Kinematical and rheological evolution of a crustal-scale ductile thrust zone, Central Metasedimentary Belt, Grenville orogen, Ontario[1]

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The Central Metasedimentary Belt boundary thrust zone is a 10 km thick, 200+ km long, stack of crystalline thrust sheets, enclosed by an anastomosing network of ductile thrust zones, formed at mid- to deep-crustal depths in the southwest Grenville orogen, Ontario. It has behaved as a coherent upper amphibolite facies thrust zone, accommodating northwestward transport of the Central Metasedimentary Belt, the largest lithotectonic entity in this part of the orogen, by coherent and contemporaneous displacements. The earliest thrusting was well under way by ca. 1.19–1.18 Ga and the boundary thrust zone was reactivated at ca. 1.08–1.05 Ga. The early thrusting records the closure of a back-arc basin within the Central Metasedimentary Belt, which closed at ca. 1.19–1.18 Ga. The younger thrusting may reflect continental collision to the southeast of the exposed Grenville and represent intraplate reactivation of the boundary thrust zone, which acted as an older, crustal-scale zone of weakness. Transverse mid- to deep-crustal thrusting was apparently contemporaneous with longitudinal (orogen-parallel) shearing at higher structural levels. The rheological behaviour of the deforming media may have influenced the localization of both the upper and lower limits of the boundary thrust zone at the time of its initiation. The upper limit coincides with a chain of relatively stiff metagabbro bodies, which may have acted as a barrier to the upward migration of fluids responsible for syntectonic nephelinization at the top of the thrust zone.

La zone de chevauchement marginale de la Ceinture métasédimentaire centrale comprend un empilement de nappes cristallines qui mesure 200+ km de long sur 10 km d'épaisseur. Les nappes sont délimitées par un réseau de cisaillements ductiles, formé à des profondeurs importantes dans la croûte continentale de la partie sud-ouest de l'orogène du Grenville, Ontario. La zone chevauchante s'est comportée en tant que structure cohérente, au faciès amphibolite, lors du déplacement vers le nord-ouest de la Ceinture métasédimentaire centrale, la plus grande unité tectonique dans toute cette partie de l'orogène. Le chevauchement à commencé avant ca. 1,19–1,18 Ga et a repris vers ca. 1,08–1,05 Ga. La phase précoce du chevauchement représente la fermeture d'un bassin marginal à l'intérieure de la Ceinture métasédimentaire centrale qui s'est achevée vers ca. 1,19–1,18 Ga. Le chevauchement tardif correspondrait à la collision continentale qui aurait eu lieu au sud-est du Grenville connu, et représenterait ainsi la réactivation d'une ancienne zone de faiblesse crustale. Une composante de chevauchement transversal aurait eu lieu en profondeur, en même temps que du cisaillement longitudinal dans les niveaux plus superficiels de l'orogène. Le comportement rhéologique des roches aurait influencé la localisation des deux limites (supérieure et inférieure) de la zone chevauchante lors de son initiation. La limite supérieure est marquée par un chaplet de corps gabbroïques qui auraient présentés une barrière aux fluides responsables de la néphélinisation dans la partie supérieure de la zone chevauchante.

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Introduction

The processes by which convergence is accommodated in the deep structural levels of the continental crust in collisional orogens are the subject of much debate, even in modern orogenic belts (e.g., Dewey *et al.* 1988, 1989). A significant contribution to this debate can be made by the study of ancient analogues where sufficient time has elapsed to allow exposure of the middle to lower crust. The Mid-Proterozoic (ca. 1.25–1.0 Ga) Grenville orogen in Ontario is deeply excavated and accessible at the appropriate structural level (Fig. 1; Moore *et al.* 1986). This part of the Grenville orogen is currently interpreted in terms of collision between the Archean Superior–Southern continent to the northwest, a magmatic arc, and an as yet unknown continent to the southeast (e.g., Windley 1986; Culotta *et al.* 1990). Regional deformation within the orogen is principally manifested in the northwestward displacement of south-east-dipping crystalline thrust sheets, separated by granulite to upper amphibolite facies ductile thrust zones (Davidson *et al.* 1982; Culshaw *et al.* 1983; Davidson 1984; Hanmer 1988; Nadeau and Hanmer 1992).

Our work has focussed upon the nature and timing of the progressive deformation within both ductile thrust zones and crystalline thrust sheets (Hanmer and Ciesielski 1984; van Breemen and Hanmer 1986; Hanmer *et al.* 1986; Hanmer 1988, 1989; McEachern 1990; Nadeau 1990; Nadeau and Hanmer 1992). We have examined the structural evolution of the lower boundary of the Central Metasedimentary Belt in Ontario (Fig. 2). There, we have been able to directly observe the scale and nature of the response of the deep-seated continental crust during continental shortening, and to assess the role of rheologically distinctive horizons in the localization, initiation, and structural evolution of thrusting.

In this paper we show that the base of the Central Metasedimentary Belt is a 10 km thick by at least 200 km long stack of lenticular crystalline thrust sheets enclosed by an anastomosing network of ductile shear zones (Fig. 2). The stack, the

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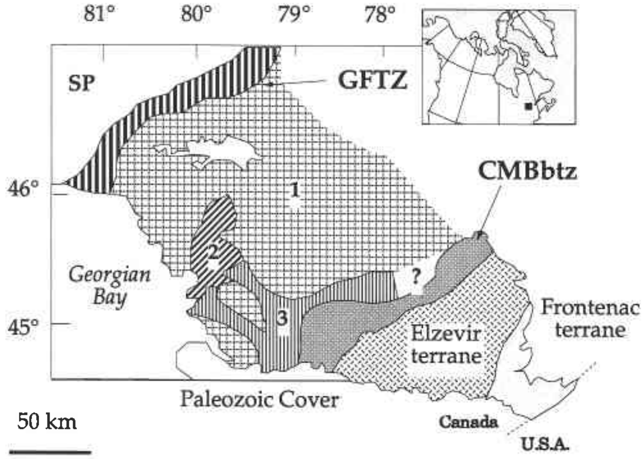

FIG. 1. Schematic illustration of the distribution of the major structural elements of the Central Gneiss Belt (1−3), of the Grenville orogen in Ontario, bounded by the Grenville Front Tectonic Zone (GFTZ) to the northwest and by the Central Metasedimentary Belt boundary thrust zone (CMBbtz) to the south. Numbers refer to structural levels in the Central Gneiss Belt mentioned in the text and indicate the stacking geometry from structurally lowest to structurally highest. Tectonic boundaries in the Central Gneiss Belt are after Wynne-Edwards (1972) and Culshaw *et al.* (1983). The Central Metasedimentary Belt is divided into the Frontenac and Elzevir terranes (after Davidson 1986). SP, Southern Province.

Central Metasedimentary Belt boundary thrust zone (referred to as the boundary thrust zone herein), is structurally distinct from the rocks above and below. We propose that it has acted as a coherent thrust zone on at least two occasions, accommodating northwest-directed thrusting of the Central Metasedimentary Belt by contemporaneous displacements all along its strike length. We suggest that the localization and initiation of the upper and lower limits of the boundary thrust zone were apparently influenced by the rheological behaviour of laterally extensive horizons of rheologically stiff gabbros and weak supracrustal rocks, respectively. Finally, we propose that the initiation of the boundary thrust zone represents the closure of a marginal basin, within the Central Metasedimentary Belt, by ca. 1.18 Ga. Subsequent thrusting at ca. 1.08−1.06 Ga may reflect major reactivation of the boundary thrust zone, which acted as an older, crustal-scale zone of weakness.

Southwest Grenville orogen

During the past 20 years, workers seeking to subdivide and classify geological elements in the Grenville orogen, especially in Ontario, have introduced a plethora of names (plutonic series, belts, domains, subdomains, terranes) that are often difficult to equate with modern tectonic terminology. Although, for completeness, we will briefly review the existing subdivisions, we prefer to employ meaningful genetic terms for those features whose structural significance can be demonstrated.

The southwestern Grenville orogen in Ontario is classically divided into three major southeast-dipping, structurally overlapping zones or belts (Fig. 1). These are, from northwest to southeast, (*i*) the Grenville Front Tectonic Zone, (*ii*) the Central Gneiss Belt, and (*iii*) the Central Metasedimentary Belt, all flanked to the southeast by the Adirondack Highlands (Wynne-Edwards 1972). In a modification of this scheme, Rivers *et al.* (1989) refer to the Central Metasedimentary Belt as an allochthonous monocyclic belt, flanked to the northwest by an allochthonous polycyclic belt and a parautochthonous belt.

The Grenville Front Tectonic Zone is structurally overlain by the Central Gneiss Belt and separates it from the pre-Grenville Southern and Superior Province rocks to the northwest. Although the Grenville Front Tectonic Zone has been recognized for the past 20 years (Wynne-Edwards 1972), it is poorly understood over much of its strike length, even in its better studied southwestern parts (Henderson 1972; La Tour 1981; Ciesielski 1988; Davidson and Bethune 1988; Indares and Martignole 1989). It appears that the Grenville Front Tectonic Zone is a broad, southeast-dipping thrust zone within which dip-slip ductile shear zones occur, each up to several hundreds of metres thick (Davidson and Bethune 1988; Green *et al.* 1988).

The overlying Central Gneiss Belt has been subdivided into a number of "domains" and "subdomains" (e.g., Davidson *et al.* 1982; Culshaw *et al.* 1983; Davidson 1984). In a modification of this scheme, the belt can also be considered in terms of three superincumbent structural levels (numbered 1−3 in Fig. 1), which are internally subdivided into individual crystalline thrust sheets, or thrust stacks (Culshaw *et al.* 1983; Nadeau 1990; Nadeau and Hanmer 1992). The individual thrust sheets are up to 5 km thick (cf. Lindia *et al.* 1983), with lateral dimensions of approximately 20 km, measured in the shear plane, perpendicular to the direction of northwestward transport. They are bounded along their bases by ductile thrust zones, up to several hundreds of metres thick. The thrust sheets are laterally discontinuous and may be deformed into broad moderate folds about gently southeast-plunging axes. Accordingly, the map pattern of the thrust zones is sinuous to curvilinear (Fig. 1). The enveloping surfaces of the thrust zones dip gently to the southeast and their foliations carry a generally southeast-plunging extension lineation (cf. Nadeau and Hanmer 1992). However, crosscutting relations between thrust sheets, contrasts in metamorphic grade, and geochronological data indicate that deformation within individual structural levels was not synchronous (Culshaw *et al.* 1983; Nadeau 1990; Nadeau and Hanmer 1992). Thus, the largest coherent structural entities within the Central Gneiss Belt are the individual thrust sheets themselves.

Thrust zone, footwall, and hanging wall

Compared with the scale of the individual thrust sheets in the underlying Central Gneiss Belt, the Central Metasedimentary Belt is the largest discrete lithotectonic entity in the Ontario Grenville orogen (Fig. 1). It is made up of marbles, quartzites, and mafic to intermediate subaqueous volcanic rocks, intruded by a variety of gabbroic, tonalitic, granitic, and syenitic plutons (e.g., Davidson 1986). Its structural base, marked by the Central Metasedimentary Belt boundary thrust zone (Hanmer and Ciesielski 1984), is the focus of our study. Except for the section between Bancroft and the Glamorgan thrust sheet, the boundary thrust zone as depicted in Fig. 2 closely approximates the area referred to as the Bancroft terrane by other workers, and its upper limit corresponds to the contact between the Bancroft and the Elzevir terranes (e.g., Davidson 1986; Fig. 1). Accordingly, the northwestern margin of the Elzevir terrane is a hanging wall, whose base we shall define (see below) at an inferred, rheologically competent horizon, represented by a string of metagabbro bodies, extending from Trooper Lake to the Ottawa River (Fig. 2). The lower limit of the boundary thrust zone is defined where it cuts across deformation structures in the underlying Central Gneiss Belt, which therefore constitutes a footwall (Hanmer 1989; Nadeau and Hanmer 1992).

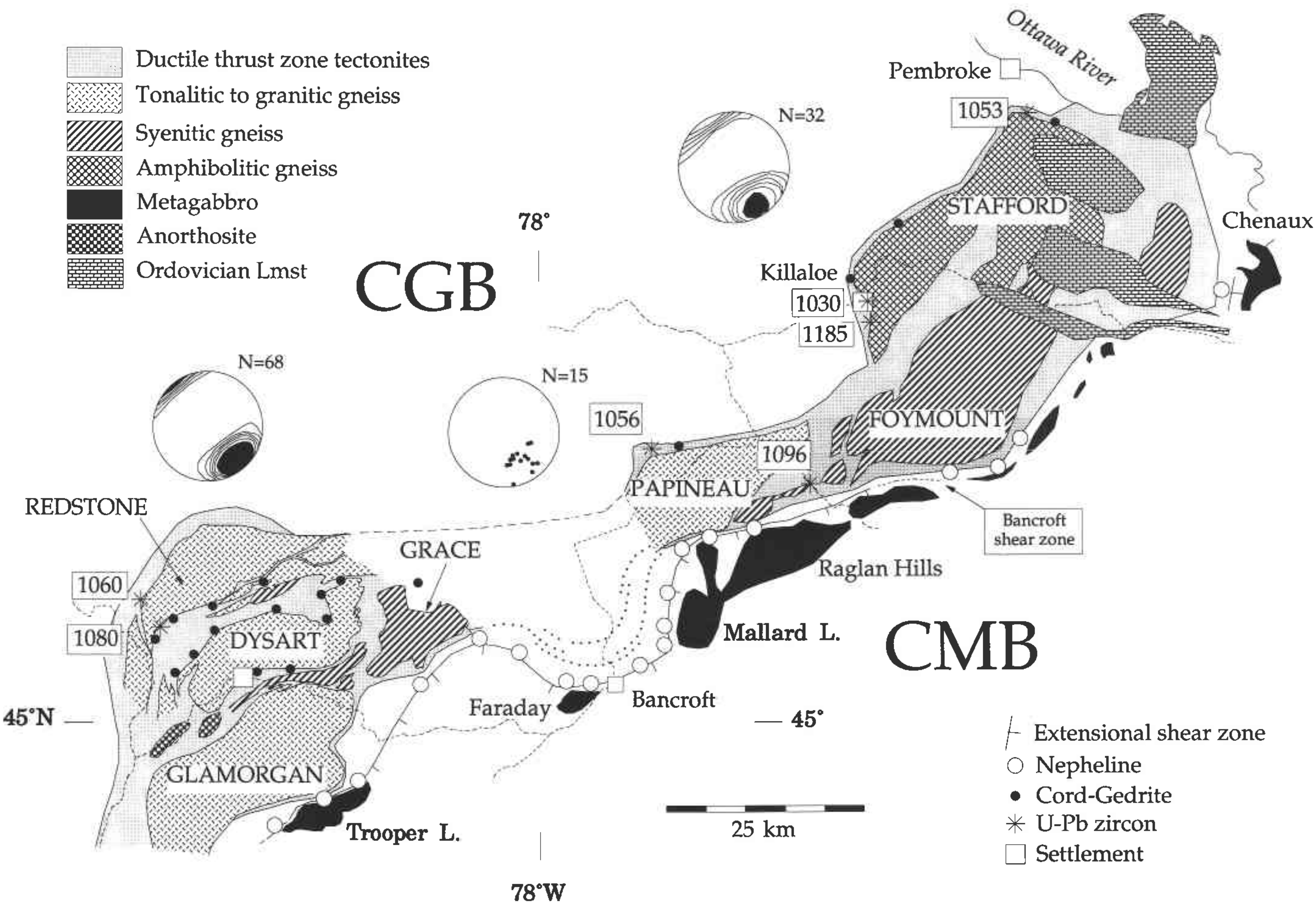

FIG. 2. The structural geology of the Central Metasedimentary Belt boundary thrust zone, a 200+ km long stack of crystalline thrust sheets surrounded by anastomosing ductile thrust zone tectonites. The two light grey portions are the Haliburton (southwest) and Killaloe (northeast) segments, separated by the Bancroft segment. The three stereonets are of finite extension lineations for each of the three segments. Foliations (not shown; see Hanmer and Ciesielski 1984; Hanmer 1988; McEachern 1990) are concordant to lithological contacts and dip monotonously at about 20° to the southeast. CGB, Central Gneiss Belt; CMB, Central Metasedimentary Belt. U−Pb zircon ages, mentioned in the text, are given here in Ma. The unlabelled square on the south margin of the Dysart thrust sheet in the Haliburton segment is Haliburton village. The ticked line is the Bancroft shear zone (indicated). The dotted lines are structural traces in the Bancroft segment. Broken lines are principal roads. Discussed in text.

Boundary thrust zone

The lower boundary of the Central Metasedimentary Belt in Ontario is marked by a gently southeast-dipping zone of planar, highly strained granoblastic tectonites, 8−10 km thick and over 200 km long, which crosses the Ottawa River near Pembroke and extends to the west-southwest, beneath the Phanerozoic cover south of Haliburton (Fig. 2) (Hanmer and Ciesielski 1984). The tectonites are divided into three principal types: (*i*) straight gneisses, (*ii*) porphyroclastic gneisses, and (*iii*) marble tectonic mélange or breccia (Hanmer 1988, 1990). The derivation of the medium-grained (1−2 mm) straight and porphyroclastic gneisses by the deformation, mechanical degradation, and transposition of megacrystic and pegmatitic granitoid protoliths, intrusive into, or intruded by, mafic protoliths has been described in detail elsewhere (Davidson *et al.* 1982; Davidson 1984; Hanmer 1988, 1990; McEachern 1990; Nadeau 1990; Nadeau and Hanmer 1992). To all intents and purposes, they are thoroughly annealed, granoblastic mylonites. Detailed description of the formation of the coarse-grained (≥2 cm) marble tectonic mélange, by flow and injection of the marble around and into bodies of gneiss and granitoid, has also been presented elsewhere (Hanmer and Ciesielski 1984; Hanmer 1988, 1990).

The straight gneisses and porphyroclastic gneisses carry a remarkably constant downdip extension lineation (Fig. 2), expressed as an alignment of minerals and intrafolial isoclinal fold hinges, or by the streaking-out of polycrystalline quartzo-feldspathic aggregates. It is also well developed in the marble tectonic mélange as a strong alignment of gneissic and plutonic inclusions, wherever the flow in the carbonate matrix was laminar. The presence of shear-sense indicators (Hanmer and Passchier 1991) demonstrates that these highly strained granoblastic tectonites are the products of northwest-directed overthrusting along the general direction of the southeast-plunging extension lineation (Hanmer 1984, 1988; Hanmer and Ciesielski 1984; McEachern 1990). However, because other, older thrust zones within the southwestern Grenville orogen are also northwest directed (Davidson 1984; van Breemen *et al.* 1986; Davidson and Bethune 1988; Nadeau 1990; Nadeau and Hanmer 1992), such kinematic consistency does not reflect the scale at which displacements were either contemporaneous or coherent.

We have undertaken detailed examination of two different parts of the Central Metasedimentary Belt boundary thrust zone in order to establish the internal geometry, the kinematic history, and the timing of deformation of the boundary thrust zone as a whole. We shall refer to these two study areas as the Haliburton (Hanmer 1988) and Killaloe (McEachern 1990) segments of the boundary thrust zone, separated from one another by the Bancroft segment (Fig. 2). The Haliburton and Killaloe segments were selected for study because of the quality of both exposure and access, and because initial reconnaissance studies (Lumbers 1982; Hanmer and Ciesielski 1984) indicated the potential for a close link between lithology and structure, which would facilitate recognition of smaller scale crystalline thrust sheets within the overall boundary thrust zone.

It is appropriate here to point out that the thrust sheets are defined according to two criteria (Hanmer 1988; McEachern 1990). On the one hand, the thrust sheets correspond to volumes of crystalline gneiss, which contrast compositionally with the highly strained ductile tectonites that enclose them. On the other hand, qualitative indicators of strain intensity, such as fold tightness, crosscutting relations, and grain size show that the magnitude of finite strain within the interior parts of the crystalline thrust sheets is much lower than that in the bounding tectonites. However, in the absence of a material discontinuity (i.e., a brittle fault or its discrete plastic equivalent) at the compositional boundary, strain gradients within the outer parts of the crystalline thrust sheets tend to blur the otherwise simple correspondence between lithology and structure. For ease of description we shall refer to the thrust sheets of the Central Metasedimentary Belt boundary thrust zone according to their lithological definition (Fig. 2), while recognizing that the outer parts of the thrust sheets are often structurally transitional to the enclosing ductile thrust zones.

Haliburton segment

The Haliburton segment of the Central Metasedimentary Belt boundary thrust zone is a stack of three crystalline thrust sheets of hornblende–biotite tonalitic composition, between which occur a large thrust sheet of clinopyroxene syenitic gneiss, a smaller slice of clinopyroxene syenitic straight gneiss, and two map-scale slices of anorthosite (Fig. 2) (Hanmer and Ciesielski 1984; Hanmer *et al.* 1986; Hanmer 1988, 1989; see also Easton 1983, 1986*a*, 1990; Easton and Van Kranendonk 1984; Culshaw 1986). The thrust sheets are lenticular, up to 5 km thick, and are separated from one another by granoblastic ductile thrust zone tectonites.

The Dysart, Grace, and Glamorgan thrust sheets are enclosed by clinopyroxene–graphite–serpentinite–phlogopite marble tectonic mélange. (Hanmer and Ciesielski 1984; Hanmer *et al.* 1986). Smaller blocks of the crystalline gneisses, varying in scale from mappable structural horses to outcrop size and smaller, were tectonically eroded from the main thrust sheets and incorporated into the flowing marble, particularly in the case of the Grace thrust sheet. All of the thrust sheets show some evidence for marginal strain gradients as one approaches the more intensely deformed, and lithologically distinctive, thrust zone tectonites. This is particularly well displayed in the case of the structurally lowest Redstone thrust sheet. The lower part of the tonalitic thrust sheet is strongly deformed compared with its interior (Hanmer 1989), and the proportion of thin intrusive granitic veins within the thrust sheet increases. The grain size is finer, the distribution of biotite is more homogeneous, and where it contains granitic or mafic material, the straightness of the banding is more pronounced. The Redstone thrust sheet is bounded along its upper and lower contacts by well-developed zones of porphyroclastic gneiss, with abundant coarse monocrystalline feldspar porphyroclasts. Polycrystalline quartz–feldspar and amphibolite inclusions, derived by the mechanical degradation of very coarse granite, pegmatite, and amphibolite, are set in a homogeneous, grey, fine-grained, quartzofeldspathic, hornblende–biotite-bearing matrix. To the south and west the porphyroclastic gneisses pass progressively along strike into equigranular straight gneiss. The structural transition represents a change in the scale of accommodation of the imposed bulk deformation (Hanmer 1987, 1989).

In addition to the straight and porphyroclastic gneisses and the marble tectonic mélange, the Redstone and Dysart thrust sheets are bounded by a distinctive lithological assemblage, which within the Ontario Grenville orogen, is confined to the Central Metasedimentary Belt boundary thrust zone. An assemblage of iron-rich cordierite–gedrite (Lal and Moorhouse 1969) and aluminous sillimanite–garnet gneisses occurs in two important zones along the upper side of the Redstone thrust sheet and along the lower boundary of the Dysart thrust sheet (Fig. 2). Smaller zones of the same lithologies also occur locally along the upper boundary of the Dysart thrust sheet. The cordierite–gedrite gneiss contains a complex metamorphic mineral assemblage including garnet, sillimanite, and kyanite. Corundum, rare staurolite, and sapphirine occur only within large grains of cordierite. The sillimanite–garnet gneiss is associated with abundant deformed and crosscutting leucocratic quartz–feldspar–garnet veins. These upper amphibolite facies mineral assemblages indicate minimum syntectonic metamorphic pressures of ca. 6 kbar (1 bar = 100 kPa) and temperatures in the range 600–650°C (Hanmer 1988). The presence of relict orthopyroxene within the Redstone thrust sheet, partly replaced by hornblende, is indicative of earlier granulite facies metamorphic conditions.

The thrust sheet boundaries and all planar structures, both within and between the thrust sheets, are coplanar and dip at about 20° to the southeast. A dip-parallel lineation (Fig. 2; Hanmer and Ciesielski 1984; Hanmer 1988), manifested as either a rodding, the elongation of streaked-out polycrystalline aggregates, or a mineral alignment, and coaxial with the bisector of rare, well-developed sheath folds (Cobbold and Quinquis 1980), is demonstrably parallel to the direction of maximum finite extension. An assemblage of mechanically independent shear-sense indicators, such as winged inclusions (Hanmer 1990), asymmetrical back-rotated pull-aparts and foliation fish (Hanmer 1986), oblique foliations in shear-parallel pegmatites (Hanmer and Passchier 1991), asymmetrical intrafolial folds with sheared-out lower limbs, and rare asymmetrical extensional shears (Berthé *et al.* 1979; White *et al.* 1980), demonstrate that the preserved deformation fabrics result principally from non-coaxial flow involving overthrusting toward the northwest, along the direction of the extension lineation. However, the deformation path was not a simple one. The $S \gg L$ symmetry of the straight gneiss and porphyroclastic gneiss tectonic banding, coupled with the presence of asymmetrical extensional shear bands (Hanmer and Passchier 1991; cf. Dennis and Secor 1987, 1990; Williams and Price 1990), "chocolate-tablet" boudinage, and the "in-plane" geometry of winged inclusions (Hanmer 1990; Hanmer and Passchier 1991), are kinematically significant. They indicate that northwest-directed thrusting was associated with a significant component of shortening across the flow plane and extension along the axis of rotation of the non-coaxial component of the flow.

The ductile thrust zone tectonites are folded by a set of open to isoclinal folds of variable size (0.1 – 10 m in wavelength), but constant S asymmetry, when looking downplunge. These folds are common and are everywhere coaxial with the southeast-plunging extension lineation, regardless of their tightness. Therefore, given their open profiles, they must have initiated in their present orientation and did not rotate significantly with respect to the extension lineation during progressive deformation. By contrast, the axes of the sheath folds initiated at a high angle to the extension lineation – shear direction (e.g., Williams and Zwart 1978; Cobbold and Quinquis 1980). Given the regional extent of their occurrence, Hanmer (1988) suggested that the asymmetry of this fold set is kinematically significant and indicates the existence of a subordinate component of topside-northeast shearing, perpendicular to the principal northwest-directed thrust vector, which generated the sheath folds. This is supported by the rare occurrence of rotated winged inclusions, whose geometry also indicates topside-northeast shearing associated with an axis of rotation, which is parallel to the regional extension lineation.

Killaloe segment

The Killaloe segment of the Central Metasedimentary Belt boundary thrust zone is composed of three map-scale allochthonous bodies: the Stafford (biotite – hornblende amphibolitic), Foymount (clinopyroxene syenitic), and Papineau (hornblende – biotite quartzofeldspathic) crystalline thrust sheets (Fig. 2) (Lumbers 1982; McEachern 1990). They are lens-like in map pattern, up to about 5 km thick, and are separated from one another by ductile thrust zones. The thrust sheets are simply stacked in two tiers, with the Foymount sheet overlying the other two. The thrust zone tectonites are an assemblage of straight and porphyroclastic gneisses and marble tectonic mélange, identical with those of the Haliburton segment (McEachern 1990).

The thrust sheets in this segment of the boundary thrust zone structurally resemble those of the Haliburton segment. As in the case of the Redstone, grey quartzofeldspathic gneisses at the base of the Papineau thrust sheet are intruded by abundant, subconcordant granitic pegmatite sheets. These are strongly deformed and mechanically refined to form a porphyroclastic gneiss containing rotated winged inclusions indicative of northwest-directed thrusting. Tectonically induced flow within the marble has pried map-scale to fist-size pieces from the main Foymount thrust sheet and incorporated them into the tectonic mélange, as in the case of the Grace thrust sheet. High-strain zones are developed within the outer parts of the lithologically defined crystalline thrust sheets. This is particularly well illustrated in the case of the Stafford thrust sheet (Fig. 2).

The interior of the Stafford thrust sheet is a homogeneous, coarse-grained, poorly to variably foliated amphibolite. Near its base, intrusive sheets of very coarse grained granite, up to several metres thick, are spatially associated with the wholesale replacement of hornblende by biotite and the development of an anastomosing phacoidal foliation. The strongest foliation is developed in shear zones composed of schistose biotite amphibolite and transposed, granoblastic, fine-grained granitic sheets, which anastomose around large lenses of coarse amphibolite, tens of metres long by several metres thick. The spacing of the anastomosing deformation zones and the size of the intervening amphibolite lenses decrease toward the base of the thrust sheet. McEachern (1990) has interpreted these observations in terms of an increase in shear strain towards the base of the thrust sheet, heterogeneous strain softening and strain rate distribution. This led to localized fracturing and granite vein emplacement, associated K-metasomatism, biotitization, and reaction softening. According to this interpretation, a feedback cycle was set up between reaction softening and further shear strain localization in the lower part of the thrust sheet. The end product of these linked processes is the formation of a hornblende – biotite quartzofeldspathic straight gneiss.

Along the base of the boundary thrust zone, directly below the Papineau and Stafford thrust sheets, are four isolated occurrences of the cordierite – gedrite and sillimanite – garnet gneisses (Fig. 2; Miller 1983; McEachern 1990). Poor exposure prevents us from tracing this map unit continuously along strike between the known outcrops, and we note that orthoamphibole has not been identified east of the Papineau thrust sheet. Nevertheless, the resemblance of these lithologies to their equivalents in the Haliburton segment, and their spatial confinement to the boundary thrust zone, justify extending the map unit as far as Pembroke (Fig. 1). As in the Haliburton segment, the metamorphic mineral assemblage indicates that thrusting occurred at upper amphibolite facies metamorphic conditions.

Structurally, the Killaloe segment closely resembles the Haliburton segment (McEachern 1990). First, the thrust sheet boundaries and all planar structures are all coplanar and dip at about 20° to the southeast. A well-developed extension lineation is oriented down the dip (Fig. 2). Second, although fewer shear-sense indicators have been observed in the less well exposed Killaloe segment, they consistently indicate overthrusting toward the northwest, along the direction of the extension lineation. Third, the S $\gg$ L symmetry of the gneissic banding in this segment indicates that thrusting was also associated with a significant component of strike-parallel extension. Fourth, local sheath folds, whose bisectors are parallel to the extension lineation, are genetically related to northwest-directed overthrusting along the direction of the lineation. As at Haliburton, a set of S folds, coaxial with the regional extension lineation, regardless of the fold tightness, is developed on a variety of scales throughout the ductile thrust zone tectonites of the Killaloe segment.

Bancroft segment

The Bancroft segment of the Central Metasedimentary Belt boundary thrust zone separates the Haliburton and Killaloe segments (Fig. 2). It is less accessible than the other two segments, and previous geological work has focussed upon the lithological and economic aspects of the area (e.g., Adams and Barlow 1910; Hewitt 1954; Hewitt and James 1955; Breaks and Thivierge 1985). Syenitic gneiss in the upper part of this segment, bounded above and below by marble tectonic mélange, can be traced into the western limit of the Foymount thrust sheet in the Killaloe segment (Anonymous 1957; Hanmer and Ciesielski 1984; McEachern 1990). Sheets of marble, charged with syenitic debris, were forcibly injected into the syenite. Foliations in the Bancroft segment, although variable in strike, dip gently ($\sim 20°$) toward the south and east, whereas the extension lineation plunges systematically at a shallow angle to the southeast (Fig. 2). There are indications that straight and porphyroclastic gneisses are developed in the structurally lower part of this segment (Breaks and Thivierge 1985), and we have observed rotated winged inclusions and attenuated intrafolial asymmetrical folds, indicative of northwest-directed thrusting along the direction of the extension lineation. These observations strongly suggest that the Bancroft segment of the boundary thrust zone shares the same structural signature as that which we have documented in the adjacent segments.

Hanging wall

The Haliburton, Bancroft, and Killaloe segments of the Central Metasedimentary Belt boundary thrust zone are all overlain by a common rock assemblage. The hanging wall immediately above the boundary thrust zone consists of metamorphosed gabbroic bodies, underlain by a mylonitic extensional shear zone and nepheline syenites, all spatially associated with marble tectonic mélange (Fig. 2). Metagabbro and spatially associated nepheline syenite also occur in the vicinity of the boundary of the Central Metasedimentary Belt on the Quebec side of the Ottawa River (Kretz *et al.* 1989). Although these rocks have long been known (e.g., Adams and Barlow 1910; Hewitt and James 1955), earlier petrologically oriented workers sought significance in the spatial association of the predominantly metasomatic nepheline rocks with carbonates and granitoids (see Gittins 1967 for review). However, from a structural perspective, one might also consider the potential significance of the *spatial restriction* of the nepheline rocks to a structural level just below that occupied by the metagabbros, the most competent rocks in the study area (Fig. 2).

Metagabbros

The Trooper Lake (Armstrong and Gittins 1970*a*, 1970*b*; Grieve and Gittins 1975), Faraday, Mallard Lake (Hewitt and James 1955), Raglan Hills (Lumbers 1982), and Chenaux (Abdurhaman 1989) metagabbros are lens-like bodies whose lower contacts dip shallowly to the southeast (see below). They constitute the major elements of a discontinuous string of metagabbroic bodies that extends from south of the Glamorgan thrust sheet to the Ottawa River (Fig. 2). For the most part, they are compositionally homogeneous and coarse grained (up to 5 cm). However, they are locally lithologically layered and variably deformed. Plagioclase is variably recrystallized and may locally preserve either the purplish hue or the monocrystalline lath-like habit of primary igneous feldspar. In the Trooper Lake body, a well-developed shape fabric, formed by aligned monocrystalline plagioclase laths, is coarsely crenulated by a set of 2 − 10 cm wide shears (Fig. 3). Within the shears, the plagioclase is generally finer (2 − 3 cm) than in the adjacent host. Some are blocky in habit, but most of the feldspar crystals are monocrystalline laths of igneous habit. Both the shape fabric and the shears are locally crosscut by diabase dykes. The point to be emphasized here is that the magmatic crystallization of the gabbros was accompanied by deformation.

In contrast to the synmagmatic deformation structures, a generally penetrative tectonic foliation of moderate intensity is expressed by flattened polycrystalline aggregates of plagioclase or hornblende in the lithologically homogeneous parts of the metagabbros. In places it is mylonitic and anastomoses about metre- to decametre-scale lenses of less foliated to isotropic metagabbro. The anastomosing zones of most intense foliation are associated with mylonitized granitic sheets, much as already described for the lower part of the amphibolitic Stafford thrust sheet within the Central Metasedimentary Belt boundary thrust zone. Both the mylonites in the metagabbro and the underlying marble tectonic mélange contain rotated inclusions, northwest-verging intrafolial folds with attenuated lower limbs and asymmetrical extensional shears, which indicate northwest-directed overthrusting (Hanmer and Ciesielski 1984, Fig. 15.5C).

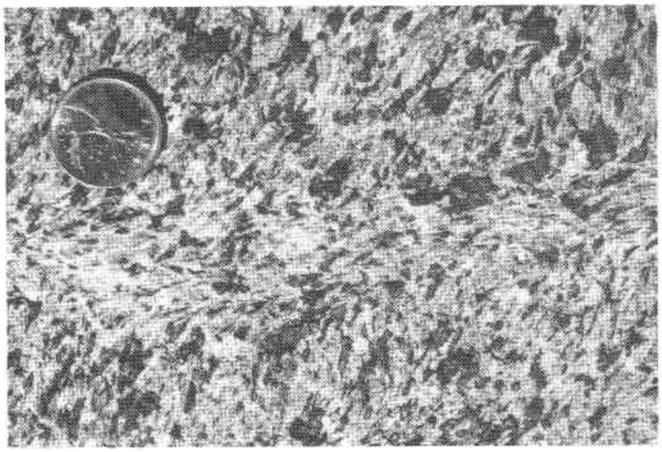

FIG. 3. A shear zone cutting the plagioclase shape fabric made of euhedral monocrystalline laths in metagabbro. Dark mineral is clinopyroxene. Note the grain size reduction within the shear zone, but also the persistence of the lath-like habit of the feldspars. Trooper Lake metagabbro. Coin is 23 mm in diameter.

Extensional mylonites

Bands of fine-grained (< 50 μm) dark grey marble mylonite, from several millimetres to several tens of metres thick, formed at the expense of the coarse-grained marble tectonic mélange. They are best developed immediately beneath the Trooper Lake, Faraday, and Mallard Lake metagabbroic bodies, but are also found beneath the Raglan Hills and Chenaux metagabbros (Fig. 2). Although their existence has long been known (Chayes 1942; Hewitt and James 1955; Hanmer and Ciesielski 1984), their kinematic significance has only recently been determined (Bancroft shear zone: van der Pluijm and Carlson 1989; Carlson *et al.* 1990). The foliation and extension lineation of the Bancroft shear zone mylonites are parallel to those of the older thrust associated tectonites of the Central Metasedimentary Belt boundary thrust zone. Shear criteria (Carlson *et al.* 1990, Fig. 8; Hanmer 1990, Fig. 4; Hanmer and Passchier 1991, Fig. 47) demonstrate that these mylonites are part of an extensional shear zone of significant strike length (Fig. 2; Carlson *et al.* 1990). According to calcite − graphite carbon isotope thermometry, and comparison with the cooling history proposed by Cosca *et al.* (1988), the extensional shearing occurred at less than 450°C, during the interval ca. 1.0 − 0.93 Ga (van der Pluijm and Carlson 1989).

Nepheline syenites

Although patchily present elsewhere in the southwestern Grenville orogen, nepheline syenites and nepheline gabbros have long been known to form a long strike-parallel band, spatially associated with the northwest margin of the Central Metasedimentary Belt (Fig. 2; e.g., Adams and Barlow 1910; Chayes 1942; Satterly 1943; Hewitt 1954, 1960; Hewitt and James 1955). Although some of the nepheline-bearing rocks are thought to be primary (e.g., Tilley and Gittins 1961; Gittins 1967), most of the nepheline syenite is metasomatic in origin, affecting protoliths of diverse composition and primary origin (e.g., Gummer and Burr 1946; Baragar 1953; Tilley 1958; Appleyard 1967, 1969, 1974). The detailed descriptions of crosscutting and replacement textures by these authors indicate that nephelinization was broadly syntectonic, as opposed to pretectonic (e.g., Appleyard 1974). Furthermore, late, isotropic patch pegmatites of nepheline cut across the structure of the host gneisses within which they are found (e.g., Hewitt and James 1955; Hewitt 1960; Appleyard 1974). These cited studies agree that the origin of the nephelinizing fluids remains unresolved, but Appleyard (1974, p. 372) ventured to suggest that the syntectonic metasomatism occurred "at a rather precise stratigraphic [sic] level within the lower parts of the Grenville supracrustal sequence." The structural significance of the timing and, more particularly, the distribution of metasomatism will be discussed below.

Discussion

In terms of geometry, lithological and rheological zonation, kinematic significance, and timing of deformation, both the Central Metasedimentary Belt boundary thrust zone and its adjacent hanging wall exhibit a remarkable consistency over a strike length of some 200 km. The consistency of the geometry is evident from the outcrop patterns of the individual thrust sheets and from the orientations of the planar and linear fabric elements (Fig. 2). It is also evident in the ubiquitous $S \gg L$ symmetry of the deformation fabrics, particularly those of the ductile thrust zone tectonites, but also those developed within the crystalline thrust sheets. The consistency of the lithological zonation requires more detailed examination.

Lithological zonation

Within the boundary thrust zone, the consistency of the lithological zonation is evident in both the thrust sheets and the bounding ductile thrust zone tectonites. The syenitic Foymount thrust sheet is the structurally highest allochthonous unit in the Killaloe segment. In the eastern part of the Haliburton segment, this structural position is occupied by the syenitic Grace thrust sheet. Other workers have already documented the continuity of the Foymount thrust sheet with similar syenitic gneisses in the Bancroft area (Hewitt 1954; Anonymous 1957). Accordingly, there is a consistent tendency for the structurally higher thrust sheets in the boundary thrust zone to be syenitic in composition, whereas the lower thrust sheets tend to be tonalitic, amphibolitic, or quartzofeldspathic (Fig. 2).

The ductile thrust zone tectonites bounding the Redstone thrust sheet, and underlying the Papineau and Stafford thrust sheets, are banded (granitic and amphibolitic) straight and porphyroclastic gneisses, whereas the structurally higher ductile thrust zone tectonites are principally marble tectonic mélange (Lumbers 1982; Hanmer 1988). Moreover, the cordierite–gedrite and sillimanite–garnet gneisses are largely confined to the lower part of the boundary thrust zone (Fig. 2). Within the hanging wall, the nepheline-bearing rocks are consistently found at the structural level immediately below the string of metagabbro bodies. Certain aspects of this laterally extensive lithological zonation are rheologically significant.

Rheological significance

Within the boundary thrust zone, rocks rich in aligned platy biotite, as well as acicular gedrite and sillimanite, occur along the base of the Killaloe segment (Fig. 2). They are also particularly well developed along the upper and lower contacts of the Redstone and Dysart thrust sheets in the Haliburton segment. Hanmer (1988) has suggested that these rocks represent a rheologically weak, schistose horizon, whose role was to localize the incipient thrust zones bounding the crystalline thrust sheets at the time of their initiation. We would extend that interpretation to explain the localization of the base of the entire boundary thrust zone. Indeed, without the presence of such a weak horizon within the gneisses, ductile shearing would perhaps have been confined to the overlying, soft carbonate tectonites.

In the hanging wall, inspection at both the map and the outcrop scale shows that the metagabbros have behaved as competent bodies set within a very soft, flowing carbonate matrix (Fig. 2). It is significant that the base of the hanging wall is spatially associated with such stiff material. However, the metagabbros, now dispersed throughout the enclosing marble, would have best exerted a rheological influence, at the scale of the boundary thrust zone, if they had originally constituted a materially continuous sheet of appropriately large strike-parallel dimensions (100–200 km). A number of observations suggest that this could have been the case.

The currently available evidence for a continuous sheet of gabbro is permissive, at best. Preliminary geochronological study (U–Pb on zircon; S. McEachern, unpublished data) indicates that the gabbros crystallized magmatically during the range ca. 1.23–1.18 Ga. Similarities in their internal synmagmatic structure and composite nature might suggest that the gabbro bodies share a common origin, but this remains to be tested by detailed petrological and geochemical studies. From a geological perspective, the localization of syntectonic nephelinization at a strictly confined structural level could reflect either (*i*) the configuration of the source of the nephelinizing agent or (*ii*) the existence of a continuous "lid," which channeled the metasomatic fluids along its base by acting as a barrier through which they could not penetrate. On the one hand, a laterally extensive evaporite horizon could have provided a geometrically appropriate source, as well as presenting a rheologically soft horizon at which to localize the base of the hanging wall during initial thrusting. However, this scenario provides no barrier to the upward flow of the nephelinizing fluids into the structurally higher levels of the Central Metasedimentary Belt. On the other hand, a materially continuous sheet of gabbro could have provided a lid. Moreover, by acting as a discrete competent horizon, it could also have localized the upper limit of the boundary thrust zone. In response to the regionally extensive general non-coaxial flow regime (northwest-directed thrusting plus an important component of strike-parallel extension), direct evidence for which is provided by the $S \gg L$ fabrics developed throughout the boundary thrust zone, such a rheologically competent sheet would be expected to extend heterogeneously during progressive deformation. The result would be that illustrated in Fig. 2, which we suggest may be an example of crustal-scale chocolate tablet boudinage (e.g, Ramsay 1967).

Kinematic consistency

The constant kinematic character of the boundary thrust zone is exemplified by the shallow southeastward plunge of the extension lineation, associated with the assemblage of consistent shear-sense criteria, which indicate predominantly northwest-directed thrusting. A second indicator of kinematic consistency is the ubiquitous set of S folds, of variable tightness, but constant orientation, parallel to the regional extension lineation. Even fairly open folds in this set are perfectly coaxial with the thrust-related extension lineation, which itself shows little or no variation in orientation. This could be taken to suggest that the finite extension lineation lay fairly close to the direction of maximum instantaneous stretching, even during the formation of the S fold set. Theoretically, it is a simple matter to generate folds of a given asymmetry, coaxial to the bulk shear direction, by inducing a short-lived, subsidiary shear couple, at a high angle to the principal one, because of a systematic, regional obliquity between the deforming surfaces and the bulk flow plane (Brun and Merle 1988; Hanmer and Passchier 1991). Were this to occur in a transpressive regime, the fold axes could form at a small initial angle to the direction of maximum instantaneous stretching. However, the scale of the boundary thrust zone, as well as the high-strain nature of the planar fabrics affected by the folding, renders this scenario highly improbable.

Alternatively, a potential recent analogue may be found in the Himalaya. Longitudinal extensional ductile shearing (topside-east) in the upper structural levels of the Tibetan Slab and its Tethyan cover was apparently contemporaneous with, and materially coupled to, Miocene transverse ductile thrusting at lower structural levels in the Slab (Main Central Thrust zone; Pecher *et al.* 1991). We speculate that the Central Metasedimentary Belt boundary thrust zone may be structurally analogous to the lower levels of the Tibetan Slab. We suggest that the set of asymmetrical coaxial S folds may reflect a regionally uniform sense of extensional shearing at structurally higher levels of the now eroded Grenvillian mountain edifice. Conversely, we predict that such asymmetrical, lineation-parallel folds should exist within the Himalayan Main Central Thrust zone. However, we note that the resultant principal direction of finite extension produced by the simultaneous operation, in the same volume of rock, of two shear couples whose flow directions are not parallel to each other, would lie between the two flow directions. From regional considerations, this does not appear to be the case in the Grenville example (e.g., Davidson 1984; Hanmer 1988; Nadeau and Hanmer 1992). Accordingly, we suggest that the northeast-directed longitudinal shear couple was very much weaker than the northwest-directed transverse shear couple, such that any resulting obliquity is too small to have been detected. Furthermore, although broadly coeval, the two shear couples may not have been strictly contemporaneous. They could have alternated in a crude ''stop–go'' fashion, or the longitudinal component could have been much shorter lived than the transverse component. Deviation of early transverse linear structures from the regional northwest–southeast trend due to longitudinal shearing would have been subsequently corrected by continued northwest-directed thrusting.

Kinematic consistency is also evident within the hanging wall. The metagabbro bodies are underlain by coarse marble tectonic mélange, whose deformation fabric, at least locally, can be shown to have been generated by northwest-directed overthrusting of the metagabbro. These coarse, high-temperature marbles were subsequently reworked to form the extensional, relatively cool, carbonate mylonites of the Bancroft shear zone along the base of the structural horizon occupied by the metagabbro bodies. The following observations suggest to us that the relatively stiff metagabbro bodies, from Trooper Lake to Chenaux (Fig. 2), directly influenced the localization of the Bancroft shear zone, albeit of relatively minor magnitude. Whereas minor occurrences of these mylonites are reported along the trace of the Bancroft shear zone (Carlson *et al.* 1990), the best-developed examples are spatially associated with the metagabbro bodies. Using published geothermobarometric data (Anovitz and Essene 1990), van der Pluijm and Carlson (1989) concluded that the Bancroft shear zone had accommodated approximately 10 km of dip-slip extensional displacement. However, without geochronological constraints on the geothermobarometric data this estimate is somewhat speculative. We suggest that the spatial association of the best-developed low-temperature mylonites with the metagabbros indicates that the mylonites were initiated during relatively minor adjustments as the dense mafic rocks slipped downslope in response to gravity. We also note that the stiff metagabbros appear to have strongly influenced the localization of the Bancroft shear zone, even though they did not constitute a materially continuous horizon at that time.

Kinematic contemporaneity

Our geological observations lead us to suggest that the Central Metasedimentary Belt boundary thrust zone operated as a structurally coherent entity throughout its 200+ km exposed strike length. Accordingly, its immediately adjacent hanging wall, i.e., the northwest margin of the Central Metasedimentary Belt, was displaced as a discrete crustal-scale thrust sheet. Using our structural studies to direct our sampling, we have determined the ages of magmatic crystallization (U–Pb in zircon) of a suite of syntectonically intruded granitic veins and pegmatites in the Haliburton and Killaloe segments of the Central Metasedimentary Belt boundary thrust zone (van Breemen and Hanmer 1986; McEachern 1990).

The presently preserved, undisturbed ductile thrust zone tectonites only represent the final distribution of loci of relatively high strain rate (Hanmer 1988). To specifically determine the time of thrusting in different parts of the boundary thrust zone, as defined by the present distribution of high-strain zones, samples were selected from outcrops of straight and porphyroclastic gneiss. In the field, intrusive veins were determined to be syntectonic where they clearly cut across the straight or porphyroclastic gneisses, but were themselves subjected to subsequent high-temperature penetrative deformation, as indicated by the formation of internal feldspar-plastic fabrics. The sampled veins are representative members of vein sets, which even within a given outcrop, show all stages of structural refinement and (or) transposition with progressive deformation. Accordingly, we can confidently take their ages of magmatic crystallization to indicate the date of at least the later stages of the ductile thrusting.

The results are summarized in Fig. 2, and the detailed data are reported elsewhere (van Breemen and Hanmer 1986; McEachern 1990). According to data obtained from strongly deformed veins, the presently preserved configuration of the boundary thrust zone is essentially the product of major ductile thrusting that occurred throughout the zone during the interval ca. 1.08–1.05 Ga. Some later thrust-related movements occurred at ca. 1.03–1.0 Ga, as evidenced by the data derived from more moderately deformed veins. The oldest age in the set, ca. 1.19–1.18 Ga, obtained from two strongly deformed samples in the Killaloe segment, indicates that the Stafford thrust sheet existed as an allochthonous entity prior to the main 1.08–1.05 Ga thrusting event (McEachern 1990).

In the Haliburton segment, the occurrence of syenitic material, tectonically intercalated between the tonalitic crystalline thrust sheets, led Hanmer (1988) to tentatively infer the existence of an earlier pile of crystalline thrust sheets, which was subsequently restacked at ca. 1.06 Ga. We suggest that the ca. 1.19–1.18 Ga thrusting at the base of the Stafford thrust sheet in the Killaloe segment is the first firm evidence of this earlier event. Moreover, U–Pb dating of sphene in, and adjacent to, the Central Metasedimentary Belt boundary thrust zone has yielded cooling ages younger than ca. 1.08–1.05 Ga, except for two ages of 1.125 and 1.152 Ga from tectonites immediately above the

Foymount thrust sheet in the Killaloe segment (Mezger *et al.* 1990; K. Mezger, personal communication, 1991). These data independently confirm that high-temperature ductile thrusting in the Killaloe segment had initiated prior to the ca. 1.08–1.05 Ga event. We also note that the restriction of syenitic lithologies to the structurally higher crystalline thrust sheets is somewhat blurred by tectonic intercalation in the southwestern part of the Haliburton segment (Fig. 2). Taken together, these observations suggest that the ca. 1.08–1.05 Ga restacking event was more intensely developed in the southwestern part of the boundary thrust zone. They also suggest that the initial ca. 1.19–1.18 Ga thrust pile comprised an upper syenitic level and a lower tonalitic to amphibolitic level (cf. Hanmer 1988).

Tectonic significance

A number of models have been proposed during the past two decades to account for the origin of the Central Metasedimentary Belt as a whole. Although some workers have proposed rift-associated tectonic models (e.g., Baer 1976; Lumbers *et al.* 1991), most interpretations represent the Central Metasedimentary Belt in terms of a magmatic arc. The principal differences between the arc models concern the continental or oceanic context of the magmatic edifice (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). Smith and Holm (1990) present a comprehensive synthesis of the available sedimentological and petrological data. They propose that the Elsevier terrane is a back-arc basin, initiated at ca. 1.29 Ga (Silver and Lumbers 1966) on the northwest side of a continental magmatic arc (southeastern Elzevir terrane), above a northwest-dipping subduction zone (Smith and Holm 1990). Accordingly, the string of hanging-wall metagabbros shown in Fig. 2 presently lies just north of the supposed back-arc basin rocks. We tentatively suggest that the synmagmatically deformed gabbros may have originally formed part of the floor of the back-arc basin. The ca. 1.19–1.18 Ga age of early thrusting in the boundary thrust zone would imply that this basin had closed by at least ca. 1.18 Ga.

Recognition that the rocks of the Grenville orogen have experienced deformation and metamorphism at various times during the interval 1.25–1.0 Ga has led previous workers to postulate a Grenville orogenic cycle, punctuated by several orogenies (Moore and Thompson 1980; Moore 1986; Rivers *et al.* 1989), or orogenic events. Recent compilations of geochronological data for the southwestern Grenville orogen (Easton 1986*b*; Nadeau 1990) have highlighted orogenic events at ca. 1.25, 1.16, 1.1, and 1.06 Ga. With the exception of the last of them, tectonic manifestations of these orogenic events have been identified on both sides of the Central Metasedimentary Belt boundary thrust zone, in the Grenville Front Tectonic Zone, in the Central Gneiss Belt, and within the Central Metasedimentary Belt itself (Fig. 1) (van Breemen *et al.* 1986; Connelly *et al.* 1987; Davis and Bartlett 1988; Green *et al.* 1988; van Breemen and Davidson 1988; Nadeau 1990; Marcantonio *et al.* 1990; Corriveau 1990; Corriveau *et al.* 1990; Lumbers *et al.* 1991). The ca. 1.19–1.18 Ga deformation and metamorphism in the Central Metasedimentary Belt boundary thrust zone fall between the ca. 1.25 and 1.16 Ga events. By contrast, the ca. 1.08–1.05 Ga deformation is only well expressed within the Central Metasedimentary Belt boundary thrust zone, and is significantly younger than tectonic events in the Central Gneiss Belt and Central Metasedimentary Belt.

Recent authors have sought to place the northwestern boundary of the Central Metasedimentary Belt into a regional tectonic context. Our results, combined with those of other workers (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; Nadeau 1990; Harnois and Moore 1991; Nadeau and Hanmer 1992), lead us to suggest that the ca. 1.19–1.18 Ga thrusting in the boundary thrust zone may reflect closure of a back-arc basin and docking of a magmatic arc with North America by ca. 1.19–1.18 Ga. Postdocking high-temperature deformation occurred at deeper structural levels within the Central Gneiss Belt at ca. 1.1 Ga (Nadeau and Hanmer 1992), but not within the overlying Elzevir terrane. By ca. 1.08–1.05 Ga, renewed, regional deformation again occurred at the northwest margin of the Central Metasedimentary Belt. One possible explanation is that continental collision occurred at about that time, at an as yet unspecified suture to the southeast of the Central Metasedimentary Belt. Intraplate thrusting was focussed upon an extensive preexisting, rheologically weak zone of well-developed crustal-scale anisotropy, i.e., thrust sheets and straight-banded tectonites of the Central Metasedimentary Belt boundary thrust zone.

Modern orogens

Until recently, it has been axiomatic that our understanding of modern orogenic belt may help elucidate the evolution of their fossil analogues. However, the processes by which convergence is accommodated in the deep-seated, as yet inaccessible, parts of the continental crust in modern collisional orogens, are the subject of continuing controversy. In the specific case of the Himalaya, the debate has focussed on the relative roles of thrusting (Seeber 1983; Coward *et al.* 1986) and lateral escape (Tapponier *et al.* 1986; Peltzer and Tapponier 1988; England and Molnar 1990; Jolivet *et al.* 1990) of discrete crustal slices and blocks, as opposed to the penetrative deformation of the continental crust (e.g., Dewey and Burke 1973; Molnar *et al.* 1987; Dewey *et al.* 1988, 1989). Both sides of the debate have their geological, as well as geophysical, proponents. However, given the young age of the mountain belt and the relatively short time span so far available for erosion, the tectonic models of crustal deformation proposed by the different schools of thought are nevertheless based largely upon indirect observation and inference. Our geological work in the deeply excavated Central Metasedimentary Belt boundary zone suggests that the debate over the role of thrusting versus that of penetrative deformation may be misplaced. Tectonic thickening in the deep-seated continental crust involves discrete ductile thrust zones on a range of scales from kilometres to hundreds of kilometres, which can accommodate both continuous (penetrative) and discontinuous deformation. Furthermore, the initial size of individual thrust sheets is in part determined by the rheological behaviour of their constituent materials. These same rheological properties, combined with regional extension of the shear plane, could lead to the disintegration of large thrust sheets into smaller ones, and a transition from discontinuous to penetrative deformation with time (see also Coward 1982).

Conclusions

(1) The Central Metasedimentary Belt boundary thrust zone is a 10 km thick by 200+ km long stack of crystalline thrust sheets, enclosed by an anastomosing network of ductile thrust zones, formed at mid- to deep-crustal depths in the Grenville orogen.

(2) It has behaved as a coherent thrust zone, accommodating northwest-directed transport of the Central Metasedimentary Belt as a single discrete thrust sheet by coherent displacements all along its strike length, initially at ca. 1.19−1.18 Ga and again at ca. 1.08−1.05 Ga.

(3) The rheological behaviour of the deforming media may have strongly influenced the location of the upper and lower limits of the boundary thrust zone at the time of its initiation.

(4) A structural horizon of metagabbro bodies may have acted as a physical barrier to the upward migration of metasomatic fluids that generated the nepheline rocks of the Central Metasedimentary Belt boundary.

(5) The gabbros were apparently thrust out of a marginal or back-arc basin within the Central Metasedimentary Belt, which had closed by ca. 1.19−1.18 Ga.

(6) Major renewed thrusting at ca. 1.08−1.05 Ga may reflect continental collision to the southeast of the exposed Grenville orogen and represent intraplate reactivation of an older, crustal-scale zone of weakness.

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