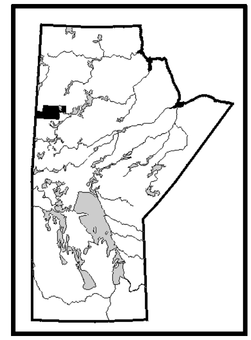

GS-19 STRUCTURAL ANALYSIS AND GEOCHRONOLOGICAL STUDIES IN THE LYNN LAKE GREENSTONE BELT AND ITS GOLD-BEARING SHEAR ZONES (NTS 64C10, 11, 12, 14, 15 AND 16), MANITOBA

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SUMMARY

The application of U-Pb geochronology and Sm-Nd isotopic studies to the continuing structural analysis of the Lynn Lake greenstone belt has elucidated the tectonometamorphic and metallogenic history of the greenstone belt. The complex volcanic tectonostratigraphy assembled during D_1 reflects the juxtaposition of ca. 1890 Ma, weakly contaminated volcanic rocks of the northern Lynn Lake belt with juvenile-arc volcanic rocks of the southern Lynn Lake belt, the age of which is currently under investigation.

The deformed, assembled greenstones acted as basement for the emplacement of successor-arc plutons that are as old as ca. 1875 to 1871 Ma. At least some of the volcanic rocks in the southern belt are therefore thought to be older than 1875 Ma, although two rhyolite bodies yield U-Pb zircon ages as young as ca. 1842 Ma and felsic plutons give ages of 1857 and 1830 to 1820 Ma. Deposition of fluvial-alluvial sediments took place prior to the ca. 1830 Ma plutonism. The greenstone belt was subsequently deformed by an intense D_2 deformation, which transposed the belt into an east-west structural trend and produced regional, dextral transpressional shear zones at ca. 1819 to 1815 Ma. This was closely followed by the peak of regional metamorphism at ca. 1814 Ma. Late synkinematic magmatism and possible associated contact metamorphism coincided with northeast-trending D_4 deformation structures at ca. 1784 Ma.

INTRODUCTION

Understanding the relationship between regional tectonic evolution and metallogeny is a fundamental aspect of successful mineral exploration. In an effort to encourage and support mineral exploration in northwestern Manitoba, a joint program focusing on the tectonic evolution of the northern Trans-Hudson Orogen (Hoffman, 1990) was initiated by the Manitoba Geological Survey (MGS) and the Geological Survey of Canada (GSC), under the auspices of the Targeted Geoscience Initiative (TGI). One component of the MGS contribution to the TGI is a program of detailed structural analysis and isotopic studies of the Lynn Lake greenstone belt, with a focus on gold metallogeny.

The structural analysis of the Lynn Lake greenstone belt was initiated in 1999 and focused on the relationship between the Johnson Shear Zone (Bateman, 1945) and gold mineralization hosted by the shear zone (*see* Beaumont-Smith and Rogge, 1999). The development of shear zones and the associated gold deposition were recognized to have a regional extent in the Lynn Lake greenstone belt. This realization resulted in the expansion of the project to include an appraisal of the deformation history and gold metallogeny of the entire belt (Beaumont-Smith, 2000; Jones et al., 2000; Ma et al., 2000). In 2001, an isotopic study of the Lynn Lake belt was added to the project, in order to provide time constraints and to integrate the structural data into a coherent tectonic framework.

GEOLOGICAL SETTING AND STRUCTURAL OVERVIEW

The Lynn Lake greenstone belt represents a Paleoproterozoic metavolcanic belt located along the northern margin of the Kisseynew sedimentary basin (Fig. GS-19-1), within the internal Reindeer Zone (Stauffer, 1984; Lewry and Collerson, 1990) of the Trans-Hudson Orogen. It comprises a diverse tectonostratigraphy collectively assigned to the Wasekwan Group (Bateman, 1945). The various metavolcanic units with differing tectonic affinities (Zwanzig et al., 1999) are structurally juxtaposed to form two east-trending, steeply north-dipping supracrustal belts.

The northern belt consists of submarine, tholeiitic, mafic metavolcanic and metavolcaniclastic rocks interpreted to represent an overall north-facing, steeply dipping succession (Gilbert et al., 1980). The north belt includes the Agassiz Metallotect (Fedikow and Gale, 1982), a tectonostratigraphic assemblage consisting of ultramafic flows (picrite), banded oxide facies iron formation and associated exhalative sedimentary rocks. The Agassiz Metallotect is a relatively narrow, strike-continuous, stratigraphic-structural unit that occurs along a significant portion of the northern belt (*see* Ma et al., 2000; Ma and Beaumont-Smith, 2001; Park et al., GS-20, this volume). The northern belt is unconformably overlain to the north by marine sedimentary rocks, the Ralph Lake conglomerate and Zed Lake greywacke (Gilbert et

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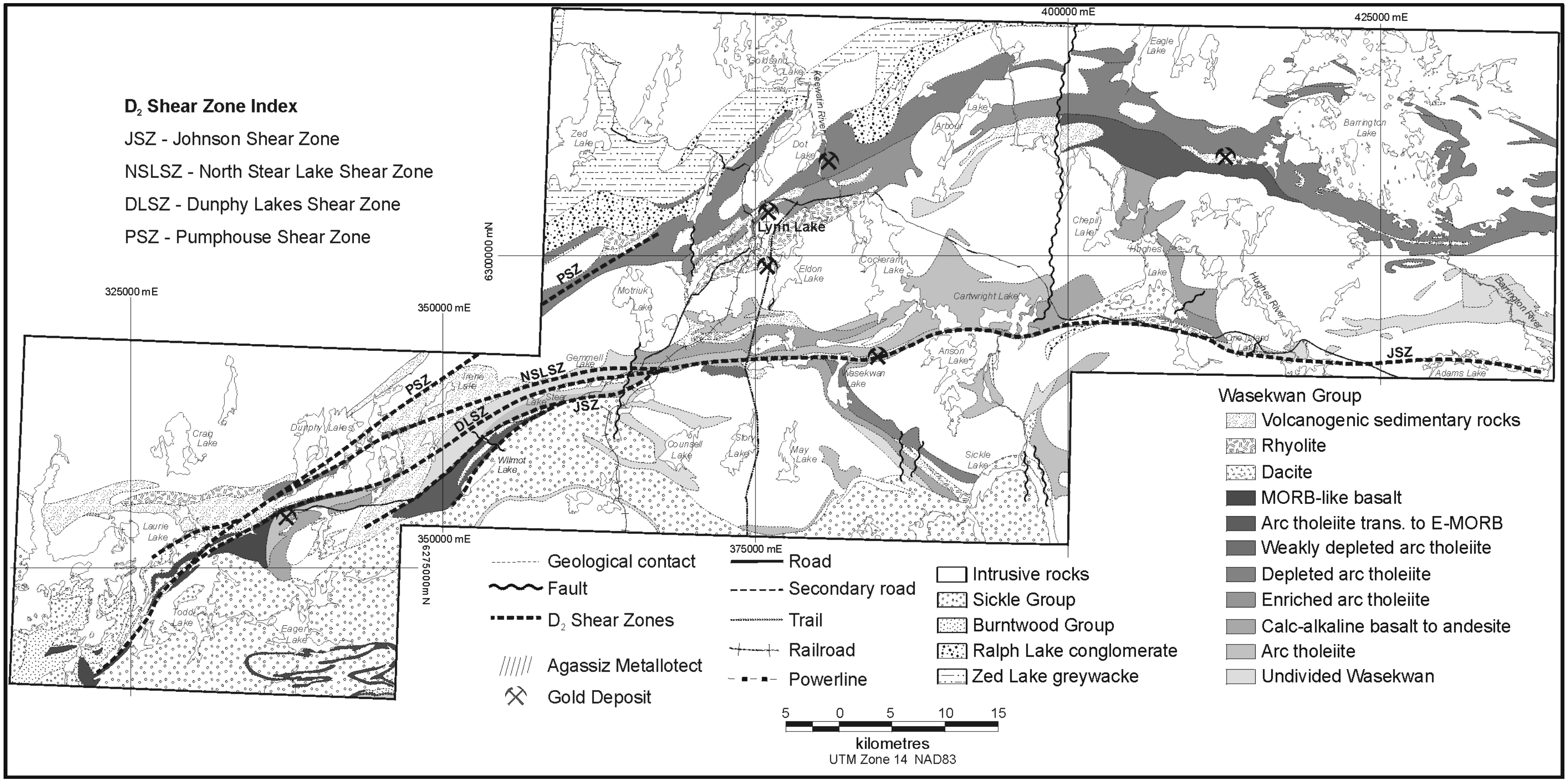

Figure GS-19-1: General geology of the Lynn Lake greenstone belt.

al., 1980), and is intruded by the 1876 Ma (Baldwin et al., 1987) to1871 Ma (Turek et al., 2000) Pool Lake intrusive suite (Manitoba Energy and Mines, 1986) and younger plutons (the post-Sickle plutons of Milligan, 1960).

The southern belt is composed of submarine, tholeiitic to calc-alkaline, metavolcanic and metavolcaniclastic rocks with minor amounts of mid-oceanic ridge basalt (MORB). The southern belt is separated from the northern belt by plutons of the Pool Lake suite. To the south, the southern belt is unconformably overlain and structurally underlain by Sickle Group (Norman, 1933) fluvial-alluvial conglomerate and arkosic sedimentary rocks. The Sickle Group also unconformably overlies the Pool Lake intrusive suite, and the basal conglomerate is characterized by a high proportion of Pool Lake–type plutonic clasts. The tectonic juxtaposition of the volcanic rocks in the northern and southern belts occurred prior to emplacement of the Pool Lake suite and deposition of the Sickle Group. The age of the Sickle Group has not been determined directly, but its composition, stratigraphic position and contact relations correlate well with the 1850 to 1840 Ma MacLennan Group in the La Ronge greenstone belt in Saskatchewan (Ansdell et al., 1999).

Two younger suites of plutons, which yield ages of 1857 Ma (Machado, unpublished data, 2000) and 1832 Ma (Turek et al., 2000), do not have clearly defined contact relations, but emplacement into the Sickle Group is suggested locally. These magmatic ages are consistent with the subdivision of intrusive rocks into early, middle and late successor-arc plutons, as applied in the Flin Flon Belt (Whalen et al., 1999), and suggest a similar complex magmatic evolution in the Lynn Lake greenstone belt.

The structural evolution of the Lynn Lake greenstone belt is the product of six regionally penetrative, ductile deformation events (D_1–D_6). The assembly of the diverse volcanic tectonostratigraphy that characterizes the greenstone belt constitutes D_1. This deformation occurred prior to intense transposition of the belt.

The D_2 transpression of the greenstone belt resulted from tight to isoclinal folding of the D_1 structural geometry. Intense D_2 deformation produced the penetrative, S_2, differentiated regional foliation present in all rock types; accordingly, D_2 deformation postdates intrusion of the youngest plutons and deposition of the Sickle Group. The plunge of F_2 folds is variable, but appears to vary systematically as a function of D_2 strain. In areas of lower D_2 strain, the plunge of F_2 is generally shallow, increasing in concert with increasing D_2 strain. The shallow F_2 plunge in areas of lower D_2 strain suggests the D_1 structural geometry was characterized by shallow dips, which were steepened to subvertical as a result of tight to isoclinal F_2 folding. The steepening of F_2 plunge corresponds to the development of D_2 shear zones throughout the greenstone belt. This reflects the evolution from fold-dominated deformation during the early stages of D_2 shortening, to dextral shear zone development during the later stages of the deformation.

This change may reflect a strain-hardening effect as F_2 folds tighten or as the succession became recrystallized and intruded by the Pool Lake suite. The initial shear zones developed on F_2 fold limbs and subsequently coalesced into broad, regional-scale shear zones. The kinematic framework of D_2 shear zones is characterized by dextral transcurrent shear fabrics and generally steep stretching lineations. This is consistent with their development in response to dextral transpression (Lin et al., 1998). Regional-scale D_2 shear zones represent an important exploration target, hosting the majority of known gold deposits in the northern and southern volcanic belts (Beaumont-Smith and Rogge, 1999; Beaumont-Smith 2000; Beaumont-Smith and Edwards, 2000; Jones et al., 2000; Ma et al., 2000).

The most significant D_2 shear zones are the Johnson Shear Zone (Bateman, 1945), located along the southern margin of the southern greenstone belt, and a similar zone within the Agassiz Metallotect (Ma et al., 2000), located in the northern belt. The Johnson Shear Zone (JSZ) is a dextral transpressive shear zone that has been delineated eastward to the Adams Lake area and westward to the Dunphy Lakes area (Beaumont-Smith et al., 2001, and references therein). The JSZ hosts the known gold deposits and occurrences in the southern belt. Similarly, D_2 shear was an important process in the development of the Agassiz Metallotect. Recognition of D_2 shear zones in both the northern and southern belts demonstrates the regional nature of the D_2 shear zone development.

The structural geometry of the greenstone belt is largely the product of D_1-D_2 interference. Subsequent deformations have had little macroscopic effect on this geometry. Fabric elements that make up the D_3 and D_4 deformations include tight S- and Z-asymmetrical chevron folds and associated crenulation cleavage, respectively. The D_3 fabrics trend northwest, whereas D_4 fabrics trend northeast. Although mesoscopic fabric elements for the respective deformations are regionally penetrative, the deformations have not significantly reoriented the regional D_2 structural trend. Exceptions to this are found in the Boiley Lake area, where the distribution of the Wasekwan Group south of the Johnson Shear Zone is controlled by a large, F_3, S-asymmetrical fold, and in the MacLellan mine area, where the Agassiz Metallotect is folded by a macroscopic, northeast-trending F_4 fold (Park et al., GS-20, this volume).

The kilometre-scale open folds of the regional east-west D_2 structural trend are the result of D_5. Mesoscopic fabrics associated with this deformation include open F_5 conjugate folds, kink bands and open crenulations. A north-trending, spaced, S_5 fracture cleavage is also common. These structures are penetrative at regional scale, but are not mesoscopically penetrative.

The final deformation (D_6) involves the brittle-ductile reactivation of D_2 shear zones, characterized by the development of pseudotachylite zones that overprint all other fabric elements. The reactivation is dominated by sinistral transcurrent movement within narrow zones in the D_2 shear zones. The T_1 fault in the footwall of the Burnt Timber gold deposit is a D_6 structure (*see* Peck et al., 1997, 1998).

Metamorphic grade in the Lynn Lake belt increases westward across the belt from upper greenschist facies to upper amphibolite facies. The relationship between deformational fabrics and metamorphic porphyroblasts indicates that metamorphism is the product of at least two thermal events. The oldest metamorphic event (M_1) reflects contact metamorphism associated with emplacement of the Pool Lake intrusive suite, as well as the younger, post-Sickle plutons. Porphyroblasts formed during M_1 are overprinted by regional D_2 fabric development. The M_1 metamorphism represents multiple, relatively closely spaced events. The regional metamorphic event (M_2) overprints M_1 porphyroblasts and is broadly coincident with D_4. The timing of M_2 does not correlate with any known magmatic event in the belt, and is characterized by the growth of low-pressure, moderate- to high-temperature assemblages.

STRUCTURAL ANALYSIS: NEW RESULTS

Structural analysis undertaken during the 2002 field season concentrated on the delineation of D_2 shear zones in the western portion of the greenstone belt and the relationship between shear-zone development in the northern and southern belts. The GIS analysis of available datasets for the Lynn Lake greenstone belt identified potential areas of D_2 shear-zone development in the Dunphy Lakes area (Rogge, 2002; Rogge et al., 2000). The postulated pattern of high "deformation potential zones" reflected the truncation of the east-trending Johnson Shear Zone by a northeast-trending structural lineament west of Dunphy Lakes. This structure corresponds to known D_2 shear-zone development within the Agassiz Metallotect in the MacLellan mine area (see Rogge, 2002), but delineation of this structure west of Sheila Lake was incomplete.

Detailed structural analysis in the Dunphy Lakes area has recently identified several D_2 shear zones. A broad area of high strain is developed along the southern shore of Dunphy Lakes (Fig. GS-19-2). This zone, the Dunphy Lakes Shear Zone (DLSZ), is characterized by a steep south to north strain gradient, culminating in a 50 m wide mylonitic core. The kinematics of the mylonite zone are consistent with those of other D_2 shear zones, with dextral transcurrent shear fabrics and a steep stretching lineation. This shear zone correlates with the Dunphy Lake Fault (Gilbert et al., 1980) and has been previously interpreted as the western extension of the JSZ (Beaumont-Smith, 2000). In this area, a D_2 shear zone is located along the contact between the Fox mine tonalite (Turek et al., 2000) and the volcanic rocks. The latter have experienced significant transposition, producing layered tectonite units and tectonic intercalation of Fox mine arc tholeiite, Hatchet Lake MORB basalt, Snake Lake dacite and Wasekwan Group epiclastic sedimentary rocks.

A second D_2 high-strain zone, the Pumphouse Shear Zone (PSZ), has been identified within the belt of supracrustal rocks that trends northeast across the central portion of Dunphy Lakes (Fig. GS-19-2). The Wasekwan Group rocks that make up this belt are intensely foliated, making the protoliths difficult to determine. Slight compositional variations of the tectonic layering developed in these rocks suggests that the original stratigraphy consisted of mafic volcanic rocks and lesser amounts of epiclastic sedimentary rocks. The layering has experienced penetrative foliation boudinage of the mafic layers, consistent with high non-coaxial strains, which appears to reflect the localization of regional D_2 deformation into the supracrustal rocks. The contacts with the Fox mine tonalite to the south and the undated intrusive rocks to the north are mylonitized and characterized by a steep strain gradient that decays rapidly away from the supracrustal belt. The dextral transpressive kinematics of this mylonite zone are similar to those of other D_2 shear zones.

The extreme glacial cover east of Dunphy Lakes prevents the delineation of the DLSZ and the PSZ immediately east of the lakes. The lack of geological constraints in this area presents a problem for correlation of the shear zones delineated in the Dunphy Lakes area and the numerous JSZ splays west of Gemmell Lake (*see* Beaumont-Smith, 2000). Geological mapping southeast of Dunphy Lakes has not identified additional high-strain zones other than a broad protomylonite zone developed along the contact between the Wasekwan and Sickle groups, immediately west of Wilmot Lake. This zone is probably the continuation of the southern splay of the JSZ. The DLSZ was interpreted to be the westward continuation of the central JSZ splay that extends along the north shore of Gemmell Lake (Beaumont-Smith et al., 2001), and there is no compelling evidence to modify this correlation.

Correlating the PSZ with previously delineated shear zones east of Dunphy Lakes is more problematic. The northeast trend of the PSZ corresponds to the structure proposed by Rogge (2002). Reconnaissance geological mapping north of Motriuk Lake has identified a broad area of high strain. This zone is developed within a thin unit of Wasekwan Group mafic epiclastic rocks separating thick units of mafic and felsic volcanic rocks, reflecting the localization of D_2 strain into the least competent member of the local stratigraphy. The correlation of the PSZ and this high-strain zone with the northeast-trending zone of high deformation potential proposed by Rogge (2002) is compelling. Alternatively,

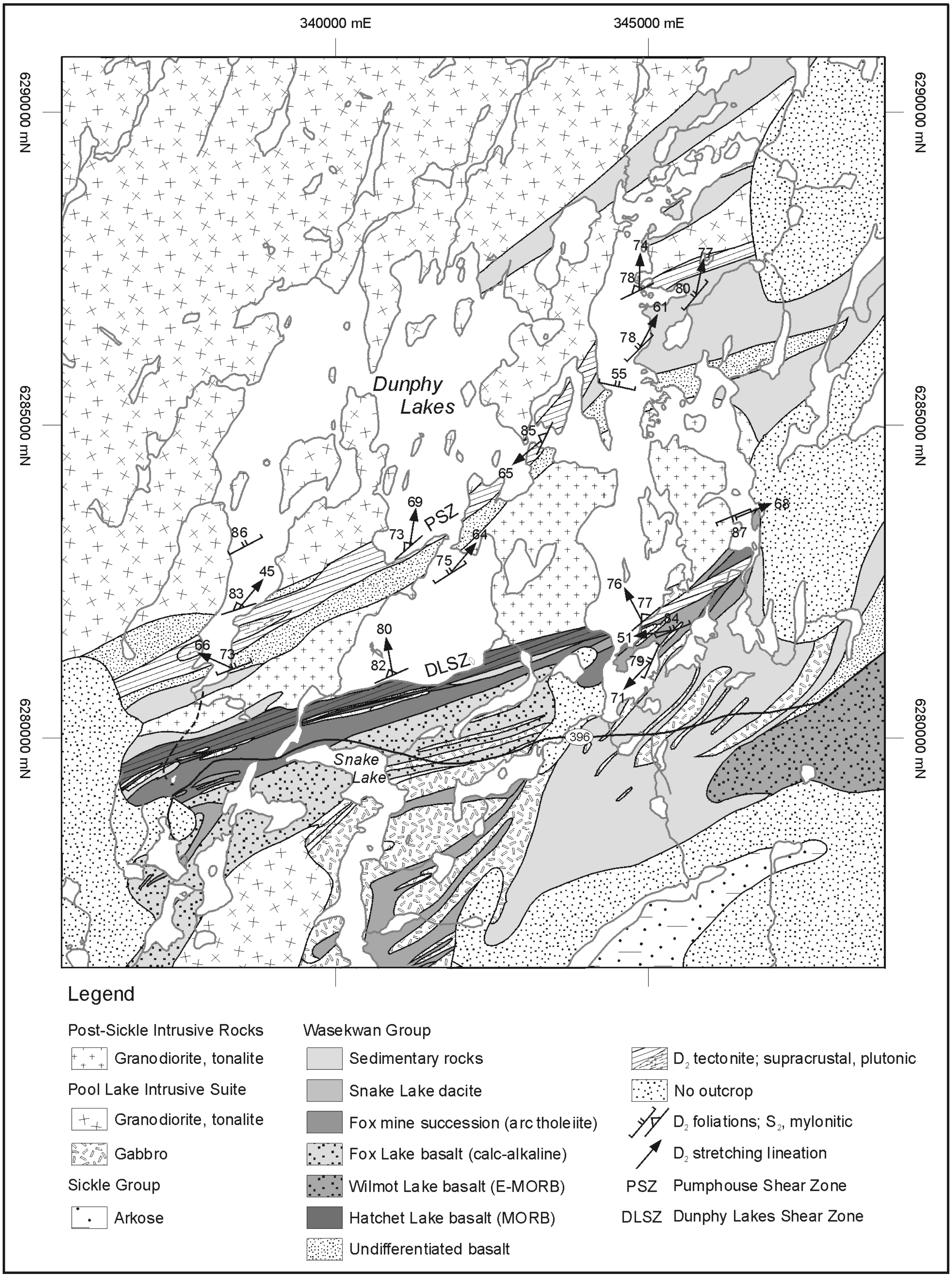

Figure GS-19-2: Detailed geology of the Dunphy Lakes area.

the westward continuation of the North Stear Lake Shear Zone (NSLSZ) has not been adequately delineated, so the relationship between it and the PSZ remains uncertain (cf. Beaumont-Smith et al., 2001). Exploration-company geophysical data clearly delineate the NSLSZ west from the Stear Lake area, where the shear zone outcrops, to the southern tip of Irene Lake (limit of geophysical coverage). Whether the NSLSZ is a splay of the PSZ or is truncated by the PSZ remains uncertain.

Identification of a large number of D_2 shear zones in the Dunphy Lakes area demonstrates that the western portion of the Lynn Lake greenstone belt hosts areas of high gold-mineralization potential. Although the relationship between the various shear zones remains poorly understood, many of the exposed portions of the shear zones host alteration favourable for gold mineralization, and several areas of gold mineralization have been identified (*see* Beaumont-Smith, 2000; Beaumont-Smith et al., 2001; Rogge, 2002).

ISOTOPE GEOLOGY

During the 2001 and 2002 field seasons, a total of 35 samples (10–20 kg of clean, relatively unaltered and homogeneous rock material) were collected for Sm-Nd isotope and/or U-Pb geochronology analysis (Fig. GS-19-3). The aim of this ongoing study is to integrate Sm-Nd isotope and U-Pb geochronological results into detailed structural and petrographic field studies, in order to establish a framework for the tectonic evolution of the Lynn Lake greenstone belt and provide a better understanding of the timing and nature of deformational, metamorphic and mineralizing events, particularly the development of D_2 gold-bearing shear zones.

Timing of felsic volcanism

Preliminary geochronological data from the northern and southern Lynn Lake belts indicate that the two belts include volcanic rocks of significantly different ages. The sample suite from the present northern belt consists of two samples from the Lynn Lake rhyolite, one from the western portion of the body and the other of the northern phase from the eastern portion of the body. Together, these two samples represent the southern and northern phases of the Lynn Lake main rhyolite body as defined by Baldwin (1983). At present, sampling of the southern belt has targeted all the major felsic volcanic bodies and a series of smaller felsic tuffaceous units intercalated with the mafic volcanic stratigraphy. Five felsic volcanic samples have been processed to date, of which two yielded datable zircons.

The age of felsic volcanism in the northern belt has been determined by U-Pb isotpic dating of prismatic euhedral, magnetic zircon crystals (0.8–1.2 Amps) from the two Lynn Lake rhyolite samples. A quartz-plagioclase–phyric sample from the eastern portion of the Lynn Lake rhyolite (northern phase) provided an age of 1892 ± 3 Ma. Two U-Pb analyses of magnetic zircon from the western, southern phase of the Lynn Lake rhyolite provided an age of 1886 ± 1.5 Ma. Most important, these new ages are significantly younger than the ca. 1910 Ma age reported by Baldwin et al. (1987), but contemporaneous with ca. 1.90 to 1.88 Ga felsic volcanism in the Flin Flon Belt (e.g., Lucas et al., 1996).

Compared to the northern belt, the age of felsic volcanism in the southern belt seems much more difficult to determine, and the following results are only preliminary. The poor zircon yield to date, particularly involving the larger, strongly sheared and silicified rhyolitic units, has necessitated additional sampling in 2002. At present, the age of southern belt volcanism is based on U-Pb age dating of two thin rhyodacite units in the Gemmell Lake (west) and One Island Lake (east) areas. Two analyses of zircon fragments from the Gemmell Lake rhyodacite sample returned an age of 1856 ± 2 Ma, and the One Island Lake rhyodacite yielded an age of ca. 1842 Ma based on a three single-grain analyses. Although these ages are preliminary and require additional analyses, they might represent the first direct evidence for the presence of units with a much younger age of volcanism in the southern belt compared to the northern belt. Nevertheless, plutons of the ca. 1876 to 1871 Ma Pool Lake intrusive suite cut across the main units in the southern belt (Gilbert et al., 1980) and indicate that an older period of volcanism is also present. Further geochronology on southern belt rhyolite and plutons with known contact relations is required to confirm this inference.

The different age ranges of felsic volcanism in the northern and southern belts are also reflected in the Sm-Nd isotope results. The five samples of felsic volcanic rocks from the southern Lynn Lake greenstone belt are relatively juvenile (T_{CR} Nd model ages (Goldstein et al., 1984) of ca. 2.1 Ga and initial ε_{Nd} of approx. +2 to +4; *see also* explanations in Zwanzig and Böhm, GS-14, this volume). This is in contrast to the two samples from the Lynn Lake rhyolite in the north belt, which yielded significantly older Nd isotopic signatures with Nd model ages of ca. 2.5 Ga and initial ε_{Nd} of approx. 0 to –0.5, clearly indicating the involvement of crust that is greater than 2.0 Ga, possibly Neoarchean. A Nd isotopic discontinuity between the northern and southern belts probably represents a cryptic crustal boundary between the belts. This correlates with differences in the trace-element geochemistry of the two belts

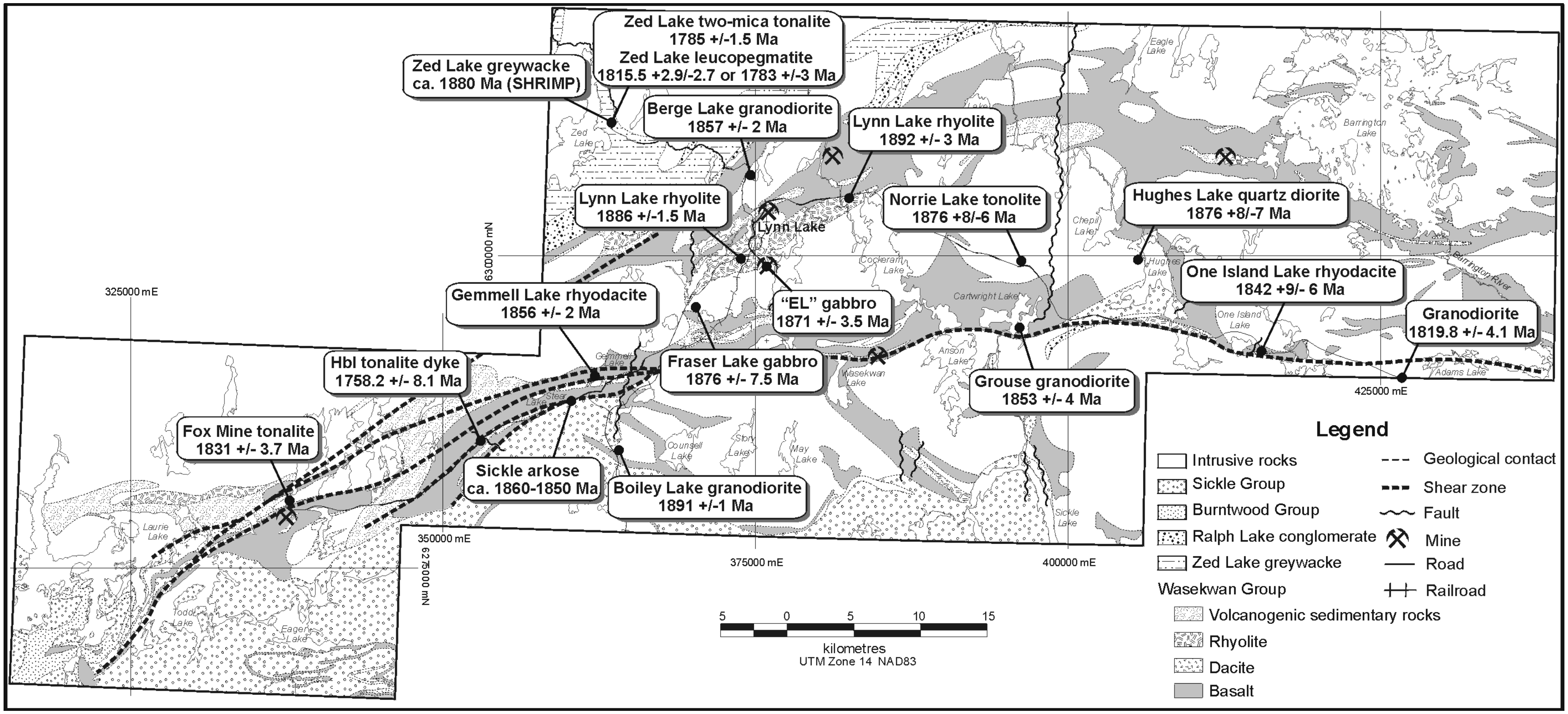

Figure GS-19-3: Locations of isotope geology samples.

(Zwanzig et al., 1999), indicating differences in the tectonic affinities of the constituent volcanic rocks. This boundary should therefore be located in the Fraser Lake–south Eldon Lake area, north of the Johnson Shear Zone (Fig. GS-19-2).

Timing of felsic plutonism

The U-Pb geochronological studies by Turek et al. (2000) and Baldwin et al. (1987) suggested multiple ages of felsic plutonism, at ca. 1876, 1871 and 1831 Ma, that postdate main accretion of the Lynn Lake greenstone belt. Additional unpublished data (Machado, pers. comm., 2000) identify another period of felsic magmatism at ca. 1857 Ma. One aim of this project was to test and complement these data with samples of additional, structurally well-constrained, felsic intrusive bodies and to characterize their geotectonic setting.

Coarse-grained, weakly feldspar-phyric granodiorite from the southeastern Lynn Lake belt forms abundant, pre- or syn-D_2 injection lits in quartz diorite. A sample of the granodiorite yielded a U-Pb zircon crystallization age of 1818.6 +1.2/–1.4 Ma that also represents a maximum age of D_2 deformation in the area (*see* below). Furthermore, the granodiorite has a T_{CR} Nd model age of ca 2.1 Ga, corresponding to ε_{Nd} of approx. +2 at 1820 Ma, which indicates a relatively juvenile crustal origin without significant involvement of older, Archean crust.

A different granodiorite body was sampled at southeastern Boiley Lake in the southern Lynn Lake belt, south of the JSZ. Outcrop and map relationships indicate that the Boiley Lake granodiorite is intrusive into the Sickle Group metasedimentary rocks north of Boiley Lake. Preliminary zircon age data from the Boiley Lake granodiorite, however, suggest a crystallization age of ca. 1891 Ma. Although the timing of felsic volcanism in the southern Lynn Lake belt is poorly constrained at present and might be as young as ca. 1850 Ma (*see* above), the 1890 Ma felsic intrusive age is similar to the ca. 1892 Ma age of felsic volcanism in the northern Lynn Lake belt. The Boiley Lake granodiorite might represent a subvolcanic intrusion, a theory which could be tested using geochemistry. Alternatively, and similar to the Flin Flon Belt, the Boiley Lake granodiorite might form part of a previously unknown, evolved-arc assemblage that predates main accretion of the Lynn Lake belt. This corresponds to the change in the geochemistry of the mafic volcanic rocks south of the JSZ, which trend toward E-MORB, as opposed to the arc tholeiites that dominate the southern part of the belt, north of the JSZ (Zwanzig et al., 1999).

Timing of tectonometamorphism

In an effort to constrain the timing of the deformational events and the emplacement of gold mineralization, U-Pb dating has been applied to samples where the detailed structural relationships between deformational fabric elements and intrusive units are known. This, in turn, constrains the timing of gold mineralization, as the emplacement of the associated alteration assemblages are hosted by D_2 shear zones and appear to result from syn-D_2 hydrothermal activity, which predates the peak of regional metamorphism and D_4 (Beaumont-Smith and Rogge, 1999).

The age of D_2 is constrained by a leucogranitic pegmatite at Zed Lake (northern margin of the northern Lynn Lake belt), which formed during (the late stage of) D_2, the main regional phase of deformation. The U-Pb zircon and monazite age data suggest that the Zed Lake leucogranitic pegmatite was emplaced at 1815 ± 3 Ma and subsequently metamorphosed at ca. 1783 ± 3 Ma.

A sample of the syn-D_4 tonalite dikes from the southwestern Lynn Lake belt, in comparison, yielded rare zircons with an age of 1758 ± 8 Ma. Together with the 1819 ± 1 Ma age of pre- or syn-D_2 granodiorite injection in the southeastern Lynn Lake belt (*see* above), it is most likely that the age of D_2 is ca. 1815 to 1819 Ma, and that the Zed Lake leucogranitic pegmatite and the tonalite dikes were formed or metamorphosed at ca. 1785 and 1758 Ma, respectively, prior to or during D_4.

To date, the age of regional metamorphism is less well constrained. The present best estimate comes from the One Island Lake rhyodacite sample (southern belt), which yielded a preliminary U-Pb age from a metamorphic zircon crystal of ca. 1814 Ma, in addition to the ca. 1842 Ma primary zircon age. These preliminary data suggest that the peak of regional metamorphism is contemporaneous with, or slightly younger than, the timing of D_2, possibly coincident with the timing of gold mineralization. The persistent recrystallization of D_2 fabrics supports the latter. The younger, possibly metamorphic zircons from the Zed Lake area may indicate one or more metamorphic events of probable contact origin associated with the emplacement of ca. 1784 Ma or younger plutons.

Sediment provenance and timing of sedimentation

Paragneiss samples from the dominantly turbiditic, volcanic-derived Wasekwan Group and Zed Lake greywacke, and from the dominantly fluvial-alluvial meta-arkosic to arenitic Sickle Group, were studied.

The Sm-Nd isotope and U-Pb zircon results indicate that a Sickle Group meta-arkose sample from the southwestern Lynn Lake belt is composed mainly of less than 2.0 Ga (1.85–1.9 Ga) and smaller amounts of pre-2.0 Ga (2.4–2.5 Ga and/or 2.6–2.7 Ga) detritus (T_{CR} Nd model age of ca. 2.3 Ga, corresponding to ε_{Nd} of +0.3 at 1850 Ma), and was probably metamorphosed at ca. 1830 Ma (platy, metamorphic zircon). Accordingly, the Sickle Group is older than 1830 Ma. The provenance of the Sickle Group is dominated by detritus shed from the Lynn Lake belt, but includes small amounts of older detritus, probably derived from crystalline rocks exposed in the hinterland of the orogen.

A Nd model age of ca. 2.1 Ga and a relatively juvenile, initial ε_{Nd} of approximately +3 for a metagreywacke from the Zed Lake area suggest source rocks that are dominantly less than 2.0 Ga but a lower proportion of pre-2.0 Ga source rocks than in the Sickle Group meta-arkose. Sensitive high-resolution ion microprobe (SHRIMP) U-Pb dating of zircons from the Zed Lake metagreywacke sample (Rayner, pers. comm., 2002) reveals that the bulk of the zircons analyzed (30 out of 33 grains) have ages around 1880 Ma. A number of zircon grains are younger than 1880 Ma, likely due to Pb loss, since they do not appear to be metamorphic. Only three zircons were found to have older ages (in the range 2300 to 2400 Ma). The dominant ca. 1880 Ma provenance suggests that the Zed Lake greywacke represents the deposition of detritus shed from the Wasekwan Group, which made up the northern Lynn Lake belt, and is older than the Burntwood Group metasedimentary rocks.

A Wasekwan Group turbiditic meta-arenite sample from the southeastern Lynn Lake belt has a juvenile Sm-Nd isotopic composition (T_{CR} Nd model age of ca 2.1 Ga corresponding to ε_{Nd} of +4 [mantle Sm-Nd isotopic composition] at ca. 1890 Ma). This is consistent with a predominantly ca. 1.89 Ga 'Wasekwan' volcanic provenance.

DISCUSSION

The relationship between the assembly of the volcanic tectonostratigraphy, at least four periods of synkinematic felsic magmatism between ca. 1876 and 1784 Ma, and two major periods of sedimentation describes a very dynamic, evolving tectonic regime (Fig. GS-19-4). Although many of the data reported here are considered preliminary, the emerging tectonic framework involves the accretion of two, possibly very different arcs to the Hearne Craton margin. The depleted arc-tholeiite and E-MORB volcanic geochemistry that dominates the rocks of the northern belt represents a possible intra-arc rift, whereas the arc-tholeiite and calc-alkaline volcanic geochemistry is typical of a juvenile island-arc setting (Zwanzig et al., 1999).

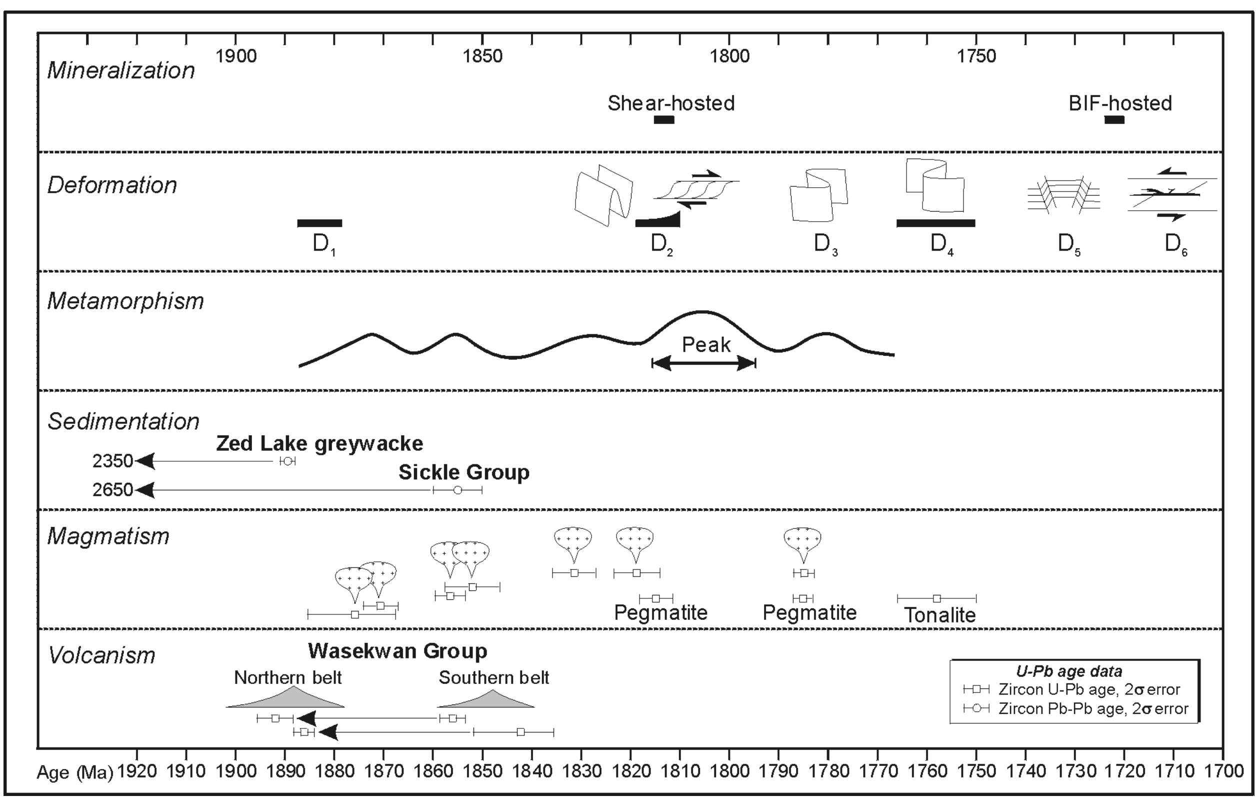

Figure GS-19-4. Time chart showing the relationship between volcanism, magmatism, sedimentation, deformation and metamorphism in the Lynn Lake greenstone belt. Abbreviation: BIF, banded iron formation.

The most significant aspect of the new data is the potential complexity involved in the assembly of the southern Lynn Lake belt. Contact relationships between the volcanic rocks and the Pool Lake suite suggest the presence of older (ca. 1890 Ma) volcanic rocks. However, U-Pb dating of the southern belt has yet to yield ages older than ca. 1855 Ma. The structural intercalation of MORB-like basaltic rocks of the Granville Lake Structural Zone in the western portion of the southern belt (Zwanzig, 2000) and transitional E-MORB basaltic rocks along the southern margin of the belt in the Wilmot Lake area suggest that the southern belt may represent a tectonic collage, assembled prior to or during accretion to the Hearne margin. This presumably occurred after ca. 1850 Ma, the youngest age of felsic volcanism in the southern belt.

The post-assembly history of the greenstone belt correlates well with the tectonic model for the northern Trans-Hudson Orogen proposed by Maxeiner et al. (2001). The focus of continued research is directed at unravelling the early history of the assembly of the volcanic tectonostratigraphy and the magmatic history of the Lynn Lake greenstone belt. This will enable the correlation of lithotectonic domains across the northern Trans-Hudson Orogen.

CONCLUSIONS

Continuing structural analysis, Sm-Nd isotopic studies and U-Pb geochronological studies of the Lynn Lake greenstone belt decipher a complex tectonic history documenting the assembly of an intricate tectonostratigraphy, comprising juvenile and contaminated volcanic-arc rocks. The northern Lynn Lake belt is mainly composed of weakly contaminated tholeiitic rocks deposited ca. 1886 to 1892 Ma, whereas the southern belt consists, in part, of juvenile tholeiitic and calc-alkaline rocks, some of which were probably deposited ca. 1856 to 1842 Ma. The main volcanic tectonostratigraphy was assembled during D_1, prior to the emplacement of ca. 1875, 1857 and 1830 Ma successor plutonic rocks, and prior to the deposition of fluvial-alluvial conglomerate and arkose along the southern margin of the greenstone belt.

The Lynn Lake greenstone belt experienced intense D_2 deformation that resulted in east-west transposition and the development of dextral transpressional shear zones at ca. 1819 Ma. Development of D_2 shear zones is penetrative throughout the greenstone belt, with the delineation of several major D_2 shear zones in the western portion of the greenstone belt. The D_2 major deformation is also responsible for the development of shear zone–hosted gold mineralization, which slightly predates the peak of regional metamorphism, estimated at ca. 1814 Ma. Accordingly, gold mineralization formed during prograde metamorphic conditions.

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