughout the East Arm of Great Slave Lake (Hoffman *et al.* 1977, Bowring *et al.* 1984). To the southeast, the mylonites are flanked by thoroughly mylonitized Early Proterozoic cordierite–sillimanite–garnet granites (1.96 Ga), and poorly foliated hornblende–biotite Snowdrift granite (Fig. 2). Strain

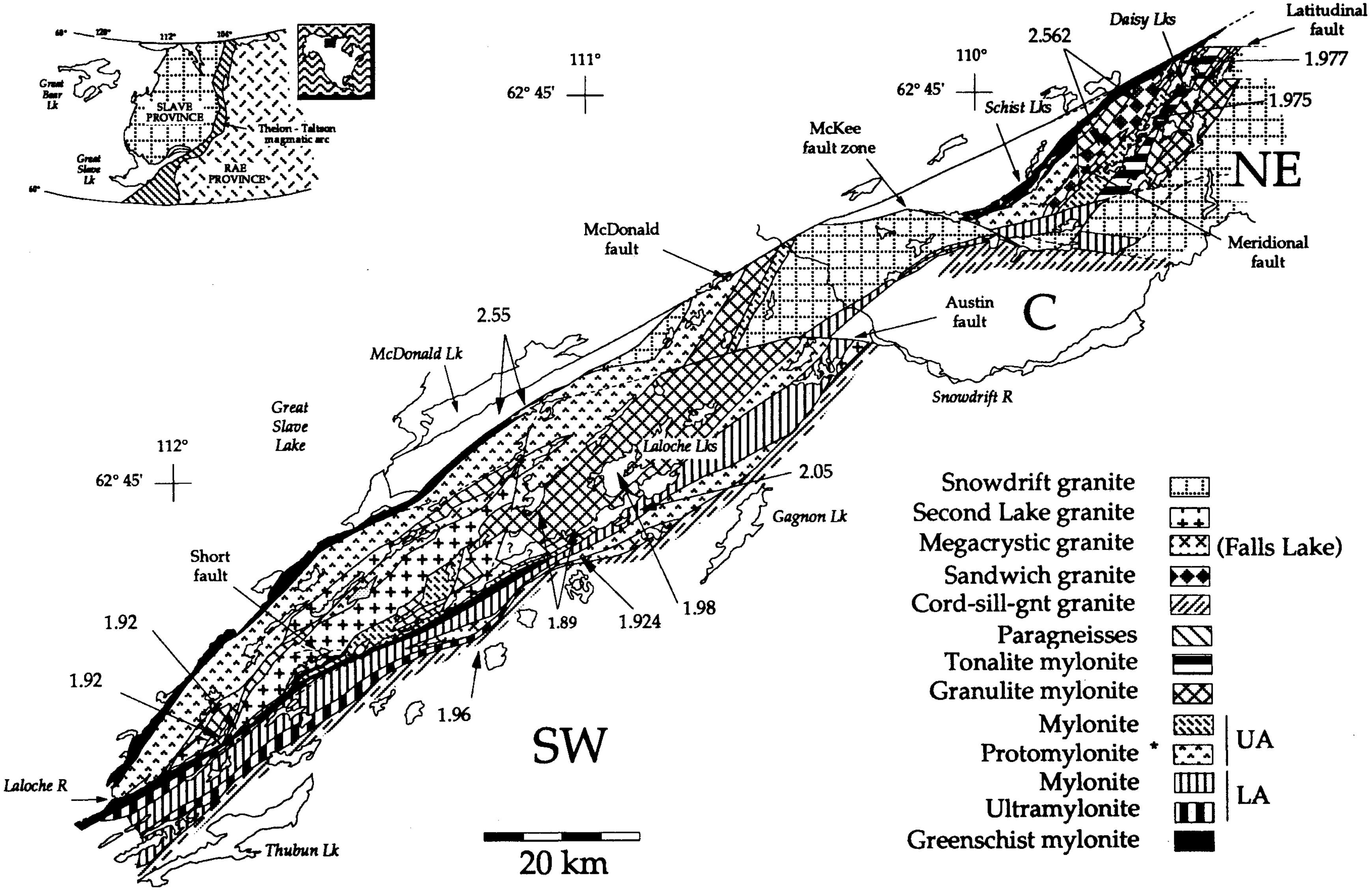

Fig. 2. Structural geological map units of Great Slave Lake shear zone. Principal geographical names, fault names and segments (Southwest, Central and Northeast; see Fig. 3) are indicated. The locations and dates (Ga) of the principal U–Pb (zircon) samples referred to in the text are identified. Granitic tectonites are labeled as mylonite, tectonites of supracrustal origin are represented as paragneisses. Geochronology for the Northeast segment is from van Breemen *et al.* (1990), while that for the Southwest segment will be presented elsewhere (Bowring & Hanmer in preparation). UA: upper amphibolite facies (*except for those located in the lower amphibolite facies belt—see Fig. 3—of the SW segment). LA: lower amphibolite facies. Inset: location of the Slave Province, Rae Province and the Thelon–Taltson magmatic arc in NW Canada.

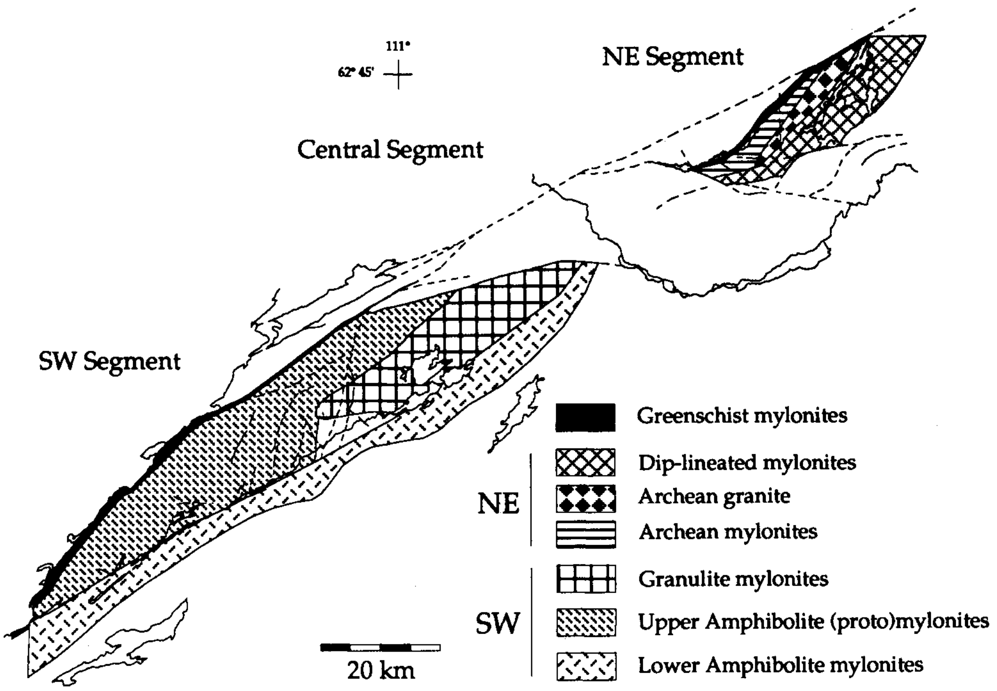

Fig. 3. Segments and constituent mylonite belts of Great Slave Lake shear zone. SW and NE in the legend are indications of the segments where the main outcrops of the principal mylonite types occur.

intensity decreases rapidly away from the shear zone such that, within 1 km the wall rock comprises an highly contorted mature metatexite to diatexite (Mehnert 1968, Brown 1973).

The shear zone is divided into mylonitic Northeast and Southwest segments and a Central segment principally composed of late to post-tectonic granite (Fig. 3). The Northeast and Southwest segments are further subdivided into longitudinal belts of mylonite, according to their metamorphic grade, relative age of mylonitization and kinematic significance. A network of greenschist facies mylonites and cataclasites, common to the Northeast and Southwest segments, reworks the older, higher grade mylonite belts. The longitudinal mylonite belts are cut by a set of discrete, mylonitic or cataclastic, transverse fault zones (Fig. 2). All the belts and fault zones are upright. For the most part, the mylonitic foliation and layering are vertical, the extension lineation is subhorizontal and shear criteria indicate a dextral sense of strike-slip displacement. However, important kinematically significant exceptions do occur (Fig. 4; see below).

The array of greenschist facies mylonites is the common structural link between the Northeast and Southwest segments of the shear zone. Consequently, we shall describe the high-grade mylonites of each of the shear zone segments before considering the greenschist facies mylonites of the shear zone as a whole. After summarizing the geological evidence for the relative timing of plutonism, metamorphism and deformation and the kinematics of flow, we shall present a summary of our geochronological data and discuss their implications for the evolution of Great Slave Lake shear zone and the Slave–Rae boundary.

LONGITUDINAL STRUCTURE

Southwest segment

Hanmer and coworkers identified a closely packed bundle of three belts of high-grade, generally strike-lineated mylonites in a segment lying to the southwest of the Austin fault (Figs. 2 and 3; for detailed descriptions see Hanmer & Lucas 1985, Hanmer & Connelly 1986). The upper and lower amphibolite facies mylonites form two belts which lie on either side of a belt of older, granulite facies mylonites.

Granoblastic orthopyroxene–garnet mylonites to protomylonites, derived from a megacrystic granite protolith (1.98 Ga), and migmatitic, porphyroclastic and mylonitic paragneisses, form a lens-like belt, 8 km wide by 50 km long (*granulite mylonites* in Figs. 2 and 3). The granoblastic mylonites and protomylonites in the southeastern half of the belt are disrupted by folding, boudinage and heterogeneous anastomosing deformation zones. Blocks of mylonite, up to 1 km across, carry an internal NW-striking layering which is transposed at their boundaries by narrow (10–100 m) NE-striking mylonite zones.

The granulite facies mylonites are flanked to the northwest by a tapered belt, up to 10 km wide, of finely porphyroclastic (1–2 mm), garnet–sillimanite, ribbon protomylonites to mylonites (*upper amphibolite protomylonites* and *mylonites* in Figs. 2 and 3). These were derived from an apparently Archean (see below) hornblende–biotite granite protolith, with very large screens of migmatitic paragneiss. To the southeast, the granulite facies mylonites are flanked by a 5 km wide belt of mylonites and ultramylonites (*lower amphibolite*

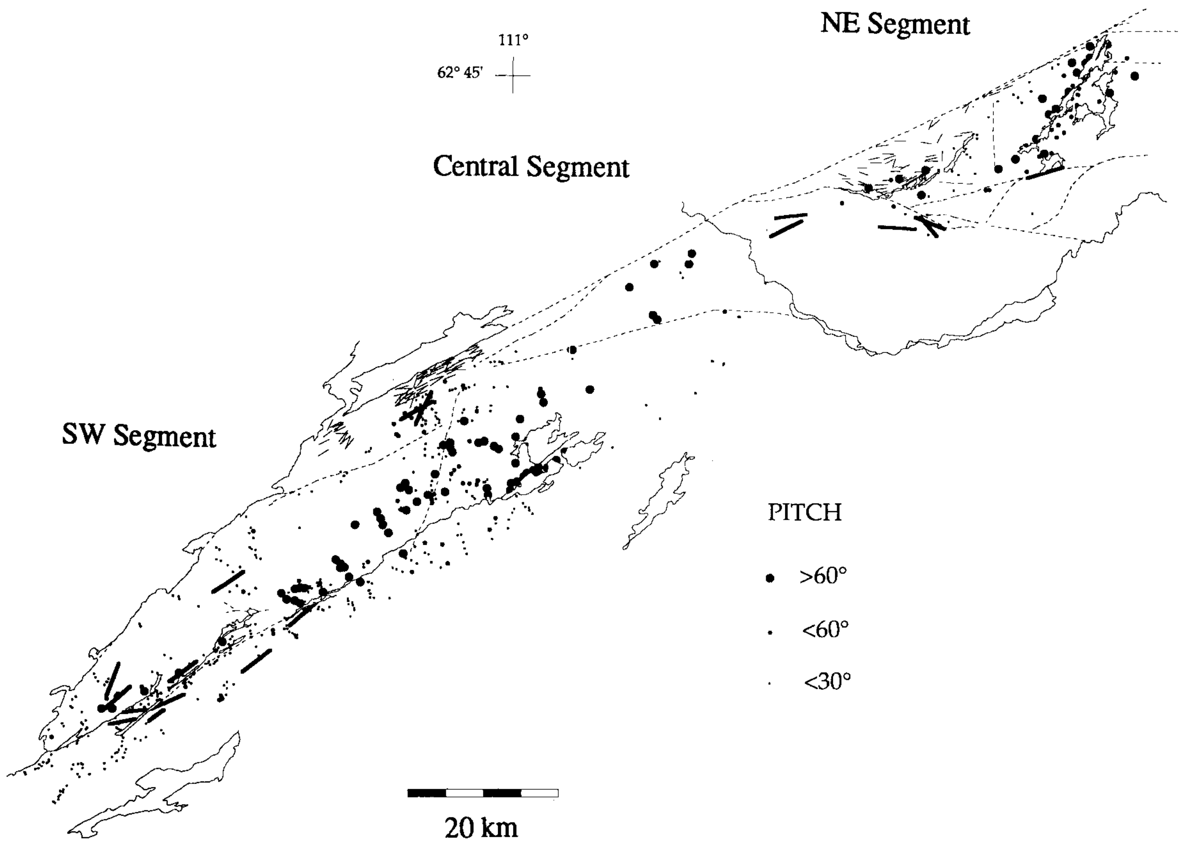

Fig. 4. Orientations and distributions of finite extension lineations (dots), late syntectonic mafic dykes within Great Slave Lake shear zone (bold lines) and closely spaced mafic dykes in the northwest wall rock (fine lines). The lineation data are presented as pitches. Note the mixed steep and shallow lineation attitudes in Laloche Lakes–Laloche River corridor. Note also the steep lineations in the Central segment and the southeast side of the Northeast segment.

protomylonites, *mylonites* and *ultramylonites* in Figs. 2 and 3), derived from Early Proterozoic (2.05–1.924 Ga) hornblende–biotite granite and rare sillimanite–muscovite paragneiss.

Northeast segment

Hanmer & Needham (1988) identified two belts of high-grade mylonite in a segment bounded by the McKee and McDonald faults (Figs. 2 and 3). A belt of strike-lineated, upper amphibolite facies protomylonites and mylonites is separated from a belt of dip-lineated, upper amphibolite to granulite facies mylonite to the southeast by the intervening Archean (2.56 Ga) Sandwich granite. The isotropic to poorly foliated, hornblende-biotite Sandwich granite intrudes the strike-lineated mylonitic rocks, whereas it represents protolith material to the dip-lineated mylonites (see below).

Strike-lineated high-grade mylonites. Ribbon protomylonites to mylonites, derived from hornblende–biotite granite, and migmatitic, mylonitic paragneisses, form a 4 km wide upright belt (*upper amphibolite protomylonites* and *paragneiss* in Fig. 2; *Archean mylonites* in Fig. 3). The mylonites carry a vertical foliation with a subhorizontal extension lineation. To the northeast, the belt is progressively cut out by greenschist facies mylonites (see below). This is reflected by a northeastward retrograde textural and metamorphic transition, wherein protomylonites are reworked to mylonites and coarsely porphyroclastic paragneiss is transformed into garnet–biotite schist. Similar textural and metamorphic transitions also occur across strike toward the northeast.

Dip-lineated high-grade mylonites. Hornblende–biotite granitic to clinopyroxene tonalitic ribbon mylonites, porphyroclastic and mylonitic paragneisses form an upright belt, 9 km wide, southeast of the strike-lineated protomylonites and mylonites (*granulite* and *tonalite mylonites*, *upper amphibolite facies mylonite* and *paragneiss* in Fig. 2; *dip-lineated mylonites* in Fig. 3). The granitic mylonites were derived from an Early Proterozoic (1.975 Ga) megacrystic protolith. The belt carries a vertical foliation with a steeply plunging, approximately dip-parallel extension lineation (Fig. 4). Along its present southeastern margin the belt has been intruded and disrupted by the Snowdrift granite (Fig. 2).

In the southeastern part of the belt, the mylonites and their protoliths are orthopyroxene granulites. Much of the orthopyroxene is partially replaced by hornblende, suggesting that the granulite facies assemblages were being replaced by upper amphibolite facies assemblages during deformation. To the northwest, the granulite and clinopyroxene tonalitic mylonites are flanked by amphibolite facies mylonites which outcrop along the chain of the Daisy Lakes (*upper amphibolite mylonite* in Fig. 2).

Central segment

The eastern part of the Central segment is underlain by the Snowdrift granite. Between the Snowdrift granite and the McDonald fault (Fig. 2) is a complex area underlain by disrupted, annealed mylonites, finely porphyroclastic protomylonites and ribbon mylonites, similar to those of the granulite and upper amphibolite facies belts of the Southwest segment. North of the Austin fault (Fig. 2), they strike at 030°, oblique to the regional 060° trend of Great Slave Lake shear zone. As in the case of the Northwest segment, the high-grade mylonitic rocks were intruded by sheets of cross-cutting Snowdrift granite, itself cut by NE–SW-trending, 0.5-1 km wide mylonite zones of uncertain metamorphic grade (Fig. 2).

GREENSCHIST FACIES MYLONITES

Lower greenschist facies chlorite-bearing mylonites and ultramylonites form an upright, 1–1.5 km wide belt along the shore of Great Slave Lake (Figs. 2 and 3). The belt extends inland southeast of McDonald Lake. The mylonites are derived from the two-mica leucogranites, paragneisses and mafic dykes of the northwestern wall rock, as well as from the reworking of the upper amphibolite facies protomylonites and mylonites to the southeast. At longitude *ca* 111°, the mylonites are cut out along strike by a 100 m wide zone of brecciation and cataclasis, itself truncated by the McDonald fault northeast of McDonald Lake (Fig. 2). Lower greenschist facies mylonites and ultramylonites re-appear in the Northeast segment as a belt, up to 2 km wide, which extends along the chain of the Schist Lakes (Figs. 2 and 3). In this segment, the belt of greenschist mylonites is truncated in the northeast by the McDonald fault and is deflected in the southwest by the McKee fault zone. Northeast of the Meridional fault (Fig. 2), the chlorite-bearing mylonites are separated from the strike-lineated, high-grade protomylonites of the Northeast segment by a 1 km wide sub-belt of upper greenschist facies, garnet–biotite ribbon mylonites derived from a mixed granitic and paragneiss protolith. As in the case of the volumetrically predominant lower greenschist mylonites, the garnet–biotite mylonites truncate and rework the strike-lineated, upper amphibolite facies protomylonites and mylonites to the southeast.

The upright belt of greenschist facies mylonites along the Laloche River (Figs. 2 and 3) is essentially identical to the lower greenschist facies mylonites and ultramylonites of the northwest margin of Great Slave Lake shear zone, just described. It is derived by the reworking and truncation of the flanking higher grade mylonites of the Southwest segment. Note, however, that the Laloche River belt does not continue to the northeast beyond the latitude of Laloche Lakes (Fig. 2).

TRANSVERSE STRUCTURE

The NE–SW grain of the Great Slave Lake shear zone is cross-cut by several approximately E–W-oriented faults of variably mylonitic character and importance. They are members of a kinematically coherent set of structures which is also detectable in the regional aeromagnetic field (Geological Survey of Canada 1981, 1984).

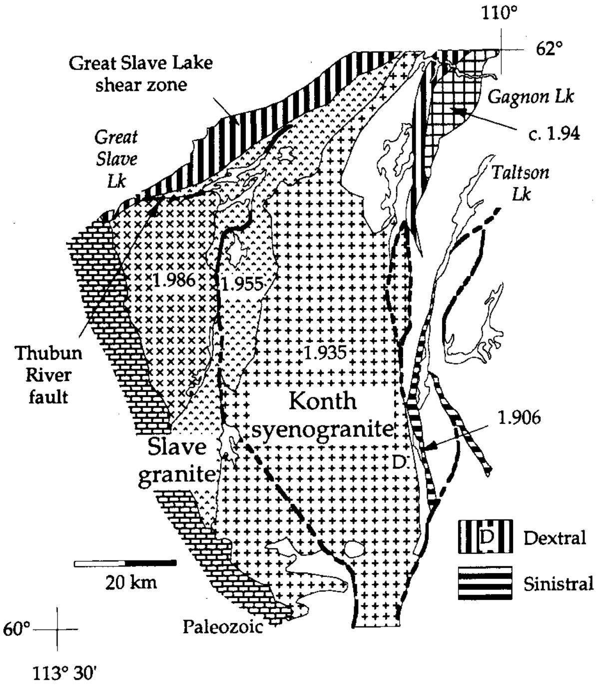

Fig. 5. Schematic geology and geochronology of the Taltson magmatic zone. Narrow mylonite zones are shown as thick black lines. Note the important sinistral and dextral mylonite zones along the east side of the granitoids (shaded) of the Taltson magmatic zone, as well as the dextral (D) mylonites in the south. U–Pb zircon dates for the granitoids are shown. After Bostock & Loveridge (1988).

The McKee fault zone truncates and deflects all of the mylonitic belts of the Northeast segment (Fig. 2). It is a narrow, E–W-trending, upright zone of anastomosing discrete fault strands, cataclasites and mylonites, initially developed as a bundle of anastomosing zones of biotite-bearing, granitic ribbon mylonite. Individual mylonite zones are upright, up to several hundreds of metres wide and carry a vertical foliation with a subhorizontal extension lineation. The mylonites and parts of the non-mylonitic wall rock are reworked by extensive zones of penetratively developed cataclasite, up to 100 m thick, which are themselves cut by dilational quartz stockworks.

Other faults in this set include the Austin, Latitudinal and Short faults (Fig. 2). The Austin fault has not been observed in outcrop, but is inferred from the regional mapping, as well as its very strong expression in the regional aeromagnetic field. The Latitudinal fault is very similar to the McKee fault zone, but is of lesser importance and scale. The strands of the Short fault, while predominantly lined with thin (1 m) zones cataclasite and fault breccia, are locally mylonitic. In the latter case, a subhorizontal lineation and rare dextral asymmetrical extensional shear bands are present. Similar dextral, strike-slip mylonites are present along the Thubun River fault zone, in the southeastern wall rock (Fig. 5) (Bostock 1988). A companion structure, oriented parallel to,

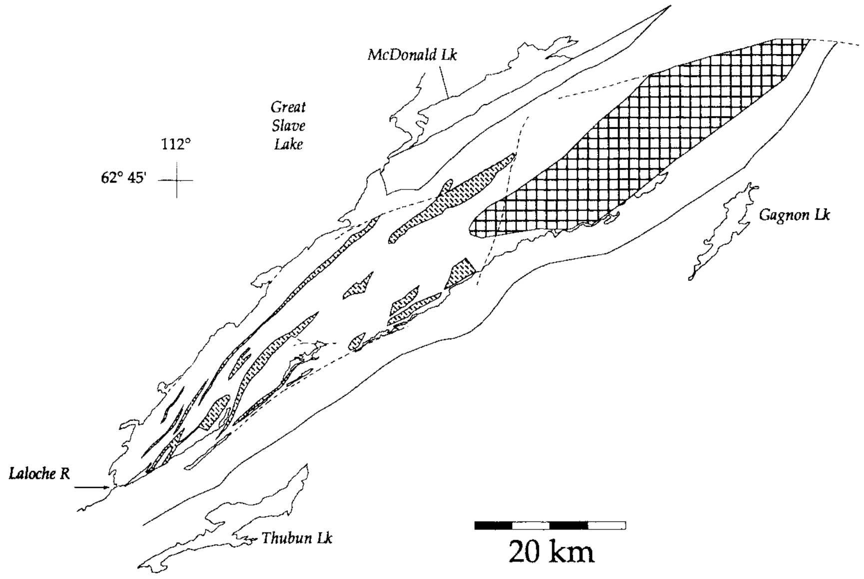

Fig. 6. Southwest segment of Great Slave Lake shear zone to illustrate the distribution of important screens of mixed paragneiss assemblage. The very abundant paragneiss screens in the granulite facies belt (Laloche Lakes) is schematically represented. Note the absence of such screens from the lower amphibolite facies mylonite belt to the southeast.

and located 25 km to the south of, the Thubun River fault, is apparent in the regional aeromagnetic field (Geological Survey of Canada 1984), but is obscured by Phanerozic deposits.

GEOMETRY AND KINEMATICS

It suffices here to briefly summarize the geometry and kinematics of Great Slave Lake shear zone. Graphical documentation and analysis of foliation and lineation orientations and of the kinematic significance of a diverse assemblage of shear-sense indicators has been presented in detail elsewhere (Hanmer 1984, 1986, 1988b, 1990, Hanmer & Lucas 1985, Hanmer & Connelly 1986, Hanmer & Needham 1988, Hanmer & Passchier 1991).

Foliations and lineations

Mylonitic foliation and tectonic banding in Great Slave Lake shear zone are generally steeply dipping or vertical. Except for the dip-lineated mylonite belt of the Northeast segment (Fig. 3) and the strip of upper amphibolite facies mylonite in the Central segment (Fig. 2), the extension lineation generally porpoises about the horizontal at all metamorphic grades (Fig. 4). The principal exception to both of these generalizations is the disrupted granulite facies mylonite at Laloche Lakes (see above) and a contiguous, narrow corridor extending along the Laloche River toward Great Slave Lake (Fig. 4). At Laloche Lakes, and on either side of the Laloche River, foliation attitudes are highly variable with respect to both dip and strike. Although not as pronounced as in the disrupted granulite mylonites already described, the heterogeneity of the deformation along the Laloche River is manifested as 50–100 m wide nests of folds which deform the mylonitic foliation and form pods wrapped around by the regional mylonitic foliation. The three-dimensional geometry of the anastomosing mylonitic foliation enclosing the stiffer pods gives rise to the observed variation in foliation attitudes (e.g. Bell 1981, 1985). This heterogeneous flow pattern is also spatially associated with strong variation in the plunge and azimuth of the associated extension lineation. Careful observation shows that the local extension lineation can be either parallel, or normal, to the local shear direction, as determined from structures such as asymmetrical extensional shear bands. Such complex flow would be expected where an aggregate of relatively stiff pods is deformed and the pods are constrained to move past each other in a locally transpressive (e.g. Sanderson & Marchini 1984) three-dimensional flow.

It is important to emphasize that the heterogeneous flow in the Southwest segment is spatially, rather than temporally, controlled because it occurs in mylonites of the granulite facies and upper amphibolite facies belts, as well as in the greenschist mylonites along the Laloche River (Figs. 3 and 4). Furthermore, it is apparently rheologically controlled because the zone of heterogeneous flow is wider in the drier, granulite facies mylonites at Laloche Lakes than in the more hydrous, lower grade assemblages along the Laloche River. Hanmer (1988a) noted that the zone of heterogeneous flow is located at the boundary between two lithologically, distinct protoliths; granites with important paragneiss screens to the northwest vs granite with few inclusions to the southeast (Fig. 6). He suggested that this boundary acted as a rheological interface within the hornblende–biotite granites and that the heterogeneous flow is the result of strain incompatibility across that interface. The importance of these observations will become apparent when we discuss the relative timing and U–Pb dating of the various mylonite belts of Great Slave Lake shear zone.

Kinematics

Deduced deformation paths in the mylonites of Great Slave Lake shear zone approximated to a general noncoaxial flow (see Hanmer 1984, 1990, Hanmer & Passchier 1991). Except for the corridor of heterogeneous flow along the Laloche River (see above), progressive deformation everywhere resulted in the rotation of the direction of maximum finite extension into the flow direction. Accordingly, with the exception of the dip-lineated mylonites to the northeast, Great Slave Lake shear zone is a strike-slip structure. Throughout most of the Southwest segment and the strike-lineated mylonites of the Northeast segment, shear-sense indicators all indicate a dextral sense of shear. Furthermore, the occurrence of (i) oblate deformation sub-fabrics, such as transposed banding and chocolate-tablet boudinage, (ii) asymmetrical extensional shear band fabrics and (iii) in-plane wings on rotated stiff inclusions (Hanmer 1990), indicates that flow was of a general noncoaxial type, rather than simple shear (Hanmer & Passchier 1991).

However, the dip-lineated mylonites of the Northeast segment are kinematically equivocal. The southeastward transition from upper amphibolite to granulite facies might most easily be explained by east-side-up non-coaxial flow along the extension lineation. However, evidence for both east-side-up and west-side-up flow has been observed. This is strongly suggestive of general non-coaxial flow in a transpressive deformation regime (Hanmer & Needham 1988). Moreover, local evidence for dextral strike-slip flow, normal to the down-dip extension lineation, has also been found; a 50 m wide strip of strike-lineated, dextral ribbon mylonite passes continuously along the west shore of north Daisy Lake (Fig. 2) into a poorly foliated, biotite megacrystic granite protolith (1.977 Ga). Contact relations with the dip-lineated mylonites are not exposed. However, the restriction of the strike-lineated mylonite to this thin granite body led van Breemen *et al.* (1990) to suggest that the sheet of granite had been intruded into the already dip-lineated mylonites, just prior to strike-slip deformation. The rectilinear, undisturbed nature of the dominant extension lineation throughout the dip-lineated mylonite belt suggests that this transcurrent component of the deformation was relatively weak and late.

RELATIVE TIMING

Longitudinal mylonites

A relative chronology of mylonitization within the Southwest and Northeast segments of Great Slave Lake shear zone can be constructed from field observations and tested by geochronological studies. The relative chronology is based principally upon cross-cutting and reworking relationships, but can be enhanced by taking account of the influence of the gross rheological contrasts within the deforming protolith on the nature of the flow (see above; Fig. 6).

In the Southwest segment, mylonites of the granulite facies belt are cross-cut by an array of veins of generally poorly foliated, white leucogranite. Some of these veins are very late with respect to mylonitization (1.89 Ga), but are spatially confined to the granulite mylonite belt. However, many other similar veins are visibly transposed as one passes progressively into the flanking lower amphibolite and upper amphibolite facies belts (Fig. 3), both of which are therefore younger than the granulite mylonites (Hanmer & Lucas 1985). The amphibolite facies mylonite belts are everywhere separated by either higher or lower grade mylonites (Fig. 3). Nevertheless, Hanmer & Connelly (1986) considered that the lower amphibolite mylonites are younger than those found at upper amphibolite because: (i) they represent lower temperature deformation; (ii) they contain reworked inclusions of migmatitic mixed paragneiss, very similar to that which occurs in the higher grade mylonites; and (iii) they are visibly derived, in part, by the reworking of the upper amphibolite facies wall rocks to the southeast. Accordingly the granulite, upper and lower amphibolite facies belts preserve progressively younger mylonites. In the Northeast segment, the relative timing of the strike-lineated and the dip-lineated mylonites is determined from their contact relations with the intervening Sandwich granite (see above; Figs. 2 and 3; Hanmer and Needham 1988).

Greenschist mylonites and transverse faults

At its western end, the trend of the McKee fault zone turns to a lower azimuth before it is truncated by the McDonald fault (Fig. 2). The McKee fault zone itself cuts out the belt of chlorite-bearing mylonites which runs along the Schist Lakes. The presence of early biotite-bearing mylonites in the McKee fault zone suggests that its mylonitic history commenced prior to the deformation represented by the belt of lower greenschist facies chlorite-bearing mylonites which it cuts. However, its later cataclastic and brittle activity continued after plastic deformation in the truncated chlorite-bearing mylonites had ceased. As already noted, the older greenschist facies mylonites in the Northeast segment formed under upper greenschist facies (garnet–biotite) conditions, and the chlorite-bearing mylonites in the Southwest segment pass along strike into cataclasites. These similarities between the longitudinal greenschist mylonites and those of the McKee fault zone led Hanmer & Needham (1988) to suggest that the McKee fault zone is a branch of the main greenschist facies mylonite belt, in which the early phase of the greenschist mylonitization history is preserved.

In addition to the McKee fault zone, the other transverse faults also cut and offset the high-grade mylonites of Great Slave Lake shear zone (Fig. 2). However, the Austin fault, the Thubun River fault, its companion structure and the Short fault do not offset longitudinal mylonite belts formed at greenschist facies (Fig. 2)

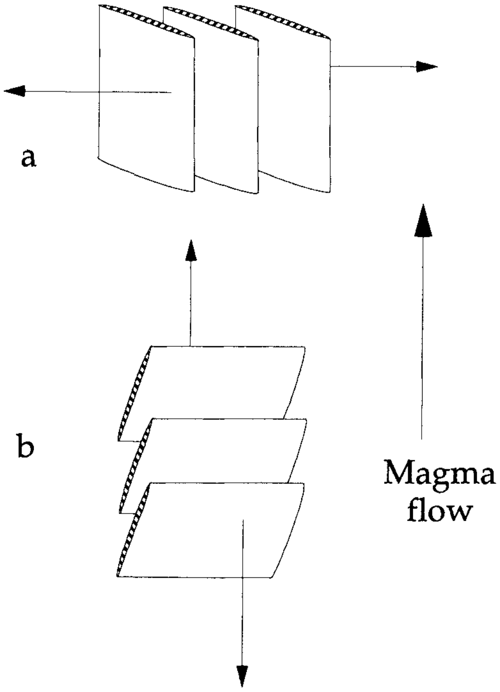

Fig. 7. Schematic illustration of the effect of the orientation of arrays of en échelon initial Riedel fractures (e.g. Logan 1979) associated with potential vertical magma flow. (a) The long dimensions of the initial en échelon components in a strike-slip array are aligned so as to favour the emplacement of magma, especially if such arrays are vertically interconnected. (b) The en échelon components in a dip-slip fracture array act as baffles to the emplacement of a vertically rising magma and do not enhance magma access. Inspired by Sibson *et al.* (1988).

(Geological Survey of Canada 1981, 1984). Taken as a whole, displacement along the array of transverse faults both pre-dates and post-dates activity in the greenschist mylonite belts. Accordingly, the longitudinal greenschist mylonites and the transverse faults form a kinematically coherent, linked network of dextral, strike-slip shear zones (e.g. Woodcock & Fisher 1986).

Syntectonic mafic dykes

The upper and lower amphibolite facies mylonite belts of the Southwest segment (Fig. 3) contain amphibolite bodies 1–100 m long. They occur as swarms of stubby pull-aparts, flattened isolated lenses or long, parallel-sided bands aligned in the mylonitic foliation of the host rock. The latter type may be discrete or pervasively fractured, dilated and sealed with pegmatitic fill. In general the origins of the first two types of body are equivocal. However, the third type is probably a set of mafic dykes. This can be demonstrated in a number of cases where apophyses, branching and cross-cutting relations are preserved (Hanmer & Lucas 1985). It is possible that the spectrum of observed inclusions reflects the progressive mechanical disruption of a swarm of mafic dykes, emplaced into the shear zone during ongoing deformation. In the particularly well exposed part of the shear zone between Great Slave and Thubun Lakes, Hanmer & Connelly (1986) determined that the upper and lower amphibolite facies mylonite belts of the Southwest segment contained approximately equal numbers of recognizable, late-syntectonic members of this dyke set (Fig. 4). The absence of post-tectonic dykes in the older, upper amphibolite facies belt suggests a causal relationship between active shearing and the locus and mechanism of dyke emplacement. Any proposed model must also account for the (i) absence of members of this dyke set from the wall rocks on the southeast side of the shear zone, as well as from the dip-lineated mylonites of the Northeast segment and (ii) their emplacement in the Snowdrift granite adjacent to, and concordant with, the McKee fault zone (Fig. 4). According to one such model (Sibson *et al.* 1988), vertical fluid flow is enhanced in a strike-slip environment (Fig. 7a), but impeded in a dip-slip regime (Fig. 7a). Applying this reasoning to dyke emplacement, one can effectively account for the distribution of syntectonic mafic dykes in Great Slave Lake shear zone; the exception being the lack of dykes in the high-grade strike-lineated protomylonites of the Northeast segment, which will be discussed below.

Late-syntectonic granites

In addition to the Snowdrift granite, two late-syntectonic, Early Proterozoic (*ca* 1.92 Ga) plutons were intruded into the mylonites of the upper amphibolite facies belt of the Southwest segment; the Second Lake and Falls Lake granites (Figs. 2 and 3) (Hanmer & Connelly 1986). Both are variably foliated, but the Second Lake granite is cut by regionally concordant, narrow (<10 m) chlorite-bearing mylonite zones. In contrast with the volumetrically dominant magnetite-bearing hornblende granites of the Southwest segment, the Second Lake granite is biotite-bearing and underlies a strong negative aeromagnetic anomaly (Geological Survey of Canada 1981).

GEOCHRONOLOGY

Isotopic dating using U–Pb in zircons from selected granites was carried out in the geochronology and isotope geochemistry laboratories of the Geological Survey of Canada and Washington University. Analytical procedures and the detailed isotopic and mineralogical systematics for the wall rocks and the Southwest segment (Bowring & Hanmer in preparation) and the Northeast segment (van Breemen *et al.* 1990) are presented elsewhere. In this presentation, we summarize the principal results (Fig. 2) and examine their geological implications for the evolution of Great Slave Lake shear zone.

Wall rocks

Two samples of two-mica leucogranite from the northwestern wall rock near McDonald Lake yield Archean ages of *ca* 2.55 Ga, similar to, but somewhat younger than, the youngest ages obtained from granites in the adjacent parts of the Slave Province (Henderson *et al.* 1987a, van Breemen *et al.* 1987a). Garnet leucogranite from the migmatitic wall rock to the southeast of the Southwest segment yields an upper intercept age of 1.96 Ga. This age is very similar to the 1.955 Ga age proposed for the Slave granite to the south and west (Bostock *et al.* 1987), which is clearly contiguous with the wall rocks analysed here (Fig. 5) (Bostock 1988).

Northeast segment

Dip-lineated protomylonite, derived from a biotite megacrystic granite protolith, which grades progressively into dip-lineated mylonites, yielded an upper intercept age of 1.976 Ga. Poorly foliated biotite megacrystic granite protolith to the thin strip of strike-lineated mylonites within the dip-lineated belt yielded an upper intercept age of 1.978 Ga, indistinguishable from that of its host. Two samples of the Sandwich granite were sampled and analysed. The first came from within the body of the granite, well removed from the contact with the country rock and yielded an upper intercept age of 2.562 Ga. A second sample from the intrusive northwestern contact zone with the strike-lineated upper amphibolite facies protomylonites and mylonites yielded a concordant fraction with an age of 2.556 Ga.

Southwest segment

Megacrystic granite protolith to mylonites of the granulite facies belt yields a ^{207}Pb–^{206}Pb model age of *ca* 1.98 Ga. Strongly foliated megacrystic granite protolith to mylonites of the lower amphibolite facies belt yields an upper intercept age of 1.924 Ga, whereas a sample of granite protomylonite from the same belt yields a poorly constrained, older upper intercept age of 2.05 ± 0.065 Ga. The late-syntectonic Second Lake and Falls Lake granites yield upper intercept ages of *ca* 1.92 Ga. Two of the young set of cross-cutting, isotropic to locally foliated leucogranite sheets from within the granulite facies belt yield ages of *ca* 1.89 Ga.

DISCUSSION

Great Slave Lake shear zone

The mylonites which constitute Great Slave Lake shear zone are principally derived from magnetite-bearing hornblende–biotite granites, clinopyroxene tonalites and included panels of mixed paragneiss. The granites, part of a magmatic arc, were emplaced during the tectonic activity of the shear zone, coeval with intrusion of mafic dykes. The spatial and temporal distribution of the mafic dykes suggests that they reached their structural level of emplacement by exploiting the active strike-slip parts of the shear zone. Non-magnetic granite was only emplaced relatively late in the shear zone history. Because we cannot constrain the timing of granite emplacement with respect to the beginning of deformation in the mylonite belts, we can only apply isotopically determined ages of magmatic crystallization (Fig. 2) to date the mylonitic fabrics *currently* preserved in the rocks. The time of on-set of mylonitization in any given belt, perhaps at higher metamorphic grade than that now preserved, remains rather loosely constrained.

Granulite facies and lower amphibolite mylonites preserved in the Southwest segment (Fig. 3) are younger than *ca* 1.98 and 1.924 Ga, respectively. Upper amphibolite to granulite facies mylonites preserved in the dip-lineated belt of the Northeast segment represent deformation which occurred within a very narrow time window at 1.977 Ga.

The upper amphibolite facies, strike-lineated protomylonites and mylonites of the Northeast segment were formed prior to the emplacement of the *ca* 2.562 Ga Sandwich granite. These mylonitic rocks bear a strong lithological, structural, metamorphic and kinematic resemblance to the upper amphibolite protomylonites and mylonites of the Southwest segment. Indeed, Hanmer & Needham (1988) initially suggested that they were two parts of the same belt. However, in the Southwest segment, Hanmer & Lucas (1985) and Hanmer & Connelly (1986) found that the upper amphibolite facies belt is younger than the post-1.98 Ga granulite facies belt. These observations raise an important question. Did upper amphibolite facies mylonitization in the SW segment occur during the Archean, or during the Early Proterozoic?

What are the geochronological constraints on the age of the upper amphibolite facies mylonitic rocks in the Southwest segment? Two large samples and several hand samples of granitic protolith, from which the protomylonites and mylonites of the upper amphibolite facies belt of the Southwest segment were derived, have all yielded preliminary Archean ages and thus do not significantly constrain the age of mylonitization (Bowring & Hanmer in preparation). Therefore, we must turn to geological considerations to resolve this issue. Three principal observations persuade us that the mylonitic fabrics preserved in the upper amphibolite facies belt of the Southwest segment formed during the interval 1.98–1.924 Ga.

(1) The field observation that cross-cutting granitic sheets in the post-1.98 Ga granulite facies belt are transposed in the adjacent upper amphibolite facies belt requires that mylonitization in the latter belt occurred post-1.98 Ga.

(2) The fabric elements of the upper amphibolite facies protomylonites and mylonites are involved in the corridor of spatially determined heterogeneous flow located along the bulk rheological interface between different parts of the hornblende–biotite granite protolith to Great Slave Lake shear zone (Fig. 6). The rheological interface did not exist until the Early Proterozoic (2.05–1.924 Ga).

(3) The SW segment is a strike-slip structure, within which significant dip-slip movements have not taken place between mylonite belts. Accordingly, an Archean, upper amphibolite facies belt, would have experienced the same metamorphic conditions as the adjacent Early Proterozoic granulite facies mylonites. Given their already partially dehydrated state, they should have developed granulite facies mineral assemblages. This has not occurred.

Kinematically, the Southwest segment of Great Slave Lake shear zone behaved as a dextral, strike-slip, mylonite zone, with a component of shortening across the shear plane (Hanmer & Connelly 1986. Hanmer 1989, 1990), throughout its post-1.98 Ga tectonic history. During cooling and uplift, the width of the zone of active shearing decreased, effectively abandoning volumes of older, higher temperature mylonites which are now preserved as longitudinal belts (Fig. 3) (see Hanmer 1988a for detailed documentation). By contrast, the contradictory shear-sense indications from the high-grade dip-lineated mylonites of the Northeast segment are indicative of strongly transpressive progressive deformation, with a subordinate component of noncoaxial flow. The apparent effect of dip-slip displacement was to raise granulite facies mylonites to their present structural level, where ambient upper amphibolite facies conditions prevailed at that time (Hanmer & Needham 1988). With further cooling to greenschist facies, the narrowing of the locus of active flow was accompanied by a kinematic switch from dip-slip to dextral strike-slip shearing.

At the scale of Great Slave Lake shear zone as a whole (Figs. 2 and 3), the longitudinal components of the anastomosing network of greenschist facies mylonites accommodated continued eastward displacement of the Slave continent. Toward the northeast, the greenschist mylonites cut across the older, curved, higher temperature mylonite belts which, according to the indentor model, had previously lain to the east of the southeast corner of the Slave continent. The transverse components of the linked network of greenschist facies mylonites served to accommodate the component of shortening across Great Slave Lake shear zone by acting as very large-scale asymmetrical extensional shear bands, synthetic to the bulk dextral shear along the trend of Great Slave Lake shear zone (Fig. 2). Analogues to such structures bound antithetically rotated crustal blocks observed in modern-day continental-scale strike-slip faults (Garfunkel & Ron 1985, Nur *et al.* 1986).

Thelon–Taltson magmatic arc

Elements of the plutonic and tectonometamorphic history of Great Slave Lake shear zone can be correlated with contemporaneous events in the Thelon magmatic arc (Henderson *et al.* 1982, Thompson & Ashton 1984, Henderson & Macfie 1985) and the Taltson magmatic zone (Bostock *et al.* 1987). From a regional perspective, Great Slave Lake shear zone lies within a corridor of 2.05–1.92 Ga granites, flanked on either side by rocks of Late Archean to Early Aphebian age (Henderson *et al.* 1987a, Bostock & Loveridge 1988). These granites link the contemporaneous biotite, hornblende and clinopyroxene bearing tonalites, granodiorites and granites of the Thelon magmatic arc to those of the Taltson magmatic zone, forming the 2500 km long Thelon–Taltson magmatic arc (Fig. 3) (Hoffman 1987). Whereas some workers have suggested that the Thelon tectonic zone extends as a structural entity into the Taltson magmatic zone (Thompson & Ashton 1984, James 1985, 1986, Thompson *et al.* 1985), others have highlighted the structural and metamorphic differences between the two (Bostock *et al.* 1987, Hoffman 1987). In the following discussion, we shall compare and contrast both the Thelon tectonic zone and the Taltson magmatic zone with Great Slave Lake shear zone in order to examine the crustal-scale deformation in the vicinity of the advancing Slave indentor.

Thelon tectonic zone

The Thelon tectonic zone closely resembles Great Slave Lake shear zone. It is a straight belt of highly deformed and transposed, granulite to upper amphibolite facies straight gneisses (Fig. 8) (Henderson *et al.* 1982, 1987b, Thompson & Ashton 1984, Henderson & Macfie 1985, 1986, James 1985, 1986, 1989, Thompson *et al.* 1985, 1986), lying within the western part of the Thelon magmatic arc (Hoffman 1987, 1988a). Its boundary with the Slave Province is marked by a lower amphibolite to greenschist facies transcurrent shear zone, up to 5 km wide (*boundary shear zone* in Fig. 8), and a 20 km wide swarm of concordant mafic dykes, spatially restricted to the area north of Bathurst fault. East of the boundary shear zone, upper amphibolite to granulite facies straight gneisses, predominantly strike-lineated in the west and generally dip-lineated in the east (Thompson *et al.* 1986, James 1989, James personal communi-

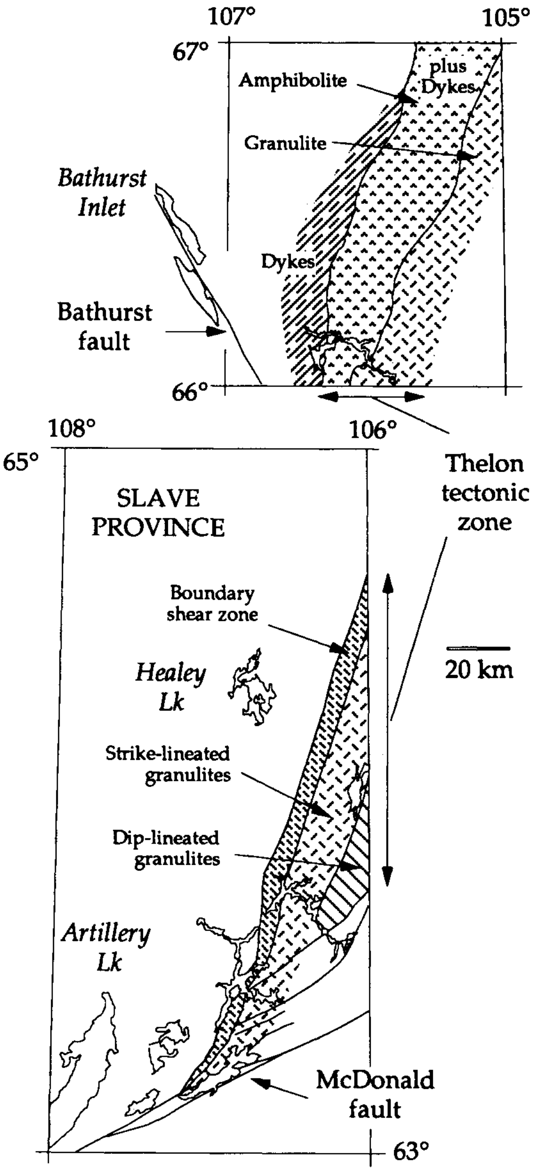

Fig. 8. Schematic representation of the structural elements of parts of the Thelon tectonic zone. The lower part is after Henderson *et al.* (1987b), the upper part is from Thompson *et al.* (1986). In the south, the straight gneisses of the Thelon tectonic zone formed at granulite facies, whereas in the north there is a western upper amphibolite facies sub-zone. In the south, the western straight gneisses are predominantly strike-lineated, whereas the eastern ones are mainly dip-lineated. In the north, the western sub-zone is strike-lineated, whereas lineations in the eastern sub-zone are somewhat more variable in plunge. The eastern limit of the Thelon tectonic zone has not yet been established on the ground. In the north, the Thelon tectonic zone is flanked to the west by a 20 km wide swarm of closely spaced mafic dykes, concordant with respect to the tectonic zone. In the northernmost reaches, the dykes also occur within the upper amphibolite facies straight gneisses ('plus Dykes'). Note that the zones of straight gneisses illustrated here are reworked by networks of discrete, east-side-up dip-slip and dextral strike-slip, lower amphibolite to greenschist facies mylonitic shear zones. See text.

cation 1985), were derived by the deformation of granite and mixed paragneiss protoliths, similar to those of Great Slave Lake shear zone.

The high-grade straight gneisses are cut by a network of discrete, steeply dipping, lower amphibolite to greenschist facies mylonite zones, hundreds of metres to several kilometres wide, of which the boundary shear zone is an important component (Thompson *et al.* 1986, James 1989). The more westerly mylonite zones tend to be strike-lineated and associated with dextral transcurrent shear. The more easterly mylonites tend to carry steeply plunging extension lineations associated with east-side-up displacements (James 1986). In view of its strike length, it is improbable that the Thelon tectonic view has been rotated to its present orientation from an initially shallow dipping configuration. Because its steep dip is unfavourable for significant reverse fault displacement (Sibson 1985), the Thelon tectonic zone and its constituent dip-lineated mylonites are likely to represent a belt of strongly transpressive progressive deformation. This suggestion is supported by the presence of contradictory indications of both east-side- and west-side-up displacements (N. G. Culshaw unpublished data).

U–Pb isotope studies of syntectonic granitoids in the Thelon tectonic zone (van Breemen *et al.* 1987b,c, James *et al.* 1988, van Breemen & Henderson 1988, Frith & van Breemen 1990, Henderson & Loveridge 1990) suggest that granulite facies metamorphism occurred during the interval 2.0–1.95 Ga. However, monazite ages indicate that ambient metamorphic temperatures had already fallen below *ca* 700°C (Parrish 1990) by *ca* 1.975 Ga (van Breemen *et al.* 1987b), indicating that granulite facies metamorphism had ceased by this time. A 1.957 Ga granite was emplaced syntectonically with respect to the network of dip-lineated mylonitic shear zones in the east, whereas the strike-lineated shear zones in the west cut granitoids as young as 1.92 Ga (van Breemen *et al.* 1987c). These data suggest that strike-slip shearing post-dated, or out-lasted, post-granulite dip-slip shearing (James 1989).

Taltson magmatic zone

In contrast to Great Slave Lake shear zone and the Thelon tectonic zone, the presence of high-grade mylonites has not been established in the Taltson magmatic zone (Fig. 5) (Bostock 1987, 1988). However, it is cut by a N–S-striking network of vertical, anastomosing, lower amphibolite to greenschist facies mylonite zones, 100 m–5 km thick (Culshaw 1984, Bostock 1988, 1989). The mylonites are related to both sinistral and dextral strike-slip movements. The western mylonite zones are of relatively minor magnitude (Fig. 5). In the more important eastern mylonite zones, dextral shearing out-lasted sinistral shearing (Bostock 1988, 1989).

Granitoids were emplaced during the interval 1.986–1.906 Ga (Fig. 5) (Bostock *et al.* 1987, Bostock & Loveridge 1988). Near concordant monazite ages suggest that ambient metamorphic temperatures in the Taltson sector were lower than 700°C by 1.935 Ga (Bostock *et al.* 1987, Bostock & Loveridge 1988, Parrish 1990). Note that this is a minimum age and quite compatible with estimates of the timing of granulite facies metamorphism from the Thelon tectonic zone (pre-1.975 Ga) and Great Slave Lake shear zone (1.975 Ga).

The principal mylonites are localized along the eastern margin of the Taltson magmatic zone (Bostock 1988, 1989). Maximum ages for the preserved sinistral and dextral mylonites are given by 1.935 Ga and *ca* 1.94–1.906 Ga granitic protoliths, respectively (Fig. 5) (Bostock & Loveridge 1988, Bostock personal communication). In the north, at Gagnon Lake (Fig. 5), a 5 km wide, lower amphibolite facies, ENE- to NNE-striking, curvilinear, dextral mylonitic belt cuts a *ca* 1.94 Ga granite protolith (Bostock personal communication). These protomylonites and mylonites are temporally, metamorphically and kinematically compatible with the lower amphibolite facies mylonites of Great Slave Lake shear zone, from which they appear to represent a curved lateral splay of limited strike-length (Bostock 1988). Further south, the mylonites are kinematically more complex. A N–S-striking, 1 km wide belt of post-1.906 Ga greenschist facies dextral mylonites is younger than the adjacent belt of similarly oriented lower amphibolite facies sinistral mylonites (Fig. 5) (Bostock 1988, Bostock & Loveridge 1988). Accordingly, we conclude that significant cooling of Taltson magmatic zone occurred during the interval *ca* 1.94–1.906 Ga, during the course of which N–S-oriented dextral shearing out-lasted N–S sinistral shearing.

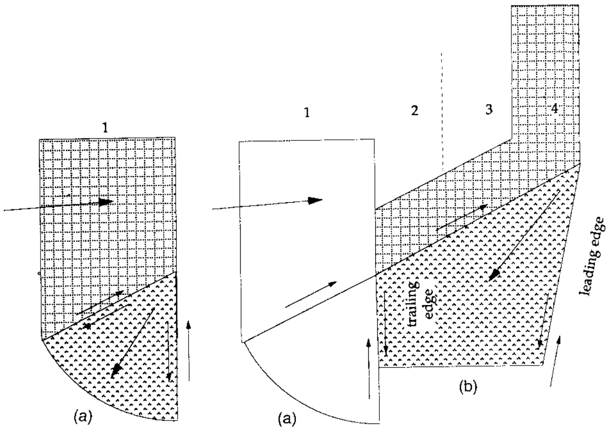

Fig. 9. A possible scenario to account for the presence of N–S-trending earlier sinistral and younger dextral strike-slip mylonites along the eastern boundary of the Taltson magmatic zone. (a) Taltson magmatic zone as a crustal wedge (shaded), bounded by the dextral Great Slave Lake shear zone and N–S-trending sinistral mylonites, escaping from the vicinity of the frontal face of the Slave indentor (squares). (b) Subsequent to the displacements illustrated in (a), the indentor has moved further east (arbitrary positions 1–4). The south-eastern margin of the indentor is illustrated as a transpressive, non-ideal transform fault. The Taltson magmatic zone is no longer an active escaping crustal wedge, the locus of escape, having migrated eastward with the frontal face of the indentor. Consequently, a new escape wedge (shaded) is bounded by sinistral shear along its leading edge and by dextral shear along its trailing and lateral edges. The trailing edge is localized by the plane of weakness presented by the older sinistral mylonites. See text.

Collision and indentation

According to one model for the Early Proterozoic tectonics of the NW Canadian Shield (Hoffman 1987, 1988a, Bowring & Grotzinger 1989, Tirrul & Grotzinger 1990), collision of the Slave and Rae continents, *ca* 1.97 Ga, was preceded by oblique convergence associated with subduction of the Slave plate beneath the Rae continent (Fig. 1). Accordingly, the granitoids of the Thelon magmatic arc and the Taltson magmatic zone would represent a continental magmatic arc at the leading edge of the Rae continent. Geological and geochronological evidence indicate that granulite facies metamorphism (pre-*ca* 1.97 Ga) occurred prior to the main continental collision event (*ca* 1.97 Ga). This suggests that heat which gave rise to the granulite facies metamorphism was advected by processes related to the construction of the magmatic arc (e.g. Windley 1981), rather than to crustal thinning by extension (James 1989) or crustal thickening by thrusting (Thompson & Henderson 1983). The granulite facies mylonitization in the Southwest segment of Great Slave Lake shear zone could represent the strike-slip component of oblique continental convergence just prior to collision (Fitch 1972, Beck 1983, Ellis & Watkinson 1987). Although the 2.0–1.97 Ga upper amphibolite to granulite facies strike-lineated straight gneisses of the western Thelon tectonic zone (Fig. 8) may also represent pre-collisional shearing, there is no direct evidence for pre-collisional strike-slip shearing in the Taltson magmatic zone.

The *ca* 1.97 Ga age of collision between the Slave and Rae continents was initially inferred from U–Pb zircon dating of ash beds at the base of the foredeep developed in front of the Bear Creek Hills foreland thrust–fold belt (Fig. 1) (Bowring & Grotzinger 1989). However, the collision event is also reflected in the kinematic history of the deep-seated mylonites of Great Slave Lake shear zone. The change in strike of the upper amphibolite and granulite facies mylonite belts, from 060° in the Southwest segment to 030° in the Northeast segment (Figs. 2 and 3), is accompanied by a kinematic change, from transpressional strike-slip, to very strongly transpressional, east-side-up, dip-slip flow. We suggest that this kinematic configuration, *ca* 1.98–1.975 Ga, reflects the proximity of the Northeast segment to the frontal face of the Slave indentor (Fig. 9) and approximately E–W shortening across a steeply eastward-dipping collisional zone (Fig. 10). This is supported by the occurrence of *ca* 1.975 Ga strike-lineated, dextral mylonites within the dip-lineated belt, reflecting minor-scale lateral displacements away from the impinging frontal face of the indentor (Fig. 10). Moreover, the dip-lineated granulites of the Thelon tectonic zone, kinematically and metamorphically comparable with the contemporaneous dip-lineated mylonites of the Northeast segment of Great Slave Lake shear zone, are readily explained by the same tectonic scenario. Accordingly, the upper amphibolite facies strike-lineated protomylonites and mylonites of the Southwest segment of Great Slave Lake shear zone represent post-collisional transpressive, non-ideal transform faulting.

In contrast to their relationship to the Thelon tectonic zone, the geometry and the kinematic framework of the high grade belts of the Southwest segment of Great Slave Lake shear zone are insensitive to proximity of the Taltson magmatic zone and, except for the lower amphibolite facies mylonites, show no evidence for linkage (Figs. 2 and 3).

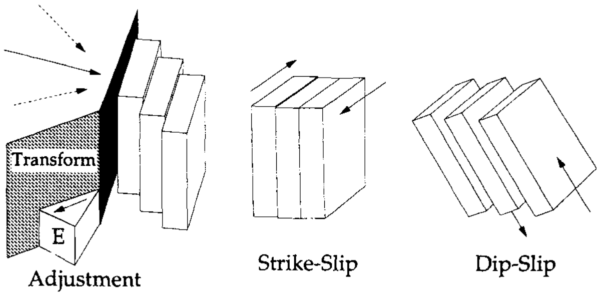

Fig. 10. Schematic illustration of displacements and rotations in the vicinity of a rigid indentor (left) pushed into a volume (right) with a anisotropy oriented obliquely with respect to the frontal face (black) of the indentor. Initial dip-slip displacement along steeply dipping planes, which strike obliquely with respect to the frontal face of the indentor, leads to steepening and rotational hardening of those planes. Further indentation can be accommodated by dextral strike-slip displacements and anticlockwise rotation of the steep shear planes. Flow in the material adjacent to the frontal face of the indentor is not sensitive to the precise orientation of the indentor displacement vector (three options shown). However, if one side of the indentor is essentially a transform fault, then the displacement vector cannot be strongly oblique to that side. If the transform side of the indentor is in fact transpressive, as opposed to purely transform, wedges of crustal material (E) close to the frontal face of the indentor may tend to escape. This may happen at the regional or the more local scales. Hence, the side of E adjacent to the model indentor could represent the local dextral strike-lineated mylonites observed to post-date the dip-lineated mylonites in the Northeast segment of Great Slave Lake shear zone, or the entire wedge could represent the Taltson magmatic zone. See text.

The linked network of strike-lineated greenschist facies mylonites throughout Great Slave Lake shear zone is clearly not continuous with either those of the Thelon tectonic zone, or with those of the Taltson magmatic zone (Figs. 2 and 3). In the Northeast segment, they cut across the higher grade dip-lineated mylonites and are not deflected with them around the southeast corner of the Slave indentor. How then, can we account for the kinematic switch from dip-slip to strike-slip flow with decreasing temperature which occurs at lower amphibolite to upper greenschist facies in both the Thelon tectonic zone and the Northeast segment of Great Slave Lake shear zone, if the two zones were not part of a linked network?

Geological evidence from both the Northwest segment of Great Slave Lake shear zone and the Thelon tectonic zone indicates that dip-slip mylonitization was accompanied by the retrogression of granulite facies mineral assemblages (Thompson *et al.* 1986, Hanmer & Needham 1988, James 1989). Together with the evidence that granulite facies metamorphism pre-dated continental collision, these observations indicate that collision related crustal thickening was accompanied by uplift and cooling, rather than heating. This can be readily explained in the context of a steeply dipping transpressive collision zone, associated with a west to east horizontal strain gradient (Fig. 11) (Coward 1983). In the Northeast segment of Great Slave Lake shear zone, the switch to strike-slip flow directly reflects progressive eastward displacement of the indentor (Fig. 9). Although, the Thelon tectonic zone trends NNE, the prevailing Archean grain in the Rae Province trends NE (Geological Survey of Canada 1987). We tentatively suggest that the kinematic switch in the Thelon tectonic zone reflects regional-scale geometrical or rotational hardening (Poirier 1980) in front of the indentor by the steepening of dip-lineated shear planes, and subsequent stress relaxation by anticlockwise rotation of steep-sided crustal blocks about vertical axes accompanied by strike-slip dextral shear between the blocks (Fig. 10) (e.g. England & Molnar 1990).

If the kinematic development of Great Slave Lake shear zone and the Thelon tectonic zone reflect continental convergence and collision, what is the tectonic significance of the marked difference in structural signature between these two zones and the Taltson magmatic zone? Several authors have suggested that the presence of N–S-trending sinistral mylonite zones in the Taltson magmatic zone might be most readily explained in terms of the southward escape of large-scale crustal blocks bounded by these sinistral mylonites and the dextral Great Slave Lake shear zone (Hoffman 1988a, Bostock 1989). However, the model can be enhanced such that it also accounts for the close spatial association and age relationships of the N–S-trending dextral and sinistral mylonites in the Taltson magmatic zone (Fig. 5). As the Slave indentor penetrated into the Rae continent, older escaped wedges of Rae crust would have been abandoned and the locus of active escape would have migrated eastward. Accordingly, the sinistral leading edge of an escaped wedge would become the dextral trailing edge of the succeeding crustal wedge as it moved away from the frontal face of the indentor. Only minor adjustments are to be expected at the trailing edge, because the principal dextral component of wedge displacement could be accommodated by Great Slave Lake shear zone itself. Geochronological data from both Great Slave Lake shear zone and the Taltson magmatic zone indicate that such displacements were still occurring along both the transform fault and the margins of the escaping crustal wedges post-1.924 Ga and, at least locally, post-1.906 Ga.

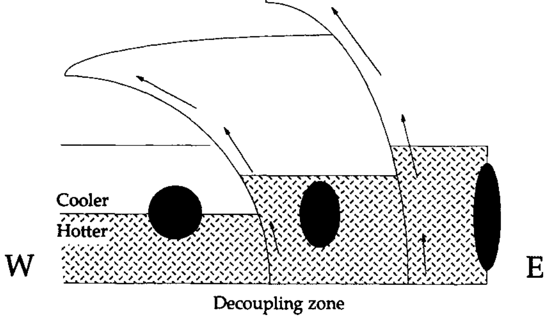

Fig. 11. Schematic illustration of the possible role of dip-slip movement on post-granulite, steeply dipping, strongly transpressive shear zones in thickening continental crust in raising hot rocks to cooler structural levels. Note the west (W) to east (E) increase in finite strain. The effects of isostatic adjustment on the geometry of the lower boundary have been ignored for graphical simplicity. Modified after Coward (1983).

Localization of faulting

Three aspects of the structural and plutonic geology of Great Slave Lake shear zone, not yet been examined in this discussion, may shed some light on factors influencing the initiation of crustal-scale shear zones and the localized persistence of deformation within them.

The trend of the Archean mylonites and the Sandwich granite in the Northeast segment (Fig. 3) suggests that they may have influenced the localization of the Early Proterozoic mylonite belts. However, the nature of that influence depends upon whether the Archean mylonites were originally part of the Slave indentor or the Rae continent. In the former case, they may have influenced the geometry of the southeastern corner of the indentor. In the latter case, they may have influenced the way in which continental impingement planed off irregularities along the original Rae continental margin.

Cross-cutting leucocratic granite veins of *ca* 1.89 Ga age are restricted to the granulite facies mylonite belt of the Southwest segment of Great Slave Lake shear zone (Fig. 2). Given that the granulite mylonites are bounded by mylonite belts whose post-granulite fabrics developed prior to *ca* 1.92 Ga, but which are not cross-cut by any granite veins, the fractures into which the veins were emplaced were apparently controlled by the material properties of the granulite mylonites. The granulites were rheologically susceptible to brittle failure, perhaps because they contained suitably oriented planes of weakness. The disrupted nature of the tectonic fabrics of the belt and the existence of older cross-cutting veins could lend some support to this interpretation.

The mafic dykes southeast of McDonald Lake and northeast of Schist Lakes are concordant with the trend of both Great Slave Lake shear zone as a whole, and of local deviations from that trend, such as the McKee fault zone (Fig. 4) (Hanmer & Lucas 1985, Hanmer & Needham 1988). The dyke swarm is more densely populated than the similarly oriented (Hearne?) dykes of the Great Slave Lake and Artillery Lake areas, northwest of the shear zone (e.g. Hoffman *et al.* 1977, Henderson & Macfie 1985, Hoffman 1988b, Henderson *et al.* 1991). On the other hand, they would appear to be analogous to the 20 km wide swarm of mafic dykes which occur concordant and adjacent to the Thelon tectonic zone, north of Bathurst fault (Fig. 8). In fact, dykes even occur within the western amphibolite facies part of the Thelon tectonic zone (Thompson *et al.* 1986). While the age of these dyke swarms remains unconstrained, we note that they are locally affected by low-temperature dextral shear (Hanmer & Lucas 1985) and that a mafic dyke was syntectonically emplaced into greenschist mylonites (Hanmer & Needham 1988). Accordingly, we tentatively suggest that emplacement of the swarms of closely spaced concordant mafic dykes may be related to low-temperature transcurrent shear (Fig. 7) along Great Slave Lake shear zone and the Thelon tectonic zone. Ductile shear zones develop by the reaction softening or strain softening of material in the immediate vicinity of initial brittle fractures (Segall & Simpson 1986, Simpson 1986). In the presence of pressurized magmas, a strike-slip en échelon array of such fractures would dilate and fill with mafic dyke material (Fig. 7). We speculate that the preserved dyke swarms are zones of post-collisional brittle failure in the eastern and southern margins of the Slave Province. They may represent a failed attempt to incorporate Slave Province material into Great Slave Lake shear zone and the Thelon tectonic zone. In other words, they were fracture zones which were subsequently abandoned when plastic deformation was localized within the shear zones to the southeast and east.

CONCLUSIONS

Great Slave Lake shear zone, Thelon tectonic zone and Taltson magmatic zone record different structural aspects of the interaction at a collisional continental boundary between the Slave and Rae continents during the interval 2.0–1.9 Ga. Early Proterozoic deformation in the Northeast segment of Great Slave Lake shear zone records the collision, post-collision indentation and transform faulting events. The Southwest segment records the pre-collisional convergence and post-collisional transform accommodation of the indentation. Deformation in the Thelon tectonic zone reflects convergence, collision and post-collisional indentation. Mylonite zones in the Taltson magmatic zone only record the post-collisional displacement of wedges of Rae continent away from the frontal face of the Slave indentor.

Granulite facies metamorphism in Great Slave Lake shear zone and the Thelon tectonic zone occurred prior to, or during the early phase of, the *ca* 1.97 Ga continental collision and may be related to construction of the Thelon magmatic arc. In the vicinity of the frontal face of the Slave indentor, the steeply E-dipping Slave–Rae collisional boundary was the site of strongly transpressive deformation. East-side-up displacements at the boundary brought granulite facies tectonites to shallower levels along steep dip-slip shear zones. Post-collisional indentation resulted in further steepening and thickening of the Slave–Rae boundary. A kinematic switch with decreasing metamorphic grade from dip-slip to dextral strike-slip tectonics occurred as the boundary zone cooled. Along the southeastern (transform) side of the indentor, this represented eastward progression of the frontal face of the indentor. In front of the indentor it may have represented rotational hardening followed by anticlockwise rotation of initially northeasterly trending crustal blocks between dextral strike-slip faults.

Magmatism and tectonic activity were genetically associated. Granitoids were emplaced within or adjacent to the locus of active shearing. Mafic dykes were emplaced either into zones of brittle fault initiation in the stiff wall rocks, or into strike-slip portions of the plastic mylonite belts. Very late granite veins were emplaced into parts of the shear zone which were rheologically particularly susceptible to brittle failure.

Acknowledgements—The field component of this study would not have been possible without the invaluable assistance of Steve Lucas, Jim Connelly and Terry Needham. Discussion and exchange of field observations with Hewitt Bostock, John Henderson, Paul Hoffman and Don James were greatly appreciated. We also thank them, as well as Shaocheng Ji, for comments which significantly improved the manuscript. This is Geological Survey of Canada contribution No. 41491.

REFERENCES

Beck, M. 1983. On the mechanism of tectonic transport in zones of oblique subduction. *Tectonophysics* **93**, 1–11.

Bell, T. H. 1981. Foliation development—the contribution, geometry and significance of progressive, bulk, inhomogeneous shortening. *Tectonophysics* **75**, 273–296.

Bell, T. H. 1985. Deformation partitioning and porphyroblast rotation in metamorphic rocks: a radical reinterpretation. *J. metamorph. Geol.* **3**, 109–118.

Bostock, H. H. 1987. Geology of the south half of the Taltson Lake map area, District of Mackenzie. *Geol. Surv. Pap. Can.* **87-1A**, 443–450.

Bostock, H. H. 1988. Geology of the north half of the Taltson Lake map area, District of Mackenzie. *Geol. Surv. Pap. Can.* **88-1C**, 189–198.

Bostock, H. H. 1989. The significance of ultramafic inclusions in gneisses along the eastern margin of the Taltson Magmatic Zone, District of Mackenzie, NWT. *Geol. Surv. Pap. Can.* **89-1C**, 49–56.

Bostock, H. H. & Loveridge, W. D. 1988. Geochronology of the Taltson Magmatic Zone and its eastern cratonic margin, District of Mackenzie. *Geol. Surv. Pap. Can.* **88-2**, 59–66.

Bostock, H. H., van Breemen, O. & Loveridge, W. D. 1987. Proterozoic geochronology in the Taltson Magmatic Zone, NWT. *Geol. Surv. Pap. Can.* **87-2**, 73–80.

Bowring, S. A. & Grotzinger, J. P. 1989. Implications of new U–Pb dating and stratigraphic correlations for current tectonic models for Wopmay orogen and Thelon tectonic zone. *Geol. Ass. Can. Miner. Ass. Can. Prog. w. Abs.* **14**, A74.

Bowring, S. A., Van Schmus, W. R. & Hoffman, P. F. 1984. U–Pb zircon ages from Athapuscow aulacogen, East Arm of Great Slave Lake. *Can. J. Earth Sci.* **21**, 1315–1324.

Brown, M. 1973. The definition of metatexis, diatexis and migmatite. *Proc. Geol. Ass.* **84**, 371–382.

Coward, M. P. 1983. Thrust tectonics, thin skinned or thick skinned, and the continuation of thrusts to deep in the crust. *J. Struct. Geol.* **5**, 113–124.

Culshaw, N. G. 1984. Rutledge Lake, Northwest Territories; a section across a shear belt within the Churchill province. *Geol. Surv. Pap. Can.* **84-1A**, 331–372.

Ellis, M. & Watkinson, A. J. 1987. Orogen-parallel extension and oblique tectonics: the relation between stretching lineations and relative plate motions. *Geology* **15**, 1022–1026.

England, P. & Molnar, P. 1990. Right-lateral shear and rotation as the explanation for strike-slip faulting in eastern Tibet. *Nature* **344**, 140–142.

Fitch, T. J. 1972. Plate convergence, transcurrent faults, and internal deformation adjacent to Southeast Asia and the Western Pacific. *J. geophys. Res.* **77**, 4432–4461.

Frith, R. A. & van Breemen, O. 1990. U–Pb zircon age from the Himag plutonic suite, Thelon Tectonic Zone, Churchill Structural Province, Northwest Territories. *Geol. Surv. Pap. Can.* **89-2**, 49–54.

Garfunkel, Z. & Ron, H. 1985. Block rotation and deformation by strike-slip faults. 2. The properties of a type of macroscopic discontinuous deformation. *J. geophys. Res.* **90**, 8589–8602.

Geological Survey of Canada. 1981. Magnetic Anomaly Map, Lockhart River, NWT, map NP 12-13 1566A, scale 1:1,000,000. Geological Survey of Canada.

Geological Survey of Canada. 1984. Magnetic Anomaly Map, Slave River, NWT, map NP 11-12-M, scale 1:1,000,000. Geological Survey of Canada.

Geological Survey of Canada. 1987. Magnetic Anomaly Map of Canada, map 1255A, scale 1:5,000,000. Geological Survey of Canada.

Gibb, R. A. 1978. Slave–Churchill collision tectonics. *Nature* **271**, 50–52.

Gibb, R. A. & Thomas, M. D. 1977. The Thelon Front: a cryptic suture in the Canadian Shield? *Tectonophysics* **38**, 211–222.

Gibb, R. A., Thomas, M. D., Lapointe, P. L. & Mukhopadyay, M. 1983. Geophysics of proposed Proterozoic sutures in Canada. *Precamb Res.* **19**, 349–384.

Hanmer, S. 1984. The potential use of planar and elliptical structures as indicators of strain regime and kinematics of tectonic flow. *Geol. Surv. Pap. Can.* **84-1B**, 133–142.

Hanmer, S. 1986. Asymmetrical pull-aparts and foliation fish as kinematic indicators. *J. Struct. Geol.* **8**, 111–122.

Hanmer, S. 1987. Textural map-units in quartzofeldspathic mylonitic rocks. *Can. J. Earth Sci.* **24**, 2065–2073.

Hanmer, S. 1988a. Great Slave Lake Shear Zone, Canadian Shield: reconstructed vertical profile of a crustal-scale fault zone. *Tectonophysics* **149**, 245–264.

Hanmer, S. 1988b. Geology, Great Slave Lake Shear Zone; parts of Fort Resolution, Snowdrift and Fort Reliance map areas, District of Mackenzie, NWT. *Geol. Surv. Can. Open-file* **1783**, 1:100,000

Hanmer, S. 1989. Initiation of cataclastic flow in a mylonite zone. *J. Struct. Geol.* **11**, 751–762.

Hanmer, S. 1990. Natural rotated inclusions in non-ideal shear. *Tectonophysics* **176**, 245–255.

Hanmer, S. & Connelly, J. N. 1986. Mechanical role of the syntectonic Laloche Batholith in the Great Slave Lake Shear Zone, District of Mackenzie, NWT. *Geol. Surv. Pap. Can.* **86-1B**, 811–826.

Hanmer, S. & Lucas, S. B. 1985. Anatomy of a ductile transcurrent shear: the Great Slave Lake Shear Zone, District of Mackenzie, NWT (preliminary report). *Geol. Surv. Pap. Can.* **85-1B**, 7–22.

Hanmer, S. & Needham, T. 1988. Great Slave Lake shear zone meets Thelon Tectonic Zone, District of Mackenzie, NWT. *Geol. Surv. Pap. Can.* **88-1A**, 33–34.

Hanmer, S. & Passchier, C. W. 1991. Shear-sense indicators: a review. *Geol. Surv. Pap. Can.* **91**.

Henderson, J. B. & Loveridge, W. D. 1990. Inherited Archean zircon in the Proterozoic Thelon Tectonic Zone: U–Pb geochronology of the Campbell granite, south of McDonald fault, District of Mackenzie, Northwest Territories. *Geol. Surv. Pap. Can.* **89-2**, 63–70.

Henderson, J. B. & Macfie, R. I. 1985. The northern Artillery Lake map area: a transect across the Thelon Front, District of Mackenzie. *Geol. Surv. Pap. Can.* **85-1A**, 441–448.

Henderson, J. B. & Macfie, R. I. 1986. Artillery Lake map area, District of Mackenzie: a transect across the Thelon Front. *Geol. Surv. Pap. Can.* **86-1A**, 411–416.

Henderson, J. B., McGrath, P. H., James, D. T. & Macfie, R. I. 1987b. An integrated geological, gravity and magnetic study of the Artillery Lake area and the Thelon Tectonic Zone, District of Mackenzie. *Geol. Surv. Pap. Can.* **87-1A**, 803–814.

Henderson, J. B., McGrath, P. H., Theriault, R. J. & van Breemen, O. In press. Intracratonic indentation of the Archean Slave Province into the early Proterozoic Thelon Tectonic Zone of the Churchill Province, northwestern Canadian Shield. *Can. J. Earth Sci.*

Henderson, J. B. & Thompson, P. H. 1980. The Healey Lake map area (northern part) and the enigmatic Thelon Front, District of Mackenzie. *Geol. Surv. Pap. Can.* **80-1A**, 165–146.

Henderson, J. B., Thompson, P. H. & James, D. T. 1982. The Healy Lake map area and the Thelon Front Problem, District of Mackenzie. *Geol. Surv. Pap. Can.* **82-1A**, 191–195.

Henderson, J. B., van Breemen, O. & Loveridge, W. D. 1987a. Some U–Pb zircon ages from Archean basement, supracrustal and intrusive rocks, Yellowknife–Hearne Lake area, District of Mackenzie. *Geol. Surv. Pap. Can.* **87-2**, 111–121.

Hoffman, P. F. 1987. Continental transform tectonics: Great Slave Lake shear zone (*ca* 1.9 Ga), northwest Canada. *Geology* **15**, 785–788.

Hoffman, P. F. 1988a. United plates of America, the birth of a craton: Early Proterozoic assembly and growth of Laurentia. *Annu. Rev. Earth & Planet. Sci.* **16**, 543–603.

Hoffman, P. F. 1988b. Geology and tectonics, east Arm of Great Slave Lake, Northwest Territories. map 1628A, scale 1:250,000 and 1:500,000. Geological Survey of Canada.

Hoffman, P. F. 1989. Precambrian geology and tectonic history of North America; In: *The Geology of North America—An overview*, Vol. A. Geological Society of America. 447–512.

Hoffman, P. F., Bell, I. R., Hildebrand, R. S. & Thorstad, L. 1977. Geology of the Athapuscow Aulacogen, East Arm of Great Slave Lake, District of Mackenzie. *Geol. Surv. Pap. Can.* **77-1A**, 117–129.

James, D. T. 1985. Geology of the Moraine Lake area and the Thelon Front, District of Mackenzie. *Geol. Surv. Pap. Can.* **85-1A**, 449–454.

James, D. T. 1986. Geology of the Moraine Lake area District of Mackenzie; Part 2: a transect across part of the Thelon Tectonic Zone. *Geol. Surv. Pap. Can.* **86-1A**, 417–421.

James, D. T. 1989. Geology of the Thelon Tectonic Zone in the Moraine Lake area, District of Mackenzie, NWT: The definition and significance of lithologic, structural and metamorphic changes across the boundary between the Slave and Churchill structural provinces. Unpublished Ph.D. thesis, Queens University.

James, D. T., van Breemen, O. & Loveridge, W. D. 1988. Early Proterozoic U–Pb zircon ages for granitoid rocks from the Moraine Lake transect, Thelon Tectonic Zone, District of Mackenzie. *Geol. Surv. Pap. Can.* **88-2**, 67–72.

Logan, J. M. 1979. Brittle phenomena. *Rev. Geophys. & Space Phys.* **17**, 1121–1132.

Mehnert, K. R. 1968. *Migmatites and the Origin of Granitic Rocks.* Elsevier, Amsterdam.

Nur, A., Ron, H. & Scotti, O. 1986. Fault mechanics and the kinematics of block rotations. *Geology* **14**, 746–749.

Parrish, R. R. 1990. U–Pb dating of monazite and its application to geological problems. *Can. J. Earth Sci.* **27**, 1431–1450.

Poirier, J.-P. 1980. Shear localisation and shear instability in materials in the ductile field. *J. Struct. Geol.* **2**, 135–142.

Reinhardt, E. W. 1969. Geology of the precambrian rocks of Thubin Lakes map area in relationship to the McDonald Fault system, District of Mackenzie, N.W.T. (75 E/12 and parts of 75 E/13 & 85 H/9). *Geol. Surv. Pap. Can.* **69-21**.

Sanderson, D. J. & Marchini, W. R. D. 1984. Transpression. *J. Struct. Geol.* **6**, 449–458.

Segall, P. & Simpson, C. 1986. Nucleation of ductile shear zones on dilatant fractures. *Geology* **14**, 56–59.

Sibson, R. H. 1985. A note on fault reactivation. *J. Struct. Geol.* **7**, 751–754.

Sibson, R. H., Robert, F. & Poulsen, K. H. 1988. High-angle reverse faults, fluid-pressure cycling, and mesothermal gold-quartz deposits. *Geology* **16**, 551–555.

Simpson, C. 1986. Fabric development in brittle-to-ductile shear zones. *Pure & Appl. Geophys.* **124**, 269–288.

Thompson, P. H. & Ashton, K. 1984. Preliminary report on the geology of the Tinney Hills–Overby Lake (W1/2) map area, District of Mackenzie: A look at the TTZ, NE of the Bathurst Fault. *Geol. Surv. Pap. Can.* **84-1A**, 415–423.

Thompson, P. H. & Henderson, J. B. 1983. Polymetamorphism in the Healey Lake map area—implications for the Thelon Tectonic Zone. *Geol. Surv. Pap. Can.* **83-8**, 2.

Thompson, P. H., Culshaw, N. G., Buchanan, J. R. & Manojlovic, P. 1986. Geology of the Slave Province and Thelon Tectonic Zone in the Tinney Hills–Overby Lake (west half) map area, District of Mackenzie. *Geol. Surv. Pap. Can.* **86-1A**, 275–289.

Thompson, P. H., Culshaw, N. G. Thompson, D. L. & Buchanan, J. R. 1985. Geology across the western boundary of the Thelon Tectonic zone in the Tinney Hills–Overby Lake (West Half) map area, District of Mackenzie. *Geol. Surv. Pap. Can.* **85-1A**, 555–572.

Tirrul, R. 1985. Nappes in the Kilohigok Basin, and their relation to the Thelon Tectonic Zone, District of Mackenzie. *Geol. Surv. Pap. Can.* **85-1A**, 407–420.

Tirrul, R. & Grotzinger, J. P. 1990. Early Proterozoic collisional orogeny along the northern Thelon tectonic zone, Northwest Territories, Canada: evidence from the foreland. *Tectonics* **9**, 1015–1036.

van Breemen, O. & Henderson, J. B. 1988. U–Pb zircon and monazite ages from the eastern Slave Province and Thelon Tectonic Zone, Artillery Lake area, NWT. *Geol. Surv. Pap. Can.* **88-2**, 73–84.

van Breeman, O., Hanmer, S. & Parrish, R. R. 1990. Archean and Proterozoic mylonites along the southeastern margin of the Slave Structural Province, Northwest Territories. *Geol. Surv. Pap. Can.* **89-3**, 55–62.

van Breemen, O., Henderson, J. B., Loveridge, W. D. & Thompson, P. H. 1987c. U–Pb zircon and monazite geochronology and zircon morphology of granulites from the Thelon Tectonic Zone, Healy Lake and Artillery lake areas, NWT. *Geol. Surv. Pap. Can.* **87-1**, 783–801.

van Breemen, O., Henderson, J. B., Sullivan, R. W. & Thompson, P. H. 1987a. U–Pb zircon and monazite ages from the eastern Slave Province, Healey Lake area, NWT. *Geol. Surv. Pap. Can.* **87-2**, 101-110.

van Breemen, O., Thompson, P. H., Hunt, P. A. & Culshaw, N. G. 1987b. U–Pb zircon and monazite geochronology from the northern Thelon Tectonic Zone, District of Mackenzie. *Geol. Surv. Pap. Can.* **87-2**, 81–93.

Windley, B. 1981. Phanerozoic granulites. *J. geol. Soc. Lond.* **138**, 745–751.

Woodcock, N. H. & Fischer, M. 1986. Strike-slip duplexes. *J. Struct. Geol.* **8**, 725–735.