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Deep-crustal break-back stacking and slow exhumation of the continental footwall beneath a thrusted marginal basin, Grenville orogen, Canada

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ABSTRACT

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Granulite to upper amphibolite facies ductile thrusting in the Central Gneiss Belt, Grenville orogen, Ontario, represents the tectonic shortening of a continental-scale footwall beneath the thrust-emplaced Central Metasedimentary Belt, during the closure of a postulated back-arc basin (ca. 1.19–1.18 Ga). Break-back stacking in the footwall occurred at mid- to deep-crustal depths (ca. 35 km) within tectonically thickened (ca. 70 km) continental crust, and culminated with renewed thrusting at the base of the overlying Central Metasedimentary Belt (ca. 1.08–1.05 Ga).

The individual mylonite belts which constitute the ductile thrust zones, and the scale of penetrative deformation of the intervening crystalline thrust sheets, are comparable with the largest known examples of high-grade thrust belts elsewhere. They reflect the large-scale thermal and rheological boundary conditions of the deformation. Flow within individual thrust zones may reflect local boundary conditions, such as the rheological behaviour of older thrust sheets and the geometry of interfaces within the thrust stack.

Restoration of the thickness of erosionally removed crustal overburden by break-back thrusting may retard the rates of exhumation and cooling of a mid- to deep-crustal thrust stack.

Introduction

Mid- to deep level thrusting plays a fundamental role in crustal thickening and mountain building in collisional orogens (Coward et al., 1982; Coward and Butler, 1985; Platt, 1986; Coward and Ries, 1986; Malinconico and Lillie, 1989). However, the nature of the structural response of tectonic plates to shortening at deep-crustal levels remains controversial, even in modern collisional orogens such as the Himalaya (e.g., Dewey and Burke, 1973; Seeber, 1983; Molnar, 1984; Burg and Chen, 1984; Hirn et al., 1984a,b; Coward et al., 1986; Tapponier et al., 1986; Mattauer, 1986; Powell, 1986; Molnar et al., 1987; Peltzer and Tapponier, 1988; Dewey et al., 1988, 1989; England and Molnar, 1990a; Jolivet, 1990). High-level linked thrust systems are thought to root at depth (e.g., McClay and Price, 1981; Butler, 1983; Coward and Ries, 1986; Malinconico and Lillie, 1989). Crustal-scale, gently dipping reflectors, imaged in seismic surveys across the initially mid-to deep-seated parts of modern and ancient orogenic belts (COCORP, LITHOPROBE, GLIMPCE), are currently interpreted by analogy with high-level thrust systems (e.g., Korsch et al., 1986; Allmendinger et al., 1987; Cook et al., 1988; Green

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et al., 1988; Pratt et al., 1989; Geis et al., 1990). Nevertheless, few field examples of this kind of mid- to deep-crustal thrust structure have been described (Le Fort, 1981; Culshaw et al., 1983; Davidson, 1984a; Coward and Butler, 1985; Hanmer, 1988; Lucas, 1989, 1990).

High-level thrust systems form under relatively low-temperature conditions. Discontinuous deformation occurs predominantly by pressure-sensitive processes, such as fracture and frictional sliding, accommodated to varying degrees by thermally activated mass transfer and dislocation glide (e.g., Engelder, 1974; Engelder and Scholz, 1976; Byerlee, 1978; Dietrich, 1978; Paterson, 1978; Logan, 1979; Sibson, 1986; Mandl, 1987; Groshong, 1988; Scholz, 1989). The resulting thrusts are discrete faults which separate the stiff, little-deformed rocks of the hanging wall and footwall. In contrast, deep-seated thrust systems which form at relatively high ambient temperatures reflect the predominance of thermally activated, and/or accommodated, deformation processes such as dislocation creep, grain boundary sliding and mass transfer, associated with metamorphically induced, reaction enhanced changes in rock rheology (Elliot, 1973; Nicolas and Poirier, 1976; White and Knipe, 1978; Rubie, 1983; Rutter, 1983; Urai, 1985; Urai et al., 1986). Deformation is associated with continuous strain gradients and results in the development of broad ductile thrust zones, as well as considerable internal ductile deformation of the thrust sheets themselves.

The Grenville orogen in Ontario, Canada (Fig. 1), has been interpreted in terms of the closure of a back-arc basin developed behind a continental magmatic arc on the southeastern margin of the Laurentian continent, and subsequent collision with a second continent outboard of the arc (e.g., Windley, 1986; Smith and Holm, 1990). The present contribution will consider the geometry, stacking order, kinematics and associated metamorphic evolution of a mid- to deep-crustal stack of thrusts (Fig. 1), located within the Laurentian continent, just beneath the thrust-emplaced back-arc basin rocks (Hanmer and McEachern, 1992). Our intention here is to illustrate the great size of individual ductile thrust zones, the penetrative nature of the deformation within individual thrust sheets and the complex nature of the flow within deep-crustal thrust systems. We will also highlight the role of the sequence of thrusting in determining the tectonothermal evolution of the thickened crust. Analogies have been drawn between the Grenville orogen and parts of the western Himalaya (e.g., Dewey and Burke, 1973; Windley, 1986). We will not discuss the relevance of these analogies here. However, we feel that our study may shed useful light on the internal structure and evolution of the deep-seated parts of ancient

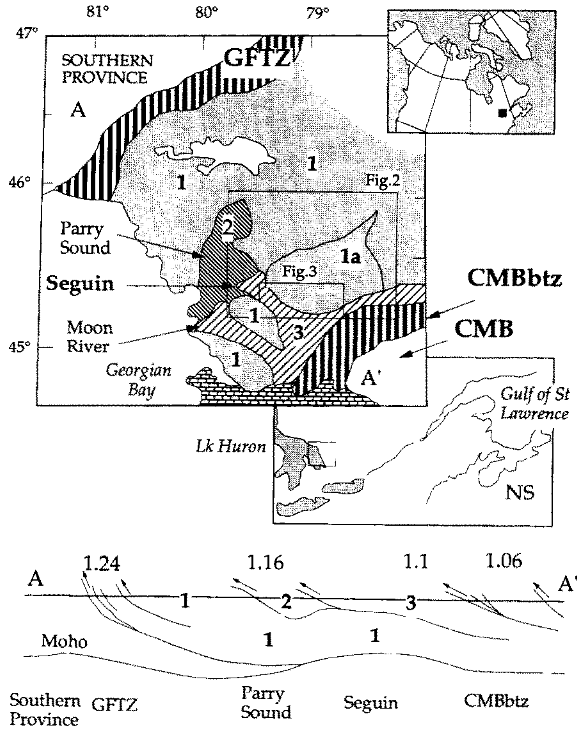

Fig. 1. Schematic illustration of the distribution of the major structural elements of the Central Gneiss Belt (*1–3*), of the Grenville orogen in Ontario, Canada, bounded by the Grenville Front Tectonic Zone (*GFTZ*) to the northwest and by the Central Metasedimentary Belt boundary thrust zone (*CMBbtz*) to the south. Numbers refer to structural levels mentioned in the text and indicate the thrust stacking geometry from structurally lowest to highest. Note that *1a* is the Novar–McClintock assembly; *3* includes both Seguin and Moon River thrust sheets. Tectonic boundaries in the Central Gneiss Belt are after Wynne-Edwards (1972), Davidson and Morgan (1981), Culshaw et al. (1983) and Davidson and Grant (1986). The schematic cross section is adapted from Davidson (1984a,b), and the ages of last major thrusting are given in Ga. Insets are locations: *NS* = Nova Scotia. The locations of Figures 2 and 3 are indicated.

and active mountain belts, formed in similar tectonic settings.

The Grenville orogen

The Grenville orogen in Ontario is the result of tectonic interaction between Laurentia, a magmatic arc and a putative continent to the southeast, during the interval 1.35–1.0 Ga (Dewey and Burke, 1973; Baer, 1974, 1976; Moore, 1986; Windley, 1986, 1989; Rivers et al., 1989). The presence of a normal crustal thickness beneath the Ontario Grenville today (Mereu et al., 1986), coupled with extensive barometric determinations of the order of 8–10 kbar at the present erosional surface (Anovitz and Essene, 1990), indicates that Grenvillian deformation resulted in a doubling of the crust. The Ontario Grenville comprises two geologically highly contrasting belts, the Central Metasedimentary Belt and the Central Gneiss Belt (Fig. 1; Wynne-Edwards, 1972). The internal part of the Central Metasedimentary Belt, a thick pile of metasediments, ca. 1.3–1.25 Ga metavolcanics and associated plutonic rocks as young as ca. 1.06 Ga, is generally interpreted in terms of a volcanic arc (Brown et al., 1975; Condie and Moore, 1977; Fletcher and Farquhar, 1982; Pride and Moore, 1983; Windley, 1986; Holm et al., 1986; Smith and Holm, 1987, 1990; Corriveau, 1990). Interpretations differ over the question of the continental or oceanic context of the arc and the polarity of subduction beneath it. Smith and Holm (1990), after reviewing the available petrological data, suggested that the Central Metasedimentary Belt includes the deformed remains of a back-arc basin which had opened and closed on the northwest side of a continental magmatic arc, developed above a northwest-dipping subduction zone.

In contrast to the Central Metasedimentary Belt, the Central Gneiss Belt (Fig. 1) is essentially composed of pre-1.4 Ga crust and does not include substantial volumes of syn-Grenvillian rocks (Nadeau, 1990). It represents the southeastern edge of the continent with which the Central Metasedimentary Belt arc docked when the back-arc basin closed at, or prior to, ca. 1.19–1.18 Ga (Smith and Holm, 1990; Hanmer and McEachern, 1992). The Central Gneiss Belt is internally divided into an assembly of crystalline thrust sheets (Figs. 1 and 2; Davidson et al., 1982; Culshaw et al., 1983), and is externally bounded by two crustal-scale, dip-slip shear zones (Fig. 1). To the northwest, the Grenville Front Tectonic Zone is a broad zone of intense ductile shearing which separates the Grenville orogen from the adjacent Archean rocks of the Superior and Southern provinces (Wynne-Edwards, 1972; Davidson, 1986; Rivers et al., 1989) and marks the northwest limit of crustal thickening within the Ontario Grenville (Green et al., 1988). To the southeast, the Central Metasedimentary Belt boundary thrust zone is a mid- to deep-crustal shear zone along which the rocks of the Central Metasedimentary Belt were emplaced on top of the Central Gneiss Belt (Hanmer and Ciesielski, 1984; Hanmer, 1988; McEachern, 1990; Hanmer and McEachern, 1992). Shear-sense criteria (Hanmer and Passchier, 1991) within the Central Metasedimentary Belt boundary thrust zone (Hanmer and Ciesielski, 1984; Hanmer, 1988; McEachern, 1990), and in the Grenville Front

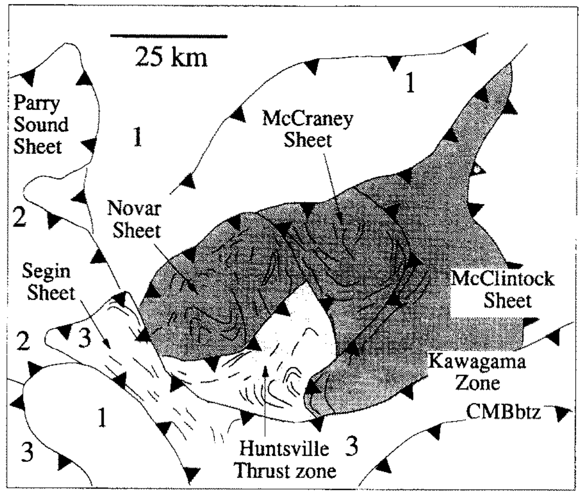

Fig. 2. Structural setting and internal structure of the study area. The Novar–McClintock thrust assembly is shaded. Foliation trajectories are given for the Novar–McClintock thrust assembly and the Seguin thrust sheet. Numbers refer to the structural levels (bold thrusts) of the Central Gneiss Belt and indicate the late-thrusting stacking sequence. *CMBbtz* is the Central Metasedimentary Belt boundary thrust zone. Structural boundaries are after Culshaw et al. (1983) and Davidson and Grant (1986).

Tectonic Zone (Davidson, 1986, 1988), indicate northwestward thrusting along southeast-plunging finite extension lineations.

The Central Gneiss belt

During the past decade, there has been a major breakthrough in understanding the geological evolution of the Central Gneiss Belt, stemming from recognition of its geologically composite nature (Fig. 1; Davidson and Morgan, 1981; Davidson et al., 1982; Culshaw et al., 1983; Davidson, 1984a,b; Nadeau, 1984). This was spurred by the observation that continuous belts of remarkably straight layered, highly strained gneisses represent ductile shear zones. A number of "domains" and "subdomains" were defined according to contrasts in their characteristic rock assemblages, metamorphism, structural trends and geophysical signatures. From study of a widespread assemblage of shear-sense indicators, their boundaries were recognised as ductile thrust zones, upon which they had been transported to the northwest as "crustal slices" and "blocks" along a generally SE-plunging finite extension lineation (Davidson, 1984b; Rivers et al., 1989). The domains and subdomains have been grouped into "structural units" (Culshaw et al., 1983) or "tectonostratigraphic units" (Fig. 1, Davidson, 1984b). In order to adequately describe the internal tectonic architecture now recognised in the Central Gneiss Belt, we shall refer to these domains and subdomains as thrust sheets, bounded by ductile thrust zones (Hanmer and Ciesielski, 1984; Hanmer, 1988; Rivers et al., 1989; Nadeau, 1990), and to the superincumbent units into which the thrust sheets have been grouped as structural levels. Structural level 1 extends southeastward from the Grenville Front Tectonic Zone to form the common footwall to the Parry Sound thrust sheet, level 2, and to the Moon River and Seguin thrust sheets, parts of level 3 (Fig. 1). The latter continues to the east where it forms the footwall to the Central Metasedimentary Belt boundary thrust zone.

Within the Central Gneiss Belt, U-Pb zircon ages of syntectonic granitic veins and pegmatites have been used to determine the age of thrusting (see below). In the field, intrusive veins were identified as syntectonic where they clearly cut across the highly strained gneisses of the ductile thrust zones, but were themselves subjected to high-temperature penetrative deformation, as indicated by the formation of internal feldspar-plastic fabrics. The sampled veins are representative members of veins sets which, even within a given outcrop, show all stages of tectonic grain size reduction and/or transposition with progressive deformation. Accordingly, one can confidently take their ages of magmatic crystallisation to indicate the date of, at least, the later stages of the ductile thrusting. Major ductile deformation in the Grenville Front Tectonic Zone both preceded and postdated the emplacement of the ca. 1.24 Ga Sudbury dykes (Bethune and Davidson, 1988; Davidson, 1988; Krogh et al., 1988), but the timing of initiation and cessation of deformation remains unconstrained. Southeast of the Grenville Front Tectonic Zone, the age of the *last major thrusting event* ranges from ca. 1.16 Ga in the Parry Sound thrust zone (Van Breemen et al., 1986), via ca. 1.1 Ga in the Seguin and Moon River thrust zones (Van Breemen and Davidson, 1990; Nadeau, 1990; Nadeau and Van Breemen, 1992b), to ca. 1.08–1.05 Ga in the Central Metasedimentary Belt boundary thrust zone (Van Breemen and Hanmer, 1986; Hanmer and McEachern, 1992). These data appear to record a progressive southeastward retreat of the principal deformation front across the southeastern half of the Central Gneiss Belt. However, the recent identification of important high-grade ductile thrusting in the Central Metasedimentary Belt boundary thrust zone at ca. 1.19–1.18 Ga (McEachern, 1990; Hanmer and McEachern, 1992) shows that the apparent southeastward progression does not necessarily relate to the time of initiation of the respective thrust zones.

Immediately below the Central Metasedimentary Belt boundary thrust zone, the Central Gneiss Belt in the Huntsville area is composed of a stack of crystalline thrust sheets, tectonically assembled within the deep crust (Fig. 2, Nadeau, 1990). Geothermobarometric measurements of ca. 8–10 kbar at greater than 700°C, obtained from annealed granoblastic tectonites sampled from various parts of the thrust stack (Grant, 1987; Manojlovic, 1987; Anovitz and Essene, 1990), are best interpreted as recording post-thrusting metamorphic conditions (Nadeau, 1990) at crustal depths in the order of 35 km. The largest and structurally highest thrust sheet is the Seguin, part of structural level 3 (Fig. 1). The footwall of the Seguin thrust sheet is formed by an assembly of thrust sheets and ductile thrust zones, part of structural level 2 (Fig. 2). Within this assembly, the Huntsville thrust zone (Nadeau, 1988) is exposed as a window, preserved between the overlying McCraney and McClintock thrust sheets and the underlying Novar thrust sheet. Even the preserved part of the Huntsville thrust zone is large, comparable in size with the adjacent crystalline thrust sheets.

It will be shown below that: (1) the Novar thrust sheet has behaved as a footwall to the Huntsville thrust zone; (2) the original hanging wall to the Huntsville thrust zone has been subsequently modified by the thrust zones which now bound the McCraney and McClintock thrust sheets; (3) the Novar–McClintock assembly has behaved as a footwall to the Seguin thrust sheet; (4) all of the above structures have behaved as a footwall to the Central Metasedimentary Belt boundary thrust zone. Delineation of the internal structure of the thrust stack is predicated upon the ability to recognise ductile thrust zones in the crystalline rocks. In what follows, we will illustrate this point with reference to the Huntsville thrust zone. Furthermore, as we shall document, with particular reference to the Seguin thrust

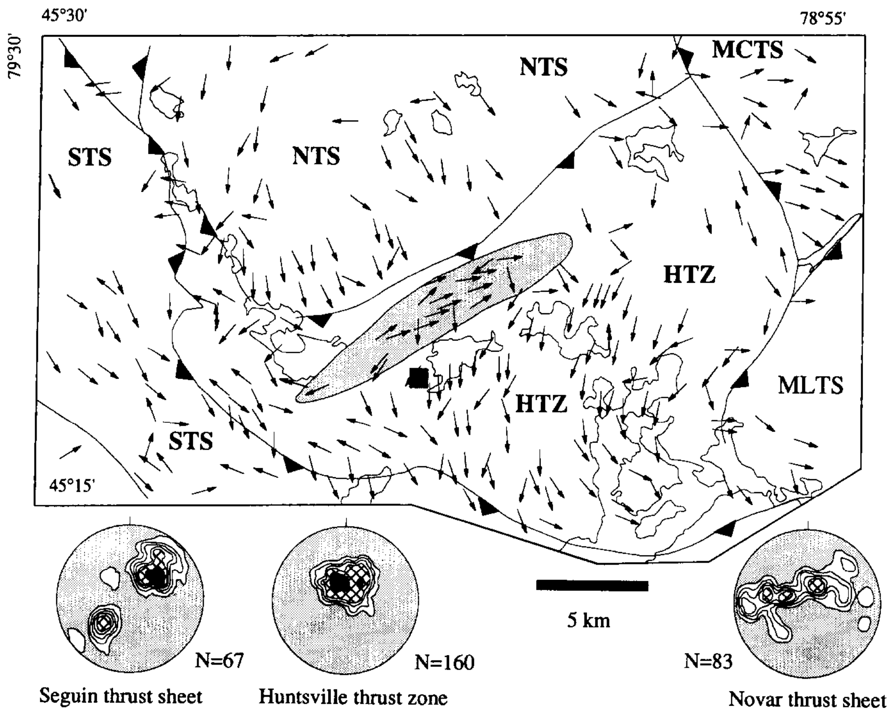

Fig. 3. Map of finite extension lineations (arrows) and foliation orientations (stereonets) within the tectonic components of the stack of thrusts within the Central Gneiss belt, beneath the Central Metasedimentary Belt boundary thrust zone. Extension lineations are shown as lines with the direction of plunge indicated by the arrow heads. Plunges range from 10 to 40°. Stereonets show normals to foliation planes. The shaded area highlights a zone of strike-parallel lineations developed in the Huntsville thrust zone, adjacent to the topographic culmination in the Novar footwall. Black square is Huntsville village. *NTS* = Novar thrust sheet; *HTZ* = Huntsville thrust zone; *MCTS* = McCraney thrust sheet; *MLTS* = McClintock thrust sheet; *STS* = Seguin thrust sheet. Discussed in text.

sheet, development of the ductile thrust zones was accompanied by significant penetrative deformation of the crystalline thrust sheets themselves.

The Huntsville thrust zone

The Huntsville thrust zone is a 5 km thick zone of penetratively developed granoblastic mylonites (Fig. 2; Nadeau, 1988). The mylonites are flaggy, strongly lineated, fine-grained gneisses, derived by the deformation of an assemblage of coarse-grained plutonic and metasedimentary protoliths, under granulite to upper amphibolite facies conditions (Nadeau, 1990). Intense deformation has resulted in complete obliteration of original cross-cutting relationships and primary features. The rocks are concordantly layered and major lithological contacts are tectonic, as indicated by the preferential development of strain gradients at the margins of the least ductile layers and the absence of detectable mineralogical and chemical variation with strain. The latter observation suggests that deformation was approximately volume-constant and that the strain gradients are associated with a component of noncoaxial flow (Cobbold, 1977). The mylonitic rocks are interleaved with, or enclose, locally preserved lenses of their moderately strained gneissic protoliths, forming a map-scale pile of transposed, shallowly southward-dipping sheets (stereonet in Fig. 3), each one 10's-100's of metres thick.

Mylonitic rock units in the Huntsville thrust zone include straight gneiss (Hanmer, 1988), mylonitic orthogneiss and paragneiss, associated with

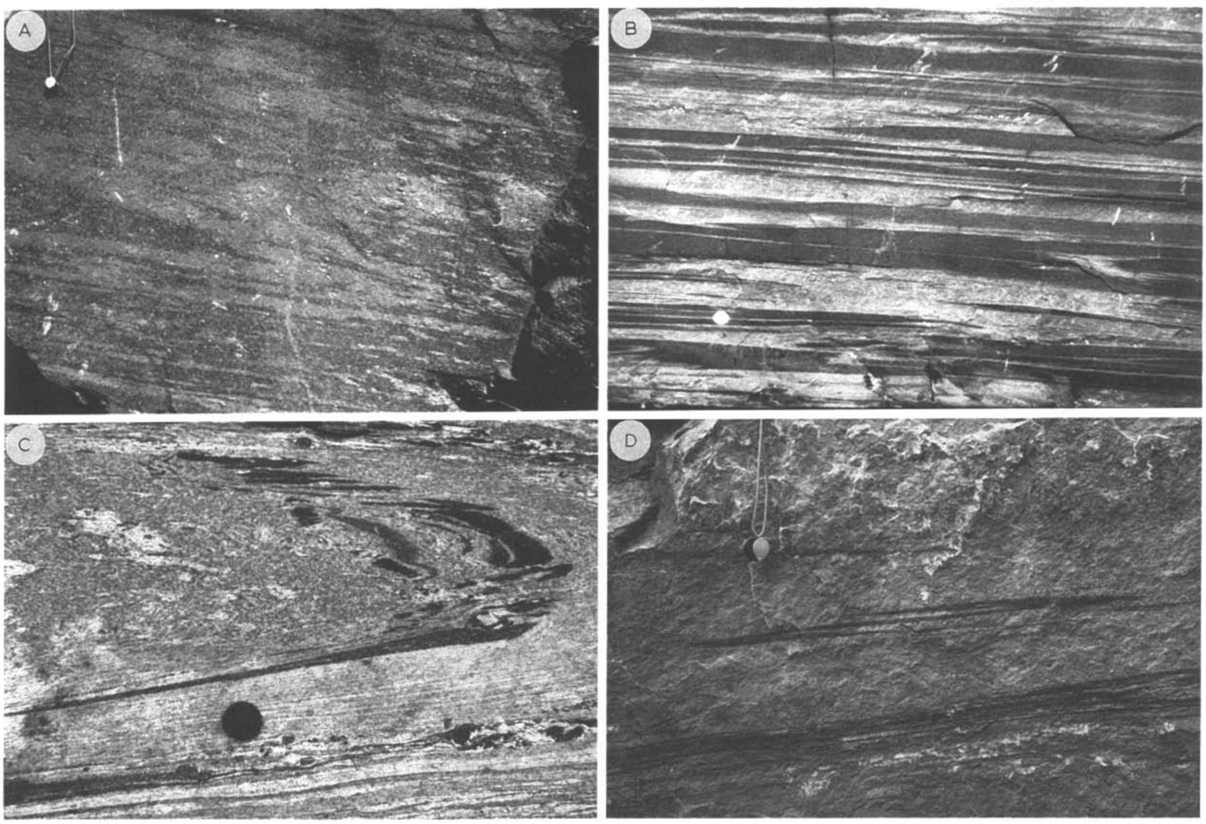

Fig. 4. Thrust zone tectonites. (A, B) Granulite facies straight gneiss derived from a mafic rock by the transposition of crosscutting granitic veins. (A) represents an intermediate stage, whereas (B) represents a penultimate stage of transposition. (C) A strain gradient and the development of straight gneiss by the transposition of mafic dykes (scale = 5 cm). (D) Boudined and isoclinally folded mafic layers in homogeneous, fine-grained, recrystallised, granulite facies orthogneiss. The axes of such folds are invariably parallel to the extension lineation. Such structures, together with the finely comminuted grain size and ribbon microstructure indicate intense and penetrative ductile strain in otherwise homogeneous felsic granulite.

boudined and transposed mafic dykes, and granitic veins and pegmatites. These will now be described in the following sections.

Straight gneiss

The following observations show that straight gneisses represent annealed mylonites and that the straight layering is of tectonic origin (Davidson et al., 1982; Hanmer, 1988). They are fine grained, flaggy S > L tectonites with a continuous, 1–10 centimetre-scale layering which is best expressed by the alternation of amphibolite and quartzofeldspathic rocks. In many places, the layering is visibly derived by the transposition and

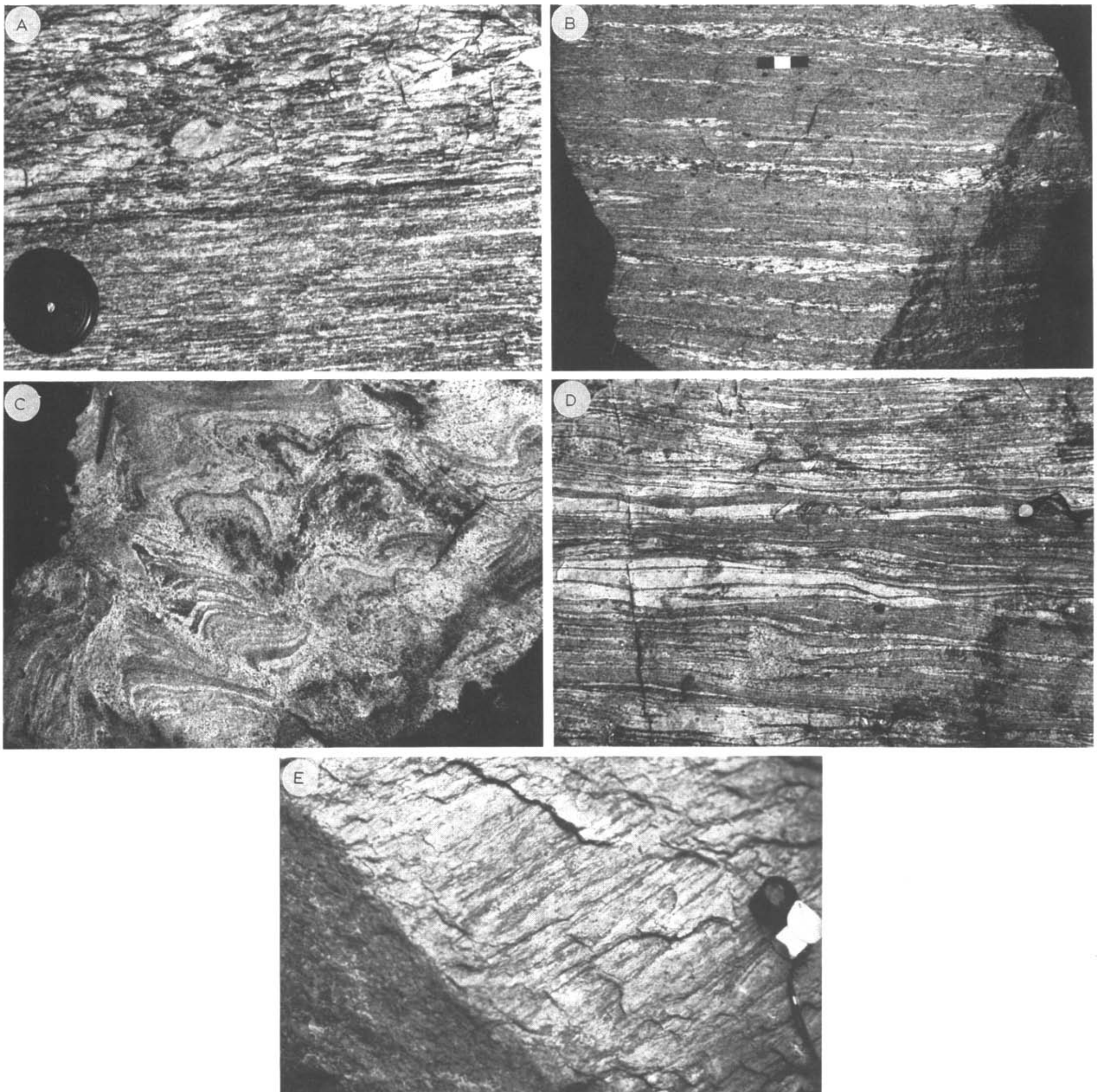

Fig. 5. (A) Mesoscopic strain gradient illustrating the progressive development of homogeneous, fine-grained, ribbon mylonite from mylonitic augen gneiss (scale = 5 cm). (B) Finely recrystallised, transposed and attenuated leucosomes in upper amphibolite facies, hornblende porphyroblastic biotite granitic gneiss. Scale bar is 3 cm long. (C) Metapelitic gneiss exhibiting relict primary layering disrupted by crosscutting and layer-parallel granitic veins. (D) Transposed equivalent of (C). (E) A typical example of the finite extension lineation developed in orthogneiss within the Seguin thrust sheet.

attenuation of pre-existing gneissic banding (Fig. 4A and B) and/or mafic and aplitic to pegmatitic granitic dykes (Fig. 4C). Subsequent to its development, the straight layering may be deformed about metre scale, tight to isoclinal intrafolial folds. Detached isoclinal fold hinges, coaxial with the extension lineation, are widespread (Fig. 4D). Such a geometrical relationship between finite strain and fold orientation has long been recognised as characteristic of high strain zones (e.g., Bell, 1978; Cobbold and Quinquis, 1980). This intense deformation is generally homogeneous at both outcrop and map scales.

The straight gneisses are fine-grained tectonites, in which grain size reduction was demonstrably the product of dynamic recrystallisation. The microstructure is annealed, with an equigranular polygonal or sub-polygonal granoblastic texture. Quartz crystals may form polycrystalline ribbons (aspect ratio 30 : 10 : 1). Grain-size typically ranges from 100 to 500 μm in the felsic layers, but tends to be coarser in the mafic bands. Such fine and uniform grain size in a now granoblastic rock suggests the existence of an earlier, even finer grained, dynamically recrystallised, mylonitic microstructure. This is supported by the existence of ca. 20 μm round quartz grains which occur both at the feldspar grain boundaries and triple points, and within the feldspar grains. Hanmer (1984) has interpreted this kind of microstructure as indicating grain growth of the feldspar component in order to reduce the surface energy of the aggregate.

Mylonitic orthogneiss

Several granitoid orthogneiss sheets are enclosed within the Huntsville thrust zone (Nadeau, 1985). In low-strain zones, the rocks show a relic igneous megacrystic structure, typically consisting of partly recrystallised, centimetre-size, stubby K-feldspar augen. Polycrystalline quartz is commonly drawn out in ribbons (50 : 10 : 1). A strong mineral foliation is invariably present and C/S fabrics (Berthé et al., 1979) are locally developed. With increasing strain in mesoscopic ductile shear zones, the augen structure is progressively obliterated, the K-feldspar augen are recrystallised and deformed into flat, sugary aggregates, or even disappear altogether (Fig. 5A). The end product is an homogeneous, strongly lineated, annealed mylonite.

Migmatitic quartzofeldspathic segregations, associated with mafic selvedges, are recrystallised, attenuated and transposed into parallelism with the overall foliation (Fig. 5B). Minor tight to isoclinal folds of mylonitic layering, coaxial with the finite extension lineation, similar to those previously described in straight gneiss, are not uncommon. Such transposed fabrics, and isoclinal folds of those same fabrics, independently suggest intense ductile strain. The strong flaggy foliation results from the parallel alignment of ribbon quartz, mafic minerals and feldspar augen or recrystallised aggregates. In low-strain areas, the stretching lineation is expressed as discrete elongate clots of recrystallised K-feldspar. In high-strain areas, it is expressed by streaked-out ribbon quartz and smeared-out trains of feldspar aggregates within the foliation.

These mylonitic orthogneiss sheets invariably show strain gradients at their margins, preserving no original contact relationships, such as intrusive apophyses or chilled margins. This, together with the strength of their L/S fabric, suggests that the concordant sheet-like shape of these bodies is, at least in part, the result of intense deformation.

Mylonitic paragneiss

Pelitic and semi-pelitic gneisses are widespread in the Huntsville thrust zone, interleaved with the mylonitic orthogneiss sheets and straight gneisses, and carrying competent calc-silicate fragments, boudined mafic and quartzofeldspathic layers, and detached fold closures as tectonic inclusions (Nadeau, 1985). These gneisses have experienced intense strain, consistent with that of the associated straight gneiss and mylonitic orthogneiss. In local low-strain zones, centimetre-scale layering, expressed by variations in the proportion of aluminosilicates, graphite and quartz, is disrupted and cut by short, discontinuous granitic veins (Fig. 5C). At higher finite strain magnitudes, these structures are attenuated and transposed parallel to the mineral foliation, contributing to a metamorphic layering and a strong planar fabric (Fig. 5D). As in the straight gneisses, the fine-grained

(100–500 μm), granoblastic microstructure and the small round quartz inclusions suggest that the dynamic grain size may initially have been even finer.

Boudined and transposed dykes

Isolated, widely separated, loaf-shaped mafic masses, 1–10's metres long, occur within the quartzofeldspathic gneisses. They are variably foliated and exhibit a strong marginal fabric concordant with that of their quartzofeldspathic matrix. Some of them are demonstrably boudined isoclinal fold hinges. Others, showing well-preserved coarse-grained igneous textures, clearly represent disrupted intrusive bodies (dykes?). Notable among them are widespread coronitic metagabbro bodies which, at the regional scale, tend to be concentrated in shear zones (Davidson and Grant, 1986; Davidson and Van Breemen, 1988). The size and the distribution of these widely separated boudins of originally intrusive mafic rocks suggest that the host tectonites have accommodated large ductile strains.

Similarly, heterogeneously deformed, crosscutting pegmatites pass progressively into sheared, mechanically disaggregated, porphyroclastic layers, concordant with the planar fabric of the wall rock. Such sheared pegmatites are widespread throughout the Huntsville thrust zone, independently demonstrating the penetrative nature of the intense deformation (Hanmer, 1988). The absence of displaced lithologic markers, or of widespread passive strain gauges, precludes estimation of the displacements across the zone.

The Novar thrust sheet

Along its northern margin, the Novar thrust sheet is underlain by a zone, several 100's of metres thick, of straight gneisses similar to those of the Huntsville thrust zone (Fig. 2; Culshaw et al., 1983). We suggest that this zone is also a ductile thrust zone and that the Novar thrust sheet represents the lowermost thrust sheet of the assembly being considered here. According to its strong aeromagnetic signature, the Novar thrust sheet extends for several kilometres beneath the northwest margin of the overlying Huntsville thrust zone (Nadeau, 1985, 1990). The tectonic fabric and internal structure of the Novar thrust sheet differ in intensity, style and orientation from those of the Huntsville thrust zone. The gneisses are distinctly coarser grained and less planar than in the Huntsville thrust zone. The primary texture of plutonic bodies is readily recognisable. Intrusive relationships of mafic dykes and relic primary layering in metasedimentary rocks are locally preserved. Layering and foliation are deformed about large (1–10 km wavelength), gently SSE-plunging folds throughout much of the southern part of the Novar thrust sheet (stereonet in Fig. 3). Although the finite extension lineation is parallel to the regional lineation trend, the planar fabric is discordant to, and truncated by, that of the Huntsville thrust zone (Figs. 2 and 3; Nadeau, 1990). Accordingly, the Novar thrust sheet has acted as an already deformed footwall with respect to the overlying thrust zone.

The McCraney and McClintock thrust sheets

The Huntsville thrust zone underlies an hanging wall, itself composed of thrust sheets separated by ductile thrust zones. Along their eastern flank, the Huntsville thrust zone and Novar thrust sheet are separated from the overlying McCraney thrust sheet by the McCraney thrust zone (Fig. 2). This thrust zone, approximately 1 km thick, comprises E- to NE-dipping straight gneisses of the type previously described for the Huntsville thrust zone, which carry a well-developed, easterly plunging finite extension lineation. The McCraney thrust zone truncates the contact between the Novar thrust sheet and the Huntsville thrust zone (Fig. 2). Therefore, the latter two structures constituted the already deformed footwall to the McCraney thrust zone. The overlying McClintock thrust sheet is separated from the McCraney thrust sheet by the shallowly SE-dipping McClintock thrust zone, a ductile shear zone with a down-dip finite extension lineation. The McClintock thrust zone reworks and/or truncates structure and lithological contacts within the underlying Huntsville thrust zone and the McCraney thrust sheet (Fig. 2). It also truncates the McCraney thrust zone. Accordingly, the structures below the McClintock thrust zone constituted an already deformed footwall at the time of emplacement of the McClintock thrust sheet. As a corollary to the foregoing, we note that the the McCraney and McClintock thrust sheets represent part of the reworked hanging wall to the Huntsville thrust zone.

The Seguin thrust sheet

The Seguin thrust sheet, part of the structural level 3 (Culshaw et al., 1983), is a gently SE-plunging synformal structure (stereonet in Fig. 3), bounded by the inward-dipping, Seguin ductile thrust zone (Figs. 1 and 2). The thrust sheet is primarily composed of migmatitic granodioritic gneisses, which grade locally into non-migmatitic amphibolite and granulite facies orthogneisses (Nadeau, 1985). The presence of orthopyroxene in granitic segregations indicates that migmatisation occurred, at least initially, under granulite facies conditions. Over a distance of 50 m, migmatitic gneisses near the base of the Seguin thrust sheet become progressively finer grained and transposed to form the annealed, mylonitic straight gneisses of the Seguin thrust zone. The mylonitic rocks of this thrust zone contrast compositionally with those of the immediately underlying Huntsville thrust zone. The Seguin thrust zone hosts a variety of tectonic inclusions, or boudins, including coronitic metagabbro and meta-anorthosite. As in the case of the Huntsville thrust zone, the isolation of these tectonic inclusions is a qualitative indicator of intense strain in the enclosing straight gneisses. From northwest to southeast, the Seguin thrust sheet successively overlies, reworks and/or truncates the structure within the Parry Sound thrust sheet, Novar thrust sheet, Huntsville thrust zone and McClintock thrust sheet (Figs. 1 and 2). Accordingly, the Seguin thrust sheet overrode an already deformed footwall comprising the Parry Sound thrust sheet and the Novar–McClintock assembly.

The emplacement of the Seguin thrust sheet was accompanied by intense internal deformation and syntectonic migmatite generation within the interior of the thrust sheet. Although similar relationships are also present in the McCraney thrust sheet (Nadeau, 1990), overprinting relationships between tectonic and metamorphic structures in the Seguin thrust sheet enable us to document the interplay between the internal deformation of the thrust sheet and its emplacement. The overall internal planar fabric of the thrust sheet, expressed by the preferred orientation of migmatitic segregations and mineral foliations, lies parallel to the straight gneisses of the Seguin thrust zone (stereonet in Fig. 3). The finite extension lineation, defined by drawn-out recrystallised K-feldspar augen, plunges shallowly to the southeast, coaxial with that in the thrust zones (Fig. 3). Open to tight minor folds, with axes parallel to the lineation, are common. Transposed migmatitic segregations, deformed about sheath folds coaxial with the extension lineation, are found within the thrust sheet, well above its base, attesting to considerable penetrative, ductile strain (Cobbold and Quinquis, 1980). Other, less deformed migmatitic segregations occur in the same outcrops as the sheath folds, indicating that deformation and migmatisation were synchronous. The orthopyroxene-bearing migmatitic segregations indicate that deformation occurred, at least in part, under granulite facies conditions. The concordance of the internal planar and linear structural elements with those of the regional array of thrust zones, suggests that deformation within the thrust sheet was kinematically related to thrusting. However, mechanical disaggregation and transposition of the migmatitic segregations as one passes into the Seguin thrust zone demonstrate that final emplacement of the Seguin thrust sheet outlasted migmatite generation and penetrative deformation within the thrust sheet.

Kinematics

Shear-sense criteria (rotated winged porphyroclasts, petrofabrics and strain insensitive shape fabrics; Nadeau, 1990; Hanmer, 1984, 1990; Hanmer and Passchier, 1991) throughout the study area are consistent with NW-directed thrusting along the SE-plunging finite extension lineation (Fig. 5E). In the absence of correlatable markers, it is not possible to estimate the displacements accommodated by the ductile thrust zones. An attempt to determine minimum structural throw using geothermobarometry was unsuccessful (Nadeau, 1990) because the observed downward increase in measured pressures is best interpreted in terms of post-thrusting re-equilibration of the metamorphic mineral assemblages. Nevertheless, a minimum displacement of 35 km for the Seguin thrust sheet is suggested by the separation between the northwest margin of the thrust sheet and the most southeasterly outcrop of its footwall (Figs. 1 and 2). However, this approach is only applicable in this case because of the lithological contrast between the thrust sheet and its footwall, and the assumption that the juxtaposition is entirely due to tectonic emplacement.

Although generally conformable with regionally extensive northwestward thrusting, tectonic flow in the Huntsville thrust zone does not conform to a simple picture, either in space or in time. Major departures of the lineation from the regional southeast trend are observed (Fig. 3). Whereas gently S- to SE-plunging lineations predominate in the upper section of the Huntsville thrust zone, the lineation azimuth swings to a NE–SW orientation at the base of the thrust zone (shaded area in Fig. 3). The following observations suggest that all of the finite extension lineations were developed synchronously within the same kinematic framework: (1) no overprinting of lineations is observed; (2) the change in the lineation direction is progressive, regionally extensive and focused upon the NE-trending segment of the interface between the Novar thrust sheet and Huntsville thrust zone; (3) mesoscopic shear-sense indicators are consistent with northwestward thrusting, even where the extension lineation is NE–SW oriented; (4) the finite extension direction deduced from quartz crystallographic fabrics in the lower level of the thrust zone (Fig. 6) is parallel to the regional NW–SE trend and normal to the anomalously NE–SW oriented mesoscopic finite extension lineation (shaded area in Fig. 3; Nadeau, 1990). The sensitivity of quartz crystallographic fabrics to changes in kinematic boundary conditions (e.g., Lister and Paterson, 1979; Lister and Williams, 1979; Lister and Hobbs, 1980; Schmid and Casey, 1986; Law, 1986; Jessell, 1988) enables them to record the orientation of the latest increments of finite strain. However, the mesoscopic finite extension lineation, attached to material points, is more likely to preserve a more significant component of the deformation history. This suggests that the last increments of finite strain recorded by the quartz petrofabrics (Fig. 6) reflect the same kinematic framework as the regional northwestward thrusting observed elsewhere within the study area.

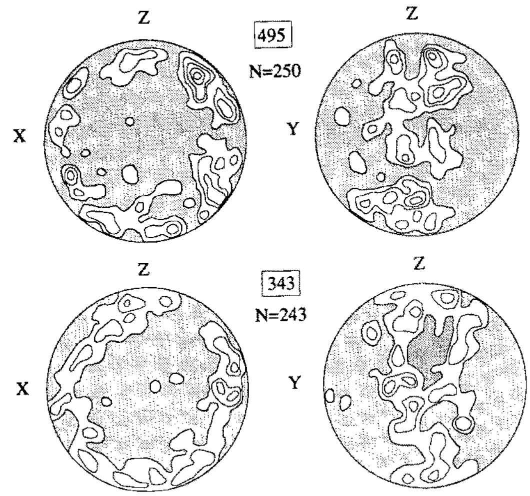

Fig. 6. Lower hemisphere projections of quartz petrofabrics measured from specimens from strike-lineated tectonites in the basal section of the Huntsville thrust sheet (shaded area in Fig. 3). X, Y and Z are the directions of the principal finite strains, taking the mylonitic foliation to approximately record the XY plane of the finite strain ellipsoid and the extension lineation as parallel to X. Note that the left hand stereograms indicate that the direction of maximum extension recorded by the petrofabrics is parallel to the intermediate (Y) direction of the finite strain ellipsoid, as recorded in the macroscopic planar and linear fabrics. Specimen numbers (boxed) and number of data points (N) are given.

Conditions of thrusting

Geothermobarometric measurements of ca. 8–10 kbar at temperatures > 700°C (Grant, 1987; Manojlovic, 1987; Anovitz and Essene, 1990) were obtained from annealed, granoblastic tectonites, sampled from the Huntsville thrust zone, and the Novar and the McClintock thrust sheets. These data are representative of the thermobarometric results obtained throughout the region between the Parry Sound thrust sheet and the Central

Metasedimentary Belt. However, as noted by Anovitz and Essene (1990), the geothermobarometric data do not reflect the existence of the geologically established crustal-scale thrust zones (see also Hanmer, 1988; Hanmer and McEachern, 1992). Accordingly, the measured pressures reflect the depths of post-thrusting re-equilibration. In view of the uplift which can be expected to follow crustal thickening (e.g., Molnar and Lyon-Caen, 1988), the depths recorded by these pressures are minimum estimates for the depth at which thrusting took place. This interpretation derives some support from the pre-thrusting emplacement of coronitic metagabbro in the Seguin thrust sheet and the Huntsville thrust zone at ca. 1.17 Ga (Davidson and Van Breemen, 1988), at depths of ca. 30–40 km (Grant, 1987).

U-Pb ages of monazite from syntectonic granitic veins and pegmatites within the Huntsville thrust zone record cooling to temperatures below ca. 700°C (Parrish, 1990) at ca. 1.065 Ga (Nadeau, 1990). U-Pb ages of zircon from the same intrusions record the magmatic crystallisation of the veins at ca. 1.08 Ga. The Seguin thrust zone is crosscut by a granitic pegmatite which has yielded a U-Pb magmatic crystallisation age from zircons at ca. 1.1 Ga (Nadeau, 1990), which, therefore, represents a minimum age for the syntectonic granulite facies metamorphic assemblages within the Seguin thrust sheet. These data imply that the rocks of the study area were held at high metamorphic temperatures throughout the interval ca. 1.1–1.065 Ga. Moreover, granulite facies metamorphism in the Parry Sound thrust sheet (Fig. 1) occurred prior to the end of thrust emplacement of the sheet at ca. 1.16 Ga (Van Breemen et al., 1986). This suggests that the Central Gneiss Belt experienced high metamorphic temperatures during an interval of ca. 100 Ma.

Discussion

Scale of intense penetrative deformation

Compared with descriptions of other high-grade thrust belts, the thickness of the penetratively developed mylonites in the Central Gneiss Belt beneath the Central Metasedimentary Belt, and the scale of penetrative deformation of the crystalline thrust sheets, are both remarkable. The thickness of the Huntsville thrust zone mylonites is comparable with that of similar structures developed at upper amphibolite to granulite facies in the Limpopo Belt, southern Africa (McCourt and Vearncombe, 1987), the Seve nappe, Scandinavia (Williams and Zwart, 1978), and in the Inverian–Nagssugtoquidian–Laxfordian of northwestern Scotland (Coward and Park, 1987) and Greenland (e.g., Grocott, 1979; Myers, 1987). However, individual mylonite belts developed in other amphibolite facies ductile thrust zones are generally much narrower, ranging from 1 to 100's m (Sibson et al., 1979; Simpson, 1982; Hodges et al., 1982; Rathbone et al., 1983; Lacassin, 1987; Butler and Prior, 1988; Hanmer, 1988; James et al., 1989; Lucas, 1989; Hanmer and McEachern, 1992), to as thick as several kilometres (Bell and Etheridge, 1976; Collins and Teyssier, 1989). To our knowledge, the only other description of a large thrusted body, whose internal synthrusting deformation is directly comparable with the penetrative development of intense planar and linear fabrics deformed by isoclinal sheath folds of the 5-km-thick Seguin crystalline thrust sheet, is that by Williams and Zwart (1978) of the Seve nappe. We suggest that the great thickness of penetrative mylonitisation in the Huntsville thrust zone reflects, in part, the high ambient metamorphic temperatures (e.g., Sibson, 1977; Hanmer, 1988). However, we note that the Central Metasedimentary Belt boundary thrust zone contains granitic mylonite belts, hundreds of metres thick, which anastomose around lenticular crystalline thrust sheets of tonalitic, amphibolitic and syenitic composition (Hanmer, 1988; Hanmer and McEachern, 1992). Accordingly, we suggest that the development of mylonite within the Huntsville thrust zone also reflects the rheological homogeneity of the deforming media and the consequent absence of strain partitioning (e.g., Bell, 1981, 1985) within the ductile thrust zone. Similarly, the rheological response of the Seguin thrust sheet implies that the partitioning of deformation between it and the bounding thrust zones was relatively subtle, at least during the early part of its emplacement history.

The obvious exception to the foregoing is the Novar thrust sheet. The discordance between its internal structure and its contacts with the superincumbent thrust zones strongly suggests that it behaved as a stiff element during the emplacement of the overlying thrust sheets and, perhaps, even during its own initial emplacement.

Flow

We have shown that mylonitic fabrics are penetratively developed within the Huntsville thrust zone during northwestward thrusting. However, the orientation pattern of the finite extension lineation does not directly reflect the regional-scale thrust direction. The greatest divergence of the lineation azimuth from the regional southeast trend occurs in the basal section of the thrust zone, opposite the NE–SW segment of the interface between the Huntsville thrust zone and the relatively stiff Novar thrust sheet. We have also shown that the finite extension lineations were developed synchronously within the same kinematic framework, irrespective of their orientation. An explanation of the lineation pattern within the Huntsville thrust zone (Fig. 3) must account for all of these observations. We suggest that the step-like Huntsville–Novar interface, oriented at an high angle to the direction of regional flow, induced a significant component of NW–SE shortening across the flow plane, which was accommodated by subhorizontal NE–SW extension. This is reflected in the progressive change in the orientation of maximum finite extension, from upper to lower structural levels within the Huntsville thrust zone. However, a component of NW–SE-oriented extension is spatially associated with the anomalous NE–SW finite extension lineation. It is expressed both as macroscopic indicators of northwestward thrusting, and as crystallographic quartz fabrics. We suggest that this association indicates the contemporaneous operation of northwestward thrusting and NE–SW maximum stretching. In other words, the flow regime adjacent to the Novar thrust sheet is one of transpression (Sanderson and Marchini, 1984), induced by the buttress effect of the relatively stiff footwall to the Huntsville thrust zone..

Breakback thrusting

We have shown that the sequence of emplacement of the Novar, McCraney, McClintock and Seguin thrust sheets involves the successive accretion of each of the thrust sheets to the footwall of the subsequent overlying hanging wall. This type of stacking sequence is similar to that commonly referred to as "out-of-sequence" thrusting (e.g., Butler, 1987; Morley, 1988; Lucas, 1989). However, this term is employed in contrast to "normal-sequence" (piggy-back) thrusting, so commonly observed in foreland fold-thrust belts where a cover sequence is generally detached from its underlying crystalline basement (e.g., Boyer and Elliot, 1980; Butler, 1982). Other workers have described natural examples of out-of-sequence thrusting (e.g., Hoffman et al., 1977; Coward, 1980; Butler, 1983, 1987; Butler and Coward, 1984; Searle, 1986; Morley, 1988; Lucas, 1989). Lacking stratigraphic control, and without a great deal more geochronological data, we are not able to determine the existence of regular-sequence thrusting in the Central Gneiss Belt. Accordingly, we prefer here to employ the term "breakback thrusting" (e.g., Morley, 1988), because it describes the location of successively active thrust zones within a reference frame given by the orientation and position of older structures and the sense of shearing.

The stacking of thrust sheets from Novar to Seguin is a breakback sequence. The relative timing of thrusting between Parry Sound and Seguin thrust sheets is also a break-back relationship. However, because they are nowhere seen in mutual contact, we do not know the relative age relationships between the emplacement of the Parry Sound thrust sheet (structural level 2) and the Novar–McClintock assembly (part of structural level 1; Fig. 2). Therefore, for the purposes of this discussion, it suffices to consider them together as components of an already deformed footwall to the Seguin thrust sheet. To the east, the Seguin thrust zone passes in continuity into the "Kawagama zone" (Fig. 2; Culshaw et al., 1983). The Kawagama zone is a belt of dip-lineated, shallowly SE-dipping straight gneisses. The Seguin–Kawagama structure (structural level 3) truncates all of the thrust zones in the underlying footwall, from Parry Sound in the west to McClintock in the east (Fig. 2). Furthermore, the ca. 1.08–1.05 Ga thrusting event in the overlying Central Metasedimentary Belt boundary thrust zone (Fig. 2) is younger than the final emplacement of structural level 3. Accordingly, all of the thrusts represented in Figure 2 are part of a break-back stacking sequence, from structural levels 1 and 2 to the Central Metasedimentary Belt boundary thrust zone.

This break-back thrusting sequence can now be placed into the geodynamic context of the Ontario Grenville. Recent interpretations of the Central Metasedimentary Belt suggest that it includes a continental magmatic arc and a back-arc basin (e.g., Brown et al., 1975; Condie and Moore, 1977; Fletcher and Farquhar, 1982; Pride and Moore, 1983; Holm et al., 1986; Smith and Holm, 1987, 1990; Harnois and Moore, 1991; McEachern and Van Breemen, 1992). The back-arc basin would have opened in the northwestern part of the Central Metasedimentary Belt at ca. 1.3 Ga (Smith and Holm, 1990) and closed by ca. 1.19–1.18 Ga (Hanmer and McEachern, 1992). Recent structural and geochronological work in the Central Metasedimentary Belt boundary thrust zone demonstrates that the Central Metasedimentary Belt was initially emplaced as a 200 + km long hanging wall over the Central Gneiss Belt by ca. 1.19–1.18 Ga (McEachern, 1990; Hanmer and McEachern, 1992). Accordingly, our present study has documented the shortening of the Laurentian margin adjacent to the proposed back-arc basin within the Central Metasedimentary Belt, i.e., a continental-scale footwall to the ca. 1.19–1.18 Ga Central Metasedimentary Belt boundary thrust zone. The thrusting within the footwall began at ca. 1.16 Ga at granulite facies with the emplacement of the Parry Sound thrust sheet, below structural level 3 (Fig. 2). We take the high metamorphic grade of this deformation to reflect the thermal perturbation caused by the earlier thrust emplacement of the overlying Central Metasedimentary Belt (McEachern, 1990; Hanmer and McEachern, 1992). The break-back sequence culminated at ca. 1.08–1.05 Ga with renewed major thrusting along the Central Metasedimentary Belt boundary thrust zone which resulted in the final emplacement of the Central Metasedimentary Belt as a discrete, coherent > 200 km long thrust sheet (Hanmer and McEachern, 1992).

Thermal consequences of breakback thrusting

Geochronological constraints indicate that, at the scale of structural levels 2 and 3 (Fig. 2), the rocks of the Central Gneiss Belt experienced temperatures in excess of ca. 700°C for between 50 and 100 Ma (ca. 1.16–1.06 Ga). Geobarometric data indicate that thrusting occurred at depths of the order of 30–40 km or more. Compared with recent studies of metamorphic thrust belts, such a long residence period at such crustal depths is apparently anomalous (Selverstone, 1985, 1988; Zeitler et al., 1989; St-Onge and Lucas, 1991). From observation of modern mountain belts, tectonic crustal thickening is accompanied by the generation of positive topography, rapid erosion and exhumation of rocks (e.g., England and Molnar, 1990b). In order to maintain approximately constant metamorphic conditions, at mid-crustal levels in a double thickness crust, the thickness of material removed by erosion, and/or extensional faulting (e.g., Platt, 1986), must be restored at a structural level above the rocks in question. In the case of our study area, constant crustal depth could have been maintained by the continued emplacement of crystalline thrust sheets at higher structural levels. This scenario is consistent with, and predictable from, our structural observations of a break-back stacking sequence. Moreover, geochronological data demonstrate that the break-back thrusting occurred over the same time interval: 1.16–1.06 Ga.

Conclusions

The post-ca. 1.16 Ga deformation of the Central Gneiss Belt of the Grenville orogen, Ontario, represents the tectonic shortening of the continental-scale footwall to the Central Metasedimentary Belt, during the closure of a possible back-arc basin. Shortening of the footwall commenced after the initial emplacement of the Central Metasedimentary Belt by ca. 1.19–1.18 Ga. Thrusting occurred in a break-back stacking sequence which culminated with renewed major deformation within the Central Metasedimentary Belt boundary thrust zone at ca. 1.08–1.05 Ga.

The individual mylonite belts which constitute the ductile thrust zones, and the scale of penetrative deformation of the intervening crystalline thrust sheets, are comparable with the largest known examples of other high-grade thrust belts. Penetrative development of intense deformation on such a large scale is, in part, a reflection of the high ambient syntectonic temperatures, but also of the absence of significant large-scale rheological contrasts within the actively deforming material. Penetrative ductile flow in high-temperature thrust zones may be complex and reflect local boundary conditions, such as the rheological behaviour of the footwall and the configuration of interfaces within the thrust stack.

Regional-scale break-back thrusting can influence the exhumation and cooling history of a mid- to deep-crustal thrust stack, developed in strongly thickened continental crust. Such a stacking sequence may restore the thickness of crustal overburden removed by erosion, and thereby contribute significantly to retarding exhumation and cooling of mid-to deep-crustal rocks.

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