Structural development of the Snow Lake Allochthon and its role in the evolution of the southeastern Trans-Hudson Orogen in Manitoba, central Canada

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Abstract: The Snow Lake Allochthon is a zone of tectonic interleaving of sedimentary rocks of an inverted marginal basin (Kisseynew Domain) with island-arc and oceanic rocks. It is located in the southeastern part of the exposed internal zone of the Paleoproterozoic Trans-Hudson Orogen in Manitoba, Canada, near the external zone (Superior collision zone or Thompson Belt), which constitutes the local boundary between the Trans-Hudson Orogen and the Archean Superior Craton. The Snow Lake Allochthon formed, was deformed, and was metamorphosed up to high grade at low to medium pressure during the Hudsonian orogeny as a result of the collision of Archean cratons -1.84–1.77 Ga. Four generations of folds (F1–F4) that formed in at least three successive kinematic frames over a period of more than 30 Ma are described. Isoclinal to transposed southerly verging F1–2 structures are refolded by large, open to tight F3 folds and, locally, by open to tight F4 folds. The axes of the F1–2 folds are parallel or near parallel to the axes of F3 folds, owing to progressive reorientation of the F1–2 axes during south- to southwest-directed tectonic transport, followed by F3 refolding around the previous linear anisotropy. A tectonic model is presented that reconciles the distinct tectono-metamorphic developments in the Snow Lake Allochthon and the adjacent part of the Kisseynew Domain on the one hand, and in the Thompson Belt on the other, during final collision of the Trans-Hudson Orogen with the Superior Craton.

Introduction

The Trans-Hudson Orogen of North America constitutes one of the continental collision zones between Archean fragments, along which the Laurentia–Baltica protocraton was assembled between 2.0 and 1.8 Ga (Fig. 1) (e.g., Lewry 1981; Green et al. 1985; Hoffman 1988, 1989; Bickford et al. 1990; Ansdell et al. 1995). The external zone of the orogen contains the reworked margins of the bounding Archean Rae–Hearne and Superior cratons including rift facies rocks (Fig. 1) (Lewry 1981; Hoffman 1988; Bleeker 1990). The eastern termination of the Snow Lake Allochthon (Kraus and Menard 1997) in the internal zone is located less than 30 km west of the external zone of the orogen (Superior collision zone or Thompson Belt). It formed, was deformed, and was metamorphosed up to high grade at low to medium pressure during the Hudsonian orogeny -1.84–1.77 Ga (e.g., Froese and Gasparrini 1975; Bailes and McRitchie 1978; Kraus and Menard 1997). The allochthon constitutes a zone
of tectonic interleaving of ~1.9 Ga island-arc and oceanic assemblages with 1.86–1.84 Ga metasedimentary rocks of the Kisseynew Domain, a former marginal basin (e.g., Zwanzig 1990; Bailes and Galley 1996, 1999; Stern et al. 1995; Connors 1996; David et al. 1996; Lucas et al. 1996; Kraus and Menard 1997). The Snow Lake Allochthon structurally overlies the 1.92–1.88 Ga Amisk collage of juvenile island-arc and oceanic assemblages to the west, structurally underlies migmatitic paragneisses and granitoids of the Kisseynew Domain to the north and east (Figs. 1, 2), and continues as the Clearwater Domain southward underneath the Phanerozoic cover (Stern et al. 1995; Connors 1996; Lucas et al. 1996; Kraus and Menard 1997). The inverted Kisseynew basin and the Snow Lake Allochthon including the juvenile volcanic assemblages are believed to constitute the upper and middle portions of a tectonic pile, respectively, that overthrust Archean basement (Sask Craton) (Bickford et al. 1990; Ansdell et al. 1995; Lucas et al. 1996). This basement is now exposed as inliers in the Glennie Domain and in the Hanson Lake Block (Fig. 1) (Chiarenzelli et al. 1998; Ashton et al. 1999). The Snow Lake Allochthon, according to the interpreted Litho-probe seismic lines 2 and 3, is not underlain by Archean crust (Lucas et al. 1993; White et al. 1994; Leclair et al. 1997). The relationships between the allochthon and its footwall assemblages cannot be resolved from the seismic reflection data and are thus unknown (e.g., White et al. 1994, their Fig. 3). In this paper, we discuss the tectono-
Fig. 2. Geological map of the Snow Lake – File Lake – Wekusko lakes area, including isograds from authors cited in Kraus and Menard (1997). Geochronological results refer to age of emplacement of granitic rocks (from Gordon et al. 1990; Bailes et al. 1991; David et al. 1996). Faults: BCSZ, Berry Creek shear zone; BeLF, Beltz Lake fault; BLF, Birch Lake fault; CBF, Crowduck Bay fault; LLF, Loonhead Lake fault; MLF, Morton Lake fault; MRF, McLeod Road fault; RLF, Roberts Lake fault; SLF, Snow Lake fault. Plutons (P), plutonic complexes (C), and gneiss domes (GD): BLC, Batty Lake; BLP, Bujarski Lake; HLGD, Herblet Lake; HLP, Ham Lake; NBP, Nelson Bay; PLGD, Pulver Lake; RiLP, Richard Lake; RLC, Rex Lake; RLP, Reed Lake; SLGD, Squall Lake; SLP, Sneath Lake; TLP, Tramping Lake; WLP, Wekusko Lake. Mineral abbreviations after Kretz (1983).
metamorphic evolution of the Snow Lake Allochthon in the context of the evolution of the southeastern Trans-Hudson Orogen during final continent–continent collision.

**Regional geology**

Four tectono-stratigraphic units are juxtaposed in the Snow Lake Allochthon: volcanic and sedimentary rocks of the Snow Lake (arc) assemblage (1), ocean-floor assemblages (2), and the younger metasedimentary rocks of the Burntwood (3) and Missi (4) groups (Figs. 2, 3). The Snow Lake assemblage comprises bimodal volcanic and associated volcanioclastic rocks that host synvolcanic granitoids and gabbros (Figs. 2, 3) (Galley et al. 1993; Bailes and Galley 1996, 1999; Stern et al. 1995; David et al. 1996). After metamorphism and polyphase deformation associated with the Hudsonian orogeny, most of the granitoids appear as gneiss domes (Fig. 2) (Bailes 1975). Burntwood and Missi group rocks appear as folded slivers in the arc and ocean-floor rocks (Figs. 2, 3). The Burntwood group comprises uniform, well-bedded greywacke turbidites that can be followed to the north and east, where they occupy the 300 km by 150 km Kisseynew Domain (Fig. 1). The contemporaneous Missi group contains nonmarine, crossbedded, arkosic metasandstones and conglomerates and forms a rim along the boundary of the Kisseynew Domain (Figs. 1, 2). Missi–Burntwood contacts are not exposed in the study area; they are, however, most likely transitional (see Syme et al. 1995). The contacts between sedimentary and volcanic rocks are, with minor exceptions, tectonic (Fig. 2). The tectono-stratigraphic sequence is intruded by mafic sills and low-angle dykes that contain an S1 fabric and hence they predate, or are coeval with, F1 compressional deformation (Table 1). Voluminous calc-alkaline 1.84–1.83 Ga plutons truncate F1 structures (Figs. 2, 3, 4a) (Gordon et al. 1990; Bailes 1992; Connors 1996; David et al. 1996). Their intrusion ages thus give an upper age limit for F1. A suite of mafic to felsic volcanic rocks which is of approximately the same age as the plutons is exposed east of Wekusko Lake (Fig. 2) (Gordon et al. 1990; Ansdel and Connors 1999). Large volumes of late kineomatic to postkinematic granitic pegmatite have intruded the higher metamorphic grade portions of the Snow Lake Allochthon and the Kisseynew Domain including its southern flank (Bailes 1975, 1985; Zwanig and Schledewitz 1992; Kraus and Williams 1994; Norman et al. 1995; Connors 1996).

**Metamorphic zones**

Froese and Gasparini (1975) divided the Snow Lake–File Lake area into metamorphic zones that are separated by reaction isograds (Fig. 2). These zones were extrapolated east towards the Superior collision zone (Thompson Belt) (Fig. 5) (see references in Kraus and Menard 1997). Temperatures at the thermal peak increase to the north in the Snow Lake–File Lake area, and to the north and east on the eastern side of Wekusko Lake, from <400°C on Wekusko Lake to 750 ± 50°C in the Kisseynew Domain (Fig. 5) (e.g., Bailes 1985; Gordon 1989; Briggs and Foster 1992; Leclair et al. 1997; Kraus and Menard 1997; Marshall et al. 1997; Menard and Gordon 1997). Metamorphic isotherms and isograds are hence approximately parallel to the Snow Lake Allochthon – Kisseynew Domain boundary (e.g., Bailes 1985; Gordon 1989; Briggs and Foster 1992; Leclair et al. 1997; Kraus and Menard 1997; Marshall et al. 1997; Menard and Gordon 1997). The metamorphic field gradient (the $P$–$T$ conditions along the presently exposed surface) is approximately isobaric; pressures associated with the thermal peak of metamorphism are 4–6 kbar (1 kbar = 100 MPa) in the Snow Lake–File Lake area and <4 kbar east of Wekusko Lake (e.g., Bailes 1985; Gordon 1989; Briggs and Foster 1992; Leclair et al. 1997; Kraus and Menard 1997; Marshall et al. 1997; Menard and Gordon 1997).

**Deformation history**

Deformation in the study area (Fig. 3) resulted in four generations of folds ($F_1$–$F_4$) and associated faults related to at least three successive kinematic frames. Associated metamorphism occurred in a single, regionally diachronous cycle (Table 1). Primary layering and the axial planes of isoclinal $F_1$ and $F_2$ structures generally verge southerly (i.e., the fold axial planes are overturned southerly) and are refolded by north-northeast-trending open $F_3$ folds with steep axial planes (Fig. 3). Steep east–west-trending $F_2$ structures are localized in and around the Squall Lake and Herblet Lake gneiss domes north of Snow Lake (Fig. 3). In large parts of the central Kisseynew Domain and on its southern flank (Fig. 1), isoclinal $F_{1,2}$ structures are also overturned southerly, but their axial planes are relatively shallow; on the southern flank of the Kisseynew Domain, $F_{1,2}$ structures are steep only in the immediate vicinity of the more rigid Amisk collage footwall (Zwanig 1990, 1999; Zwanig and Schledewitz 1992; Norman et al. 1995). Compared with the Snow Lake Allochthon, structures in the Amisk collage are different in style, orientation, and age (Table 1) (e.g., Ansdel and Ryan 1997; Ryan and Williams 1999). Deformation in the Amisk collage was mainly accommodated along steeply dipping shear zones, many of which predate Missi and Burntwood sedimentation (e.g., Ansdel and Ryan 1997; Ryan and Williams 1999). A structural correlation between the Amisk collage and the Snow Lake Allochthon was attempted by Ryan and Williams (1999).

Individual generations of structures in the Amisk collage and on the southern flank of the Kisseynew Domain were assigned to discrete deformation events based on the different kinematic frames in which these structures developed (Norman et al. 1995; Ryan and Williams 1999). Thus, structures formed by a progressive strain in a constant kinematic frame belong to a single deformation event. It is, however, problematic to apply the deformation-event concept to the Snow Lake Allochthon (cf. Connors 1996), because there $F_1$ and $F_2$ structures, although they may have formed in the same kinematic frame, are separated by 15–35 Ma (David et al. 1996). Conversely, overlapping geochronological ages (within error; see below) indicate that $F_2$, $F_3$, and possibly $F_4$ structures formed in successive, continuously changing kinematic frames during a single orogenic cycle rather than in discrete episodes. These changes in kinematic frames are related to the final collision of the internal Trans-Hudson Orogen with the Superior plate (see below). We therefore cautiously designate the structures of the Snow Lake
Fig. 3. Simplified geological map of the Threehouse synform area at Snow Lake (SL). The letters A–E refer to structural domains. ABA, Anderson Bay anticline. Faults: ABSZ, Anderson Bay shear zone; BCSZ, Berry Creek shear zone; BSZ, Bartlett shear zone; SL, town of Snow Lake. Abbreviations of the granitoid plutons as in Fig. 2. Additional structural data were taken from Harrison (1949) and Froese and Moore (1980). A three-dimensional view of this area is given in Fig. 6.
Allochthon to deformation events in Table 1, however only for the purpose of correlation with adjacent areas.

**F₁–F₄ structures and metamorphic fabrics**

**F₁**

The Burntwood group rocks contained an S₀-parallel S₁ that was generally destroyed during the formation of an S₂ (Kraus and Williams 1998). Where locally present, S₁ is defined, depending on metamorphic grade, by chlorite or muscovite aligned parallel to bedding (Kraus and Menard 1997; Kraus and Williams 1998). S₁ is also preserved as straight to curved inclusion trails in porphyroblasts of garnet, staurolite, and biotite (Kraus and Menard 1997; Kraus and Williams 1998). An L₁ is only preserved in syn-F₁ bedding-parallel quartz veins; it is defined by stretched quartz aggregates, which plunge approximately parallel to minor fold axes. Three orders of F₁ folds appear in the Burntwood group, with amplitudes of centimetres to metres, tens to hundreds of metres, and kilometres (Fig. 3). Small F₁ folds are rare. When several small F₁ folds are exposed in one outcrop, they are variably flattened and have in common that bedding is markedly thicker in the hinge than in the limbs. Many of the small F₁ folds were dismembered by shearing along their limbs and (or) axial planes, so a change in younging direction and, commonly, a set of quartz veins tracking the shear planes are the only evidence of F₁ folding. Z- and S-asymmetrical folds and, locally, doubly verging fold pairs (sensu Holdsworth and Roberts 1984) occur on both limbs of larger F₁ structures. All larger folds were also dismembered and hence have not preserved their asymmetry. F₂ features were not identified in the strongly recrystallized Missi group rocks.

In hydrothermally altered volcanic rocks that are now semipelitic chloritic schists, S₁ constitutes a fabric parallel to primary layering (where no S₂ is developed) and is defined by aligned chlorite. S₁ is also contained as straight inclusion trails in porphyroblasts of garnet, staurolite, and biotite. These porphyroblasts are, however, not indicative of metamorphic grade (Zaleski et al. 1991). In unaltered volcanic rocks, an S₁ schistosity parallel to primary layering is defined by aligned amphibole in the mafic rocks and by aligned biotite in the felsic rocks. Minor and intermediate F₁ folds are rare in the volcanic rocks and are developed mainly along lithological contacts. Younging criteria and consistently dextral S₁/S₂ and S₁/S₀ asymmetry (see below) suggest that the volcan stratigraphic sequence is possibly repeated between the Snow Lake “fault” and the Berry Creek shear zone across a crustal-scale F₁ anticlinal structure, termed the An-

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Table 1. Summary of the tectono-metamorphic history of the Snow Lake Allochthon.

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anastomosing system, which includes the poorly exposed Anderson Bay shear zone and Bartlett shear zone (Figs. 3, 6). F1 movement along the Berry Creek shear zone resulted in a less than 100 m wide zone of moderately to steeply northwesterly dipping felsic to mafic tectonites that contain northeasterly plunging stretched clasts (Fig. 7a) with an aspect ratio of at least 25:11:5:1. Burntwood group rocks adjacent to the shear zone are relatively weakly strained and young away from it (Fig. 3).

The Snow Lake fault constitutes the southern termination of the isoclinally folded Burntwood group sliver at Snow Lake (Figs. 2, 3, 6) (see Kraus and Williams 1998). It extends as the Loonhead Lake fault to the west through the File Lake area, where it appears to form the boundary between the Amisk collage and the southern flank of the Kisseynew Domain (Zwanzig 1999; Connors 1996). The Snow Lake fault is cut off in the east by the younger McLeod Road (reverse) "fault" (see later in this paper). At Snow Lake, the Snow Lake fault is inferred from local variations in strike of the rocks in the hanging wall (Burntwood group) relative to the rocks in the adjacent footwall (arc and minor ocean-floor rocks), and from abundant quartz–carbonate brecciation. Clasts in metabasalts of the immediate footwall show a high degree of stretching parallel to their northeasterly plunge. The Snow Lake fault is stitched by an undeformed, syn- to post-peak metamorphic pegmatite at the southern end of Squall Lake and by the ~1.830 Ga post-F1 Ham Lake pluton northeast of File Lake (Fig. 2) (Connors 1996; David et al. 1996). South of the town of Snow Lake, the fault cuts off an F1 syncline in the Burntwood group at a low angle (Fig. 3). The early movement on the Snow Lake fault is thus constrained to have occurred after the initiation of F1 folding.

The Berry Creek shear zone and the Snow Lake fault possibly originated as shallow thrusts during early south to southwest movement of thick sedimentary nappes derived from the Kisseynew Domain onto the Snow Lake arc (see below; cf. Connors 1996).

F2

An S2 crenulation cleavage or schistosity, which envelopes porphyroblasts, is ubiquitous in the metasedimentary rocks and in the hydrothermally altered volcanic rocks (Figs. 4b–4d) (Kraus and Menard 1997; Menard and Gordon 1997; Kraus and Williams 1998). Growth of the porphyroblasts in the Snow Lake Allochthon commenced, independent of final metamorphic grade, in the early stages of F2 (Kraus and Menard 1997; Menard and Gordon 1997; Kraus and Williams 1998). Depending on metamorphic grade, the S2 cleavage septa in the Burntwood group rocks are defined by muscovite and (or) chlorite. Garnet and biotite grains occupy the microlithons, and the latter are crystallographically and dimensionally aligned parallel to S2. Many biotite grains are pulled apart, the stretching directions being parallel and at a high angle to the moderately to steeply northeasterly plunging fold axes and S0/S2 lineation. Locally, the porphyroblasts...
have pressure shadows parallel to $S_2$ at a high angle to the northeasterly plunging $F_2$ fold axes and $S_0/S_2$ intersection lineation; these pressure shadows developed early during $F_2$ folding (Kraus and Williams 1998). Quartz veins that formed during the subsequent stages of $F_2$ folding are ubiquitous (Fig. 4b). These veins, which are broadly parallel to $S_2$, locally cut across peak metamorphic staurolite and are crenulated by $F_3$ (Kraus and Williams 1998). In Missi group rocks and weakly altered rhyolite, $S_2$ is defined by dimensionally oriented biotite. In fluid-altered volcanic rocks at Anderson Lake (Fig. 3), an $S_1$ is overgrown by kyanite and staurolite porphyroblasts, and both are folded into decimetre-scale asymmetrical folds, which have an axial-plane domainal $S_2$ cleavage (Fig. 4d).

A domainal $S_2$ cleavage is typically only developed in volcanic rocks that contain phyllosilicates (which define the septa). In the granitic gneiss domes and in the surrounding high-grade rocks, in contrast, $S_2$ constitutes a gneissosity that is parallel to lithological contacts and that wraps around garnet grains (Fig. 4f). A composite $L_{1-2}$ in the volcanic rocks is defined by stretched quartz aggregates, elongate amygdales and clasts, rods, aligned grains of amphibole (Fig. 4e), kyanite, staurolite, and disk-shaped garnets. Where $S_2$ is present, $L_{1-2}$ is parallel to the $S_0/S_2$ intersection. Minor $F_2$ folds are rare and are flattened to a lesser degree than $F_1$ structures; such folds are mainly Z-asymmetrical and have a well-developed axial-plane $S_2$ (Figs. 4b, 4d). One large $F_2$ structure, the curvilinear McLeod Lake fold, has been identified (Fig. 3). It is cored by Missi sandstones that young towards the axial plane and appears to be laterally continuous from Snow Lake to File Lake (see Connors 1996). The significance of this structure is discussed below.

$F_2$ shear zones

The McLeod Road “fault” and the poorly defined Birch Lake “fault” are major, possibly contemporaneous reverse structures, which are, with exceptions, parallel or at low angles to primary layering (Figs. 3, 6). Both structures are parallel to one another and have arcuate traces curved more than 180° due to $F_2$ and $F_3$ refolding (see below) (Figs. 3, 6).

The moderately northerly to easterly dipping McLeod Road fault, to the north between Snow Lake and Squall Lake, cuts up section across the axial plane of the McLeod Lake fold (Fig. 3). The McLeod Road fault constitutes an $F_2$ thrust, which dismembered the McLeod Lake fold during tightening (Kraus and Williams 1998). Basalt in the immediate hanging wall is impregnated with synkinematic carbonate in a zone 20 cm to several metres wide. The carbonate is strongly foliated and contains transposed minor folds with axes parallel to the stretching lineation, plunging moderately to the northeast (Fig. 7c). The shear zone fabric in the hanging wall anastomoses around porphyroblasts and lenticular low-strain domains of decimetre scale. In the footwall, high strain is concentrated in decimetre-wide zones containing northeasterly plunging shear folds, up to 20 m beneath the contact. Fault-related folds in the hanging wall and footwall overprint peak-metamorphic porphyroblasts. We interpret the direction of movement along the McLeod Road fault, based on the northeasterly plunging stretching lineation (Fig. 7c), as being top to the southwest. This transport direction is roughly preserved after polyphase deformation, as indicated by the approximate parallelism of the stretching lineation (Fig. 7c) with the $F_2$–$F_3$ linear features in the core of the Threehouse synform at Snow Lake (Fig. 8). Thus, the local $F_4$ overprinting is weak here (see below).

The Snow Lake fault and the Berry Creek shear zone were reactivated as thrusts during $F_2$. The $F_2$ manifestation of the Berry Creek shear zone truncates the 1837±6 Ma (David et al. 1996), post-$F_1$ Tramping Lake pluton (Figs. 2, 3, 7b). The transition from undeformed to a subvertical ultramylonitic foliation at the granite margin is only a few centimetres wide. The width of the $F_2$ deformation zone is indeterminate.
owing to lack of outcrop, but is in excess of 30 cm. Ultramylonitic portions mainly comprise fine-grained, dynamically recrystallized quartz and feldspar. The quartz, however, is granoblastic due to later static recrystallization. Shear bands indicate a south- to southwest-directed hanging-wall transport. Subsidiary centimetre- to decimetre-wide steep ductile shears parallel to the main structure are developed within the otherwise unfoliated pluton. Sericite-rich domains in the subvertical S₂ mylonitic foliation at the pluton margin are crenulated by F₃ and contain a faint, steep northeasterly plunging L₃ crenulation lineation (Fig. 7b). The latest movement along the shear zone system is manifest by pseudotachylyte, which, in one location, cuts across the mylonitic fabric of the Anderson Bay shear zone and is deflected in a sinistral sense.

The northeastern segment of the Berry Creek shear zone (north of the Tramping Lake pluton; Figs. 2, 3), on the other hand, was not reactivated during F₂. It is overprinted by the regional S₂ that envelopes peak metamorphic porphyroblasts containing S₁, the L₂ crenulation lineation, small Z-asymmetrical F₂ folds, and small symmetrical F₃ folds (Figs. 4e, 7a). L₂ plunges parallel to the clasts stretched during F₁ (Fig. 7a). Calculated P–T estimates of ~550°C at 4.1 kbar on both sides of the structure also suggest no major postpeak metamorphic offset (Kraus and Menard 1997). The local F₃ manifestation of the Berry Creek shear zone appears to run offshore in Wekusko Lake within the Burntwood group metaturbidites, as indicated by linear geoelectrical anomalies (Hudson Bay Mining and Smelting, unpublished data). The segment of the Berry Creek shear zone exposed in the study area is, as a whole, openly folded by F₃, causing its trace to be sinusoidal (Fig. 3).

Evidence of F₂ movement along the Snow Lake fault is given by a reversal in S₀/S₂ asymmetry across the structure from sinistral in the northern hanging wall (Kraus and Williams 1998) to dextral in the southern footwall; younging is to the north on both sides of the structure. The dextral S₀/S₂ asymmetry and a corresponding Z-asymmetry of minor F₂ folds remain constant across the inferred Anderson Bay anticline and in outcrops on Wekusko Lake (Fig. 3). This implies that the Snow Lake fault dismembered a major F₂ fold.

F₃ and F₄

F₃ produced polyharmonic folds at all scales throughout the Snow Lake – Reed Lake areas. The largest structure appears to be the Reed Lake fold (Syme et al. 1995), which contains the File Lake antiform (see Connors 1996) and the Threehouse synform (Figs. 2, 3). These large open folds are symmetrical. The Threehouse synform contains parasitic open F₃ folds a few hundreds of metres in scale that are generally discontinuous along their axial planes. Rare minor F₃ folds are open to tight and overprint S₂, trains of peak metamorphic porphyroblasts, and their F₂ pressure shadows (Fig. 4g). F₃ crenulations of S₂ are developed locally around the Threehouse synform, where S₂ is at low angles to bedding; these crenulations deform S₂-parallel quartz veins and pressure shadows on staurolite and biotite (Kraus and Williams 1998). Pervasive S₃ is only developed in low-grade rocks of the Burntwood group at Reed Lake (Syme et al. 1995) (Fig. 2).

Approximately east–west-trending F₄ folds overprint F₃ structures, giving rise to dome and basin to mushroom interference structures (Figs. 2, 3). The variations in the geometry of the interference patterns arise from a regional variation in the orientations of primary layering prior to F₃ folding. F₄ folds occur only locally in approximately north–south-trending domains such as the west limb of the Threehouse synform. Here, no cleavage is associated with the F₄ structures, but Connors (1996) report a local S₃ in F₄ fold hinges north of File Lake (Fig. 2). In unfavourably oriented domains, for example on the eastern Threehouse limb, biotites in fluid-altered volcanic rocks, aligned with their (001) faces along a north-northeasterly-trending moderate to
steep $S_0$, are kinked. Similarly, conjugate kink bands in north-northeasterly-trending steep $S_2$ occur on the islands of northwest Wekusko Lake (Fig. 4h).

**Structural domains**

Five structural domains (A–E) are distinguished in the study area based on the orientation of F1–F3 structural elements (Fig. 3). Domain A (Fig. 3) essentially coincides with the staurolite + biotite zone. It is characterized by coaxial F1–F3 structures and coplanar F1 and F2 structures (Fig. 8). F1–F2 fold axes and all other linear features plunge into the northeast quadrant of the stereonet. The poles to the moderately to steeply dipping $S_0$, $S_2$ and axial planes of F3 crenulations plot on approximately the same great circle. This great-circle girdle dips to the south-southwest and is perpendicular to the local axis of the F3 Threehouse synform. Similar geometrical relationships apply to large portions of the Kisseynew Domain, where the great-circle girdles are slightly steeper (Bailes 1975; Zwanzig 1990).

Domain B (Fig. 3) is transitional between A and C and contains the axial plane of the F2 McLeod Lake fold and the McLeod Road fault. Both structures are overprinted by F3 and F4 (Fig. 3). In domain B, all planar elements plot approximately on a great circle that dips to the southwest (Fig. 9), more steeply than the girdle in domain A (Fig. 8), owing to local F3 refolding of the axial plane of the F2 fold about a steep axis. The pole to this girdle is parallel to the local F3 and F4 axes and coincides with the southwesterly transport direction along the McLeod Road fault (see earlier in this paper; Fig. 7c).

Domain C (Fig. 3) contains a segment of the McLeod Lake fold in Missi group metasandstones that young towards the hinge area along the Missi–Burntwood boundary, poles to $S_0$ yield a moderately northeasterly plunging major F2 axis (southern part of domain C in Fig. 10a). The fold hinge is approximately horizontal in the northern part of domain C (Fig. 10a).

Domain D (Fig. 3) hosts the Squall Lake gneiss dome, a culmination within an F1–F4 dome and basin interference structure (Fig. 11). The gneissosity in granitic orthogneiss dips gently towards the margins of the dome. A shallow mineral lineation appears to track a smoothly curved north-northeast- to south-southwest-plunging F3 axis.

Domain E (Fig. 3) defines a steep belt exposed on islands in northwestern Wekusko Lake (Fig. 12). The variations in $S_0$ trend results from large-scale open F3 refolding. Linear features are steeper than in domains A and B, but they also plunge into the northeast quadrant.

**Discussion of deformation**

The curvilinear McLeod Lake fold

Although the hinge of the F2 McLeod Lake fold is dismembered by the McLeod Road fault in parts of domains B and in domain A, we can estimate the hypothetical local orientation of the major F3 axis by assuming parallelism of the hinge with the $S_0$/$S_2$ lineation. This information is important for the reconstruction of the McLeod Lake fold prior to overprinting. In its present, refolded state, the McLeod Lake fold changes its character from a synform in domain C to an antiform in domains B and A (Fig. 3), the hinge being curved through $-60^\circ$. In the northern part of domain C, the hinge is subhorizontal and becomes progressively steeper northeasterly, plunging in the southern part (Fig. 10a) and through domains B and A (Figs. 8, 9).

In an attempt to reconstruct the original geometry and attitude of the F3 McLeod Lake fold, we restore its orientation prior to F3 and F2 refolding. We assume an initial south to southwest overturning, as reported for large F2 structures from the southern flank of the Kisseynew Domain (Fig. 1) (Zwanzig and Schledewitz 1992; Norman et al. 1995; Zwanzig 1999). The unfolding procedure is described in Fig. 10b. After unfolding, the F2 axial plane dips moderately to the north-northwest, and the local hinge segments plunge moderately west-northwest to steeply north-northeast (Fig. 10b). Similar orientations of F2 axial planes and F2 fold axes in the Cleunion Lake – Kississing Lake area (Fig. 1) were interpreted as being evidence for a local south- to southwest-directed tectonic transport (Norman et al. 1995). Thus, if there is any validity in our simple unfolding procedure, the McLeod Lake fold was a south- to southwest-overturned, synformal, curvilinear structure, and possibly a sheath fold. It is similar in geometry and attitude to the large F3 structures on the southern flank of the Kisseynew Domain (e.g., Zwanzig and Schledewitz 1992; Zwanzig 1999).

**Parallelism of linear features, tectonic transport, and kinematic frames**

Except in domains C and D, the axes of the isoclinal F1 and F2 folds are broadly parallel to the axes of the open F3 folds, and their azimuth approximately coincides with the inferred south- to southwest-directed tectonic transport. Coaxiality of earlier tight structures with later open structures has been reported from many orogenic belts (e.g., Knill...
and Knill 1958; Bryant and Reed 1969; Escher and Watterson 1974; Skjernaa 1980; Meneilly and Storey 1986; Norman et al. 1995). In those cases, parallelism of the early fold axes was attributed to progressive rotation of the fold hinges towards the direction of tectonic transport in a constant kinematic frame, associated with high shear strains \((\gamma > -10)\) (e.g., Knill and Knill 1958; Bryant and Reed 1969; Escher and Watterson 1974; Skjernaa 1980; Meneilly and Storey 1986; Norman et al. 1995). The occurrence of sheath folds is regarded as evidence that this process has operated (Williams and Zwart 1977). The lack of sheath folds, however, does not exclude this possibility (Mawer and Williams 1991; Williams and Compagnoni 1983; Norman et al. 1995).

We believe that, in the study area, the near-parallelism of the linear \(F_{\text{1–2}}\) features and also the chocolate-block boudinage pattern (see above) were achieved by rotation in response to shearing within thick thrust sheets during south-to-southwest-directed nappe emplacement. Boudins that formed by stretching in a direction parallel to the lineation were linear features and were, as straining continued, rotated and further boudinaged in the same way as the \(F_{\text{1–2}}\) fold hinges.

Although the orientation of the minor \(F_{\text{1–2}}\) fold axes alone is far from conclusive evidence for the direction of \(F_{\text{1–2}}\) tectonic transport, our interpretation is consistent with the tectonic transport inferred for the region (Bailes 1975; Zwanzig 1990; Norman et al. 1995; Connors 1996; Lucas et al. 1996). The \(F_{\text{1–2}}\) inferred tectonic transport direction is also in agreement with the interpreted movements along the \(F_{\text{1–2}}\) faults and shear zones (see above).

\(F_{\text{3}}\) straining, however, was too low to account for the rotation of the axes of the late open folds into parallelism with the axes of the earlier isoclinal folds in a constant kinematic frame, since there is generally a direct relationship between fold tightness and fold axis orientation (Mawer and Williams 1991). Hence, the isoclinal \(F_{\text{1–2}}\) folds and the open \(F_{\text{3}}\) folds must have developed in different kinematic frames. The contrasting dip directions of the axial planes and the systematic variation in tightness of \(F_{\text{1–2}}\) and \(F_{\text{3}}\) structures is good evidence that this was the case.

There are three possible ways for the open \(F_{\text{3}}\) folds to have formed, and they all have in common that the orientation of the \(F_{\text{3}}\) axes was controlled by the linear anisotropy developed during the earlier folding (Cobbold and Watkinson 1981; Watkinson and Cobbold 1981). (1) The tectono-stratigraphic packages were deformed merely by approximately east–west shortening, so the deformation path was coaxial. (2) Flow continued to be south to southwest directed during \(F_{\text{3}}\), and a component of approximately east–west shortening was superimposed so that the noncoaxial de-
formation path became constrictional (cf. Meneilly and Storey 1986). (3) The large F₃ folds simply resulted from draping of the foliation around the previous linear anisotropy and around the large plutons during south- to southwest-directed flow. This latter scenario does not involve shortening perpendicular to the axial plane, but it requires that the F₂ and F₃ folds formed simultaneously. We eliminate possibility 3, however, based on the different mineral assemblages and thus the different metamorphic grades associated with F₂ and F₃ (Kraus and Menard 1997; Menard and Gordon 1997). Possibility 2 cannot be ruled out, but it is not likely, as there are no shear-sense indicators for south- to southwest-directed F₃ flow and no L₃ stretching lineation parallel to the F₃ axes. Further, the tectono-stratigraphic packages were already moderately to steeply dipping prior to F₃ (see below), and their attitude therefore inhibited large-scale tectonic transport. The low strains associated with F₃ and the orthorhombic symmetry of the large F₃ folds suggest a bulk coaxial strain path, and thus model 1 is the most realistic one.

**Ages of deformation and metamorphism**

Based on U–Pb geochronology of zircon, monazite, and titanite, the thermal peak of metamorphism in the southern Trans-Hudson Orogen was interpreted as having occurred between 1.820 and 1.805 Ga (e.g., Gordon 1989; Gordon et al. 1990; Machado et al. 1990; Ansdell and Norman 1995, and references therein; Bleeker et al. 1995; David et al. 1996, and references therein). A thermal anomaly, interpreted as the product of prolonged, extensive granitoid intrusion in the Kisseynew Domain, was the source of high-grade metamorphism at moderate pressures in the Kisseynew Domain, its southern flank, and the northern and eastern portions of the Snow Lake Allochthon (Bailes et al. 1976, 1985; Bailes and McRitchie 1978; Gordon 1989; Kraus and Menard 1997; Menard and Gordon 1997). The anomaly, which resulted in rapid, possibly isobaric heating from ~550–600°C to ~750–800°C, developed either late-F₂ or post-F₂, and outlasted F₃ (Kraus and Menard 1997; Menard and Gordon 1997). In contrast, rocks metamorphosed not higher than staurolite grade (Tmax ≤ 580°C) south of Snow Lake were not affected by the thermal anomaly; these rocks reached metamorphic peak conditions early during F₂ and show evidence of cooling during F₃ (Kraus and Menard 1997; Menard and Gordon 1997; Kraus and Williams 1998). Titanite in rhyolite from Anderson Lake (Tmax = ~550°C at 5 kbar) yielded an age of 1812 ± 5 Ma, interpreted as dating the local syn-F₂ peak of metamorphism (Fig. 3) (Zaleski et al. 1991; David et al. 1996). Zircon overgrowths in samples from the Herblet Lake gneiss dome (Fig. 2) (Tmax = 700–800°C at 5–6 kbar; Menard and Gordon 1997) yielded interpreted syn-peak metamorphic ages of 1807 ± 7, 1807 ± 3, and 1803 ± 2 Ma (David et al. 1996). By comparison, monazite ages, interpreted as dating the cooling of ~1815 Ma Kisseynew Domain granitoids, are 1806 ± 2 and 1804 ± 2 Ma (Gordon 1989; Gordon et al. 1990). Titanite grains in gneisses from the southern flank of the Kisseynew Domain at Sherridon near Kississing Lake (Tmax = 660°C at 5 kbar; Froese and Goetz 1981) yielded interpreted cooling ages of 1808 ± 2, 1805 ± 5, and 1804 ± 3 Ma (Fig. 1) (Ashton et al. 1992). In summary, the ~1.805 Ga ages date the onset of cooling in the high-grade rocks and thus give a minimum age for F₃.

Late kinematic to postkinematic pegmatites with interpreted crystallization ages of ~1.805–1.760 Ga are widespread in the high-grade rocks of the southern Trans-Hudson Orogen (e.g., Hunt and Zwanzig 1993; Ansdell and Norman 1995; Parent et al. 1995, 1999; Chiarenzelli et al. 1998). Undated pegmatites north of File Lake (Fig. 2) were interpreted as being broadly coeval with F₃ (Connors 1996). The majority of the ages interpreted as dating the intrusion of the deformed pegmatites at the southern flank of the Kisseynew Domain scatter around 1.790 Ga; some of the older ages

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may reflect inheritance of near-peak metamorphic monazites (Parent et al. 1995, 1999). Although speculative, it is possible that F3 followed F1 within a few million years.

Hornblende and biotite in Missi rocks from near the eastern shore of Wekusko Lake (Fig. 2) yielded minimum cooling ages between 1764 ± 11 and 1747 ± 11 Ma (Marshall et al. 1997), consistent with cooling ages in the Amisk collage (Hunt and Roddick 1992, 1993; Fedorowich et al. 1995). It appears that cooling during uplift was buffered to some extent by hot fluids accompanying the 1.805–1.760 Ga pegmatites. Widespread hydrothermal activity outlasting deformation of the Snow Lake Allochthon was noted by various workers (Marshall et al. 1997; Menard and Gordon 1997; Kraus and Williams 1998). Cooling rates increased rapidly once this activity had ceased (cf. Gordon 1989; Marshall et al. 1997; Leclair et al. 1997).

Structural features between Wekusko Lake and the Setting Lake fault zone

In the Snow Lake Allochthon east of Wekusko Lake (Figs. 2, 5), isoclinal F1 and F2 structures verge northwest- erly and are coplanar with tight to isoclinal F3 structures (Connors et al. 1999). The F3 folds are generally tighter than the F3 folds west of Wekusko Lake (Bailes 1985; Connors et al. 1999). Steep F3 shear zones (e.g., Crowduck Bay “fault”; Fig. 2) (Connors et al. 1999), some of which may be reactivated F1-2 thrusts, trend approximately parallel to the large F1-F2 fold hinges and also parallel to the Setting Lake fault zone (Fig. 1), which is the boundary between the internal and the external Trans-Hudson Orogen (Bailes 1985; Bleeker 1990, and references therein; Connors et al. 1999).

A moderately north-dipping structure (Roberts Lake “fault”; Fig. 2) was interpreted as a late, south-translating thrust (Connors et al. 1999). In the part of the Kisseynew Domain that is sandwiched between the Snow Lake Allochthon and the Thompson Belt (Fig. 1) (hereafter referred to as southeasteren Kisseynew Domain), the axial planes of large, steep S-asymmetrical (?) F4 folds are overturned in the same way as the axial planes of F3 folds within the Thompson belt, indicating a component of sinistral shear with respect to the boundary of the Thompson Belt (Bailes 1985; Bleeker 1990).

Structural correlation of the Snow Lake Allochthon and the Thompson Belt

The thermal peak of metamorphism in the Thompson Belt of the external Trans-Hudson Orogen outlasted the local F2 and was broadly coeval with the thermal peak of metamorphism in the internal zone of the orogen (Bleeker 1990; Machado et al. 1990; Bleeker and Macek 1996). A granite, believed to be associated with the thermal peak of metamorphism, was dated at 1822 ± 3 Ma (zircon), and a paleosome from Sasagiu Rapids (Fig. 1) yielded an age of 1809 ± 14 Ma (zircon) (Machado et al. 1990; Bleeker et al. 1995). F3 folds formed at the inception of widespread pegmatite intrusion and associated retrogressive fluid activity spanning ~1.786–1.765 Ga; the Setting Lake fault zone was active during approximately the same time interval (Lewry 1981; Green et al. 1985; Bleeker 1990; Machado et al. 1990; Machado and David 1992; Bleeker et al. 1995; Bleeker and Macek 1996). It thus appears that the F2 and F3 structures in the Snow Lake Allochthon were broadly coeval with the F2 and F3 structures in the Thompson Belt, respectively.

An evolutionary model for the southeastern Trans-Hudson Orogen

The Snow Lake arc assemblage originated as an outboard accretionary complex during continental convergence at ~1.9 Ga (Sterne et al. 1995; David et al. 1996). Crustal contamination suggests that the arc was built on continental crust (rifted fragment or subsided margin of the Superior plate?) (Sterne et al. 1995). Alternatively, the arc was built on oceanic crust, providing that magma generation involved melting of continental sediments on the subducting slab. Burntwood group turbidites, derived from the eroding arc and from the Archean continental margins, were deposited at 1.86–1.84 Ga on oceanic crust in the marginal Kisseynew basin (Ansdell et al. 1995; Machado and Zwanzig 1995; David et al. 1996). Coevally, Missi group sediments were deposited in fluvial–alluvial fans at the basin margin and on the Amisk collage (Stauffer 1990; Lucas et al. 1996). The following ~1.84–177 Ga Hudsonian orogeny resulted from the final collision of the Amisk collage, Snow Lake Allochthon, and Kisseynew basin with the Archean Sask craton (preserved as windows in the Hanson Lake Block) and Superior craton (Fig. 1) (e.g., Lewry 1981; Bickford et al. 1990; Bleeker 1990; Ansdell et al. 1995; Lucas et al. 1996, 1997). During the final closure of the “Manikewan” ocean, thick, internally deformed, ductile F3 nappes derived from the collapsing Kisseynew basin as a tapering thrust wedge. The sedimentary nappes were emplaced along shallow northerly dipping structures (Snow Lake fault, Berry Creek shear zone) onto and interleaved with arc and ocean floor rocks, causing 12–15 km of crustal thickening (Fig. 13a) (Kraus and Menard 1997, and references therein). The nappe pile was intruded by calc-alkaline plutons after F1 (Gordon et al. 1990; David et al. 1996). Both this intrusive activity and coeval arc volcanism have been interpreted as relating to southward subduction of Kisseynew basin oceanic crust (Gordon et al. 1990; Ansdell and Connors 1995; Ansdell et al. 1995, 1999).

Crustal thickening by folding and overthrusting continued during F2 (Menard and Gordon 1997; Kraus and Menard 1997). The general northeasterly plunge of F1-2 linear features suggests south- to southwest-directed F1-2 tectonic transport (Fig. 13a). We believe that F1-2 structures in the allochthon were initially recumbent and were steepened during F3 by a component of northeast–southwest shortening superimposed on the south- to southwest-directed flow; further steepening was achieved by later F3 refolding around the previous fold axes. The F3 shortening may have resulted from both the topographic relief of the footwall assemblages (Connors 1996) and the inhibition of forward propagation on detachments. For example, the eastern termination of the rigid Amisk collage (footwall to the Morton Lake “fault” zone) may have acted as a steep oblique-lateral ramp (Figs. 2, 13a). Contemporaneous deformation in the Amisk collage, already characterized by steep anisotropy, was con-
centrated along steep northerly trending shear zones (Ryan and Williams 1999).

Following the development of the Snow Lake Allochthon, the Superior plate collided with and shallowly underthrust the internal Trans-Hudson Orogen, probably in a sinistral-oblique sense (Fig. 13b) (e.g., Lewry 1981; Green et al. 1985; Bleeker 1990). During underthrusting, the allochthon and the southeastern Kisseynew Domain were folded by large-amplitude, upright F₃ structures (Fig. 13b), which are symmetrical in the Snow Lake – File Lake area. Uplift of the Snow Lake Allochthon may have commenced at that time. East of Wekusko Lake, F₁–₂ thrusts were steepened and reactivated and new structures formed (e.g., Crowduck Bay fault; Fig. 2). It is, however, uncertain whether the eastward-increasing tightness of the large F₃ folds was achieved by F₃ or F₄ strains, or by both. Contemporaneously, in the Thompson Belt, recumbent F₃ folds developed within a nappe pile that translated passive-margin sediments easterly onto the Superior plate (Bleeker 1990; Bleeker and Macek 1996).

After a possible vergence reversal and a short-lived period of westward overthrusting of the Superior plate onto the internal Trans-Hudson Orogen, the Thompson Belt, now being uplifted, was transformed into a zone of sinistral transpression during the local F₄ (Fig. 13c) (Green et al. 1985; Bleeker 1990; Bleeker and Macek 1996). Transpression caused steepening of the recumbent earlier structures and the development of west-vergent, near-upright, high-amplitude, doubly-plunging F₃ folds in the Thompson Belt (Bleeker 1990; Bleeker and Macek 1996). At the same time, F₄ folds developed in the Snow Lake Allochthon. The area between Wekusko Lake and the Setting Lake fault zone (Fig. 5) was likely affected by the transpressional deformation (Connors and Ansdell 1994a, 1994b; Connors 1996). Here, the steep F₃ structures were tightened, and further dismembered along a series of reactived shear zones (e.g., Crowduck Bay fault; Fig. 2); F₄ movement along the northeasterly trending, steep structures was possibly sinistral strike slip or oblique slip (Connors et al. 1999). Coevally, favourably oriented packages were overthrust southerly along shallowly north-vergent structures (e.g., Roberts Lake fault; Fig. 2). In the Snow Lake – File Lake area, in contrast, the F₁–₂ kinematic frame was approximately restored during F₄. There, only favourably oriented northerly trending domains in highly metamorphosed rocks north of Snow Lake were deformed by large F₄ folds. These rocks were still relatively hot and thus more susceptible to straining than the colder, more rigid rocks to the south. The temperature gradient between the high-grade and the low-grade rocks was maintained during this stage of cooling, possibly owing to syn-F₄ and younger pegmatites, and to associated retrogressive fluids, that affected all high-grade domains of the southern Trans-Hudson Orogen (e.g., Bleeker 1990; Connors 1996; Chiarenzelli et al. 1998). The generally steep attitude of the tectono-stratigraphic units inhibited further south- to southward-directed thrusting in the Snow Lake Allochthon.

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