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Late Neoarchean thick-skinned thrusting and Paleoproterozoic reworking in the MacQuoid supracrustal belt and Cross Bay plutonic complex, western Churchill Province, Nunavut, Canada

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Abstract

In the western Churchill Province, Canadian Shield, Neoarchean supracrustal and plutonic rocks, intruded by Paleoproterozoic mafic dykes and granitic rocks, comprise the MacQuoid supracrustal belt and the structurally overlying Cross Bay plutonic complex. They form part of the northwestern Hearne subdomain that occupies an intermediate position between the continental Rae domain to the north and west, and the oceanic central Hearne subdomain to the south and east. New geological mapping and supporting geoscience are compatible with the presence of 2550–2500 Ma, southeast-directed, mid-crustal, thick-skinned thrusting that juxtaposed the plutonic complex over the supracrustal belt. The structural contact between the MacQuoid supracrustal belt and the Cross Bay plutonic complex potentially represents a fundamental boundary between isotopically distinct crustal blocks.

The ∼2190 Ma MacQuoid mafic dyke swarm cuts across Neoarchean deformation fabrics, but records ∼1.9 Ga, deep-crustal, regional metamorphism that affected both the supracrustal belt and the plutonic complex. Other Paleoproterozoic deformation events that occurred at ∼1850–1810 Ma are of local extent and appear to be relatively minor manifestations of more important events elsewhere, related to the Trans-Hudson orogen.

Keywords: Neoarchean; Structural geology; Thick skinned thrusting; Tectonomagmatic reworking; Western Churchill; Nunavut; Canadian Shield

1. Introduction

The western Churchill Province is one of the largest, yet poorly known fragments of Archean crust in the world (Fig. 1). Internally divided into Rae and Hearne domains by the Snowbird tectonic zone (Hoffman, 1988; cf. Jones et al., 2002), it is flanked to the northwest and southeast by the ∼2.0–1.9 Ga Thelon and ∼1.9–1.8 Ga Trans-Hudson orogens, respectively. Neoarchean supracrustal belts (Fig. 2) are interpreted to have developed at less than 2740–2680 Ma in a broad, oceanic suprasubduction environment in the Hearne domain (Sandeman et al., 2001, 2004a, 2004b; Hanmer et al., 2004), as well as on older, extended continental crust in the Rae (Zaleski et al., 2001).

Miller and Tella (1995) originally suggested that the supracrustal rocks of the Hearne domain could be divided into an older greenstone belt (∼2690 Ma) to the southeast, flanked by younger greenstone belts (∼2660 Ma) to the north and west. Although new data (Davis et al., 2004) have disproven the rationale for this model, the concept of distinct central and northwestern

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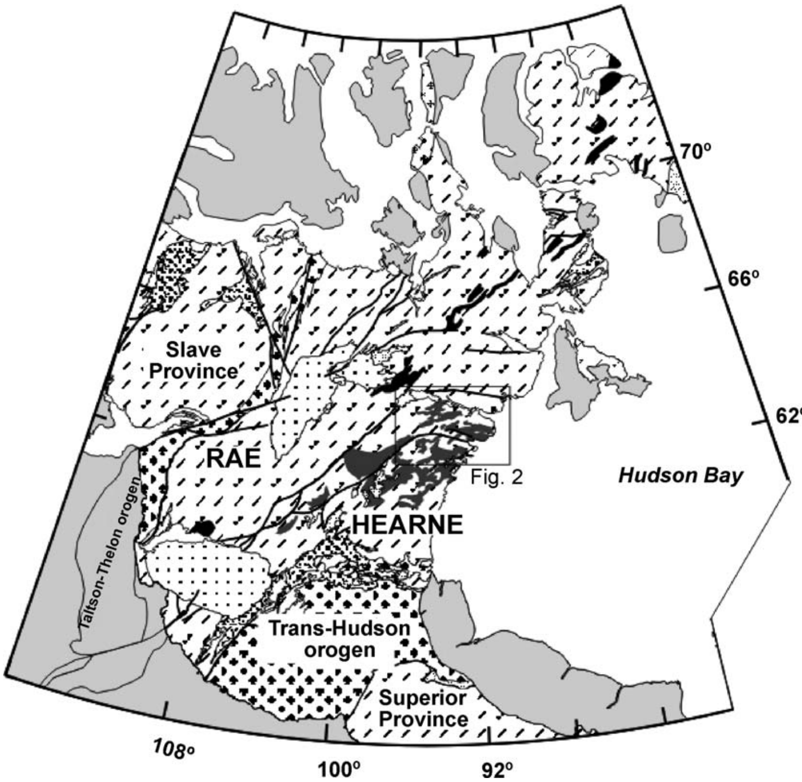


Fig. 1. Major tectonic components of the western Canadian Shield showing the western Churchill Province (WCP) divided into Rae and Hearne domains by the Snowbird tectonic zone (STZ). The location of Fig. 2 is indicated.

Hearne subdomains (Fig. 2), each with its characteristic Neoarchean to Paleoproterozoic tectonometamorphic history, is now well supported (Davis et al., this issue). The central Hearne subdomain includes the Central Hearne supracrustal belt (Hanmer et al., 2004), whereas the northwestern Hearne subdomain includes the Yathkyed (MacLachlan et al., 2000), Angikuni (Aspler et al., 2000) and MacQuoid supracrustal belts (Hanmer et al., 1999a, 1999b; Davis et al., this issue, this study).

In the Rae domain, the predominant regional tectonothermal events appear to have occurred at ∼2350 Ma and ∼1850–1800 Ma, with local deformation and metamorphism at ∼1900 Ma (e.g. Mills et al., 2000; Sanborn-Barrie et al., 2003; Carson et al., 2004; Williams and Hanmer, in press). The central Hearne subdomain experienced regional metamorphism and deformation at ∼2680 Ma, after which it remained thermotectonically inert until ∼1830 Ma (Hanmer et al., 1999a, 1999b, 2004; Davis et al., 2000, 2004; Sandeman, 2001). In contrast to both the above, the Yathkyed, Angikuni and MacQuoid supracrustal belts of the northwestern Hearne subdomain recorded regional metamorphism and deformation at ∼2660–2640 Ma and ∼2550–2500 Ma, and were further reworked by regional tectonothermal events at ∼1900 Ma and ∼1830 Ma (Berman et al., 2000, 2002a, 2002b; Stern and Berman, 2000; MacLachlan et al., 2000; MacLachlan and Relf, 2000; Aspler et al., 2002; Davis et al., this issue, this study).

This contribution focuses on the structural evolution of the MacQuoid–Cross Bay segment of the northwestern Hearne subdomain (Hanmer et al., 1999a, 1999b; Tella et al., 2001), where the MacQuoid supracrustal belt and the overlying Cross Bay plutonic complex lie along strike from the Yathkyed belt (Figs. 2 and 3; MacLachlan and Relf, 2000; MacLachlan et al., 2000). New data from the Western Churchill NATMAP Project (Hanmer and Relf, 2000) highlight the geochronological and isotopic distinctiveness of the plutonic complex. The plutonic complex contains rocks that are older, and have a distinctly more evolved Neoarchean isotopic signature, than parts of the stratigraphy in the underlying supracrustal belt (see Davis et al., 2000, this issue; Sandeman et al., 2000a, 2000b, 2001). These new data raise the possibility that the plutonic complex was tectonically juxtaposed over the supracrustal belt, potentially by thrusting related to that already identified in the Yathkyed belt (MacLachlan et al., 2000; MacLachlan and Relf, 2000).

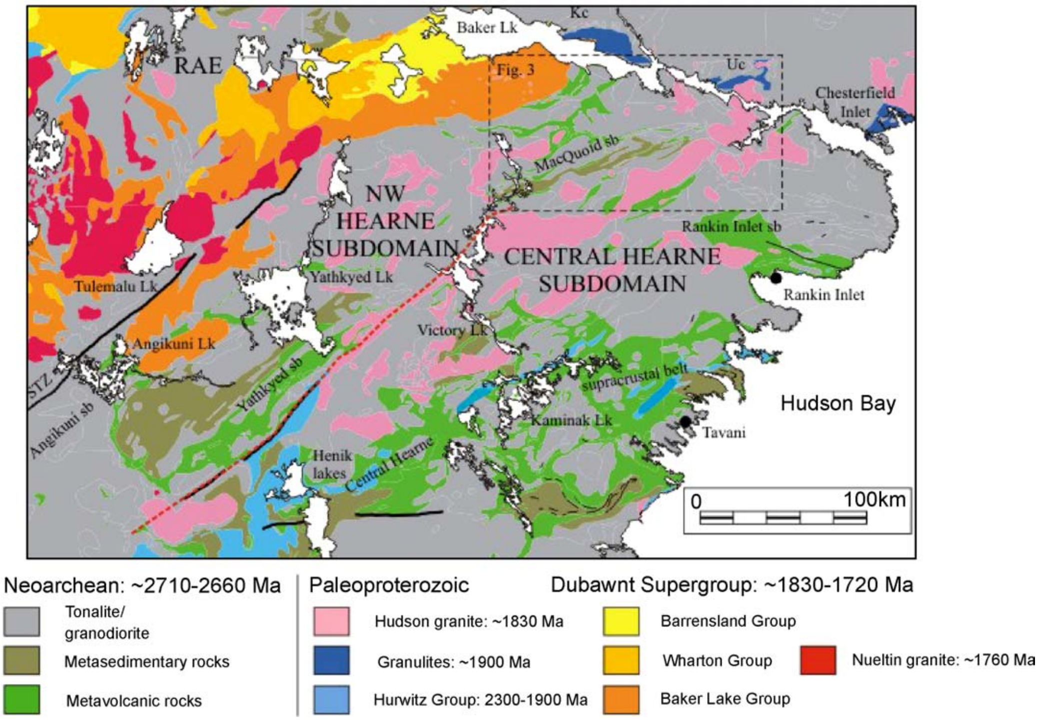


Fig. 2. Major lithological units of the central and northwestern subdomains of the Hearne domain. Box indicates the location of Fig. 3; continuous heavy black lines are faults. Kc, Kramanituar Complex; Uc, Uvauk Complex; sb, supracrustal belt; STZ, Snowbird tectonic zone. The dashed red line is the boundary between the Central and Northwestern Hearne subdomains; the location of its northeastern portion is tentative (after Paul, 2001).

The principal intent of this contribution is to examine the potential compatibility of our recent geological mapping with this hypothesis.

It is increasingly apparent that the western Churchill Province has been subjected to widespread tectonothermal reworking at ∼1850–1830 Ma (e.g. Zaleski et al., 2001; Aspler et al., 2002; Peterson et al., 2002; Sanborn-Barrie et al., 2003; Mahan et al., 2003; Carson et al., 2004), including the Cross Bay plutonic complex (Davis et al., 2000, this issue). Therefore, the second aim of this contribution is to present the geological evidence for this reworking in the MacQuoid–Cross Bay area. Dates cited in the following sections are U–Pb zircon and presented in detail in Hanmer et al. (this issue; see also Table 1), unless otherwise indicated.

2. Geology

2.1. MacQuoid supracrustal belt

The MacQuoid supracrustal belt comprises two main components (Hanmer et al., 1999a, 1999b; Davis et al., this issue; Tella et al., 2001): a predominantly metasedimentary homocline (MH in Fig. 3), overlain to the west by a succession of volcanic and minor sedimentary rocks. The metasedimentary homocline is predominantly composed of moderately (average 40–50°) north–northwest-dipping, concordant panels of semipelite and psammite, with subordinate mafic volcanic rocks, gneissic tonalite, and tonalitic plutons. Amphibolite-facies regional metamorphism and deformation (∼650–700 °C at ∼0.5 GPa) occurred at ∼2550–2500 Ma (Table 1; Berman et al., 2000; Stern and Berman, 2000).

The principal volcanic belt (<2745–2682 Ma; PVB in Fig. 3), which directly overlies the western part of the metasedimentary homocline, is also intruded by tonalitic plutons and gneisses (2685–2678 Ma; Table 1). It has a concave-east, arcuate geometry, comprising a laterally extensive, moderately north dipping, eastern branch (EB in Fig. 2) that passes westwards into a “knot” of regional-scale, upright, moderately east–northeast-plunging folds (Ryan et al., 1999, 2000a, 2000b). The eastern branch wraps around three lobate, ∼2685–2655 Ma, tonalitic plutons (Table 1), between which its lower boundary is pinched into a pair of cusp-like keels (#1 in Fig. 3a). The mafic rocks of the eastern branch are predominantly fine

Table 1
Summary of geological history

	MacQuoid supracrustal belt	Big lake shear zone	Cross Bay plutonic complex
<1805 Ma	*South Channel fault*		
	Unconformity	*Unconformity*	*Unconformity*
∼1815–1805 Ma			Post-kinematic granite veins
	ENE trending folds		*Deformation, ENE trending folds, lineation and coaxial S-plunging folds*
∼1830 Ma	Post-kinematic monzogranite plutonism	Stitching monzogranite pluton	Monzogranite plutons and sheets
∼1840 Ma	Granitic plutonism		
∼1900 Ma	*Static metamorphism,* ∼1.0 GPa	*Static metamorphism,* ∼1.0 GPa, *minor strike-slip?*	*Metamorphism,* ∼1.2–1.0 Gpa, *S-plunging folds?*
∼2190 Ma	MacQuoid dykes	MacQuoid dykes	MacQuoid dykes
∼2550–2500 Ma	*D2 deformation, transposition,* ∼0.5 GPa	*Granulite mylonite,* ∼1.2–1.3 GPa, *minor dextral strike-slip on main segment*	*Deformation*
∼2670–2655 Ma	Volcanism and tonalite plutonism		
	D1 deformation?		
∼2685–2680 Ma	Volcanism, tonalite plutonism, and sedimentation		
∼2695 Ma			South Channel granite and veins
			Deformation, foliation, lineation and coaxial S-plunging folds
2701 Ma			Diorite, tonalite and mafic intrusions
<2740 Ma	Volcanism		

grained, layered (1–5 cm), hornblende-garnet amphibolite schists, with rare, relict pillow structures (Fig. 4). The latter suggest that the mafic schists have accommodated significant strain.

Other volcanic rocks occur further north, but their relationship to the principal volcanic belt is equivocal. Along South Channel (SC in Fig. 3), southwest-trending intermediate volcanic rocks, flanked by east-trending semipelitic rocks (Bowell Island in Fig. 3) appear to be truncated by the South Channel fault (SCf in Fig. 3). Between these rocks and the principal volcanic belt, northwest-trending volcanic rocks surrounded by intrusive tonalite and granite are structurally discordant with respect to the rest of the volcanic belts (#2 in Fig. 3a). The point to retain here is that the volcanic rocks north of the principal volcanic belt may not represent a coherent lithological or stratigraphic unit.

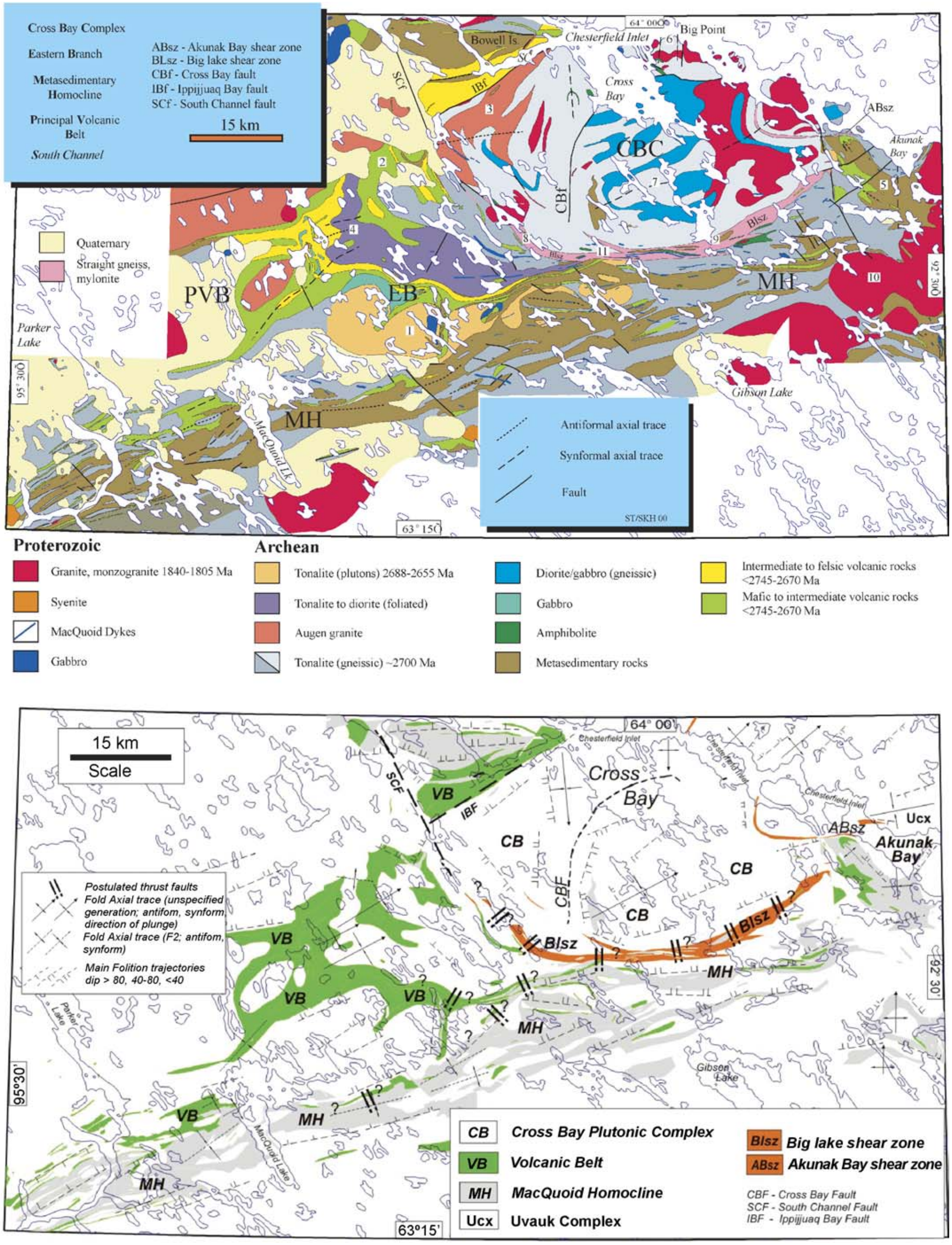
2.2. *Cross Bay plutonic complex*

An ovoid plutonic complex composed of orthogneisses and plutonic rocks straddles Cross Bay, north and east of the metasedimentary homocline and the volcanic belts, respectively (Fig. 3; Hanmer et al., 1999; Tella et al., 2001; Davis et al., this issue). Its central part is made of tonalite to gabbro, intruded to the east and west by granitic plutons. The tonalites are principally gneissic, and superficially similar to those in the MacQuoid supracrustal belt. East of the Cross Bay fault (CBf in Fig. 3), the tonalite intrudes map-scale panels of diorite (2701 ± 2 Ma), gabbro and amphibolite (Hanmer et al., this issue; Table 1).

The western part of the Cross Bay plutonic complex is intruded by the variably to strongly foliated South Channel granite (∼2692 Ma; #3 in Fig. 3a; Table 1). The eastern part of the plutonic complex, and the eastern end of the metasedimentary homocline, are intruded by variably foliated to massive, ∼1830–1815 Ma, equigranular, monzogranite plutons and map-scale sheets (Hanmer et al., 1999a; Tella et al., 2000).

2.3. *MacQuoid dykes*

The MacQuoid dykes, a swarm of 070–120° trending, subvertical, mafic sheets, 2–20 m thick (locally 500 m), were injected into the MacQuoid–Cross Bay area at ∼2190 Ma (Table 1; Tella et al., 2001). They form part of a large swarm of similarly oriented dykes that extends throughout the northwestern Hearne subdomain as far as Angikuni Lake (see Fig. 2 for locations; Eade, 1986; Tella et al., 1997; LeCheminant et al., 1997). In both the MacQuoid



(a)

(b)

Fig. 3. (a) Geology of the MacQuoid supracrustal belt and Cross Bay plutonic complex (after Tella et al., 2001). Numbers correspond to locations referred to in the text, and abbreviations are defined in the top-left inset. (b) Structural elements of A.

supracrustal belt and the Cross Bay plutonic complex, the dykes were emplaced after Neoarchean deformation of their wallrocks (Hanmer et al., 1999a, 1999b), but recorded regional metamorphism at ∼1900 Ma (Table 1; Berman et al., 2000; Stern and Berman, 2000).

3. Structure

3.1. MacQuoid supracrustal belt

Deformation fabrics in the eastern part of the metasedimentary homocline and the eastern branch of the

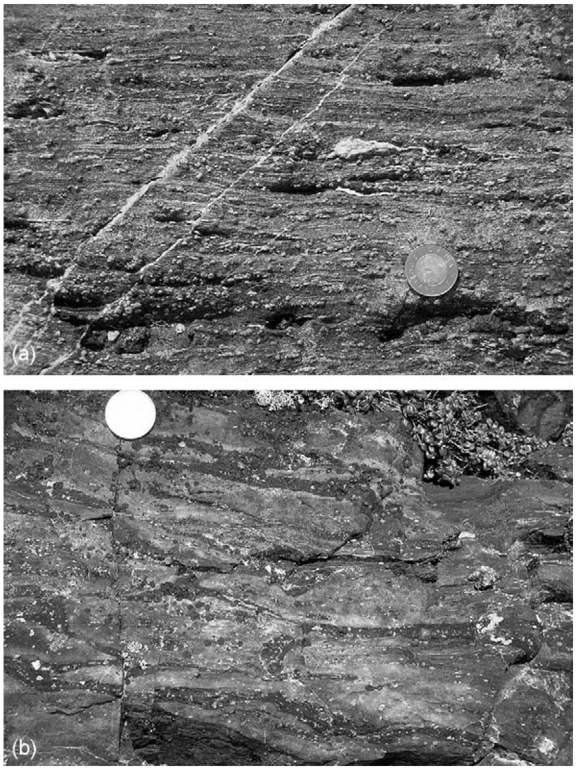


Fig. 4. Amphibolite of the principal volcanic belt. (a) Flattened, attenuated pillows with dark selvedges. Coin for scale. (b) Compositionally laminated amphibolite schist studded with millimetric garnet. Coin for scale.

principal volcanic belt (Fig. 3b) are geometrically simple, comprising a single, penetratively developed, moderately (40–50°) northwest-dipping, layer-parallel foliation, and a moderately to steeply pitching, north to northeast-plunging extension lineation that is particularly well developed in the principal volcanic belt (Hanmer et al., 1999b; Tella et al., 2001). The axes of isolated, isoclinal folds of the foliation are lineation-parallel. However, multiple fabrics and associated folds, are present in the western parts of both the sedimentary homocline and the principal volcanic belt. From detailed inspection (see Ryan et al., 1999, 2000a, 2000b), the first foliation (S1) preserved in the west was folded, crenulated, and transposed into a second foliation (S2), axial planar to F2 folds. This structural sequence can be progressively traced into the single foliation preserved in the eastern volcanic and sedimentary rocks, which is therefore a composite S1/S2 transposition fabric, compatible with the scarcity of relict pillow structures in the volcanic rocks (Fig. 4). Similarly, the extension lineation, developed parallel to F2 fold axes, is L2. Both D1 and D2 fabrics are associated with garnet–staurolite–andalusite–sillimanite, or garnet–hornblende–plagioclase metamorphic assemblages that form the planar and linear fabric elements, and/or are enclosed by the foliation. The S2 transposition foliation and the lower boundary of the eastern branch of the principal volcanic belt wrap around three subjacent ~2685–2655 Ma tonalitic plutons, and form the upright keels that separate them (#1 in Fig. 3a; Table 1). Although the interiors of the plutons are not strongly deformed, strain gradients are developed at their margins, indicating that the tonalitic plutons were already in place when the S2/L2 transposition fabrics developed. Berman et al. (2000) identify this event with mid-crustal (~0.5 GPa) metamorphism dated at ~2550–2500 Ma (Table 1; see also Stern and Berman, 2000). Non-deformed ~2190 Ma MacQuoid dykes cut and provide a minimum age for the S_2/L_2 structures. Similar fabric elements are present north of the principal volcanic belt. However, because of insufficient outcrop, intervening granitoid plutons, and the possibility of tectonic discontinuities (e.g. the South Channel fault; Fig. 3), direct correlation of metamorphic and structural histories is not possible. In the western part of the principal volcanic belt, both S1 and S2 are deformed about the “knot” of regional-scale, upright, moderately east–northeast-plunging folds (#4 in Fig. 3a), interpreted by Ryan et al. (2000a, 2000b) as Paleoproterozoic structures.

South of Akunak Bay, a northwest trending, map-scale panel of semipelite and amphibolite, with minor iron formation, is abruptly discordant with the eastern end of the metasedimentary homocline (#5 in Fig. 3a; Tella et al., 2000, 2001). Lithologically, it appears to be part of the eastern homocline, separated from it by a locally mylonitic, discrete, north–south-trending fault. The layer-parallel, apparently S2 foliation within the panel is folded by upright, strike-parallel folds that are cut by vertical, northwest-trending, ~2190 Ma MacQuoid dykes, and subsequently deformed about a north–south trending, map-scale, neutral fold (Fig. 3; Tella et al., 2000).

3.2. Cross Bay plutonic complex

The Cross Bay plutonic complex is characterized by a penetrative, layer-parallel foliation and extension lineation (Hanmer et al., 1999b; Tella et al., 2000, 2001). In the absence of metamorphically sensitive mineral assemblages, the crystal–plastic behaviour of feldspars suggests that deformation occurred under amphibolite-facies conditions. The foliation varies from steeply dipping and north–south-trending north and west of the Cross Bay fault, to moderately south–southeast- or north–northwest-dipping in much of the rest of the complex (Fig. 3b). In contrast, except for the southern part of the complex where it is poorly preserved, the extension lineation consistently plunges shallowly (∼20°) to the south throughout. Locally, foliation and layering in the gneisses are deformed about isolated, open to tight, meter- to map-scale upright folds, coaxial with the pervasive extension lineation. In the western part of the plutonic complex, the variably deformed ∼2692 Ma South Channel granite (Davis et al., this issue) cross-cuts the foliation and the extension lineation in the adjacent tonalitic gneiss. Therefore, at least the initial north–south structural grain west of Cross Bay appears to have developed during the Neoarchean, prior to the deposition of at least part the MacQuoid supracrustal belt (Table 1). However, the ∼2190 Ma MacQuoid dykes cut the foliated South Channel granite, as well as ∼2700 Ma (Davis et al., this issue) tonalitic gneisses in the central and eastern parts of the plutonic complex. Therefore, the foliation and extension lineation in the latter rocks must have formed during the interval 2695–2190 Ma, after emplacement of the South Channel granite but prior to injection of the MacQuoid dykes. Given the regional deformation history, these fabric elements must also be Neoarchean in age (however, see below).

In the western part of the Cross Bay plutonic complex, the MacQuoid dykes are deformed by steeply plunging folds with north–south axial traces. Locally, the dykes also contain garnet and clinopyroxene that record deep-crustal metamorphic conditions (800–700 °C at 1.0–1.2 GPa), similar to those derived for the static metamorphism in the metasedimentary homocline at ∼1900 Ma (690–660 °C at ∼1.0 GPa), and equated by Berman et al. with this event (Table 1; Berman et al., 2000; Stern and Berman, 2000). However, shallowly south-plunging, open folds localized in tonalitic gneisses at the north end of Big Point peninsula involve weakly foliated ∼1830 Ma monzogranite sheets with a coaxial extension lineation that are cut by non-deformed 1807 Ma granite sheets (#6 in Fig. 3a; Davis et al., this issue). Accordingly, a component of the south plunging extension in the Cross Bay plutonic complex developed during the ∼1830–1805 Ma Paleoproterozoic reworking of the gneisses of the Cross Bay plutonic complex (Table 1) and was associated with east–west shortening.

South and east of Cross Bay, the map pattern in the plutonic complex is dominated by moderately (up to ∼40°) east–northeast plunging, upright folds of layering and foliation in the tonalites and diorites (#7 in Fig. 3a; Hanmer et al., 1999b). On geometrical grounds, Ryan et al. (1999, 2000a) suggested that they correlate with the “knot” of regional-scale, upright, moderately east–northeast-plunging folds in the principal volcanic belt (#4 in Fig. 3a) that deform S1 and S2, which they interpreted as Paleoproterozoic in age.

In summary, similarly oriented fabric elements within the Cross Bay plutonic complex formed both in the Neoarchean and the Paleoproterozoic, highlighting the difficulty in ascertaining the relative timing of deformation structures, especially extension lineations and associated folds, on the basis of field geometry alone. Furthermore, in the absence of cross-cutting relationships, the relative ages of the Paleoproterozoic south- and east–northeast plunging fold structures remains equivocal. It is for these reasons that we hesitate to apply generational terminology (Dn etc.) within the Cross Bay plutonic complex.

3.3. Shear zones

Big lake shear zone (Blsz in Fig. 3) is a steeply dipping belt of S > L porphyroclastic straight gneisses (annealed mylonite) and ribbon mylonites, up to 2 km thick, that marks the southern margin of the Cross Bay plutonic complex and separates it from the supracrustal rocks to the south and west (Hanmer et al., 1999b; Ryan et al., 2000b, 2000c; Tella et al., 2001). To the west, the main segment branches into several, laterally discontinuous mylonitic splays, the largest of which, the western segment, progressively curves to the northwest (#8 in Fig. 3a), and projects along what is now the South Channel fault. The main segment of the shear zone is principally composed of steeply north-dipping, strike-lineated, amphibolite-facies (garnet–hornblende–plagioclase), straight gneisses that preserve evidence for dextral strike-slip movement (Fig. 5).

The western segment differs significantly from the rest of the shear zone (Hanmer et al., 1999b; Ryan et al., 2000b, 2000c). It is predominantly composed of spectacular, steeply southwest-dipping, granulite-facies (750–800 °C at 0.8–0.9 GPa), ribboned and laminated, garnet–clinopyroxene–hornblende mylonites (Fig. 6) derived from localized synkinematic mafic and anorthositic protoliths, plus tonalite and granitic sheets. Evidence for synkinematic emplacement of the magmatic rocks onto the shear zone is two-fold: (i) localization of the compositionally anomalous anorthosite within the shear zone, plus (ii) the variable state of deformation of the granitoid rocks, where less deformed sheets clearly intrude more strongly mylonitic equivalents, within the shear zone. The mylonites carry an oblique, moderately northwest-plunging extension lineation, and are kinematically complex with evidence for both sinistral and dextral shear-sense. Note that the

Fig. 5. Multiple dextral shear-sense indicators on horizontal outcrop surface, in annealed straight gneiss of the main segment of the Big lake shear zone. Note the 'Z' asymmetry of folds (lower field), the clockwise rotation of their axial planes with progressive tightening of the fold profile (centre field), and the concentrations of amphibole (dark) in the upper-right and lower-left pressure shadows of a round amphibolite inclusion in straight gneiss (upper field). Coin for scale.

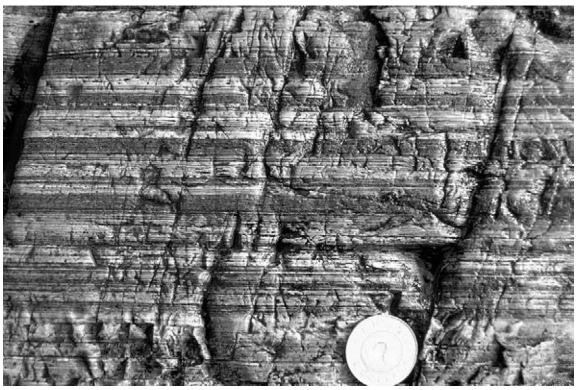


Fig. 6. Finely laminated tonalitic (light) and amphibolitic (dark) ribbon mylonite in the western segment of the Big lake shear zone. Coin for scale.

present dip of this segment of the Big Lake shear zone may not correspond to its original orientation. The lithological association of protoliths and the metamorphic grade are anomalous for the MacQuoid–Cross Bay area. However, their setting is similar to that of mylonites and synkinematically emplaced magmatic protoliths of the Kramanituar and Uvauk complexes on the north side of Chesterfield Inlet (Fig. 2), where localized granulite-facies conditions are also regionally anomalous and attributed to heat advection by mantle-derived, synkinematic melts (Tella et al., 1993; Sanborn-Barrie, 1999; Sanborn-Barrie et al., 2001; Mills et al., 2000; Mills, 2001).

Imprecise dating of the fabric-forming granulite-facies mineral assemblages in the western segment of the shear zone at ∼2500 Ma is supported by non-deformed ∼2190 Ma MacQuoid dykes that cut across the mylonitic fabric, as well as an 1832 ± 4 Ma massive monzogranite that stitches its wallrocks (just northwest of #8 in Fig. 3a; Table 1; Ryan et al., 2000b, 2000c). Post-kinematic MacQuoid dykes also occur within the main east–west segment of the shear zone. However, a few dykes in the main segment are deformed and disrupted, and at one location (#9 in Fig. 3a) the annealed straight gneiss fabric wraps around a high amplitude (1.5 m), joint-controlled, 90°, primary jog in the margin of a 10 m thick dyke (Hanmer et al., 1999b). Therefore, some reworking of the main segment of the shear zone occurred during the Paleoproterozoic (Table 1), although it appears to be localized.

The Akunak Bay shear zone (ABsz in Fig. 3) outcrops discontinuously along the western shore of Akunak Bay as a relatively narrow (∼200 m), northwest trending corridor of steeply dipping, strike-lineated, anastomosing strands (∼10 m wide; Tella et al., 2000, 2001). It is composed of annealed, tonalitic–amphibolitic–granitic, S > L straight gneiss of uncertain shear-sense that appears to mark the northern boundary of the Akunak Bay supracrustal panel (Fig. 3). The main strand of the shear zone is under water, hence it is not apparent in Fig. 3. Outcrop is poor, but the shear zone appears to extend discontinuously westward into the eastern subaerial part of the Cross Bay complex, where Paleoproterozoic ∼1830–1805 Ma monzogranites include and intrude an arcuate, map-scale raft of tonalitic gneiss that contains a curved band or train of similar straight gneiss. Note that this subaerial segment of the shear zone is disproportionately represented in Fig. 3a. Along the shoreline of Akunak Bay, the shear zone is cut by a post-kinematic ∼2190 Ma MacQuoid dyke, thereby allowing that it may be contemporaneous with the principal displacements along the Big lake shear zone.

3.4. Granite domes

South and east of the metasedimentary homocline, closely packed ∼1840 Ma (van Breemen et al., unpublished data), equigranular to megacrystic granite plutons charged with wallrock xenoliths are separated from one another by septa of metasedimentary and minor mafic volcanic rocks (#10 in Fig. 3a, Table 1). Layer-parallel foliation in the wallrocks is concordant to, and dips steeply away from, the generally isotropic interiors of the plutons. Biotite-garnet ± sillimanite metasedimentary rocks in the septa are very coarsely recrystallised, with abundant feldspar porphyroblasts and poorly preserved shape fabrics, indicative of slow cooling from elevated temperatures. Given the absence of geometrically comparable cross-folding in the adjacent MacQuoid homocline, we interpret these features in terms of a set of plutonic domes that were forcibly (cf. Paterson and Fowler, 1993) emplaced into a shallowly dipping panel of foliated metasedimentary rocks. The latter flowed around the plutons to form a generally concordant carapace that was pinched between them. This suggests that the carapace was originally part of a flat lying panel, a southeastern extension of the metasedimentary homocline.

4. Discussion

Geochronological, isotopic and geological evidence (Sandeman et al., 2000a, 2000b, 2006; Davis et al., this issue, this study) suggests that the MacQuoid supracrustal belt and the Cross Bay plutonic complex were tectonically juxtaposed. Given the presence of older rocks over younger, a stratigraphic relationship can be excluded. Accordingly, the questions we address here include: (i) does the geological evidence support the hypothesis of tectonic juxtaposition?; if so, (ii) what does it tell us regarding the kinematic context and the timing of the event?; and (iii) does it correlate with other tectonic events elsewhere in the northwestern Hearne subdomain? As we will show, the answer to all three questions is affirmative.

4.1. Metasedimentary homocline—volcanic belt contact

Detailed geological mapping has highlighted aspects of the MacQuoid supracrustal belt that are compatible with low-angle, Neoarchean shearing. First, the metasedimentary homocline is a crustal-scale panel, 150 km long by >20 km wide (Fig. 3), within which

Fig. 7. Decussate amphibole blades (dark) set in a plagioclase-rich matrix (light) in an intermediate composition rock of the principal volcanic belt.

a ∼2550–2500 Ma S2 foliation has a uniformly moderate northward dip, with a locally steeply pitching extension lineation (L2). Second, S2 is a transposition foliation throughout the eastern 100 km of the homocline, i.e. where it approaches and underlies the Cross Bay plutonic complex. Similar observations apply to the principal volcanic belt, where the S2 transposition fabric extends throughout the 60 km long eastern branch. Third, the eastern branch of the principal volcanic belt strikes east–southeast into the upper boundary of the metasedimentary homocline, where it changes to an east-northeast trend and abruptly thins from ∼5 km wide down to several hundred meters (Fig. 3a). Any increase in fabric intensity with thinning is obscured because the rocks are annealed and the fabrics are commonly overprinted by coarse hornblende crystals (<10 cm, mostly <5 cm) randomly oriented within the foliation plane (Fig. 7). Without widespread strain markers it is not possible to demonstrate unequivocally that the change in thickness of the eastern branch is a function of strain. Nevertheless, the foregoing geological observations would be readily explained if the contact between the metasedimentary homocline and the overlying volcanic belt was a ductile tectonic boundary, i.e. a moderately dipping D2 shear zone. If this interpretation is valid, the lower state of strain in the western part of the principal volcanic belt suggests either that displacement on this contact was relatively limited, or that it was partitioned between a localized ductile shear zone and a more extensive discrete (brittle?) fault. Correlation with the S2 transposition foliation in the metasedimentary homocline implies that displacement along this contact occurred at ∼2550–2500 Ma (Table 1).

4.2. Supracrustal belt—plutonic complex contact

Sandeman et al. (2000, 2001) have shown that the isotopic signature of the Cross Bay plutonic complex (εNd < +1 to < −3 at ∼2700 Ma) contrasts with that of the volcanic and associated tonalitic rocks of the MacQuoid supracrustal belt (εNd +3.6 to +1.8 at ∼2700 Ma). Accordingly, they interpret the former as an exotic fragment of (continental?) crust, possibly related to the Rae domain. The enveloping surface to Paleoproterozoic east–northeast-trending, upright, open folds south and east of Cross Bay is flat lying, as evidenced by preservation of the dish-like geometry of the central part of the complex (#7 in Fig. 3a). Accordingly, prior to this folding event, foliation in this part of the plutonic complex was part of a flat-lying panel. As discussed by Davis et al. (this issue), this foliation is probably contemporaneous with the ∼2.55–2.50 Ga tectonometamorphic events. The metasedimentary homocline projects beneath the plutonic complex at ∼45°, and is separated from it by the Big lake shear zone that, along its main segment, is steeply northward dipping at the surface. However, prior to the east–northeast-trending folding, the shear zone had a shallower northward dip, placing the older, isotopically more evolved gneisses of the Cross Bay plutonic complex over the younger, isotopically juvenile, MacQuoid supracrustal belt. Therefore, one possible interpretation is that the western segment of the shear zone could have acted as a lateral ramp to a south–southeast directed, late Neoarchean thrust, penecontemporaneous with peak metamorphism in the supracrustal footwall at ∼2.55–2.50 Ga. In this scenario, the main segment of the shear zone would originally have represented the leading edge of the thrust. By association, coupled with an admittedly simplistic interpretation of the kinematic significance of the dip-parallel extension lineation (cf. Lin et al., 1998), the anomalously high strain localized in the eastern branch of the principal volcanic belt could also be related to shearing, potentially reflecting the same inferred south–southeast-vergent thrusting. However, kinematic observations require that the main segment of the Big lake shear zone was subsequently reworked as a dextral strike-slip structure prior to intrusion of the ∼2190 Ma MacQuoid dykes, presumably during the late Neoarchean (Table 1).

4.3. Late Neoarchean thick-skinned thrusting (∼2500 Ma)

Despite the lack of kinematic information, the accumulated geological, geochronological and isotopic evidence in the MacQuoid–Cross Bay area is consistent with inferred assembly of the MacQuoid supracrustal belt and the Cross Bay plutonic complex by late Neoarchean thick-skinned thrusting at ∼2550–2500 Ma. The thrusting scenario is not proven. However, it offers a viable framework to explain: (i) the moderate to shallow, regional north to north–northwest dip of the Neoarchean foliations and the steep pitch of associated extension lineations in the eastern part of the metasedimentary homocline and the eastern branch of the principal volcanic belt, (ii) geometrical discordances between the metasedimentary homocline, the principal volcanic belt, and the Cross Bay plutonic complex, (iii) the development of Neoarchean transposition fabrics within and above the metasedimentary homocline where it enters the footwall beneath the Cross Bay plutonic complex, and (iv) the abrupt juxtaposition of an older, isotopically distinct panel above the north-dipping MacQuoid supracrustal belt. It could also explain the difference in metamorphic pressures recorded in the metasedimentary homocline (∼0.5 GPa) and the western segment of the Big lake shear zone (0.8–0.9 GPa; Table 1). Assuming that they are coeval within the ∼2550–2500 Ma time window, rocks in the hangingwall of the shear zone would have been uplifted as they were thrust over the metasedimentary homocline. A map-scale cut-off of the inferred ductile thrust zone at the contact between the metasedimentary homocline and the volcanic belts (#11 in Fig. 3a) indicates that it is older than thrusting at the base of the plutonic complex. This is at least suggestive of a break-back thrusting sequence (e.g. Nadeau and Hanmer, 1992), emplacing the Cross Bay plutonic complex over an already deformed, or actively deforming footwall.

Late Neoarchean, southeast-vergent, ductile thrusting also occurs elsewhere along strike within the northwestern Hearne subdomain. In the northeastern part of the Neoarchean Yathkyed supracrustal belt (Fig. 2; MacLachlan et al., 2000; MacLachlan and Relf, 2000), the northwest-dipping Tyrrell shear zone is a mylonitic fault, up to 2 km thick, that initially acted as a southeast-vergent thrust whose hangingwall was stratigraphically overturned and penetratively deformed at ∼2660–2640 Ma. The upper part of this hangingwall was subsequently reactivated at a structurally higher level as a southeast-vergent ductile thrust zone at ∼2570–2500 Ma that separates structurally overlying hotter rocks from underlying cooler rocks (Fig. 8; MacLachlan et al., 2000; MacLachlan and Relf, 2000). The late Neoarchean thrusting in the Yathkyed belt is southeast-vergent and thick-skinned, with a break-back polarity, as inferred here for the MacQuoid–Cross Bay area. These two areas lie along the southeastern margin of the northwestern

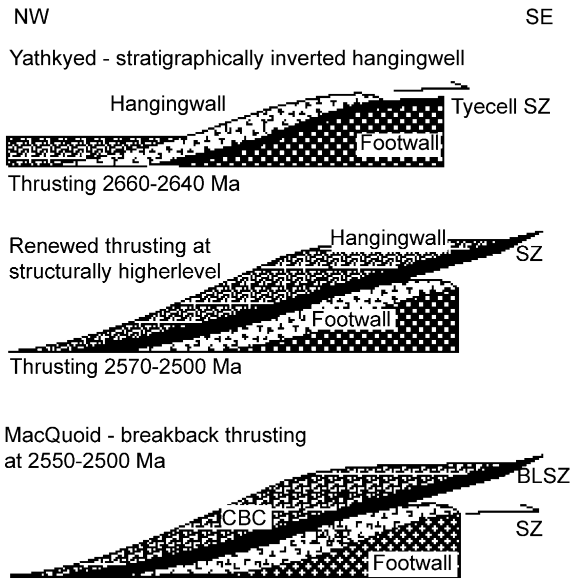


Fig. 8. Simplified structural evolution of polyphase, thick-skinned thrusting in the Yathkyed supracrustal belt at 2660–2640 Ma and 2570–2500 Ma (see MacLachlan et al., 2000; MacLachlan and Relf, 2000). The lower part of the figure represents the principal elements of the Cross Bay plutonic complex overthrusting an already deformed (thrusted) footwall containing the MacQuoid supracrustal belt at ~2.55–2.50 Ga.

Hearne subdomain, and we suggest that they are segments of the same thick-skinned thrust system (Fig. 8). We speculate that the pre-transposition deformation (S1) preserved in the western part of the principal volcanic belt in the MacQuoid supracrustal belt could reflect the ~2660–2640 Ma events recorded in the Yathkyed belt. If valid, this would allow that initial juxtaposition of the rocks of the Cross Bay plutonic complex and the metasedimentary homocline might have occurred at this time.

Outside of the northwestern Hearne subdomain, the period 2550–2500 Ma is a period of tectonic and magmatic quiescence in the rest of the western Churchill Province. Recent geochronological studies have highlighted evidence for: (i) zircon growth in the Snowbird tectonic zone, just north of the Athabasca Basin (Fig. 1; Baldwin et al., 2003), (ii) monazite growth in the Angikuni supracrustal belt (Fig. 3; Berman et al., 2002b), (iii) granulite-facies metamorphism localized within the Uvauk complex (Fig. 2; Mills et al., 2000; Mills, 2001), and (iv) minor shearing in the Kramanituar complex (Fig. 2; Sanborn-Barrie et al., 2001), during this time interval. However, these areas are either part of, or potentially adjacent to, the northwestern Hearne subdomain. In summary, given that late Neoarchean, ~2550–2500 Ma regional tectonometamorphism has not been identified in other parts of the western Churchill Province, it appears to represent a defining characteristic of the northwestern Hearne subdomain. Boundary conditions for this event remain unconstrained at present.

4.4. Fundamental tectonic boundary?

Detailed mapping and supporting geoscience studies have established that the MacQuoid supracrustal belt and the structurally overlying Cross Bay plutonic complex can be discriminated on the basis of lithology, isotopic signature and tectonothermal history. Key observations are that the plutonic complex contains rocks that are older and have a more evolved Neoarchean isotopic signature than parts of the stratigraphic succession in the underlying supracrustal belt. According to Sandeman et al. (2000, 2001; see also Sandeman et al., 2004a), the isotopic signature of the volcanic and associated tonalitic rocks of the MacQuoid supracrustal belt (εNd +3.6 to +1.8 at ~2700 Ma) places it in a relatively short-lived (principally ~2710–2680 Ma) oceanic regime that extended across the Hearne domain and included the Central Hearne supracrustal belt (Fig. 2). In contrast, they interpret the isotopic composition of the Cross Bay plutonic complex (εNd < +1 to <−3 at ~2700 Ma) in terms of interaction with older, more evolved crust, possibly affiliated with the continental Rae domain. Potentially, the Big lake shear zone that separates the plutonic complex from the supracrustal belt could represent a fundamental, crustal-scale tectonic boundary between oceanic and continental crust, reworked at ~2550–2500 Ma. We speculate that this boundary might represent initial accretion of Rae and Hearne domains, a partial test of which could involve demonstration of the isotopic affinity of the Cross Bay plutonic complex and parts of the Rae.

4.5. Paleoproterozoic reworking (2190–1810 Ma)

Paleoproterozoic tectonothermal reworking in the MacQuoid–Cross Bay area occurred with four distinctly different styles (Table 1):

(i) post-2190 Ma localized reactivation of the Big lake shear zone;
(ii) ~1900 Ma regional metamorphism at lower crustal depths (Berman et al., 2000);
(iii) ~1840 Ma pluton-induced deformation of the eastern metasedimentary homocline;
(iv) ~1830–1810 Ma south-trending folds and coaxial extension.

The three deformation-related events do not involve high strains, or major displacements. Rather, they appear to be local attempts to accommodate relatively minor shortening across unspecified, local boundaries (Ryan et al., 2000b, 2000c). The disharmonic responses of the Cross Bay plutonic complex and the MacQuoid supracrustal belt to Paleoproterozoic reworking would have been favoured by weak coupling across the Big lake shear zone. We speculate that the deflection of annealed amphibolite-facies straight gneiss around a joint-controlled apophysis on a MacQuoid dyke in the main segment of the shear zone could be related to minor strike-slip displacements during the ∼1900 Ma metamorphic event when temperatures were suitably elevated. By similar reasoning, the south-trending folding of MacQuoid dykes in the Cross Bay plutonic complex might also have occurred at that time (Table 1). Finally, there is no evidence for ductile reactivation of the Big lake shear zone at ∼1830 Ma, although it was reworked in the west by the brittle South Channel fault after deposition of the Dubawnt Supergroup (post-1840 Ma; Rainbird et al., 2003; Table 1), raising the possibility of cryptic, brittle decoupling during the folding event at ∼1830–1810 Ma.

These events are of the same vintage as extensive folding and thrusting recorded by the <2450–1910 Ma Hurwitz Group that sits unconformably on the Neoarchean rocks of the Hearne domain (e.g. Aspler and Chiarenzelli, 1997; Aspler et al., 2001, 2002), and the bloom of Hudson monzogranite plutons that straddles the Rae-Hearne boundary (Fig. 2). They are also contemporaneous with widespread regional deformation and metamorphism that affected much of the northern part of the Rae domain between Baker Lake and Melville Peninsula (Figs. 1 and 2; Zaleski et al., 2001; Sanborn-Barrie et al., 2003; Carson et al., 2004). All the foregoing cited authors agree that these events are genetically related to the Trans-Hudson orogen, whose western internides are located well to the south of our study area (see Fig. 1).

5. Conclusions

Mid-crustal, southeast-directed, thick-skinned ductile thrust zones, localised at the contacts between the metasedimentary homocline and the principal volcanic belt of the MacQuoid supracrustal belt, and between the supracrustal belt and the older, overlying Cross Bay plutonic complex, are late Neoarchean D2 structures (∼2.55–2.50 Ga). The latter is marked by the kinematically complex Big lake shear zone. A similar, contemporaneous structural signature has been identified along strike in the Yathkyed supracrustal belt, indicating that this event is widespread within, and characteristic of, the northwestern Hearne subdomain.

Neoarchean deformation fabrics in both the MacQuoid supracrustal belt and the Cross Bay plutonic complex are cut by the ∼2190 Ma MacQuoid mafic dyke swarm that acts as a time marker separating Neoarchean events from ∼1.9 Ga, deep-crustal, regional metamorphism that affected both the supracrustal belt and the plutonic complex. Other Paleoproterozoic deformation events that occurred at ∼1840–1810 Ma, including open folding in the plutonic complex and forceful pluton emplacement, appear to represent relatively mild, localised, far-field manifestations of regional events related to the Trans-Hudson orogen.

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