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A THERMAL GRADIENT AT CONSTANT PRESSURE: IMPLICATIONS FOR LOW- TO MEDIUM-PRESSURE METAMORPHISM IN A COMPRESSIONAL TECTONIC SETTING, FLIN FLON AND KISSEYNEW DOMAINS, TRANS-HUDSON OROGEN, CENTRAL CANADA*

JÜRGEN KRAUS[1]

Department of Geology, University of New Brunswick, P.O. Box 4400, Fredericton, New Brunswick E3B 5A3

THOMAS MENARD[1]

Department of Geology and Geophysics, University of Calgary, Calgary, Alberta T2N 1N4

Abstract

The Kisseynew Domain, a major tectonostratigraphic domain in the internal Trans-Hudson Orogen of central Canada, originated as a sedimentary basin filled with turbidite of the Burntwood Suite at 1.86–1.84 Ga. Rocks of the Kisseynew Domain were thrust southwestward over, and structurally interleaved with, the 1.9-Ga Snow Lake island arc and ocean-floor assemblage during convergence of the Trans-Hudson Orogen and the Archean Superior craton at ~1.84–1.81 Ga. At Snow Lake, the zone of tectonic interleaving records polyphase deformation (F_1–F_4) and is characterized by a northerly increase in peak temperatures of metamorphism at 4–6 kbar pressure. The central Kisseynew Domain was metamorphosed at uniform high grade of 750 ± 50°C and 5–6 kbar. Pressure and temperature calculations on 13 representative samples from the Snow Lake area in a critical temperature window of 500–700°C (staurolite and sillimanite zones) yield evidence for two successive thermal regimes that varied in time and intensity from south to north. F_1 (1.842–1.835 Ga) was the major burial event, and may have been followed by thermal relaxation in a 15–35 Ma time interval between F_1 and F_2. Garnet commenced growing during early F_2 (1.82–1.805 Ga) at ~500°C and ~4 kbar throughout the study area. During F_2, temperature and pressure in the staurolite zone increased by ~50°C and 1–1.5 kbar to peak conditions. In the staurolite zone, cooling had commenced by the time of F_3, as indicated by chlorite-grade F_3 assemblages. In the sillimanite zone, however, metamorphism was related to a thermal event in the Kisseynew Domain. Here, peak conditions continued until after F_3 (duration of more than 10 Ma), on the basis of isograds and isotherms cross-cutting large F_3 structures. We suggest that low- to medium-pressure, high-temperature metamorphism in the Kisseynew Domain resulted from heat advected by sheets of ~1.815 Ga peraluminous granitic rocks. The solidi of the granitic rocks buffered the peak conditions of metamorphism. A possible cause for the mobilization of granitic magma in the lower crust of the Kisseynew Domain was high basal heat-flow resulting from convective thinning of the lithosphere during thickening of the crust.

Keywords: Trans-Hudson Orogen, Kisseynew Domain, Flin Flon Domain, metamorphism, metamorphic zones, geothermobarometry, P–T–t path, thermal anomaly, granite intrusions, heat advection, polyphase deformation, Snow Lake, Manitoba.

Sommaire

Le domaine de Kisseynew, domaine tectonostratigraphique majeur dans la zone interne de la ceinture orogénique trans-hudsonienne, au Manitoba, a pris naissance comme séquence de turbidites (suite de Burntwood) dans un bassin sédimentaire à 1.86–1.84 Ga. Ces roches ont été chevauchées vers le sud-ouest et imbriquées avec la séquence d'arc insulaire et de roches océaniques du lac Snow (1.9 Ga) au cours de la collision impliquant la ceinture trans-hudsonienne et le socle archéen de la Province du Supérieur, à 1.84–1.81 Ga. Au lac Snow, la zone d'imbrication témoigne d'une déformation polyphasée (F_1–F_4) et d'une augmentation vers le nord de la température maximale de métamorphisme à une pression de 4 à 6 kbar. Le domaine de Kisseynew central a été métamorphisé à un degré uniformément élevé, 750 ± 50°C et 5–6 kbar. La pression et la température calculées pour treize échantillons représentatifs de la région du lac Snow, dans un intervalle critique de 500–700°C (zones à staurolite et à sillimanite), révèlent la présence de deux régimes thermiques successifs, variables dans leur intensité et dans leur durée du sud vers le nord. La déformation F_1 (1.842–1.835 Ga) correspond à l'enfouissement majeur, et a peut-être été suivi d'une décontraction thermique dans un intervalle 15–35 Ma entre F_1 et F_2. Le grenat est apparu au stade précoce de F_2 (1.82–1.805 Ga) à environ 500°C et 4 kbar partout dans la région étudiée. Au cours de F_2, la température et la pression de la zone de la staurolite ont augmenté d'environ 50°C et 1–1.5 kbar, jusqu'aux conditions maximales. Dans la zone à staurolite,

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[1] *E-mail addresses*: kraus@unb.ca, tmenard@geo.ucalgary.ca

un refroidissement a débuté avec l'avènement de F_3, comme le témoigne le développement d'assemblages F_3 à chlorite. Dans la zone à sillimanite, cependant, le métamorphisme est attribuable à un événement thermique dans le domaine de Kisseynew. Là, les conditions maximales ont continué pour plus de 10 Ma, après la fin de F_3, comme le révèlent les isogrades et les isothermes qui recoupent les structures majeures de F_3. A notre avis, le métamorphisme de faible à moyenne pression et de température élevée du domaine de Kisseynew est le résultat de l'advection de chaleur associée à la mise en place de masses stratiformes de roches granitiques peralumineuses à environ 1.815 Ga. Le solidus de ces magmas a tamponné les conditions maximales du métamorphisme. La mobilisation de magma granitique dans la croûte inférieure serait l'expression d'un flux de chaleur dû à l'amincissement de la lithosphère au cours de l'épaississement de la croûte.

(Traduit par la Rédaction)

Mots-clés: ceinture orogénique trans-hudsonienne, domaine de Kisseynew, domaine de Flin Flon, métamorphisme, zones métamorphiques, géothermobarométrie, tracé P–T–t, anomalie thermique, intrusions de granite, advection de chaleur, déformation polyphasée, lac Snow, Manitoba.

INTRODUCTION

Low-pressure, high-temperature metamorphism is characteristic of extensional tectonic settings, such as the Pyrenean massifs in France (Zwart 1962, 1967, 1969, Wickham & Oxburgh 1985, 1987), but it also occurs in compressional regimes, for example the Slave province, in the Northwest Territories (Thompson 1989), and some Proterozoic inliers in Australia (*e.g.*, Wyborn *et al.* 1988, Sandiford & Powell 1991, Reinhardt 1992). High-temperature metamorphism in middle to upper levels in the crust results from heat advection by intruding magmas (Wells 1980, Lux *et al.* 1986, Warren & Ellis 1986, Bohlen 1987, Barton & Hanson 1989, De Yoreo *et al.* 1989a, b, 1991, Harley 1989, Loosveld 1989, Sandiford & Powell 1991). The generation of magmas in the lower crust requires anomalously high basal heat-flow, which is commonly manifested in rift settings by condensed isograds caused by lithospheric thinning (McKenzie 1978, Thompson 1981). In compressional settings, anomalously high basal heat-flux can be produced by a variety of tectonic processes, including crust–mantle delamination, convective removal of the lithospheric mantle root, and simultaneous thickening of the crust and thinning of the lithosphere (Bird 1979, Houseman *et al.* 1981, Loosveld & Etheridge 1990, Platt & England 1993).

The Paleoproterozoic Kisseynew Domain, a 300 by 150 km wide lithotectonic domain in the interior Trans-Hudson Orogen of Manitoba and Saskatchewan (Fig. 1), was metamorphosed at low- to medium-pressure, high-temperature conditions during compressional deformation (Gordon 1989, Gordon *et al.* 1990, 1993). Peak metamorphic grade was uniformly high in the central Kisseynew Domain, and much lower at the north and south flanks (Fig. 2) (Bailes & McRitchie 1978, Froese & Moore 1980, Gordon 1989, and references therein, Perkins 1991, Briggs & Foster 1992, Gordon *et al.* 1993, 1994a, b).

Our work has concentrated on the southern boundary zone of the Kisseynew Domain (Fig. 1). Detailed structural work is presented in Kraus & Williams (1998). Microstructural studies and geothermobarometry are used in this study to interpret the tectonometamorphic history of the area. A companion paper by Menard & Gordon (1997) presents detailed metamorphic P-T-t paths calculated from hydrothermally altered volcanic rocks of the Snow Lake assemblage (see below).

GEOLOGICAL SETTING

The Snow Lake area hosts three distinct tectonostratigraphic assemblages. The first comprises Fe-rich, Al-poor distal metaturbidites of the Burntwood Suite, which were deposited on oceanic crust in the marginal Kisseynew basin at 1.86–1.84 Ga, probably during rifting (Fig. 3) (Bailes 1980b, Zwanzig 1990, Ansdell & Norman 1995, Machado & Zwanzig 1995, David *et al.* 1996). The second is the correlative fluvial-deltaic Missi Suite (*e.g.*, Stauffer 1990, Ansdell 1993, Ansdell & Norman 1995, Ansdell *et al.* 1995). The third is the Snow Lake assemblage, comprising ~1.9-Ga island arc rocks and synvolcanic intrusions, which are coeval with the amalgamated island arc and oceanic assemblages of the Amisk Collage to the west, although the Snow Lake assemblage and the Amisk Collage are chemically and structurally distinct (Stern *et al.* 1995a, b, Lucas *et al.* 1996, David *et al.* 1996, Ryan & Williams 1996). Sandstones of the Missi Suite may have unconformably overlain the Kisseynew basin margin, and now sit unconformably on the Snow Lake assemblage rocks east of Wekusko Lake (K.A. Connors & K.M. Ansdell, pers. commun., 1996). The southern boundary zone of the Kisseynew Domain around Snow Lake comprises a series of folded tectonic slices of sedimentary and volcanic rocks (Fig. 3). We define this zone of tectonic interleaving, which also includes ocean-floor assemblages at the northeast side of Reed Lake and east of Wekusko Lake (Fig. 3) (Syme *et al.* 1995), as the Snow Lake Allochthon. The Snow Lake Allochthon terminates to the west against the Amisk Collage along the Morton Lake fault (Fig. 3) (Stern *et al.* 1995a, b, Lucas *et al.* 1996). To the northwest of the Morton Lake fault, the continuation of the Snow Lake

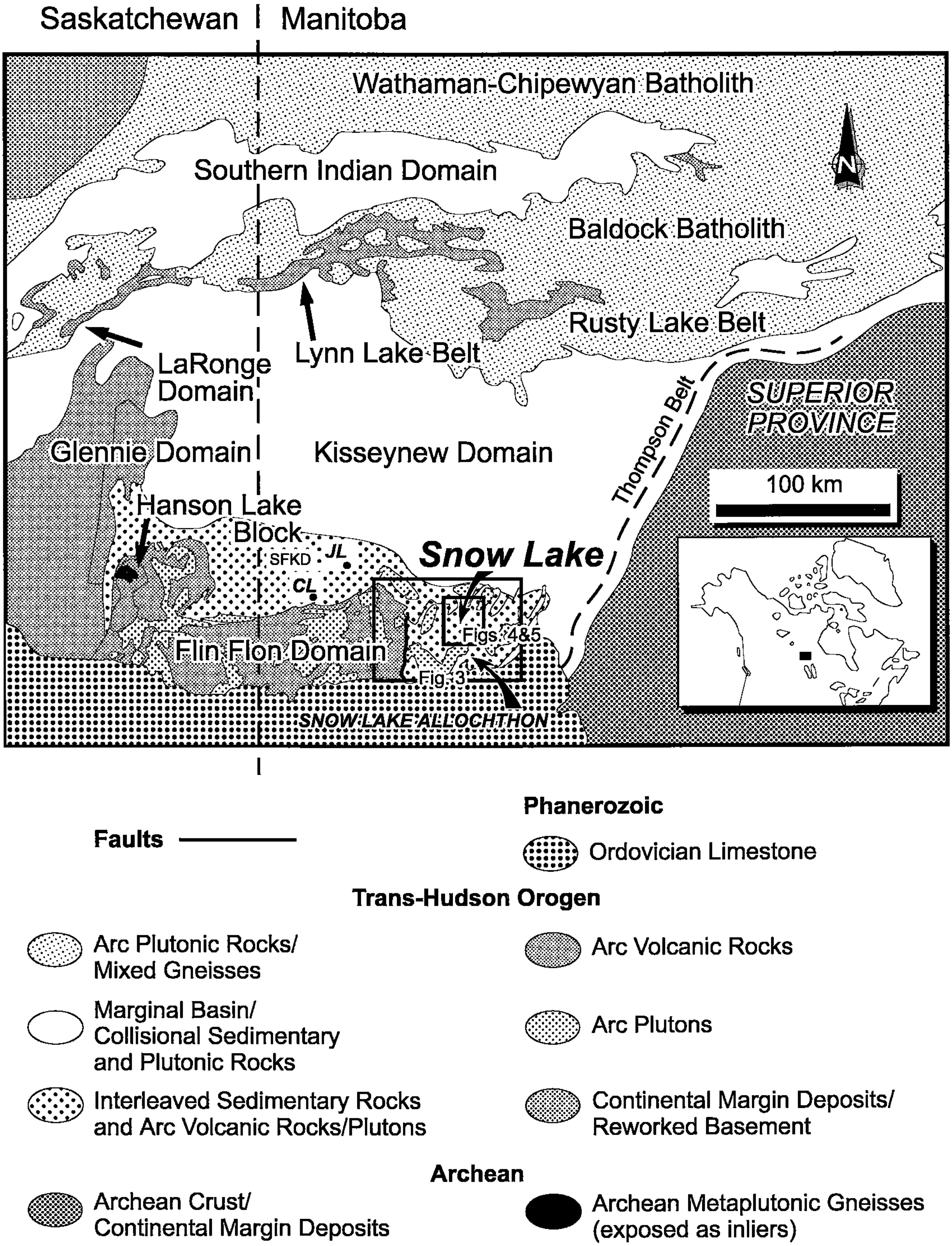

FIG. 1. Lithotectonic domains of the interior Trans-Hudson Orogen, and location of the Snow Lake area at the southern margin of the Kisseynew Domain. CL: Cleunion Lake, JL: Jungle Lake.

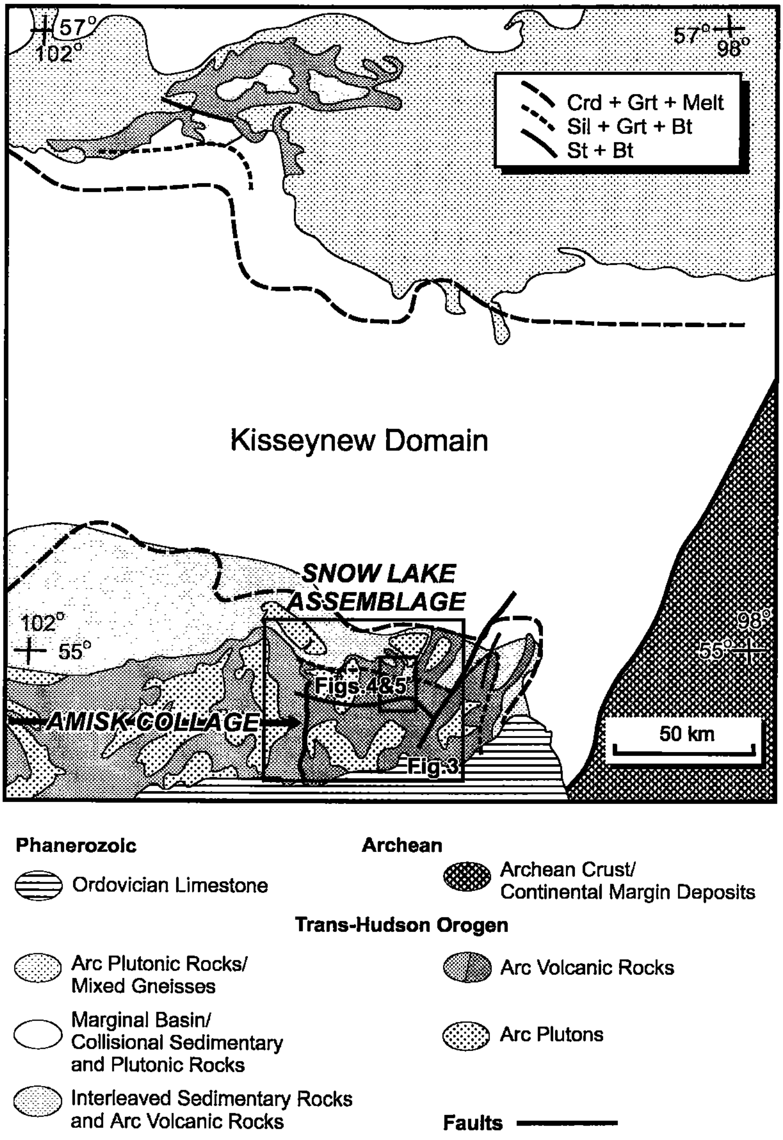

FIG. 2. Metamorphic isograds of the southeastern Trans-Hudson Orogen, after Gordon (1989). Mineral symbols after Kretz (1983).

Allochthon comprises interleaved rocks of the Kisseynew Domain and the Amisk Collage, and is called the south flank of the Kisseynew Domain (Zwanzig 1990, Ansdell & Norman 1995, Norman *et al.* 1995). The Amisk Collage and the Snow Lake Allochthon constitute the central and eastern segments of the Flin Flon – Glennie Complex, respectively (Lucas *et al.* 1996, 1997).

Tectonic imbrication occurred during two phases of deformation (F_1, F_2) that obscured the precise location of the southern boundary of the Kisseynew Domain and structurally underlying assemblages (Kraus & Williams 1998, Connors 1996, David *et al.* 1996, E. Froese, pers. commun., 1996). Kisseynew sedimentary rocks were transported generally to the southwest during north–south convergence of the forming Trans-Hudson Orogen and the Superior craton at 1.84–1.81 Ga (Zwanzig 1990, Zwanzig & Schledewitz 1992, Norman *et al.* 1995, Ansdell *et al.* 1995, Lucas *et al.* 1996, David *et al.* 1996, Kraus & Williams 1998).

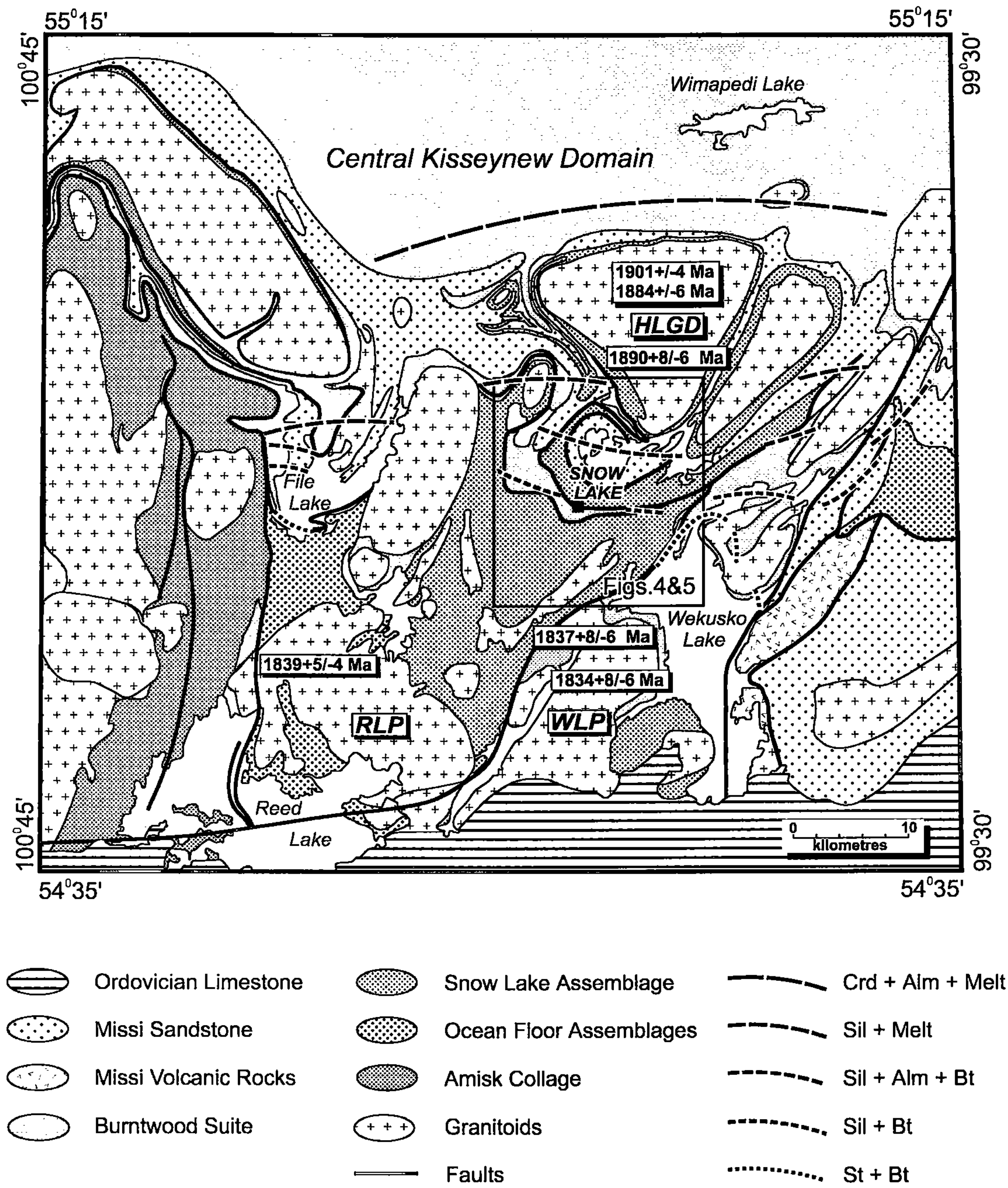

FIG. 3. Snow Lake – File Lake – Wekusko Lake map, including isograds of Froese & Gasparrini (1975), Bailes & McRitchie (1978), Gordon (1981), Gordon & Gall (1982), and Zaleski & Gordon (1990). HLGD: Herblet Lake gneiss dome, WLP: Wekusko Lake pluton, RLP: Reed Lake pluton. Geochronological results refer to age of emplacement of granitic rocks, as cited in text.

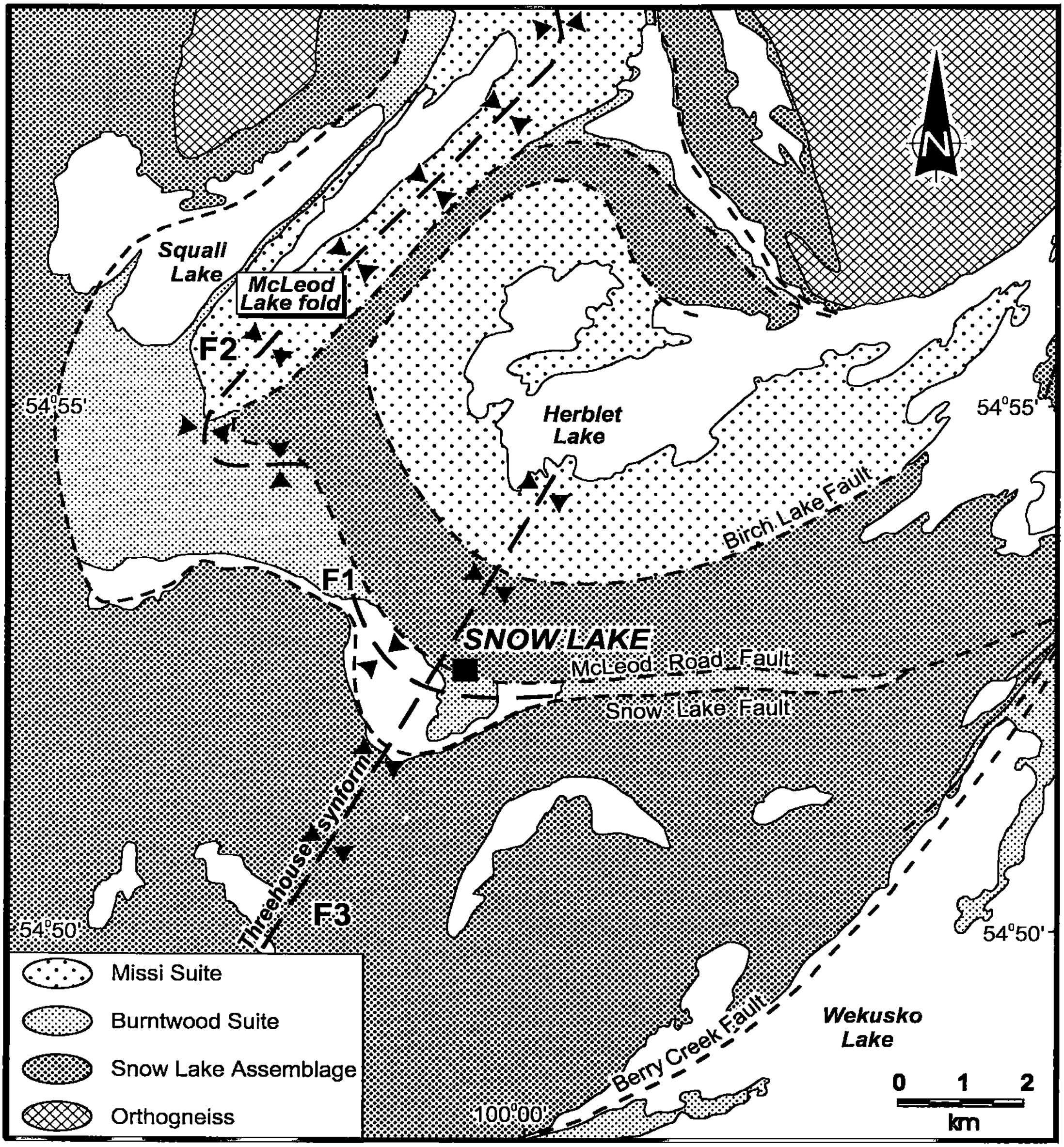

FIG. 4. Simplified structural map of the study area.

In the Snow Lake area, F_1 produced, transposed, and dismembered isoclinal ductile folds at all scales (Fig. 4). Large-scale F_1 structures were truncated by late F_1 movements on the Snow Lake and Berry Creek faults. F_1 deformation may have spanned a narrow time-interval, which is bracketed by the ages of the youngest detrital zircon in samples of the Burntwood Suite (1845 ± 2 Ma and 1842 ± 2 Ma; Machado & Zwanzig 1995), and the age of the post-F_1 Wekusko Lake granite (1841 ± 5 Ma, 1837 +8/-6 Ma and 1834 +8/-6 Ma; Gordon *et al.* 1990, David *et al.* 1996) (Fig. 3).

During F_2 at 1.82–1.805 Ga (Gordon *et al.* 1990, Ansdell & Norman 1995, Parent *et al.* 1995, David *et al.* 1996), the tectonostratigraphy was folded into southwest-verging, tight to isoclinal curvilinear structures, such as the McLeod Lake fold (Fig. 4), with moderately to steeply dipping axial planes, producing a regionally pervasive axial plane S_2 fabric (Kraus &

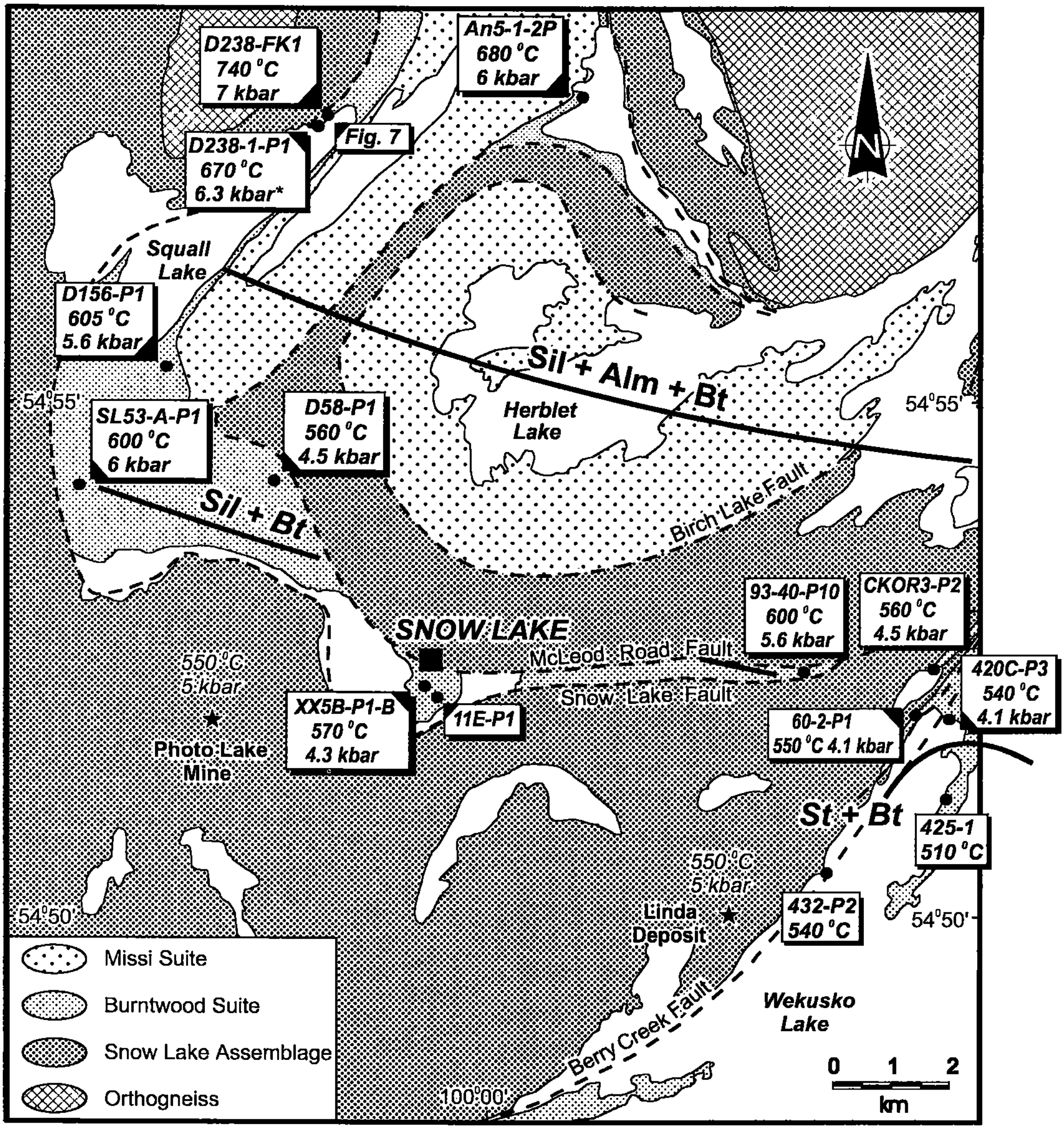

FIG. 5. P–T data [calculated with TWQ 1.02, except * calculated with the database of Hoisch (1990); see Table 5] and isograds of the study area. For comparison, P–T estimates calculated with conventional thermobarometers are given in Table 5.

Williams 1998). F_2 structures were dismembered by late F_2 reverse faults, such as the McLeod Road and Birch Lake faults (Fig. 4), that had southwest-directed transport (Kraus & Williams, in press).

F_3 produced the large-scale, open, upright, north-northeast-trending Threehouse synform (Fig. 4) (Kraus & Williams 1998) during broadly east–west shortening at ~1.8 Ga (Ansdell & Norman 1995, Connors 1996). Folding was associated with sinistral oblique collision of the Trans-Hudson Orogen with the Superior province (Hoffman 1988, Bleeker 1990, Connors 1996, Kraus & Williams, in press).

East–west-trending F_4 cross folds north of Snow Lake overprint favorably oriented, large-scale F_3 folds, yielding dome-and-basin to mushroom-shaped interference patterns (Fig. 3). These domes are cored by synvolcanic intrusions, which are now orthogneisses (Bailes 1975, Gordon *et al.* 1990, David *et al.* 1996).

We interpret the F_4 folds to result from renewed north–south convergence.

Metamorphic grade increases to the north, from sub-biotite grade at Wekusko Lake to lower granulite grade in the central Kisseynew Domain (Figs. 2, 3, 5). Metamorphic zones in the Snow Lake Allochthon extend ~70 km eastward from File Lake to east of Wekusko Lake (Figs. 2, 3) (Harrison 1949, Froese & Gasparrini 1975, Bailes & McRitchie 1978, Gordon 1981, Gordon & Gall 1982, Bailes 1985, Zaleski & Gordon 1990).

North of the study area, sillimanite-bearing migmatites in the vicinity of the Herblet Lake gneiss dome (Fig. 3) were produced by partial melting (Bailes 1975, Bailes & McRitchie 1978). The entire central Kisseynew Domain, north of the Snow Lake gneiss domes, is characterized by the widespread occurrence of quartz + plagioclase + biotite + garnet + cordierite + sillimanite + graphite assemblages, which suggest uniform peak conditions of metamorphism at temperatures of 750 ± 50°C and pressures <5.5 ± 1 kbar (Figs. 2, 3) (Bailes & McRitchie 1978, Gordon 1989, W.D. McRitchie, pers. commun., 1996). Two suites of voluminous granitic rocks are present. An older calc-alkaline suite dated at 1.84–1.83 Ga (*e.g.*, the Wekusko Lake pluton; Fig. 3), which also intruded the Amisk Collage and the Snow Lake Allochthon, and a younger peraluminous suite, which is restricted to the central Kisseynew Domain (Gordon 1989, Gordon *et al.* 1990). The younger suite occurs as sheets of granitic rocks, which have been interpreted as the products of *in situ* anatexis during low- to medium-pressure, high-temperature metamorphism (Bailes 1975, Bailes & McRitchie 1978, Gordon 1989, Zwanzig 1990, Gordon *et al.* 1993). These granitic rocks have yielded ages of 1816 +23/–12 and 1814 +17/–11 Ma (Gordon 1989, Gordon *et al.* 1990), which are generally regarded as the age of the thermal peak of metamorphism in the Kisseynew Domain.

Regional metamorphism began during F_1 and has been interpreted to have peaked during the early stages of F_2 at 1.82–1.805 Ga along the southern flank of the Kisseynew Domain (Zwanzig & Schledewitz 1992, Ansdell & Norman 1995, Norman *et al.* 1995, Parent *et al.* 1995, David *et al.* 1996). Similarly, in the Snow Lake Allochthon, zircon ages interpreted to date the peak of metamorphism range between 1807 ± 7 Ma and 1803 ± 2 Ma (David *et al.* 1996).

METHODS

In a related study, the structure of the Snow Lake area was determined by detailed structural mapping of over 1000 outcrops (Kraus & Williams 1998). In the present study, approximately 200 thin sections of rocks of the Burntwood and Missi suites were examined petrographically in order to examine regional variations in metamorphic grade. Specific targets were variations (1) across the Berry Creek and Snow Lake faults (Fig. 4), (2) across the tectonostratigraphy, and (3) along strike of the tectonostratigrahy (between the Snow Lake and McLeod Road fault; Fig. 4). Metaturbidite samples from thirteen representative locations (Fig. 5) were analyzed with an electron microprobe to determine mineral compositions for geothermobarometry

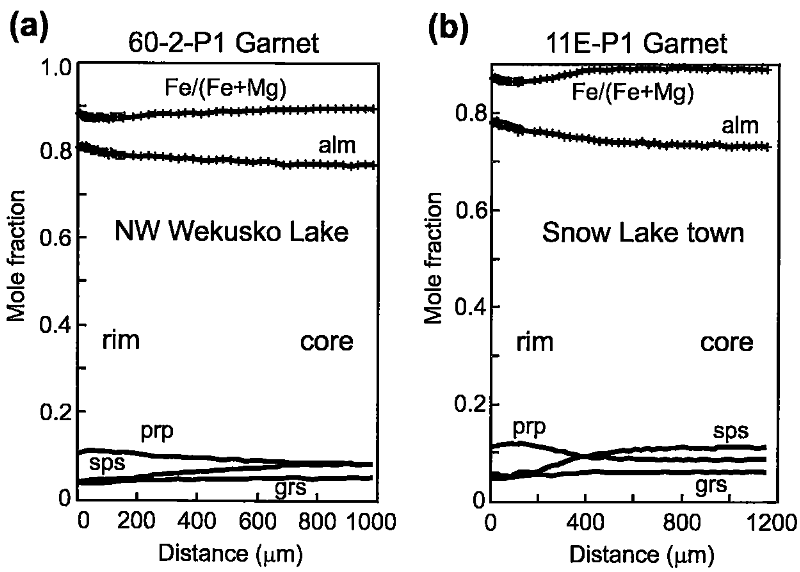

FIG. 6. Compositional profiles across two grains of garnet in metaturbidite of the Burntwood Suite. For sample locations, see Figure 5.

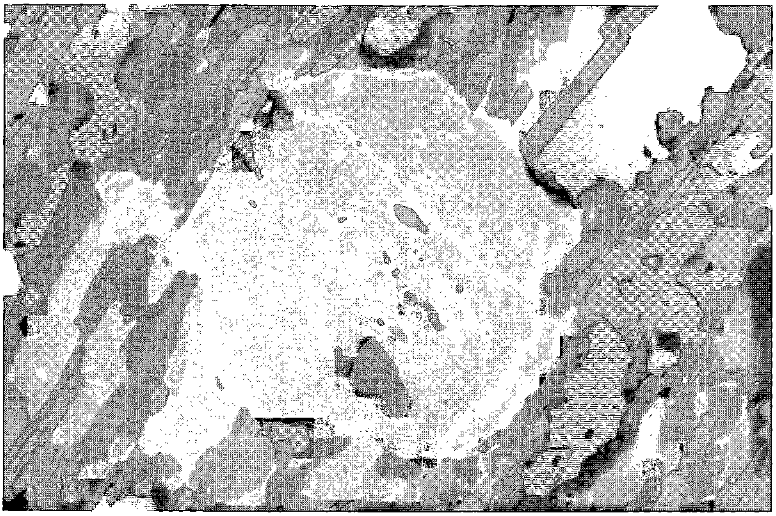

FIG. 7. Orthogonal porphyroblast-matrix relationships between internal S_1 foliation in garnet and external S_2 from the sillimanite zone. Similar relationships are evident in the staurolite zone (Kraus & Williams 1998) and the migmatite zone of the central Kisseynew belt (Menard & Gordon 1997), indicating that the beginning of garnet growth everywhere predated the development of S_2, and thus F_2 folding (Kraus & Williams 1998). Sample D238–1–P1. For sample location, see Figure 5. Width of field of view: 2.7 mm.

(Tables 1–4). In a parallel study of the metamorphism of hydrothermally altered rocks of the Snow Lake assemblage, Menard & Gordon (1997) examined more than 200 additional samples from the Snow Lake area and 10 samples from the Kisseynew gneisses.

Electron-microprobe analysis

The chemical analysis of selected samples was performed on the fully automated JEOL 733 Superprobe at the University of New Brunswick using an acceleration voltage of 15 keV. The maximum counting times were 20 s with a beam current of 50 nA for garnet line traverses, and 40 s at 10 nA for phyllosilicates and plagioclase. Sillimanite was assumed to have an ideal composition. Two detailed compositional profiles across garnet were acquired for zoned crystals (Fig. 6). Spacing of analyses in the traverses was reduced in steps from 25 μm in the core of grains to 2 μm at the rim.

Geothermobarometry

Temperatures and pressures (Fig. 5) were calculated with the TWQ 1.02 program using thermodynamic data from Berman (1988, 1991), and activity models for garnet (Berman 1990, Berman & Koziol 1991), biotite (McMullin *et al.* 1991), amphibole (Mäder *et al.* 1994), and plagioclase (Fuhrman & Lindsley 1988). The results were compared to temperatures and pressures obtained from conventional geothermobarometry using the following assemblages: garnet + biotite (Kleemann & Reinhardt 1994), garnet + biotite + plagioclase + quartz (Hoisch 1990), and garnet + biotite + muscovite + plagioclase (Hodges & Crowley 1985, Powell & Holland 1988, Hoisch 1990). Note that the data-set of Powell & Holland incorporates the model of Hodges & Spear (1982) for garnet. For compositionally zoned grains of garnet, minimum Fe/(Fe + Mg) values near the rim were used to obtain maximum recoverable temperatures, which likely are underestimates of the peak temperatures (Spear 1991). For homogenized grains of garnet in high-grade rocks, peak conditions were estimated using core compositions. In all cases, compositions of matrix biotite near (but not in contact with) the garnet (both separated by matrix quartz) were used for temperature calculations, together with the assumption of an infinite reservoir of biotite. Only samples without retrograde chlorite were used, in order to minimize the effect of retrograde net-transfer reactions on the calculated P–T conditions (Spear 1991, 1993, Spear & Florence 1992). The construction of metamorphic P–T–t paths from zoned garnet in the metapelites was found to be impossible in the samples

TABLE 1. GARNET COMPOSITIONS USED FOR P–T CALCULATIONS

Sample	432 -P	425 -P1	420C -P3	60-2 P-1	CKOR3 -P2	93-40 -P10	XX5 -P1-B	SL53 -A-P1	D58 -P1	D156 -P1	D238- -1-P1	D238 -FK1	AN5 -1-2P
SiO_2 wt%	38.07	37.76	37.49	37.25	37.95	37.69	35.86	38.79	38.31	38.66	38.63	38.52	38.25
TiO_2	0.05	0.11	0.04	n.d.	0.02	0.08	0.09	n.d.	n.d.	0.04	n.d.	n.d.	0.04
Al_2O_3	21.35	20.97	21.19	21.43	21.28	21.29	21.00	21.78	21.45	21.57	21.79	21.56	21.55
MgO	1.76	1.19	2.44	2.86	3.08	3.16	2.76	4.40	2.98	3.95	4.62	4.66	4.40
FeO	32.03	24.95	37.84	35.33	35.85	32.76	37.97	33.40	37.97	36.78	35.13	34.10	35.02
MnO	3.83	8.80	0.72	2.16	1.99	2.63	0.21	0.62	0.46	0.87	0.69	1.22	1.49
CaO	4.64	7.40	1.76	1.52	1.66	2.35	1.30	3.61	1.69	1.48	2.13	1.34	1.46
Total	101.73	101.18	101.48	100.55	101.83	99.96	99.19	102.60	102.86	103.35	102.99	101.40	102.21

n.d.: not detected.

TABLE 2. BIOTITE COMPOSITIONS USED FOR P–T CALCULATIONS

Sample	432 -P	425 -P1	420C -P3	60-2 -P1	CKOR3 -P2	93-40 -P10	XX5 -P1-B	SL53 -A-P1	D58 -P1	D156 -P1	D238 -1-P1	D238 -FK1	AN5 -1-2P
SiO_2 wt%	35.52	35.36	35.31	35.65	35.29	35.10	34.50	37.05	35.56	35.75	36.41	34.87	35.44
TiO_2	1.81	1.83	1.53	1.92	1.51	1.98	2.02	1.48	1.39	1.48	1.93	1.86	1.77
Al_2O_3	18.07	17.91	19.20	19.18	19.64	19.12	19.61	16.85	19.25	19.63	18.47	19.22	19.55
Cr_2O_3	0.03	0.06	0.01	0.07	n.d.	0.04	0.09	0.01	0.05	n.d.	0.08	0.08	0.02
MgO	8.91	9.44	9.15	10.03	10.33	10.50	9.16	13.26	9.85	10.96	11.28	10.38	10.43
FeO	20.99	21.14	20.45	19.10	18.88	18.75	20.81	16.94	19.35	18.51	18.31	19.91	19.03
MnO	0.06	0.17	0.01	0.01	0.01	0.10	n.d.	n.d.	n.d.	n.d.	n.d.	0.05	n.d.
K_2O	9.18	9.12	8.90	8.79	8.71	8.15	7.81	8.56	8.74	8.63	8.24	8.16	8.54
CaO	0.05	0.07	0.03	n.d	0.01	0.03	0.07	n.d.	0.03	n.d	0.03	n.d.	n.d.
Na_2O	0.10	0.02	0.28	0.34	0.20	0.31	n.d.	0.01	0.29	0.25	0.31	0.14	0.20
Total	94.72	95.12	94.87	95.09	94.58	94.08	94.07	94.16	94.51	95.21	95.06	94.67	94.98

n.d.: not detected.

TABLE 3. MUSCOVITE COMPOSITIONS USED FOR P–T CALCULATIONS

Sample	420C -P3	60-2 -P1	CKOR3 -P2	93-40 -P10	XX5 -P1-B	SL53-A -A-P1	D58 -P1	D156 -P1
SiO_2 wt%	47.93	48.41	44.23	44.78	44.75	47.88	44.87	45.70
TiO_2	0.21	0.18	0.54	0.57	0.26	n.d.	0.41	0.30
Al_2O_3	33.92	33.97	34.90	35.18	36.27	35.18	36.25	36.90
Cr_2O_3	n.d.	0.01	n.d.	0.01	n.d.	n.d.	n.d	n.d.
MgO	0.34	0.46	1.01	0.74	0.41	0.69	0.51	0.35
FeO	0.65	0.86	2.65	1.30	0.67	0.83	1.01	0.84
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CaO	0.35	n.d.	n.d.	0.03	n.d.	n.d.	n.d.	n.d.
Na_2O	2.79	1.39	1.06	1.22	1.54	0.47	1.31	1.77
K_2O	7.78	8.30	8.68	9.10	7.68	9.96	9.06	8.63
Total	93.97	93.58	93.07	92.93	91.58	95.01	93.42	94.49

n.d.: below detection limit.

studied owing to the lack of either compositionally zoned plagioclase or appropriate inclusions in garnet.

Variations in Metamorphic Grade

Chlorite and chlorite + biotite zone

Metaturbidite samples of the Burntwood Suite on the islands near the west side of Wekusko Lake (Figs. 3, 5) contain the assemblage chlorite ± biotite + quartz + plagioclase, with minor muscovite, ilmenite, rutile, pyrrhotite, tourmaline, magnetite, zircon, and monazite. Chlorite (locally with some muscovite) defines an S_2 slaty cleavage to a crenulation cleavage that cross-cuts large (hundreds of meters) F_1 folds. Locally, a bedding-parallel S_1 chlorite fabric is preserved. At higher grades, S_2 wraps around biotite porphyroblasts. A distinct garnet + biotite zone is not developed (Froese & Gasparrini 1975), but garnet appears in and near calc-silicate pods near the staurolite isograd. These grains of garnet have high Ca and Mn contents and grew at lower temperatures than in typical samples of the Burntwood Suite (*cf.* Spear 1993).

Staurolite + biotite zone

Staurolite-grade metaturbidites occur between Wekusko, Snow and Squall lakes (Figs. 3, 5); they contain the assemblage staurolite + biotite + garnet + muscovite + quartz + plagioclase + graphite ± chlorite, and accessories as above. Locally, retrograde chlorite partially replaced biotite and the rim of garnet grains, and retrograde chlorite and muscovite replaced the rim of staurolite grains. Although there is no marker fabric associated with the retrograde chlorite, this replacement

TABLE 4. PLAGIOCLASE COMPOSITIONS USED FOR P–T CALCULATIONS

Sample	420C -P3	60-2 -P1	CKOR3 -P2	93-40 -P10	XX5 -P1-B	SL53 -A-P1	D58 -P1	D156 -P1	D238 -1-P1	D238 -FK1	AN5 -1-2P
SiO_2 wt%	62.42	61.89	62.52	60.20	62.33	46.00	61.80	62.82	59.82	62.34	63.26
Al_2O_3	24.60	24.35	24.83	25.98	23.18	35.52	24.36	24.10	26.41	24.30	25.00
FeO	n.d.	0.02	0.02	0.19	0.12	0.04	0.23	0.08	n.d.	0.20	n.d.
CaO	5.48	5.35	5.71	6.96	4.89	18.29	5.64	5.05	7.69	5.32	5.58
Na_2O	8.76	8.74	8.88	8.07	8.84	1.35	7.72	9.29	7.56	9.01	8.86
K_2O	0.07	0.23	0.17	0.08	0.07	0.05	0.07	0.08	0.22	0.07	0.11
Total	101.33	100.58	102.13	101.48	99.43	101.25	99.82	101.42	101.70	101.24	102.81

n.d.: not detected.

occurred most likely during F_3, in agreement with chlorite aligned parallel to a local S_3 in hydrothermally altered rocks of the Snow Lake assemblage of comparable peak metamorphic grade (Menard & Gordon 1997). The majority of the samples examined of any metamorphic zone, however, lack a pervasive retrograde overprint. Prograde muscovite grew during F_1, defining a bedding-parallel S_1 that was subsequently crenulated and differentiated into an S_2 domainal cleavage during F_2 (Kraus & Williams 1998, *cf.* Briggs & Foster 1992, Connors 1996). S_2 anastomoses around porphyroblasts of garnet, biotite, and staurolite. Trails of inclusions in porphyroblasts of garnet, staurolite and biotite are straight or preserve open F_2 crenulations of S_1, and thus document early increments of S_2 development (Fig. 7). A detailed description of porphyroblast–matrix relationships is given by Kraus & Williams (in press).

Grains of almandine-rich garnet 1–3 mm in diameter are compositionally zoned (Fig. 6), with Fe/(Fe + Mg), X_{Grs}, and X_{Sps} decreasing from core to rim, and X_{Alm} and X_{Prp} increasing, which may indicate growth during minor heating and loading (Spear *et al.* 1991). Nevertheless, the magnitude of the zoning in garnet is small (Fig. 6), indicating that the changes in temperature and pressure during growth were minor. The rims of garnet grains display patterns of zoning that suggest minor modification by diffusion during cooling (Fig. 6). The lack of significant diffusional homogenization of garnet is in agreement with modeling by Spear (1988), which predicts that grains of garnet with a radius greater than 1 mm will not experience significant modification of zoning at a thermal maximum of 585°C (following 65 Ma of heating).

Laths of porphyroblastic biotite up to 2 mm long are aligned with (001) along S_2 (Kraus & Williams 1998). The biotite is compositionally homogeneous. Prograde chlorite is preserved only as inclusions in porphyroblasts, suggesting that chlorite was consumed during growth of the porphyroblasts. Euhedral, poikiloblastic, texturally zoned grains of staurolite are up to 14 cm long and contain abundant inclusions of partially corroded garnet, suggesting that garnet was consumed during the staurolite-forming reaction.

Grains of matrix plagioclase are generally small (≤0.25 mm), untwinned and nearly unzoned (<2.5 wt% variation of CaO + Na_2O from core to rim). Compositions vary from An_{20} to An_{36} in typical rocks of the Burntwood Suite, and from An_{86} to An_{89} in rare calc-silicate pods and layers. Within an individual thin section, An contents of plagioclase rims vary by only 0.4 to 2.1 wt%. A second generation of albite-twinned plagioclase occurs in F_2 pressure shadows of staurolite and biotite, and thus postdates growth of the peak metamorphic assemblage.

Sillimanite + biotite and sillimanite + garnet + biotite zones

Above the sillimanite isograd north of Snow Lake (Figs. 3, 5), metaturbidites of the Burntwood Suite contain the assemblage biotite + garnet + sillimanite + quartz + plagioclase + graphite ± muscovite ± staurolite, and the accessories listed above. In rocks that record peak temperatures of approximately 600°C immediately south of Squall Lake (Fig. 5), small volumes of fibrous sillimanite nucleated on plagioclase–plagioclase or plagioclase–quartz grain boundaries. Locally, nodules of fibrolitic sillimanite (Faserkiesel) replaced biotite (*cf.* Vernon 1987, Foster 1991). Sillimanite fibers that replaced muscovite in the S_2 septa were subsequently kinked by F_3 or F_4, and display recovery features, such as migrated kink-band boundaries and rare growth of new grains parallel to the kink planes, as is common in micas (Etheridge & Hobbs 1974, Williams *et al.* 1977). In rocks exceeding calculated temperatures of 650°C, muscovite and staurolite are not observed in the matrix. Locally, staurolite relics and biotite are armored by large grains of twinned plagioclase, suggesting that the breakdown of staurolite may have involved the formation of plagioclase. Here, an early generation of garnet is texturally identical to the garnet in the staurolite zone (see above), but a second generation of garnet (up to 5 mm in diameter) is texturally zoned, with inclusions of quartz, plagioclase, biotite and sillimanite in the core, and fibrolitic sillimanite close to the rim. Both

TABLE 5. MINERAL ASSEMBLAGES AND CALCULATED P–T ESTIMATES

			P (kbar)*						T (°C)**	
Sample	Assemblage	TWQ P–T	1-Mg	1-Fe	2-Mg	2-Fe	3	4	K & R	TWQ
432–P2	Grt+Pl+Bt+Chl+Qtz								540 – 550	540 – 550
425–1	Grt+Pl+Bt+Qtz								505 – 515	500 – 510
420C–P3	St+Grt+Pl+Bt+Ms+Qtz	4.1 – 540	4.2	4.2	4.0	4.1	4.5	4.4	545 – 555	
60–2–P1	St+Grt+Pl+Bt+Ms+Qtz	4.1 – 550	4.1	4.1	4.1	4.2	4.6	4.5	550 – 560	
CKOR3–P2	St+Grt+Pl+Bt+Ms+Qtz	4.5 – 560	4.4	4.5	4.4	4.4	4.7	4.6	555 – 565	
93–40–P10	Sil+St+Grt+Pl+Bt+Ms+Qtz	5.6 – 600	5.8	5.8	5.6	5.6	5.7	5.9	585 – 595	
XX5B–P1–B	St + Grt + Pl + Bt + Ms + Qtz	4.3 – 570	4.7	4.7	4.5	4.6	4.8	4.6	565 – 575	
SL53–A–P1	St+Grt+Pl+Hbl+Bt+Ms+Qtz	6.0 – 600	6.1	6.1	6.1	6.1	4.4	4.9	595 – 605	
D58–P1	St+Grt+Pl+Bt+Ms+Qtz	4.5 – 560	4.4	4.6	4.5	4.8	4.9	4.9	555 – 565	
D156–P1	St+Grt+Pl+Bt+Ms+Qtz	5.6 – 605	4.8	4.6	5.0	4.6	5.4	4.7	590 – 600	
D238–1–P1	Grt+Pl+Bt+Qtz		6.3†	6.4†					645 – 655	660 – 670
D238–FK1	Sil+Grt+Pl+Bt+Qtz	7.0 – 740	7.5	7.1					705 – 715	
AN5–1–2P	Sil+St+Grt+Pl+Bt+Qtz	6.0 – 680	6.4	6.3					655 – 660	

Methods: 1-Mg: Hoisch (1990), Mg end-member (R1); 1-Fe: Fe end-member (R2); 2-Mg: Hoisch (1990), Mg end-member (R5); 2-Fe: Fe end-member (R6); 3: Powell & Holland (1988); 4: Hodges & Crowley (1985); K&R: Kleemann & Reinhardt (1994); TWQ: TWQ version 1.02 (Berman 1991).
Notes: *P calculated at T from TWQ. ** T calculated at 4 and 6 kbar. † at 670°C.

generations of garnet are compositionally homogeneous, except for strongly retrograde rims up to 100 μm wide, which yielded calculated temperatures up to 180°C lower than the core. These higher-grade rocks do not contain staurolite and muscovite, but a second generation of prismatic sillimanite parallel to S_2.

Spatial distribution of P–T

Calculated temperatures and pressures are shown in Figure 5 and Table 5. In the staurolite zone, between the Berry Creek and McLeod Road faults, recorded apparent peak conditions of metamorphism are constant at approximately 540–570°C and 4.1–4.3 kbar, in agreement with 550°C and 5 kbar reported for altered rocks of the Snow Lake assemblage at the Linda and Photo Lake deposits (Fig. 5) (Zaleski *et al.* 1991, Menard & Gordon 1997).

At the northwestern end of Wekusko Lake, rocks of the Burntwood Suite are exposed on both sides of the Berry Creek fault without an apparent discontinuity in metamorphic grade (Fig. 5). Calculated temperatures and mineral assemblages in rocks immediately east of the Berry Creek fault indicate progressively lower grades toward the south (Fig. 5). There also is no detectable change in metamorphic grade across the Snow Lake fault. Temperature increases to 600°C along strike of the tectonostratigraphy, between Snow and Squall lakes. Along the eastern shore of Squall Lake, temperatures increase to the north along strike, up to >670°C at the north end (Fig. 5). Thus at Squall Lake, the isotherms crosscut the F_3 Threehouse synform, in agreement with the regional pattern of isograds (Figs. 3, 5) (Froese & Gasparrini 1975, Zaleski & Gordon 1990). Further north, above the cordierite + garnet isograd, cordierite + garnet + sillimanite assemblages suggest uniform temperatures of 750 ± 50°C throughout the central Kisseynew Domain (Bailes & McRitchie 1978, Gordon 1989). The lack of earlier kyanite and the presence of cordierite + garnet indicate that pressures never exceeded 6 kbar (Gordon 1989, Spear & Cheney 1989).

As mentioned above, the calculated temperatures for compositionally zoned grains of garnet may be slightly below the peak temperatures. In the staurolite zone, for example, temperatures are constrained by the absence of sillimanite to have been less than 580–590°C (Spear & Cheney 1989). Thus the associated calculated pressures also may be erroneously low, because temperatures and pressures are positively correlated in the Snow Lake Allochthon, which experienced heating during loading (Menard & Gordon 1997, this study). On the other hand, pressures calculated for higher-grade rocks that contain compositionally homogenized garnet (T > 650°C) may be slightly overestimated, if the calculated temperatures are overestimated, because calculated pressures are temperature-dependent (positive slopes of K_{eq} lines). The calculated temperatures in such high-grade rocks depend on the corrections of the barometers for the nonideality of (Mg, Fe)–^{VI}Al mixing and (Mg, Fe)–Ti mixing in biotite, which give temperatures up to 100°C lower than those obtained without correction (Spear 1993, Spear & Kohn, unpubl. manuscript). Taking these factors into account, it seems realistic to assume that pressures associated with peak temperatures of metamorphism were similar throughout the study area at 5–6 kbar, consistent with pressures determined for the central Kisseynew Domain (Bailes & McRitchie 1978, Gordon 1989).

DISCUSSION

Timing of peak metamorphism: evidence of diachronism

Garnet commenced growing in metaturbidite and altered volcanic rocks everywhere in the study area prior to F_2 folding, as shown by identical porphyroblast–matrix relationships (Kraus & Williams 1998, Menard & Gordon 1997, this study). The temperature of inception of garnet growth was ~500°C, as determined from petrogenetic grids (*e.g.*, Spear & Cheney 1989) and from calculations of P–T–t paths (Menard & Gordon 1997). In a companion paper, Menard & Gordon (1997) report increases in temperature and pressure during garnet growth in samples from the Photo Lake deposit (equivalent to the staurolite zone in the metapelites; Fig. 5) of up to 50°C and 1–1.5 kbar to peak conditions during the development of S_2 (Fig. 8). Thus, taking into consideration that S_2 predated or coincided with early stages of F_2 folding in strongly anisotropic micaceous rocks (Kraus & Williams 1998), peak conditions of metamorphism were reached in the staurolite zone early during F_2 folding.

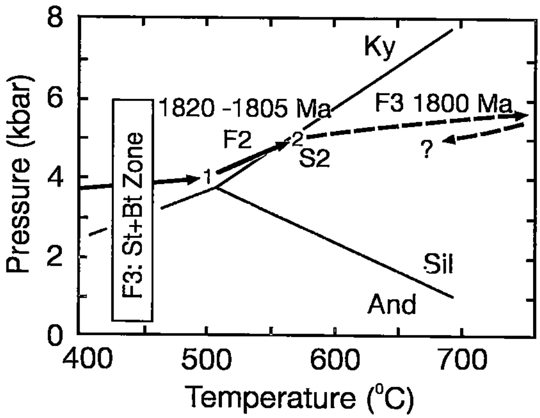

FIG. 8. Calculated P–T–t paths for the Snow Lake Allochthon and the Central Kisseynew Domain. Garnet commenced growing at 1; S_2 formed at 2. For further explanation, see text.

Metamorphism during F_3 in the staurolite zone was a retrograde event at greenschist-facies conditions. In contrast, metamorphism during F_3 in the sillimanite zone and above was the peak metamorphic event, as shown by isotherms and isograds that cross-cut large F_3 structures, such as the Threehouse synform and the File Lake antiform 20 km to the west (Fig. 3) (Froese & Gasparrini 1975, Bailes & McRitchie 1978, Bailes 1980a, this study). At File Lake, the staurolite isograd follows bedding in the Burntwood Suite metaturbidite around most of the F_3 File Lake antiform, whereas the sillimanite and partial melting isograds clearly cross-cut the structure (Fig. 3) (Bailes & McRitchie 1978, Bailes 1980a; see also Connors 1996). Thus, peak metamorphism was diachronous within the Snow Lake Allochthon.

Possible mechanisms of low- to medium-pressure metamorphism

Rocks above staurolite grade had a significantly higher thermal gradient (tighter spacing of isotherms) during peak metamorphism than did rocks at lower metamorphic grades, indicating a thermal anomaly centered in the Kisseynew Domain (Gordon 1989, Ansdell *et al.* 1995, Menard & Gordon 1997). The thermal anomaly started after the development of S_2, as demonstrated by syn-S_2 garnet from Snow Lake and from the Kisseynew Domain, even though the Kisseynew samples later reached much higher temperatures (Fig. 8; Menard & Gordon 1997).

Several models have been proposed to explain the thermal anomaly in the Kisseynew domain: (1) Production of radiogenic heat in the crust was unlikely the sole cause, because it would have required a period of at least 60 Ma to produce the observed high-temperature metamorphism (Gordon 1989, Gordon *et al.* 1993). (2) Conductive response to a high basal heat-flow (England & Richardson 1977, England & Thompson 1984) could have resulted in the elevated local thermal gradients during metamorphism (Gordon 1989, Gordon *et al.* 1993, Ansdell & Norman 1995). High basal heat-flow was related by Ansdell & Norman (1995) to the influx of astenospheric material to the base of the continental lithosphere as a result of crust–mantle delamination (Bird 1979, Bird & Baumgardner 1981, Kay & Kay 1993) or convective removal of the root of the mantle lithosphere (Platt & England 1993). (3) Heat advection by the 1.84–1.83 Ga granitic plutons has been proposed as the cause (Gordon 1989, Gordon *et al.* 1993, Ansdell *et al.* 1995). However, rocks surrounding the 1.84–1.83 Ga Reed and Wekusko Lake plutons (Fig. 3) (Gordon *et al.* 1990, David *et al.* 1996; R.A. Stern, pers. commun., 1996) are of low metamorphic grade (chlorite assemblages), suggesting that calc-alkaline magmatism contributed little to the overall heat-budget. Discussion of these alternatives follows, starting with observable features in the ~1.815 Ga granitic suites and migmatites of the central Kisseynew Domain.

Third-order mechanisms – granite emplacement versus in situ anatexis?

The widespread migmatites and younger peraluminous granitic rocks in the central Kisseynew Domain are generally considered to have been generated by *in situ* anatexis, and thus as the product rather than the (local) cause of high-temperature metamorphism at moderate pressures (Bailes 1975, Bailes & McRitchie

1978, Gordon 1989, Zwanzig 1990, Gordon *et al.* 1990, 1993). In contrast, there is compelling evidence presented below that heat was advected by the ~1.815 Ga granitic plutons and thus caused local anatexis of the metasedimentary country-rocks at the presently exposed level of the crust. This disparity has important implications for possible crust–mantle interactions (first-order mechanism) that may have led to the production of granitic magmas at lower levels in the crust (second-order mechanism).

The origin of the younger peraluminous granitic rocks and the migmatitic paragneisses in the Kisseynew Domain can be constrained by the following lines of evidence. (1) Peraluminous granitic, granodioritic and minor trondhjemitic orthogneisses occupy *ca.* 40% of the area around Wimapedi Lake, immediately north of the Snow Lake gneiss domes (Fig. 3) (Bailes 1975), and up to 75% of the central Kisseynew Domain (Zwanzig 1990). (2) The granitic rocks contain xenoblasts of garnet and cordierite, and locally biotite, and up to 15% xenoliths of fertile rocks of the Burntwood Suite (Bailes 1975, Bailes & McRitchie 1978, Gordon 1989, Zwanzig 1990). (3) Most of the large bodies of granitic rocks are tabular, concordant with bedding or parallel to the axial planes of large folds, but they also occur as dykes and stocks (Bailes 1975, Gordon 1989, Zwanzig 1990, W.D. McRitchie, pers. commun., 1996). (4) The contacts between the granitic rocks and the metasedimentary rocks range from sharp, intrusive for the larger bodies to diffuse for the smaller bodies (Bailes 1975, Gordon 1989, Zwanzig 1990). (5) There are no large volumes of noritic restite located at the margin of larger bodies of granitic rocks. Restite rich in garnet + cordierite + biotite occurs only as small-scale domains associated with ubiquitous small stringers of granitic material (A.H. Bailes, pers. commun., 1996). (6) The paragneisses of the Burntwood Suite are compositionally uniform in the Kisseynew Domain, and muscovite was not present in the rocks prior to partial melting (Froese & Gasparrini 1975, Bailes & McRitchie 1978, Gordon 1989, W.D. McRitchie, pers. commun., 1996, this study). (7) The peak temperatures recovered from the paragneisses are also uniform at 750 ± 50°C at pressures of 6 kbar or less (Bailes & McRitchie 1978, Gordon 1989).

These observations lead to the following implications: (1) The tabular geometry of the bodies of granitic rocks and their field relationships with the metasedimentary country-rocks in the Kisseynew Domain imply the emplacement of magma along zones of primary and structural anisotropy (*cf.* Collins & Sawyer 1996). Paragneisses of relatively uniform compositions in the entire Kisseynew Domain had uniform solidus temperatures, which cannot account for selective *in situ* generation of stratiform sheets of granitic rocks with locally sharp contacts to country rock and xenoliths. Local gradational contacts can be explained by anatexis of country rocks, which were already relatively hot at the time of emplacement of granitic rocks (see above), resulting from heat transferred from the granitic rocks. (2) If the peraluminous granites were generated by *in situ* anatexis and remained at the site of generation, there should be large volumes of restite associated with them, which are not observed. Further, magmas created by high melt-fractions (>30–40%) generally have low viscosities, and are thus easily extracted to form plutons at higher levels in the crust (Wickham 1987, and references therein). Therefore, the voluminous granitic rocks exposed in the Kisseynew Domain likely were produced at lower structural levels. (3) The mineral assemblages in the granitic rocks, in particular the presence of cordierite and the lack of muscovite, implies H_2O-undersaturated conditions during magma generation (Barbarin 1996). Recent experiments on the dehydration melting of similar assemblages suggest that the production of large proportions of melt would require a temperature significantly greater than 850°C at 3–15 kbar (Le Breton & Thompson 1988, Vielzeuf & Holloway 1988, Patiño Douce & Johnston 1991, Gardien *et al.* 1995, Patiño Douce & Beard 1995). In comparison, more hydrous assemblages started melting at ~750°C, and melt fractions increased from 28 to 60% between 825 and 960°C at 10 kbar (Gardien *et al.* 1995). Thus, the metamorphic temperatures in the Kisseynew Domain appear too low to account for high melt-fractions during "dry" *in situ* anatexis.

The available data can be interpreted as suggesting that intruding bodies of granitic magma, of tabular geometry, were the local heat-source for a regional-scale contact metamorphism, similar to that in northern New England (Lux *et al.* 1986, De Yoreo *et al.* 1989b). In contrast to conductive heating, heat advection by intrusions is rapid (*e.g.*, Lux *et al.* 1986, Barton & Hanson 1989), and thus commonly produces near-isobaric P–T paths, which is consistent with observations of Gordon (1989) and Menard & Gordon (1997). The solidus temperatures of the large volumes of granitic magma in intrusive sheets may have buffered the peak temperatures of metamorphism, resulting in partial melting of the paragneisses at a uniform metamorphic grade and thus in uniform mineral assemblages throughout the Kisseynew Domain (*cf.* Waters 1986, De Yoreo *et al.* 1989b).

Second-order mechanisms: implications for melting of the lower crust

The peraluminous composition of the ~1.815 Ga granitic rocks and the abundance of paragneiss xenoliths indicate that the granitic magma may have been generated from rocks of the Burntwood Suite. The site of magma generation was most likely the base of the sedimentary pile, where temperatures exceeded the solidi of the sediments, but not of the mafic rocks of the underlying oceanic crust. One reason that advection of heat, and thus high-temperature metamorphism, were restricted to the Kisseynew Domain could be that metasediments did not occur at greater depth in the adjacent Flin Flon – Glennie Complex and Lynn Lake Belt (Fig. 1). Consequently, ~1.815 Ga granitic suites are absent in these domains, suggesting that low- to medium-grade metamorphism there was triggered mainly by conductive relaxation.

First-order mechanisms: possible crust–mantle interactions

The question remains concerning which tectonic processes generated a deep-seated heat-source that caused the melting of metasedimentary rocks at the base of the Kisseynew Domain. Three possible first-order mechanisms are simultaneous thickening of the crust and thinning of the mantle lithosphere (Houseman *et al.* 1981, Loosveld 1989, Loosveld & Etheridge 1990), delamination of the lithosphere (Bird 1979, Bird & Baumgardner 1981, Kay & Kay 1993), and convective removal of the base of the thickened lithosphere (Platt & England 1993), all of which lead to intrusion of asthenospheric mantle into or accretion beneath the crust (Furlong & Fountain 1986, Huppert & Sparks 1988). Such heat addition can be sufficient for dehydration melting of the lower crust to form granitic magma (Huppert & Sparks 1988, Atherton 1993). Although the thermal effects of the mechanisms are equivalent, their geological consequences differ (Loosveld & Etheridge 1990, Platt & England 1993). Delamination and removal of the root of the lithospheric mantle are typically followed by a short period of extension and uplift (*e.g.*, Bird 1979, Sacks & Secor 1990, Nelson 1992, Kay & Kay 1993, Platt & England 1993), for which there is no structural or petrological evidence in the southern Trans-Hudson Orogen. Thickening of the crust and thinning of the lithosphere, commencing during F_1, thus appears to be the most plausible first-order mechanism for low- to medium-pressure, high-temperature metamorphism in the Kisseynew Domain.

Duration of the thermal anomaly

In the Wimapedi Lake area of the Kisseynew Domain (Fig. 3), the ~1.815 Ga granitic rocks have been deformed by a minimum of two phases of folding, as inferred from Bailes (1975). The earliest generation of folds affecting the intrusions constitute south-verging, recumbent structures that most likely correlate with the F_2 folds in the Snow Lake Allochthon and the south flank of the Kisseynew Domain at Jungle and Cleunion lakes, *ca.* 70 km and 90 km to the west-northwest and west of Snow Lake, respectively (Fig. 1). Elsewhere in the central Kisseynew Domain, some bodies of granitic rocks appear as dyke complexes axial-planar to south-verging (F_2) folds (Zwanzig 1990). In the high-grade rocks at Squall and File lakes, however, metamorphic isograds cross-cut north-northeast-trending, open F_3 folds (Fig. 3). It thus appears that granitic magmatism in the Kisseynew Domain outlasted at least F_2, and that temperatures in the Kisseynew Domain and the high-grade rocks of the Snow Lake Allochthon must have remained elevated until after F_3.

Wells (1980) calculated that a thermal anomaly produced by sill-like batholiths lasts as long as 10–20 Ma, if heat transfer occurs purely by conduction (although heat advection by late-stage fluids would accelerate relaxation of isotherms; Huppert & Sparks 1988). In comparison, monazite ages of the ~1815 Ma granitic rocks in the central Kisseynew Domain (which date the cooling through 725 ± 25°C; Parrish 1990) are 1806 ± 2 and 1804 ± 2 Ma (Gordon 1989, Gordon *et al.* 1990). At Jungle Lake (Fig. 1), several phases of syn-F_2 diatexites and leucosomes, interpreted as having formed near the peak of metamorphism, range from 1812 ± 4 Ma to 1789 ± 2 Ma (Parent *el al.* 1995). Near Cleunion Lake (Fig. 1), syn-F_2 gneissic veinlets were dated at 1818 ± 5 Ma, coeval with peak conditions of metamorphism of greater than 580°C at ~5 kbar (Ansdell & Norman 1995, Norman *et al.* 1995). Syn-F_3, near-peak metamorphic pegmatites yielded crystallization ages from 1801 ± 3 Ma to 1799 ± 3 Ma (Norman & Williams 1993, Ansdell & Norman 1995). Westward migration of the F_3 deformation away from the Trans-Hudson – Superior suture zone could explain why sillimanite-grade conditions were associated with local F_2 in the Jungle Lake and Cleunion Lake areas, and with local F_3 in the Snow–File lakes area, at the same time.

Thus, the duration of high-temperature metamorphism in the Kisseynew Domain may be explained by emplacement of many sheet-like bodies in multiple stages. This scenario can also account for the extent of the thermal effects into the Snow–File lakes area for more than 10 km south of the southernmost exposed granitic rocks (Lux *et al.* 1986, De Yoreo *et al.* 1989a). Such magmatic pulses are compatible with the longevity of a thermal anomaly at the base of the crust caused by thickening of the crust and thinning of the mantle lithosphere owing to convection (Houseman *et al.* 1981, Loosveld 1989, Loosveld & Etheridge 1990). In summary, although the geochronological data do not date metamorphism directly, they are consistent with the persistence of near-peak conditions of metamorphism for 10 Ma in the Kisseynew Domain and the higher-grade rocks of the Snow Lake Allochthon.

SUMMARY AND CONCLUSIONS

Structural mapping, microstructural studies, and geothermobarometry on representative samples suggest that the Snow Lake Allochthon is characterized by syntectonic, diachronous, low- to medium-pressure metamorphism resulting from thermal relaxation after thickening of the crust and granitic plutonism in the adjacent Kisseynew Domain. Because the peak conditions of metamorphism in the staurolite zone were coeval with early F_2 (Kraus & Williams 1998), F_1 likely was the major episode of burial, and was followed by thermal relaxation. Burial during F_1 was probably rapid (1–5 Ma), as indicated by the youngest population of detrital zircon in the Burntwood Suite, and by the age of granitic suites that cross-cut F_1 structures (Gordon *et al.* 1990, Machado & Zwanzig 1995, David *et al.* 1996). The ductile style of F_1 structures suggests thickening of the crust by thick, internally deformed nappes (thick-skinned tectonics). Conditions of metamorphism increased within 15–35 Ma, from chlorite grade to staurolite grade, possibly isobarically. During F_2, temperature and pressure increased only slightly, suggesting relatively minor thickening of the crust after F_1 (Menard & Gordon 1997). A thermal anomaly developed during or after F_2, which led to high-grade metamorphism in the northern part of the study area and the adjacent Kisseynew Domain. Peak conditions in the high grade-rocks prevailed for some 10 Ma until after F_3, whereas mineral assemblages in staurolite-grade rocks indicate cooling during F_3.

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