Structural development of the Snow Lake Allochthon and its role in the evolution of the southeastern Trans-Hudson Orogen in Manitoba, central Canada[1]

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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 (F_1–F_4) that formed in at least three successive kinematic frames over a period of more than 30 Ma are described. Isoclinal to transposed southerly verging F_{1-2} structures are refolded by large, open to tight F_3 folds and, locally, by open to tight F_4 folds. The axes of the F_{1-2} folds are parallel or near parallel to the axes of F_3 folds, owing to progressive reorientation of the F_{1-2} axes during south- to southwest-directed tectonic transport, followed by F_3 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.

Résumé : L'Allochtone de Snow Lake est une zone d'imbrication tectonique de roches sédimentaires d'un bassin marginal inversé (Domaine de Kisseynew) avec des roches d'arc insulaire et océaniques. Il occupe la portion sud-est de la zone interne paléoprotérozoique exposée de l'orogène trans-hudsonien, au Manitoba, Canada, près de la zone externe (zone de collision du Supérieur ou ceinture de Thompson), représentant la limite locale entre l'orogène trans-hudsonien et le craton archéen du Supérieur. L'Allochtone de Snow Lake une fois formé, fut subséquemment déformé et métamorphisé jusqu'a un degré élevé, sous conditions de pression variant de faible à modérée durant l'orogénie trans-hudsonienne, en réponse à la collision des cratons archéens survenue vers ~1,84–1,77 Ga. On décrit quatre générations de plis (F_1–F_4) qui résultent d'au moins trois événements cinématiques successifs échelonnés sur une période de plus de 30 millions d'années. Des structures isoclinales à transposées, avec vergence vers le sud, F_{1-2}, sont replissées par de grands plis ouverts à fermés, F_3, et localement par des plis ouverts à fermés, F_4. Les axes des plis F_{1-2} sont parallèles ou quasi parallèles aux axes des plis F_3, résultant de la réorientation progressive des axes des plis F_{1-2} durant le transport tectonique dirigé vers le sud-ouest, et suivi du replissement F_3 autour de l'anisotropie linéaire antécédente. On présente un modèle tectonique qui harmonise, d'une part les développements tectonométamorphiques dans l'Allochtone de Snow Lake avec la portion adjacente du Domaine de Kisseynew, et d'autre part avec la ceinture de Thompson durant le stade terminal de la collision de l'orogène trans-huronien avec le craton du Supérieur.

[Traduit par la Rédaction]

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

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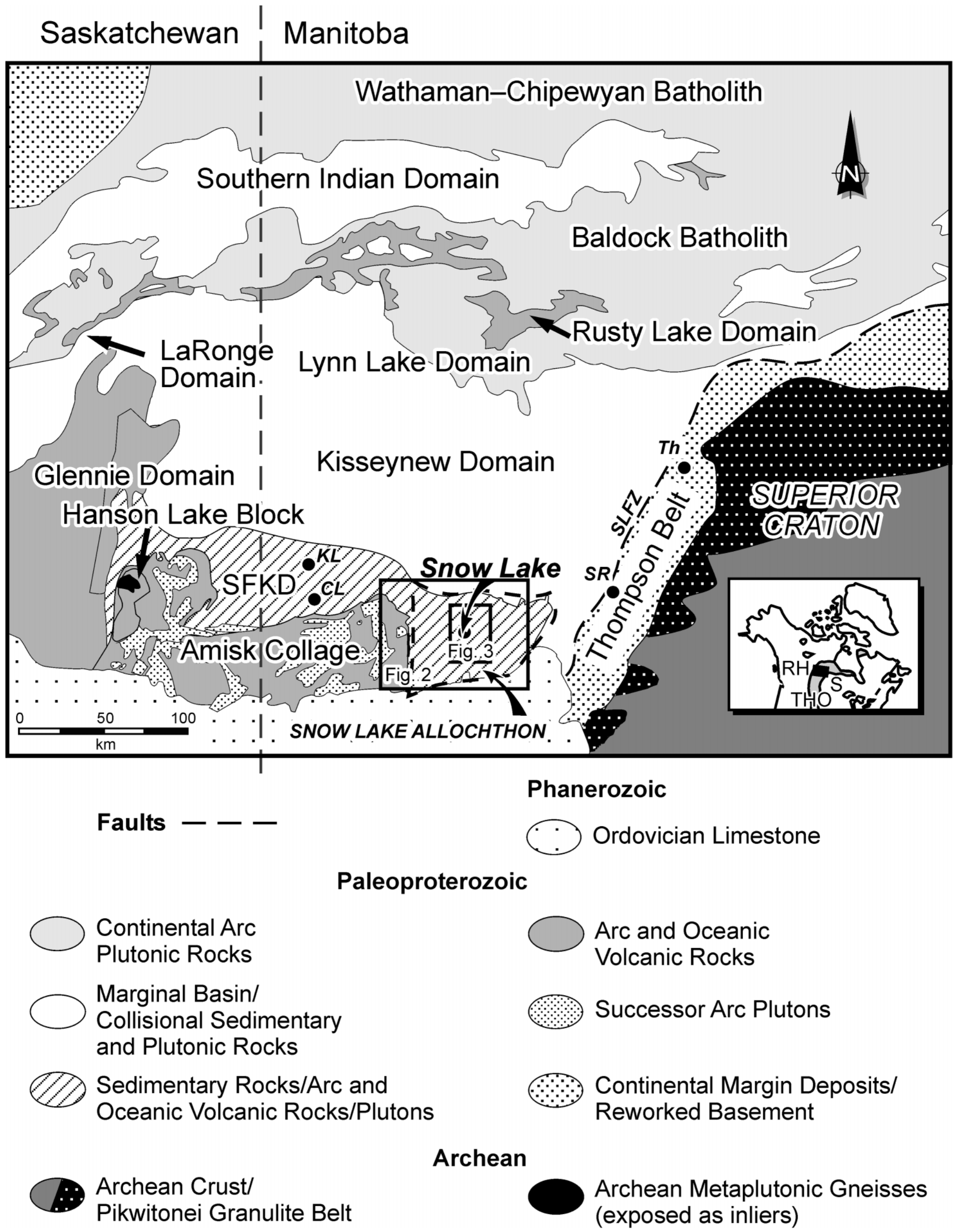
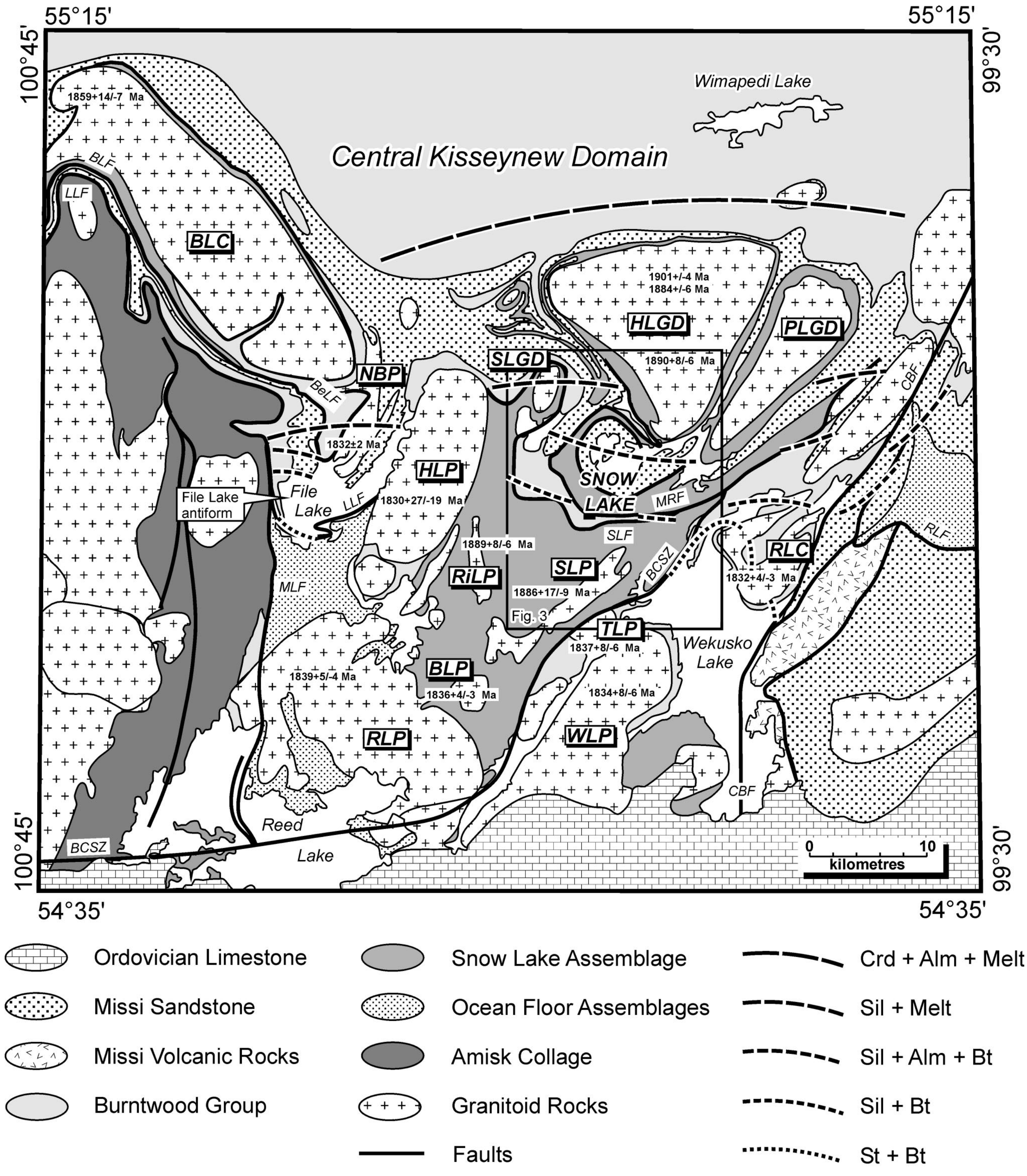
Fig. 1. Lithotectonic domains of part of the Trans-Hudson Orogen and location of the Snow Lake area at the southern margin of the Kisseynew Domain (after Hoffman 1988). CL, Cleunion Lake; KL, Kississing Lake; SFKD, southern flank of Kisseynew Domain; SLFZ Setting Lake fault zone; SR, Sasagiu Rapids; Th, Thompson. Inset map: RH, Rae–Hearne; S, Superior craton; THO, Trans-Hudson Orogen.



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; Leclair et al. 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 Lithoprobe 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 volcaniclastic 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 S_1 fabric and hence they predate, or are coeval with, F_1 compressional deformation (Table 1). Voluminous calc-alkaline 1.84–1.83 Ga plutons truncate F_1 structures (Figs. 2, 3, 4*a*) (Gordon et al. 1990; Bailes 1992; Connors 1996; David et al. 1996). Their intrusion ages thus give an upper age limit for F_1. 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; Ansdell et al. 1999). Large volumes of late kinematic 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; Zwanzig and Schledewitz 1992; Kraus and Williams 1994; Norman et al. 1995; Connors 1996).

Metamorphic zones

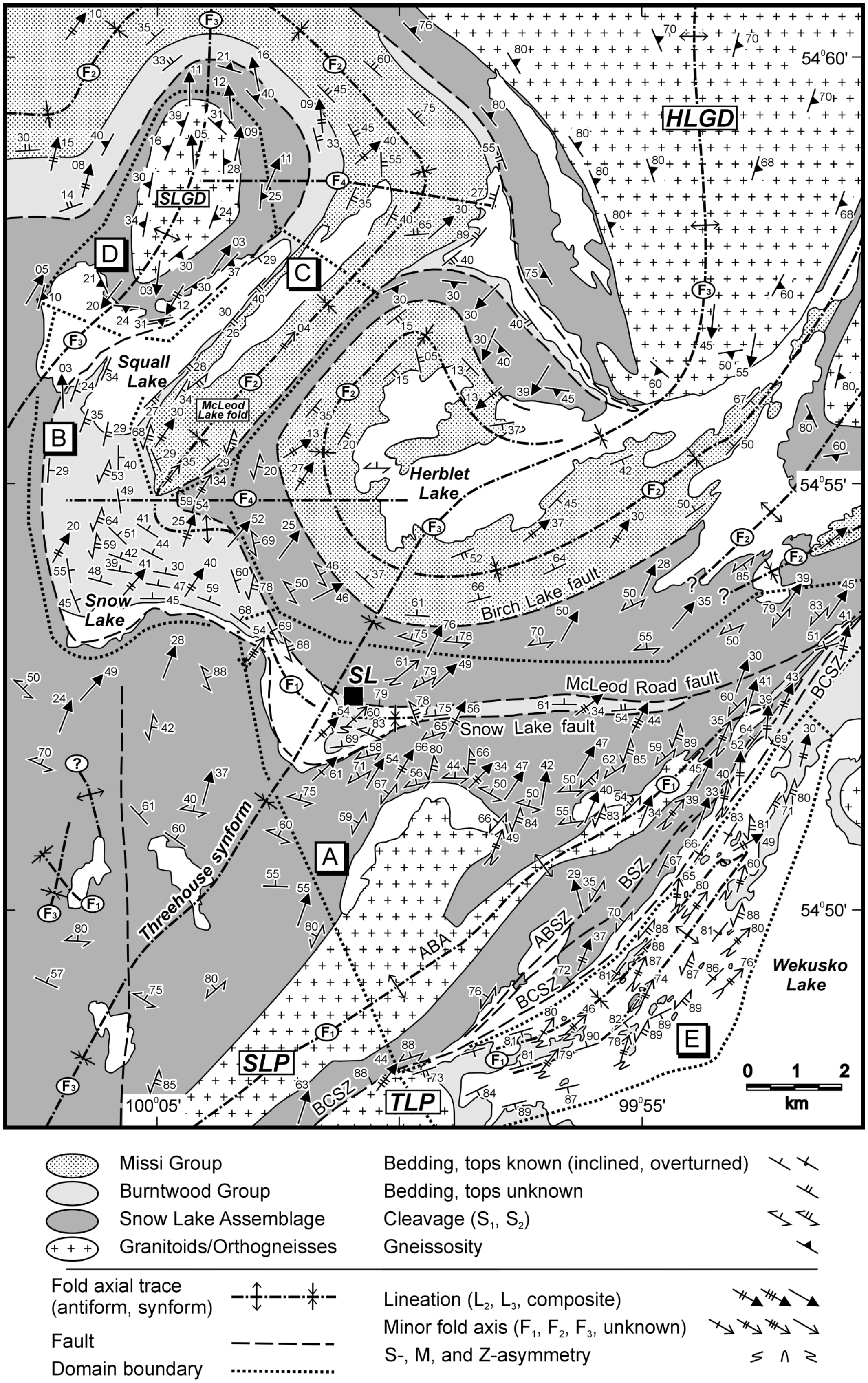
Froese and Gasparrini (1975) divided the Snow Lake – File Lake area into metamorphic zones that are separated by reaction isograds (Fig. 2). These zones were extrapolated east towards the Superior collision zone (Thompson Belt) (Fig. 5) (see references in Kraus and Menard 1997). Temperatures at the thermal 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_4 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 (Zwanzig 1990, 1999; Zwanzig 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., Ansdell 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., Ansdell 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.



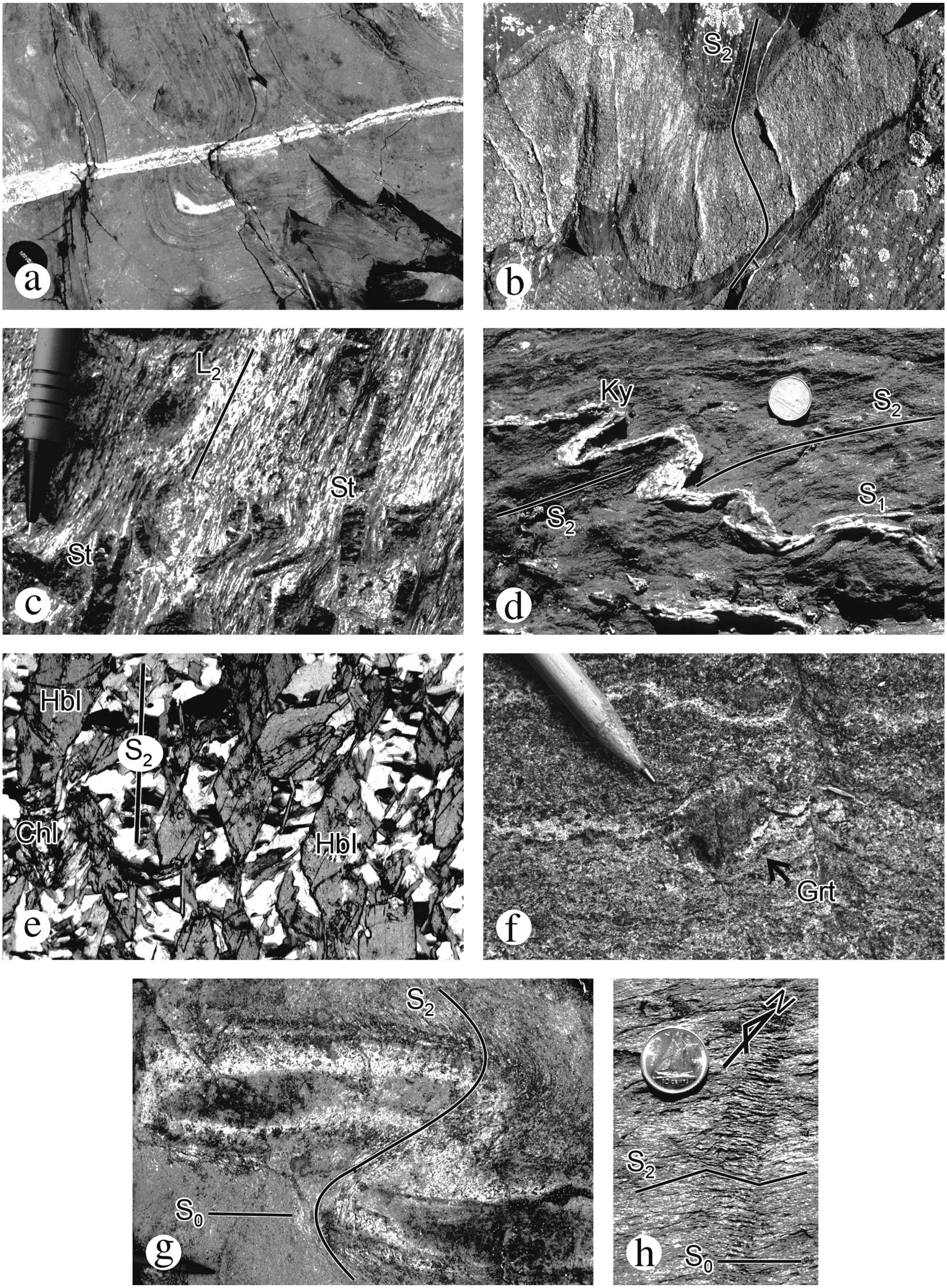

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Fig. 4. (*a*) F_1 fold in Burntwood group crosscut by undeformed, local apophysis of the Wekusko Lake granite. (*b*) Z-asymmetric F_2 fold with refracted axial-plane S_2; Burntwood group in domain E. Note the S_2-parallel quartz veins that developed in response to S_2-parallel slip (see Kraus and Williams 1998). Pencil tip in the upper right for scale. (*c*) L_2 crenulation lineation on S_0 tightened around grains of staurolite (St); Burntwood group in domain B. (*d*) S_1 overgrown by train of kyanite (Ky) and folded by a Z-asymmetrical F_2; hydrothermally altered pillow basalt in domain A. An axial-plane S_2 is well developed. (*e*) Photomicrograph of basal sections of ferroan-pargasitic hornblende porphyroblasts (determined with microprobe) aligned parallel to S_2 in mafic volcanic rock, overprinting the F_1 Berry Creek shear zone on northwestern Wekusko Lake. View is down the northeasterly plunge of the hornblende c axes, which are parallel to the composite $L_{1–2}$ (see Fig. 7*a*). Long edge of photomicrograph is 3.7 mm. Chl, chlorite; Hbl, hornblende. (*f*) Gneissosity enveloping garnet (Grt) in highly metamorphosed felsic volcanic rock; domain D. (*g*) Zoned calc-silicate boudin and crosscutting S_2, both overprinted by Z-asymmetrical F_3 fold; Burntwood group in domain A. Note the opposite asymmetry of S_2 and F_3 with respect to S_0. Pencil in the lower left for scale. (*h*) F_4 kink band overprinting S_2; Burntwood group in domain E.

Allochthon to deformation events in Table 1, however only for the purpose of correlation with adjacent areas.

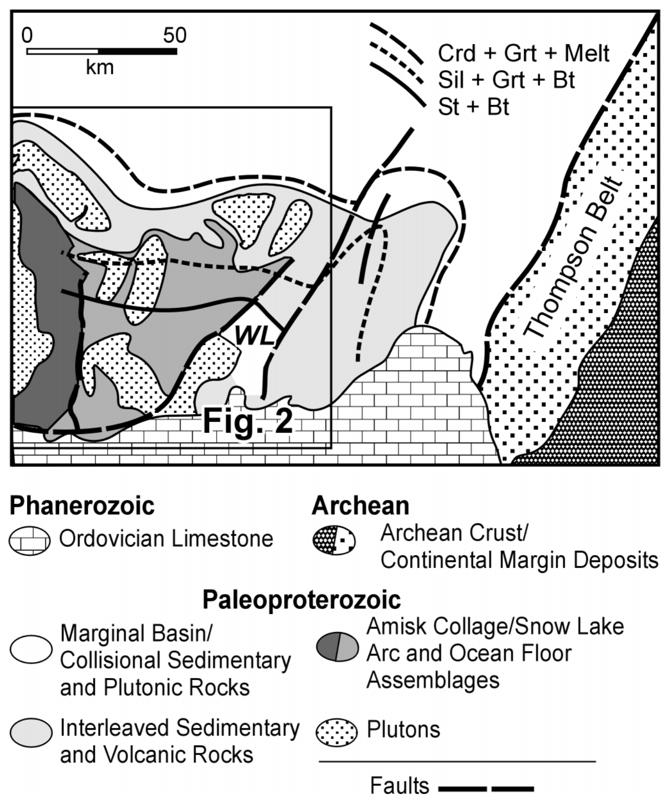
F_1–F_4 structures and metamorphic fabrics

F_1

The Burntwood group rocks contained an S_0-parallel S_1 that was generally destroyed during the formation of an S_2 (Kraus and Williams 1998). Where locally present, S_1 is defined, depending on metamorphic grade, by chlorite or muscovite aligned parallel to bedding (Kraus and Menard 1997; Kraus and Williams 1998). S_1 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 L_1 is only preserved in syn-F_1 bedding-parallel quartz veins; it is defined by stretched quartz aggregates, which plunge approximately parallel to minor fold axes. Three orders of F_1 folds appear in the Burntwood group, with amplitudes of centimetres to metres, tens to hundreds of metres, and kilometres (Fig. 3). Small F_1 folds are rare. When several small F_1 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 F_1 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 F_1 folding. Z- and S-asymmetrical folds and, locally, doubly verging fold pairs (sensu Holdsworth and Roberts 1984) occur on both limbs of larger F_1 structures. All larger folds were also dismembered and hence have not preserved their asymmetry. F_1 features were not identified in the strongly recrystallized Missi group rocks.

In hydrothermally altered volcanic rocks that are now semipelitic chloritic schists, S_1 constitutes a fabric parallel to primary layering (where no S_2 is developed) and is defined by aligned chlorite. S_1 is also contained as straight inclusion trails in kyanite, staurolite, garnet, and biotite. These porphyroblasts are, however, not indicative of metamorphic grade (Zaleski et al. 1991). In unaltered volcanic rocks, an S_1 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 F_1 folds are rare in the volcanic rocks and are developed mainly along lithological contacts. Younging criteria and consistently dextral S_0/S_2 and S_1/S_2 asymmetry (see below) suggest that the volcanistratigraphic sequence is possibly repeated between the Snow Lake "fault" and the Berry Creek shear zone across a crustal-scale F_1 anticlinal structure, termed the Anderson Bay anticline (Fig. 3) (Walford and Franklin 1982). The sequence appears to be structurally thinned on the lower limb of this structure near the Berry Creek shear zone. In all rocks, the pre- to syn-F_1 mafic dykes–sills and the primary layering are boudinaged in two directions, parallel and at a high angle to the lineation (see later in this paper), yielding a chocolate-block pattern. S_1 is deflected around the boudins. S_2, where present, overprints the boudins.

Fig. 5. Metamorphic isograds of the Snow Lake Allochthon (after Gordon 1989). Mineral abbreviations after Kretz (1983). WL, Wekusko Lake.



F_1 shear zones

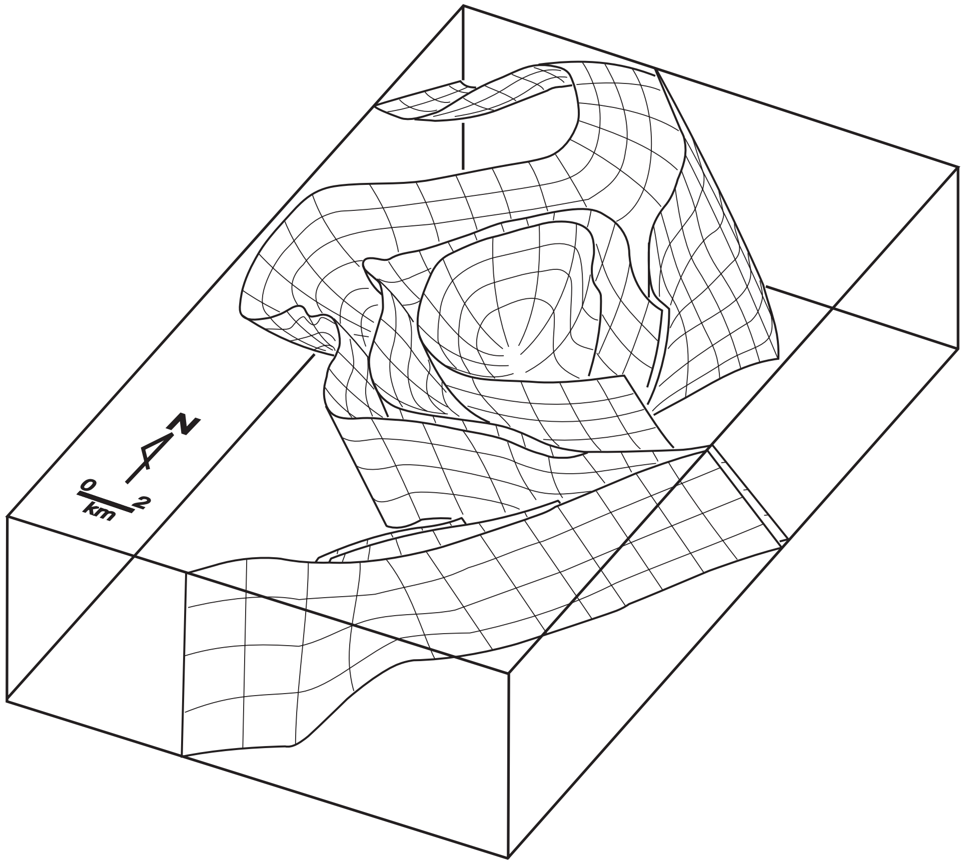
The Berry Creek shear zone and the Snow Lake – Loonhead Lake (ductile) "fault," both of which juxtapose metasedimentary rocks and volcanic rocks (Figs. 2, 3, 6), originated as subhorizontal, low-angle (with respect to primary layering) reverse structures during F_1.

Northeastwards, along the northwestern margin of Wekusko Lake (north of the Tramping Lake pluton; Figs. 2, 3), the Berry Creek shear zone becomes part of an

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

Deformation		Structures	Magmatic events		Metamorphism	Age (Ga)	Tectonic setting
		—	Arc volcanism; synvolcanic intrusions (*Sneath Lake tonalite, Herblet Lake and Squall Lake granitoids*)	Hydrothermal activity	Synvolcanic fluid alteration	1.892–1.886	Subduction of Archean crust
D_1 & D_2		—				1.860–1.840	Sedimentation of Burntwood and Missi groups
D_3	Hudsonian orogeny: F_1	Isoclinal folds and low-angle shear zones; S_1 parallel to primary layering	Mafic dykes and sills Calc-alkaline granitoid suite (e.g., *Wekusko Lake pluton*); Missi volcanic suite		Prograde metamorphism: chl grade, $T < 400°C$, $P = 4$ kbar	1.840–1.830	Overthrusting of Kisseynew Domain over Snow Lake Allocation during N–S convergence; crustal thickening: 12–15 km
D_3	F_2	Isoclinal curvilinear folds; regional S_2; low-angle shear zones and reactivation of F_1 shear zones			Thermal peak in st zone and lower: $T < 580°C$, $P = 4–6$ kbar	1.820–1.810	Continued N–S convergence; minor crustal thickening
D_4	F_3	NNE-trending open to tight folds (*Threehouse synform*); crenulations of S_2			Cooling in st zone; thermal peak in sil zone and higher: $T = 600–800°C$, $P = 5–6$ kbar	1.810–1.805	E–W shortening: W-directed underthrusting of Superior plate
D_5	F_4	Local WNW–ESE-trending folds; kink bands	Pegmatites in sil zone and higher			1.800–1.765	Renewed N–S convergence; exhumation during orogen-parallel movements
		Brittle features			Blocking temperatures of hbl and bt reached	1.765–1.745	Ongoing exhumation

Fig. 6. Block diagram of the area depicted in Fig. 3 showing the ductile "faults" and shear zones. These structures are generally parallel to or at a low angle to primary layering, so they are representative reference surfaces.



anastomosing system, which includes the poorly exposed Anderson Bay shear zone and Bartlett shear zone (Figs. 3, 6). F_1 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. 7*a*) 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).

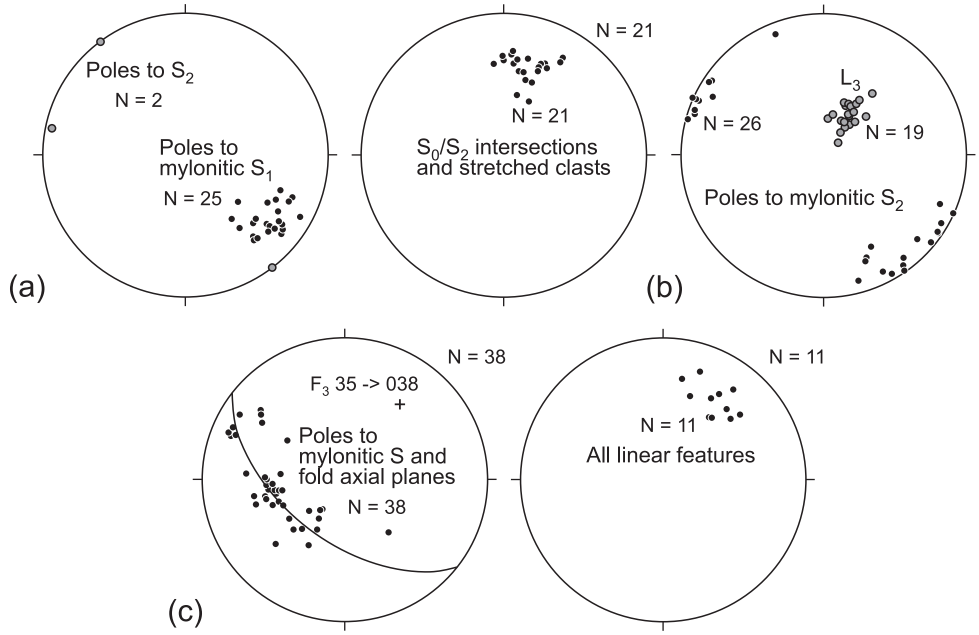
The Snow Lake fault constitutes the southern termination of the isoclinally folded Burntwood group sliver at Snow Lake (Figs. 2, 3, 6) (see Kraus and Williams 1998). It extends as the Loonhead Lake fault to the west through the File Lake area, where it appears to form the boundary between the Amisk collage 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-F_1 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 F_1 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 F_1 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).

F_2

An S_2 crenulation cleavage or schistosity, which envelops porphyroblasts, is ubiquitous in the metasedimentary rocks and in the hydrothermally altered volcanic rocks (Figs. 4*b*–4*d*) (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 F_2 (Kraus and Menard 1997; Menard and Gordon 1997; Kraus and Williams 1998). Depending on metamorphic grade, the S_2 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 S_2. 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 S_0/S_2 lineation. Locally, the porphyroblasts

Fig. 7. Equal-area projections (lower hemisphere, Schmidt net) of structural data for major shear zones (Fig. 3). (*a*, *b*) Berry Creek shear zone, Northwest Wekusko Lake segment (*a*) and Tramping Lake pluton segment (*b*). (*c*) McLeod Road fault zone in **domain B**.



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. 4*b*). 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. 4*d*).

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. 4*f*). 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. 4*e*), 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. 4*b*, 4*d*). 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.

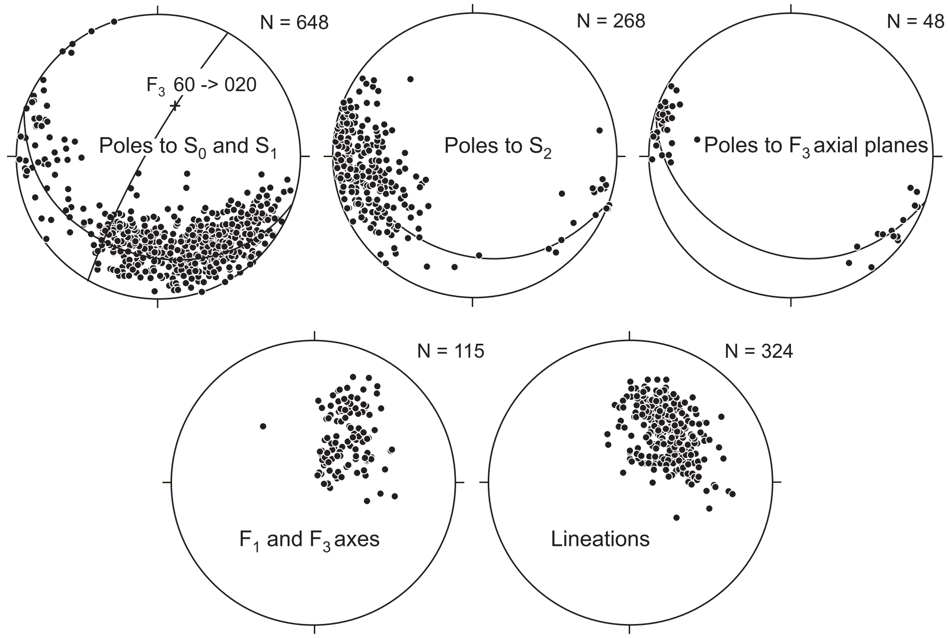
F_2 shear zones

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_3 and F_4 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. 7*c*). 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 sheath 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. 7*c*), 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. 7*c*) with the F_1–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^{+8}_{-6} Ma (David et al. 1996), post-F_1 Tramping Lake pluton (Figs. 2, 3, 7*b*). 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

Fig. 8. Equal-area projections (lower hemisphere, Schmidt net) of structural data for **domain A** (Fig. 3).



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 S_2 mylonitic foliation at the pluton margin are crenulated by F_3 and contain a faint, steep northeasterly plunging L_3 crenulation lineation (Fig. 7*b*). 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 F_2. It is overprinted by the regional S_2 that envelops peak metamorphic porphyroblasts containing S_1, the L_2 crenulation lineation, small Z-asymmetrical F_2 folds, and small symmetrical F_3 folds (Figs. 4*e*, 7*a*). L_2 plunges parallel to the clasts stretched during F_1 (Fig. 7*a*). 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 F_2 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 F_3, causing its trace to be sinusoidal (Fig. 3).

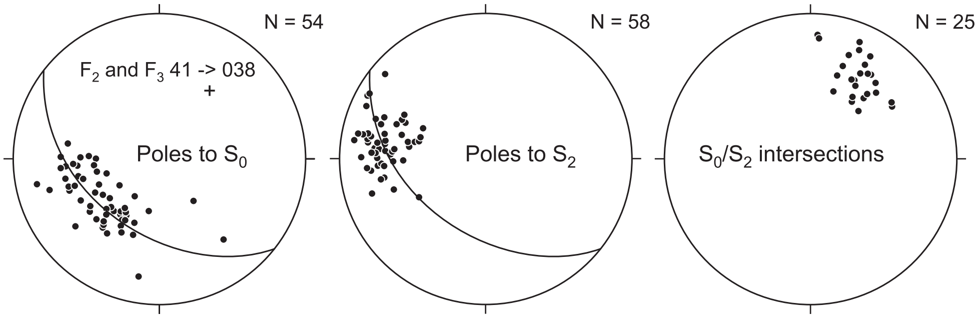
Evidence of F_2 movement along the Snow Lake fault is given by a reversal in S_0/S_2 asymmetry across the structure from sinistral in the northern hanging wall (Kraus and Williams 1998) to dextral in the southern footwall; younging is to the north on both sides of the structure. The dextral S_0/S_2 asymmetry and a corresponding Z-asymmetry of minor F_2 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 F_2 fold.

F_3 and F_4

F_3 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 F_3 folds a few hundreds of metres in scale that are generally discontinuous along their axial planes. Rare minor F_3 folds are open to tight and overprint S_2, trains of peak metamorphic porphyroblasts, and their F_2 pressure shadows (Fig. 4*g*). F_3 crenulations of S_2 are developed locally around the Threehouse synform, where S_2 is at low angles to bedding; these crenulations deform S_2-parallel quartz veins and pressure shadows on staurolite and biotite (Kraus and Williams 1998). Pervasive S_3 is only developed in low-grade rocks of the Burntwood group at Reed Lake (Syme et al. 1995) (Fig. 2).

Approximately east–west-trending F_4 folds overprint F_3 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 F_4 folding. F_4 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 F_4 structures, but Connors (1996) report a local S_4 in F_4 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

Fig. 9. Equal-area projections (lower hemisphere, Schmidt net) of structural data for **domain B** (Fig. 3).



steep S_2, are kinked. Similarly, conjugate kink bands in north-northeasterly-trending steep S_2 occur on the islands of northwest Wekusko Lake (Fig. 4*h*).

Structural domains

Five structural domains (A–E) are distinguished in the study area based on the orientation of F_1–F_3 structural elements (Fig. 3). Domain A (Fig. 3) essentially coincides with the staurolite + biotite zone. It is characterized by coaxial F_1–F_3 structures and coplanar F_1 and F_2 structures (Fig. 8). F_1–F_3 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 F_3 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 F_3 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 F_2 McLeod Lake fold and the McLeod Road fault. Both structures are overprinted by F_3 and F_4 (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 F_4 refolding of the axial plane of the F_2 fold about a steep axis. The pole to this girdle is parallel to the local F_2 and F_3 axes and coincides with the southwesterly transport direction along the McLeod Road fault (see earlier in this paper; Fig. 7*c*).

Domain C (Fig. 3) contains a segment of the McLeod Lake fold in Missi group metasandstones that young towards the axial plane. In the hinge area along the Missi–Burntwood boundary, poles to S_0 yield a moderately northeasterly plunging major F_2 axis (southern part of domain C in Fig. 10*a*). The fold hinge is approximately horizontal in the northern part of domain C (Fig. 10*a*).

Domain D (Fig. 3) hosts the Squall Lake gneiss dome, a culmination within an F_3–F_4 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 F_3 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 F_3 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

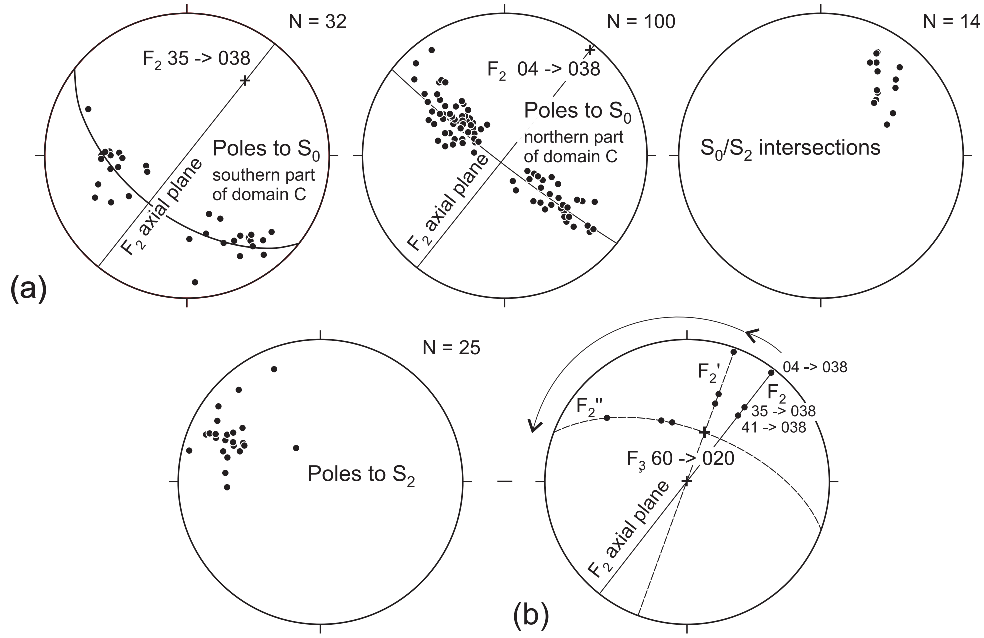
Although the hinge of the F_2 McLeod Lake fold is dismembered by the McLeod Road fault in parts of domains B and in domain A, we can estimate the hypothetical local orientation of the major F_2 axis by assuming parallelism of the hinge with the S_0/S_2 lineation. This information is important for the 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°. In the northern part of domain C, the hinge is subhorizontal and becomes progressively steeper northeasterly, plunging in the southern part (Fig. 10*a*) and through domains B and A (Figs. 8, 9).

In an attempt to reconstruct the original geometry and attitude of the F_2 McLeod Lake fold, we restore its orientation prior to F_3 and F_4 refolding. We assume an initial south to southwest overturning, as reported for large F_2 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. 10*b*. After unfolding, the F_2 axial plane dips moderately to the north-northwest, and the local hinge segments plunge moderately west-northwest to steeply north-northeast (Fig. 10*b*). Similar orientations of F_2 axial planes and F_2 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 F_2 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 F_1 and F_2 folds are broadly parallel to the axes of the open F_3 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

Fig. 10. (*a*) Equal-area projections (lower hemisphere, Schmidt net) of structural data for **domain C** (Fig. 3). (*b*) Schematic unfolding of the McLeod Lake fold. The local fold hinge segments for **domains B and C** (taken from Figs. 9, 10*a*) plot on a steep northeasterly trending great circle that coincides with the axial plane of the McLeod Lake fold in **domain C**. Variation in plunge angles of the individual hinge segments reveals the curvilinear character of the structure. The effects of F_3 and F_4 are removed by rotating the axial plane of the McLeod Lake fold first 18° anticlockwise around a steep F_4 axis into position F_2' (in which the axial plane contains the F_3 Threehouse synform axis), and subsequently 90° anticlockwise around the F_3 Threehouse axis into position F_2''.



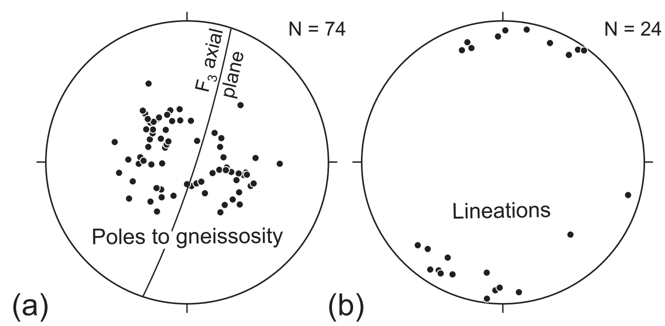
and Knill 1958; Bryant and Reed 1969; Escher and Watterson 1974; Skjernaa 1980; Meneilly and Storey 1986; Norman et al. 1995). In those cases, parallelism of the early fold axes was attributed to progressive rotation of the fold hinges towards the direction of tectonic transport in a constant kinematic frame, associated with high shear strains (γ > ~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_{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_{1-2} fold hinges.

Although the orientation of the minor F_{1-2} fold axes alone is far from conclusive evidence for the direction of F_{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_{1-2} inferred tectonic transport direction is also in agreement with the interpreted movements along the F_{1-2} faults and shear zones (see above).

F_3 straining, however, was too low to account for the rotation of the axes of the late open folds into parallelism with

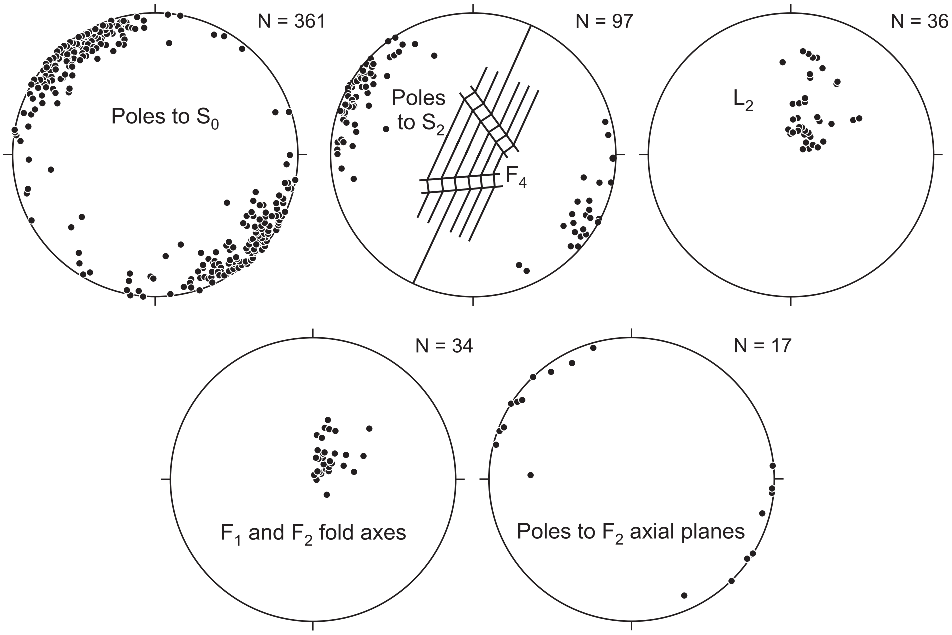
Fig. 11. Equal-area projections (lower hemisphere, Schmidt net) of structural data for **domain D** (Fig. 3).



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_{1-2} folds and the open F_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_{1-2} and F_3 structures is good evidence that this was the case.

There are three possible ways for the open F_3 folds to have formed, and they all have in common that the orientation of the F_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_3, and a component of approximately east–west shortening was superimposed so that the noncoaxial de-

Fig. 12. Equal-area projections (lower hemisphere, Schmidt net) of structural data for **domain E** (Fig. 3).



formation path became constrictional (cf. Meneilly and Storey 1986). (3) The large F_3 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 F_2 and F_3 folds formed simultaneously. We eliminate possibility 3, however, based on the different mineral assemblages and thus the different metamorphic grades associated with F_2 and F_3 (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 F_3 flow and no L_3 stretching lineation parallel to the F_3 axes. Further, the tectono-stratigraphic packages were already moderately to steeply dipping prior to F_3 (see below), and their attitude therefore inhibited large-scale tectonic transport. The low strains associated with F_3 and the orthorhombic symmetry of the large F_3 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 1975, 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-F_2 or post-F_2, and outlasted F_3 (Kraus and Menard 1997; Menard and Gordon 1997). In contrast, rocks metamorphosed not higher than staurolite grade (T_{max} ≤ 580°C) south of Snow Lake were not affected by the thermal anomaly; these rocks reached metamorphic peak conditions early during F_2 and show evidence of cooling during F_3 (Kraus and Menard 1997; Menard and Gordon 1997; Kraus and Williams 1998).

Titanite in rhyolite from Anderson Lake (T_{max} = ~550°C at 5 kbar) yielded an age of 1812 ± 5 Ma, interpreted as dating the local syn-F_2 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) (T_{max} = 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 (T_{max} = 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 F_3.

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 F_4 (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 may reflect inheritance of near-peak metamorphic monazites (Parent et al. 1995, 1999). Although speculative, it is possible that F_4 followed F_3 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 F_1 and F_2 structures verge northwesterly and are coplanar with tight to isoclinal F_3 structures (Connors et al. 1999). The F_3 folds are generally tighter than the F_3 folds west of Wekusko Lake (Bailes 1985; Connors et al. 1999). Steep F_3 shear zones (e.g., Crowduck Bay "fault"; Fig. 2) (Connors et al. 1999), some of which may be reactivated $F_{1–2}$ thrusts, trend approximately parallel to the large F_1–F_3 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 southeastern Kisseynew Domain), the axial planes of large, steep S-asymmetrical (? F_4) folds are overturned in the same way as the axial planes of F_3 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

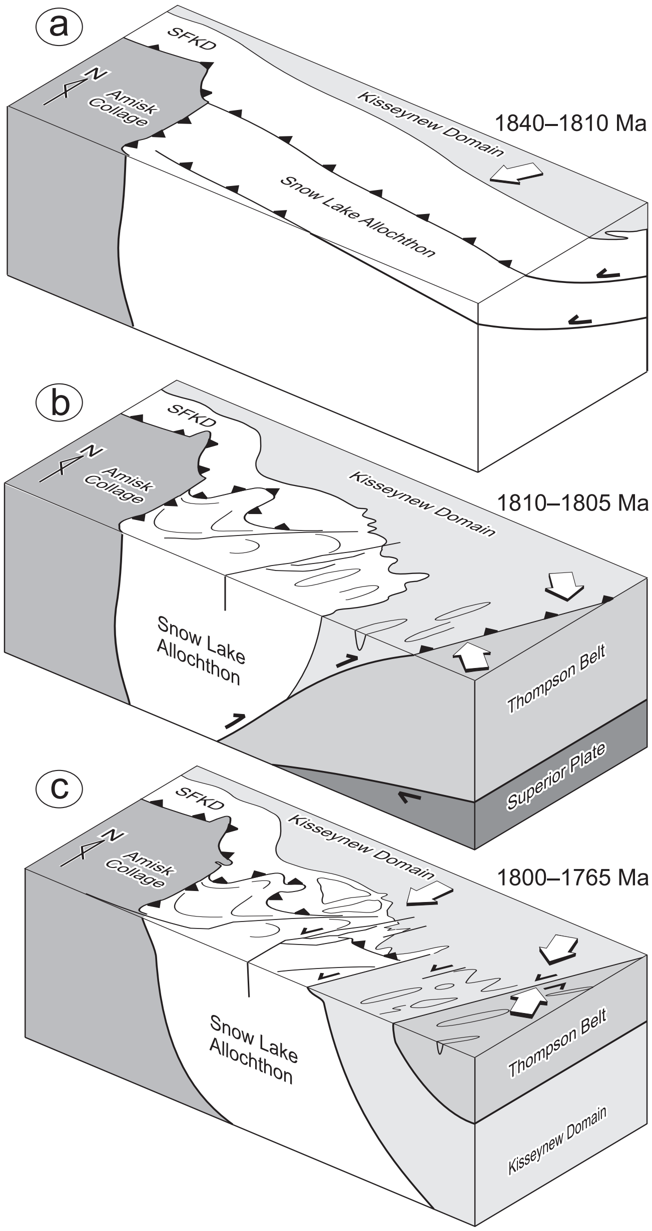
The thermal peak of metamorphism in the Thompson Belt of the external Trans-Hudson Orogen outlasted the local F_2 and was broadly coeval with the thermal peak of metamorphism in the internal zone of the orogen (Bleeker 1990; Machado et al. 1990; Bleeker and Macek 1996). A granite, believed to be associated 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). F_3 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 F_3 and F_4 structures in the Snow Lake Allochthon were broadly coeval with the F_2 and F_3 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 (Stern 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?) (Stern 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 F_1 nappes derived from the collapsing Kisseynew basin as a tapering thrust wedge. The sedimentary nappes were emplaced along shallowly 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. 13*a*) (Kraus and Menard 1997, and references therein). The nappe pile was intruded by calc-alkaline plutons after F_1 (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 F_2 (Menard and Gordon 1997; Kraus and Menard 1997). The general northeasterly plunge of $F_{1–2}$ linear features suggests south- to southwest-directed $F_{1–2}$ tectonic transport (Fig. 13*a*). We believe that $F_{1–2}$ structures in the allochthon were initially recumbent and were steepened during F_2 by a component of northeast–southwest shortening superimposed on the south- to southwest-directed flow; further steepening was achieved by later F_3 refolding around the previous fold axes. The F_2 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, 13*a*). Contemporaneous deformation in the Amisk collage, already characterized by steep anisotropy, was con-

Fig. 13. Sequence of block diagrams that schematically depict the tectonic development of the southeastern Trans-Hudson Orogen. See text for explanation. SFKD, southern flank of Kisseynew Domain.



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. 13*b*) (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 F_3 structures (Fig. 13*b*), 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, F_{1-2} 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 F_3 folds was achieved by F_3 or F_4 strains, or by both. Contemporaneously, in the Thompson Belt, recumbent F_2 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 F_3 (Fig. 13*c*) (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 F_3 folds in the Thompson Belt (Bleeker 1990; Bleeker and Macek 1996). At the same time, F_4 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 1994*a*, 1994*b*; Connors 1996). Here, the steep F_3 structures were tightened, and further dismembered along a series of reactivated shear zones (e.g., Crowduck Bay fault; Fig. 2); F_4 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 northerly dipping structures (e.g., Roberts Lake fault; Fig. 2). In the Snow Lake – File Lake area, in contrast, the F_{1-2} kinematic frame was approximately restored during F_4. There, only favourably oriented northerly trending domains in highly metamorphosed rocks north of Snow Lake were deformed by large F_4 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-F_4 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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