Source and tectono-metamorphic evolution of mafic and pelitic metasedimentary rocks from the central Quetico metasedimentary belt, Archean Superior Province of Canada

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Abstract

A study of the sedimentary rocks in the Jean Lake area of the Quetico metasedimentary belt, Superior Province, Canada, was conducted in order to evaluate the origin, source, and evolution of rocks, including mafic rocks previously mapped as ultramafic rocks. Bulk compositions of these sedimentary rocks show a mixing between two end members: quartzo-feldspathic sedimentary rock and komatiitic basalt. High CaO and MgO contents of the rocks suggest a proximal source of the komatiitic basalt.

The rocks in the study area record a pressure-temperature (P–T) path with three tectono-metamorphic stages. The first stage formed staurolite (500–700 °C) under medium P–T (MP–MT) metamorphic conditions shortly after the sedimentation. The second stage yielded the biotite–sillimanite–garnet assemblage under the peak conditions of 0.6 ± 0.1 GPa and 700 ± 70 °C during transpressional deformation. The third stage, low P–medium T (LP–MT; 0.37 ± 0.1 GPa, 540 ± 80 °C) metamorphism, was associated with regional south-southeast compression and its timing is constrained by a new U–Th–Pb monazite age of 2667 ± 20 Ma. Combining the regional deformation events, we suggest the burial metamorphism of sediments up to MP–MT conditions during the D1 deformation at 2698–2689 Ma. They attained the peak metamorphic conditions during the regional transpressive D2–D3 deformation (2689–2671 Ma), and retrograded to LP–MT condition during the south-southeast compression (regional D4) at 2671–2667 Ma.

A short time span between the sedimentation and MP–MT metamorphism accompanied by D1 deformation suggest that the Quetico sedimentary rocks formed in an acretionary prism. The studied rocks in the Jean Lake area deposited close to the Wabigoon Subprovince, transported towards the south, and buried up to ∼20 km by rapid underthrusting to amphibolite facies conditions. Subsequent dextral transpressive deformation (D2–D3) during the oblique docking of the Quetico belt to the Wabigoon Subprovince to the north and Wawa greenstone belt from the south resulted in the displacement of the sedimentary rocks to the west from the original depositional site. The P–T–time path of the Quetico sedimentary rocks is similar to that in modern arc accretion prisms, except for a high geothermal gradient of ∼30 °C/km recorded in the former compared to ∼10 °C/km.

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in modern counterparts. The high temperature gradient in the Archean accretionary prism explains the lack of high-pressure metamorphic rocks, such as blueschist, that are common in modern accretionary prisms.

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Keywords: Archean tectonics; Archean accretionary prism; Provenance; Geochemistry of clastic sedimentary rocks; Geothermal gradient; Archean subduction zone; Monazite geochronology

1. Introduction

The Archean Superior Province contains linear arrays of granite-greenstone belts and metasedimentary belts. Successive accretion of volcanic arcs in late Archean time (Card, 1990; Thurston and Chivers, 1990; Fig. 1) is considered as a likely process for the formation of the Archean craton. This is supported by “frozen subduction zones” shown in seismic reflection profiles across the Canadian Shield (Ludden et al., 1993; Calvert et al., 1995; Cook et al., 1999). Metasedimentary belts between greenstone belts may represent accretionary prisms developed during subduction and collision of arcs (Percival, 1989; Card, 1990; Williams, 1990).

The Quetico Subprovince is one such metasedimentary belt in the western Superior Province and is bounded by two volcanic belts: the Wabigoon Subprovince to the north and by the Wawa Subprovince to the south (Figs. 1 and 2). The Quetico Subprovince consists mostly of turbiditic Qtz-rich metasedimentary rocks with minor banded iron formations. In this monotonous sedimentary belt, a lens of “ultramafic rocks”, ∼2 km long and 400 m wide, crops out around Jean Lake, north of Lake Superior (Williams, 1988, 1991; Fralick et al., 1992; Fig. 2).

![Fig. 1. Simplified geological map of the Superior Province with the location of the study area shown in Fig. 2, modified after Card (1990).](image-url)
work suggested that this lens formed by weathering of ophiolites or serpentinite diapirs (Williams, 1988, 1990; Fralick et al., 1992). Ophiolites and serpentinite diapirs are common in modern convergent margins (Nicolas, 1989; Fryer et al., 1995), but their occurrence of Archean age is in debate (e.g. Hamilton, 1998; De Wit, 1998). Therefore, the “ultramafic” unit in the Quetico belt may provide information relevant to Archean tectonics, and may shed additional light on the evolution of the Quetico metasedimentary belt. In addition, the deformation of the Quetico sedimentary rocks are poorly dated and the prograde path of their metamorphic evolution has not been studied. Such information will contribute to better understand-
ing of the origin of the belt. Consequently, a study of the “ultramafic” rocks was conducted with four objectives: (i) to characterize the “ultramafic” rocks, (ii) to evaluate their origin and source, (iii) to examine the structural and metamorphic evolution of the “ultramafic” and surrounding rocks, and (iv) to compare the Quetico belt with modern accretionary prisms.

2. Geological setting

The Quetico Subprovince is a linear belt of dominantly Qtz-rich, turbiditic metagraywackes (Ojakangas, 1985) that has a relatively consistent width of about 70 km and extends approximately 1200 km from longitude 70°–96° W (Williams, 1991). The timing of sedimentation is constrained by the youngest U–Pb age of zircon in the sedimentary rocks and the oldest age of intrusions. These ages are significantly different between the northern and southern parts of the belt. The youngest age of zircon in the northern part is 2698 ± 3 Ma, whereas that in the southern part is 2688.5 ± 1.5 Ma (Zaleski et al., 1999; Davis et al., 1990). The oldest U–Pb zircon intrusion age is 2696 ± 2 Ma in the northern Quetico belt and 2670.5 ± 2 Ma for granitic intrusions in the south Quetico belt (Percival and Sullivan, 1988; Percival, 1989; Williams, 1991). The data suggest that the southern part deposited significantly later than the northern part.

Igneous rocks are minor in the Quetico belt. They include rare felsic volcanic rocks (Williams, 1991), tonalitic–granodioritic intrusions dated at 2696 Ma (Davis et al., 1990; Zaleski et al., 1999), a suite of carbonate-bearing alkaline complexes of 2680 ± 1 Ma (Hattori and Percival, 1999; Lasen et al., 2000), and large aluminous granites dated between ~2670 and ~2650 Ma (Percival, 1989; Williams, 1991).


3. Occurrence of mafic rocks in the Jean Lake area

3.1. Distribution and lithology

The “ultramafic” lens is well exposed on the shores and on small islands in the Jean Lake (Fig. 2). Rocks contain millimeter-size clasts of quartz aggregates and exhibit sedimentary textures, including climbing ripples, flame structure, loading, and stratification (Fig. 3a–c). These observations suggest a sedimentary origin of these rocks. Bedding and foliation strike east, dip mainly south (between 85° north and 50° south) and the “ultramafic” lens is surrounded by quartz–feldspathic metasedimentary rocks similar to those in the remainder of the Quetico belt.

The quartz–feldspathic rocks consist of two lithological units, a volumetrically dominant semi-aluminous unit (45–25 vol.% of Qtz, 50–30 vol.% of Pl, 35–10 vol.% of Bt, <35 vol.% of Chl, <10 vol.% of Ms, and <10 vol.% of Grt, abbreviations are from Kretz, 1983) and a minor aluminous unit (40–25 vol.% of Qtz, 40–30 vol.% of Pl, 30–20 vol.% of Bt, 30–5 vol.% of St, <10 vol.% of Chl, <5 vol.% of Ms, and <5% of Grt). The “ultramafic lens” contains three units: Bt–amphibole-rich unit, amphibole-rich unit, and felsic unit. All units contain more than 10 vol.% of felsic minerals (Qtz and feldspars), and are thus, mafic rather than ultramafic. The biotite–amphibole-rich unit is made of 40–30 vol.% of Qtz, 50–70 vol.% of Pl, 5–10 vol.% of amphibole, 40–5 vol.% of Bt, and <40 vol.% of Chl. The amphibole-rich unit contains 90–40 vol.% of amphibole, 20–5% of Qtz, 20–5% of Pl, and <10 vol.% of Chl.

The felsic unit is identical to the surrounding quartz–feldspathic rocks. Unit names instead of rock names are used in this manuscript because the unit names can best characterize the studied rocks which are highly heterogeneous in mineral abundance.

The decameter to meter lenses of amphibole-rich rock unit occurs parallel to the bedding and form high relief on the weathered surface of the hosting, biotite–amphibole-rich unit. South and north of the Jean Lake, sedimentary rocks are cut by granitic dykes and sills, which originated from the voluminous, two mica granite of 2670–2650 Ma (Percival, 1989; Williams, 1991).
3.2. Chemical composition and source of the mafic sedimentary rocks

The mafic sedimentary rocks show significant compositional variations (Table 1). The contents of SiO$_2$, Al$_2$O$_3$, MgO, and Cr vary from 53.9 to 65.5 wt.%; 9.8 to 17.9, 3.0 to 12.5 wt.%, and 149 to 866 ppm, respectively (Table 1). The contents of Al, Ti, Mg, and Cr form linear arrays with high correlation coefficients ($r > 0.95$), suggesting that they were immobile after sedimentation (Rollinson, 1993). Linear arrays of Ca, Ga, V, Zn, Co, and Ni against these immobile elements suggest that they too were relatively immobile.

The plot of two immobile elements, such as TiO$_2$ versus Al$_2$O$_3$, shows a linear array between the felsic and amphibole-rich units and the biotite–amphibole-rich unit plot between the two units (Fig. 4).

There are three possible causes to form the linear arrays: constant sum effect, hydraulic sorting, and mixing of two end-members. The constant sum effect is rejected because elemental ratios, such as Ti/Mg and Al/Mg, also show linear arrays (Fig. 5a). Hydraulic sorting from a single source was suggested for the cause of compositional variation of sedimentary rocks from the area between the Beardmore and Jean Lake area (Fralick and Kronberg, 1997). The high Cr contents (up to 866 ppm) imply contribution of chromite to the sedimentary rocks. Moreover, the positive correlation between Cr and Ca contents requires contributions of Ca-minerals and chromite with similar proportion to the sediments; an unlikely condition because these two minerals have very different densities.

![Fig. 3](image)

(a) Sedimentary structures of the quartzo-feldspathic turbiditic rock. Pen is 14 cm long. Bedding in the middle section under the erosion surface, cross lamination in the top portion of the photograph, and grain size sorting in the bottom part of the outcrop. (b) Loading structure, bedding, and sedimentary grading in the biotite–amphibole-rich unit and amphibole-rich unit. (c) Photomicrograph showing a quartz-rich clast in the amphibole-rich unit. This section was made in the amphibole-rich unit of the sample shown in (b).

![Fig. 4](image)

Fig. 4. TiO$_2$ vs. Al$_2$O$_3$ wt.% of mafic rocks in the Jean Lake area. Circles correspond to the rocks from the biotite–amphibole-rich unit; triangle: rocks from the amphibole-rich unit; square: felsic unit in the mafic lens and quartzo-feldspathic rocks surrounding the mafic lens.
Table 1

Bulk chemical composition of representative rocks from the Jean Lake area

<table>
<thead>
<tr>
<th></th>
<th>Amphibole-rich unit</th>
<th>Biotite–amphibole-rich unit</th>
<th>Felsic unit and host rocks</th>
<th>Kom. basalt</th>
<th>Harz</th>
<th>Lherz</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>58.20</td>
<td>59.19</td>
<td>54.43</td>
<td>63.14</td>
<td>57.10</td>
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<td>13.17</td>
<td>13.00</td>
<td>11.84</td>
<td>14.13</td>
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<td>Fe₂O₃</td>
<td>6.65</td>
<td>6.33</td>
<td>9.21</td>
<td>7.23</td>
<td>6.64</td>
<td>8.64</td>
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<td>MnO</td>
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<td>0.16</td>
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<tr>
<td>MgO</td>
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<td>2.08</td>
<td>2.08</td>
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<tr>
<td>CaO</td>
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<td>3.28</td>
<td>3.28</td>
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<tr>
<td>K₂O</td>
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<td>Na₂O</td>
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<td>3.08</td>
<td>2.04</td>
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<td>P₂O₅</td>
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<td>0.19</td>
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<tr>
<td>LOI</td>
<td>1.80</td>
<td>1.30</td>
<td>1.90</td>
<td>0.50</td>
<td>4.70</td>
<td>1.30</td>
</tr>
</tbody>
</table>

Oxides are in wt.% and other elements in ppm.

* Total Fe expressed as Fe₂O₃.

** Loss on ignition.

Uranium contents were all below detection limit of 1 ppm.

Kom. basalt: komatiitic basalt from Wabigoon (Ayer, 1999).

Harz: harzburgite from the Mariana fore arc (Yamamoto et al., 1992).


Note: All values are normalized to 100%.

Fig. 5. (a) Weight ratios of Al₂O₃/MgO vs. TiO₂/MgO for rocks from the mafic lens in the Jean Lake area. The regression line (solid line) of sedimentary rocks from the mafic lens compared to various mixing lines between an average composition of the Quetico quartzofeldspathic sedimentary rocks (host rocks and felsic-unit rocks) and possible ultramafic and mafic rocks; komatiitic basalt (dashed line; Ayer, 1999), komatiite (dash-dot line; Ayer, 1999) in the southern Wabigoon Subprovince, harzburgite (dash-double-dot line; Yamamoto et al., 1992) and lherzolite (dotted line; Ballantyne, 1992). (b) Weight ratios of CaO/TiO₂ vs. Cr/Ti for rocks from the mafic lens in the Jean Lake area. Line types same as in (a).

Linear correlations of elements are thus attributed to the mixing of end-member components. The amphibole-rich unit contains high MgO (up to 12.5 wt.%), Cr (up to 866 ppm), and high PGE (20 ppb Pt, 17 ppb Pd, 3 ppb Ir). These values are all greater than those of most basaltic igneous rocks, suggesting a contribution from more Mg-rich rocks such as komatiitic basalts, komatiites, and mantle peridotites. The latter two possibilities are rejected because of too low CaO contents of komatiites and mantle peridotites to account for our samples (Fig. 5b). This leaves komatiitic basalts as the source. The amount of the komatiitic basalts is calculated to be nil in the felsic unit and up to ~90% in the amphibole-rich unit using the lever rule (Fig. 5b).

4. Deformation in the Jean Lake area and correlation with regional deformation

Four tectono-metamorphic events have been recognized in the Quetico Subprovince (Williams, 1991, and references therein). The first regional deformation (D1) included slumping, and recumbent folding shortly after sedimentation (2698 to ~2690 Ma; Table 4; Sawyer, 1983; Williams, 1991; Zaleski et al., 1999). This event was followed by mainly dextral strike-slip shearing (D2) which resulted in the formation of a Subprovince scale west-trending vertical planar fabric, and lineations plunging east at 10–30° (Williams, 1991). The subsequent deformation was also transpressive (D3), and produced upright folds (F3) that affected both the bedding and the earlier-formed planar fabrics (Sawyer, 1983; Williams, 1991). The D4 deformation corresponds to minor shearing under a south-southeast compression (Sawyer, 1983).

The study area shows only three stages of deformation. The first stage of deformation in the Jean Lake area produced a foliation defined by Bt and St. The foliation was later folded during the main deformation event in the area, producing decimeter-scale tight folds with an axial trace of N95°S. The later foliation, which is defined by Bt, Grt, and Sil, is locally oblique to this main axial surface direction in hinges of larger-scale upright folds (Fig. 2). Similar upright folds are common in the Quetico belt and considered to have formed under the regional transpressive D3 deformation (part 4; Williams, 1991), suggesting that this later foliation and upright folds in the Jean Lake area are most like a product of the regional D3 deformation. It implies that the earlier foliation in the Jean Lake area formed during either D1 or D2 regional deformation.

The final deformation in the Jean Lake area produced local shear zones, which are defined by Chl ± Bt ± amphibole, strike N110°, dip steeply to the south (Fig. 2). They cut earlier planar and folded fabrics (pre-S3 and S3) and bedding planes and east-trending pre-S3 and S3 foliations rotate into the shear planes, forming hectometer-scale dextral sigmoidal Z-folds, which also suggest the presence of dextral strike-slip motion are located on both flanks of F3 folds suggesting that they are not drag folds, but folds posterior to F3 (F4). Narrow (up to 1 cm width) Quartz veins...
show Z-folds inside the shear zones and the folding likely developed during this dextral shearing (Fig. 6a).
The geometry of centimeter-wide kink bands of Chl is also consistent with dextral shearing along a N110° direction. Pye (1964) identified N50° sinistral shear planes in the Jean Lake area (Fig. 2) and these planes may be a conjugate with the observed N110° dextral shear plane under south-southeast compression. Sawyer (1983) documented similar conjugate ductile shears and semi-brittle deformation features which were formed under a south-southeast compression in the Kashabowie area (~200 km south-east of the Jean Lake area). This defines the regional D4 deformation.

5. Metamorphism in the Jean Lake area

5.1. Mineralogy and mineral chemistry

We conducted a detailed thermobarometric study of the felsic and amphibole-rich units, which represent the end members of the compositional variations in the study area.

5.1.1. Aluminous unit

The rocks are defined by the following minerals: Qtz + Pl + Bt + St ± Grt ± Sil ± Chl ± Ilm ± Tur. Both Qtz and Pl, ranging from 0.05 to 0.1 mm in size, are anhedral in shape and Pl is homogenous in composition (An25–27). Bt (up to 2 mm in length) defines the main S3 foliation and is also consistent with dextral shearing along a N110° direction. Pye (1964) identified N50° sinistral shear planes in the Jean Lake area (Fig. 2) and these planes may be a conjugate with the observed N110° dextral shear plane under south-southeast compression. Sawyer (1983) documented similar conjugate ductile shears and semi-brittle deformation features which were formed under a south-southeast compression in the Kashabowie area (~200 km south-east of the Jean Lake area). This defines the regional D4 deformation.

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This unit consists of Qtz + Pl + Bt ± Grt ± Chl ± Ilm ± Tur. Anhedral Qtz (0.05–0.1 mm) is predominant and Pl crystals (0.05–0.2 mm) are anhedral and commonly contain Qtz inclusions. Individual grains do not show compositional zoning, with An0.26–0.31 and An0.16 in sample 244 and An0.22–0.25 in sample 262b (Table 2). Bt (0.1–0.3 mm) defines the main S3 foliation and re-crystallized in D4 shear bands (Fig. 6b). Some Bt surrounding Grt define delta micro-structures. Bt grains in contact with Grt have similar compositions as those in the matrix, with XFe of 0.48–0.49.

Euhedral Grt grains (0.3–1.2 mm) contain rounded Qtz inclusions and were rotated by the late dextral strike-slip shear deformation (D4), suggesting crystallization before or during D4. In sample 262b, Grt crystals (~1 mm) have Alm component ranging from 0.78 to 0.84, whereas Grt (~0.7 mm) in sample 244 contains between 0.67 and 0.73 Alm component (Table 2). Sps, Grs, and Prp components vary from 0.17 to 0.13, 0.06 to 0.04, and 0.12 to 0.10, respectively. The Grs and Sps components and XFe decrease from core to rim. The 80 μm-thick rims record a reverse trend; an increase in Sps and XFe and a decrease in Prp component (Fig. 7).

This bell-shaped zoning pattern suggests that the zoning is a result of crystal growth without subsequent modification by diffusion (Fig. 7). The reverse zoning in the outer rim implies a consumption of Grt during retrogression (Spear, 1993).

Anhedral Kfs crystals identified in one sample occur around Grt porphyroblasts in association with Chl; Spear (1993) suggested the following reaction:

\[ \text{Grt} + \text{Bt} + \text{H}_2\text{O} = \text{Kfs} + \text{Chl} \]
Fig. 6. (a) Photomicrograph showing a Z-folded quartz vein parallel to D4 shear planes in the semi-aluminous unit (sample 219). Thin section perpendicular to the shear plane. (b) Photomicrograph of a sample of the semi-aluminous unit (sample 262b). S2–S3 foliation defined by biotite is transposed into dextral C4 shear bands, and cut by C4 shear planes. Thin section perpendicular to the foliation plane. (c) Photomicrograph of a sample of the aluminous unit (sample 251). Staurolite (St) and earlier formed biotite (Bt) are folded by D3 event. A new foliation (S2–S3) defined by biotite (Bt2) develops parallel to the axial plane. Note the folded grain of St at the nose of the tight fold. (d) Photomicrograph of a sample of quartz-feldspathic unit (sample 251). Staurolite is replaced by garnet (Grt), biotite (Bt), and sillimanite (Sil).
Table 2

Representative mineral compositions used for the P-T estimate for the peak and LP-HT metamorphic events

<table>
<thead>
<tr>
<th>Sample</th>
<th>Garnet</th>
<th>Biotite</th>
<th>Plagioclase</th>
<th>St</th>
<th>Amiphole</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>H9251</td>
<td>251</td>
<td>262b</td>
<td>244</td>
<td>251</td>
<td>262b</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>H9253</td>
<td>244</td>
<td>Rim</td>
<td>251</td>
<td>262b</td>
<td>244</td>
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<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>H9273</td>
<td>268a</td>
<td>252a</td>
<td></td>
<td>244</td>
<td></td>
</tr>
</tbody>
</table>

| SiO₂  | 36.82  | 36.81  | 36.87       | 35.74| 35.38 |
| TiO₂  | 0.05   | 0.05   | 0.01        | 0.01| 0.06  |
| Al₂O₃ | 21.31  | 20.66  | 21.35       | 21.52| 21.61 |
| FeO   | 0.05   | 0.05   | 0.01        | 0.01| 0.06  |
| MgO   | 0.05   | 0.05   | 0.01        | 0.01| 0.06  |
| CaO   | 0.05   | 0.05   | 0.01        | 0.01| 0.06  |
| Na₂O  | 0.05   | 0.05   | 0.01        | 0.01| 0.06  |
| K₂O   | 0.05   | 0.05   | 0.01        | 0.01| 0.06  |

Total 784 analyses were obtained with the Cameca CAMEBAX SX100 electron microprobe of the Blaise Pascal University in Clermont-Ferrand. Operating conditions included 15 kV accelerating voltage, 10 nA beam current, 1–2 μm beam diameter, 20 s counting time. Standards are natural silicates (albite, wollastonite, orthoclase, quartz) for calibration of Na, Ca, K, Si, and oxides for calibration of Fe, Mg, Al, Ti, and Mn. Reported values are averages of three repeated analyses of the same point. Abbreviations: X alm = almandine component, X prp = pyrope component, X grs = grossular component, X sps = spessartine component, X ab = albite component, X an = anorthite component, X or = orthoclase component.

Footnotes:
(1) 251: staurolite-bearing unit; 262b and 244: staurolite-free unit; 268a and 252a: amphibole-rich unit.
(2) Total Fe expressed as FeO. Fe 2⁺ and Fe 3⁺ were calculated using the method described in Holland and Blundy (1994).
(3) spessartine reversal zoning.
(4) retrograde product.
Fig. 7. Compositional zoning of representative garnet crystals. Garnet grains in sample 262b and sample 244 are from the semi-aluminous unit, and garnet in sample 251 from the aluminous unit. Alm = 100 × Fe/(Fe²⁺ + Mg + Ca + Mn), Prp = 100 × Mg/(Fe²⁺ + Mg + Ca + Mn), Grs = 100 × Ca/(Fe²⁺ + Mg + Ca + Mn), Sps = 100 × Mn/(Fe²⁺ + Mg + Ca + Mn) (abbreviations are from Kretz, 1983).

Fig. 8. Photomicrograph showing amphibole crystals in samples 252a (a) and 268a (b). Black lines represent the compositional profiles reported in Figs. 9 and 10.
Chl crystallized along the dextral shear planes that formed during the south-southeast compression (D4) and replaces Bt under retrograde greenschist facies conditions.

5.1.3. Amphibole-rich unit

This unit consists of amphibole + Pl + Qtz + IIm + Chl + Tur. Euhedral to subhedral amphibole (0.05–1 mm), the predominant mineral (Fig. 8), does not display a preferred orientation. The amphibole grains plot near the boundary between magnesiohornblende and Tr, following the classification of Leake et al. (1997). They display an array of solid solutions between Tr–Ts and Tr–Prg (Fig. 9), suggesting a typical Act–Hbl exchange reaction (Spear, 1993). Individual grains in sample 268a show an enrichment in Hbl component whereas those in sample 252a show an enrichment of Tr component towards the rims (Fig. 10). The P–T evolution corresponding to the zoning pattern will be discussed in Section 5.3.

Pl grains are subhedral to anhedral in shape and range in size from 0.1 to 0.6 mm. They contain inclusions of Qtz and amphibole, and show Ca-increase towards the rims ranging from An0.32 to An0.38 in sample 252a and from An0.33 to An0.32 in sample 268a.

5.1.4. The felsic unit

This unit contains Qtz+Pl+IIm±Chl±Tur. It is similar in texture and mineralogy to the semi-aluminous unit, but this contains higher abundance of quartz than the latter.

5.2. Thermobarometric methods

The felsic unit contains low MnO and TiO2 contents and feldspar is the only Na- and Ca-bearing phase. Therefore, we used KFMASH petrogenetic grids of Spear and Cheney (1989), together with the aluminosilicate triple point by Richardson et al. (1969) to represent the metamorphic assemblages. We used two Fe–Mg exchange thermometry; Grt-St thermometry of Perchuk (1989) and Grt-Bt thermometry (GARB thermometer) of Pigage and Greenwood (1982) and Williams and Grambling (1990). The classical GARB thermometers of Ferry and Spear (1978) and Indares and Martignole (1985) were not used because Bt compositions in our samples are outside the range accepted for the thermometry (Bt with (AlV+Ti)/[(AlV+Ti)+Mg+Fe]<0.15). Furthermore, the Ti–Al substitution of Bt in our samples is different from Bt used by Indares and Martignole (1985). Pressures were estimated using the Grt–Pl–Si–Qtz (GASP) barometer of Hodges and Crowley (1985) and Koziol and Newton (1988), and the empirical Grt–Pl–Qtz (GPQ) barometer of Hoisch (1990). We also used the THERMOCALC computer program by Powell et al. (1998) in order to identify possible reactions among given end-member phases and to calculate the pressure and temperature of sub-systems.

The amphibole-bearing rocks contain low TiO2, MnO, and K2O and plot on the ACFM+SiO2 petrogenetic grid defined by Spear (1981). Complementary P–T estimates of the amphibole-rich unit were obtained using the thermometer based on the cation exchange reaction of edenite (Holland and Blundy, 1994). Thermobarometers based on the contents of Ti and Al(VI) in amphibole were not used because of the absence of Zo and Ep in our samples (Raase, 1974; Plyusnina, 1982; Hammarstrom and Zen, 1986).

5.3. Metamorphic conditions

5.3.1. Aluminous unit

The KFMASH P–T grid of Spear and Cheney (1989) suggests early formation of St at 500–700°C.
The second metamorphic stage is characterized by the assemblage of Sil + Grt + Bt, suggesting a metamorphic condition on the high temperature side of Reaction 1, greater than 600 °C (Spear and Cheney, 1989; Fig. 11). The absence of Kfs and the occurrence of Sil suggest temperatures lower than 800 °C and pressures between 0.2 and 1 GPa. Sample 251 yielded 628 ± 35 °C and 0.64 ± 0.16 GPa using the THERMOCALC program. Taking the uncertainties into consideration, these results are comparable to the estimated P–T based on mineral assemblages (Fig. 11). Furthermore, the values are consistent with the temperature, 600 ± 70 °C, estimated using the GARB thermometer, and also to the 625 ± 50 °C and 0.54 ± 0.09 GPa using the independent GS thermometer and the GASP barometer (Table 3a).

5.3.2 Semi-aluminous unit

The assemblage of Grt + Bt + Pl suggests a metamorphic temperature greater than 500 °C (Spear and Cheney, 1989), as Grt can not crystallize on the low temperature side of the reaction:

\[ \text{Fe–Cld + Ann + Qtz = Alm + Ms + H}_2\text{O} \quad (3) \]

In sample 262b, the GARB thermometer and GPQ barometer on Grt cores yielded 610 ± 70 °C and 0.56 ± 0.12 GPa. In sample 244, Grt cores yielded 585 ± 70 °C and 0.71 ± 0.08 GPa, whereas rims with Pl and Bt show a retrograde condition of 540 ± 80 °C and 0.25 ± 0.11 GPa (Table 3b).

5.3.3 Amphibole-rich unit

The absence of clinopyroxene suggests a maximum temperature of 780 °C, following the reaction on the ACFM + SiO₂ petrogenetic grid (Fig. 11; Table 3). The Hbl form under amphibolite facies conditions. Therefore, the zoning from Hbl cores to Tr rims (sample 252a) likely reflects the retrograde path. The Hbl core and co-existing Pl yield a temperature of 700 ± 70 °C.

Amphibole grains without any evidence of retrogression in sample 268a record a prograde path, in-
Fig. 11. Solid lines and curves show P–T grid for pelites in the KFMASH system (modified from Spear and Cheney, 1989). Dash-dot and dashed curves represent the disappearance of Hbl and appearance of Cpx in mafic rocks in the ACFM+SiO₂ system (modified from Spear, 1981). Dash-dot curves are under fo₂ conditions buffered by quartz–fayalite–magnetite and dashed curves under fo₂ conditions buffered by hematite–magnetite. “Cpx” refers to clinopyroxene and other abbreviations of minerals are from Kretz (1983). Shaded fields with 1, 2, and 3 correspond to the P–T condition based on the mineral assemblages formed during the three tectono-metamorphic events in the Jean Lake area (M1–D1, M3–D3, and M4–D4, respectively; Table 4). Hatched area with 1–3 corresponds to the P–T stability field where metamorphic assemblages 1 and 3 are both stable. Solid circles are the estimated P–T values of the M3 and M4 events. Note that the P–T fields defined by metamorphic mineral assemblages agree well with the P–T values calculated by mineral assemblages. The error bars represent uncertainties of geothermobarometric calculations (Table 3a and b).

Increasing Hbl component towards the rims. The rims yielded a maximum temperature of 600 ± 70 °C using the Hbl–Pl thermometer.

In summary, the rocks in the Jean Lake area underwent a metamorphic event starting with the crystallization of St in the aluminous unit and the increasing Hbl component in amphiboles in the amphibolite unit. The peak metamorphic P–T, 0.61 ± 0.10 GPa and 700 ± 70 °C, was attained during the D3 regional transpressive deformation. This was followed by retrograde metamorphism forming Chl and Tr during D4 south-southeast compression at 0.25 ± 0.11 GPa, 540 ± 80 °C (LP–MT event).

6. Tectono-metamorphic evolution of the study area and the Quetico belt

The first regional deformation (D1) took place shortly after sedimentation and involved burial of sediments, producing moderate P–moderate T (MP–MT) metamorphism (Tabor et al., 1989; Pan and Fleet, 1999). This tectono-metamorphic event was followed by mainly strike-slip deformation (D2). This deformation and subsequent deformation (D2–D3) mainly involved strike-slip deformation, implying no significant change in P conditions. Therefore, the crystallization of St and Hbl-rich amphiboles under the MP–MT metamorphism in the Jean Lake area occurred before the regional D2 transpressional deformation. It is most likely related to the regional D1.

The second stage of deformation in the Jean Lake area, which corresponds to regional D2–D3, is accompanied by peak metamorphism conditions. The final deformation in the Jean Lake area includes local shearing during south-southeast compression, accompanied by LP–MT metamorphism. The metamorphic condition, 0.25 ± 0.11 GPa and 540 ± 80 °C in the Jean Lake area, is in agreement with the conditions expected from the Subprovince-wide metamorphism. There is a systematic increase in metamorphic conditions, from 0.2 GPa and 500 °C at the USA–Canada border in the west to 0.5–0.6 GPa and 780 °C in the eastern part of the Quetico belt (Pirie and Mackasey, 1978; Percival,
Table 3
(a) Thermobarometric estimates of different lithological units from the Jean Lake area

<table>
<thead>
<tr>
<th>Unit</th>
<th>Sample</th>
<th>Assemblage</th>
<th>Tr-Ed</th>
<th>P-T Grid$^a$</th>
<th>GS$^b$</th>
<th>GAR$^b$</th>
<th>GASP$^b$</th>
<th>GPI$^c$</th>
<th>THERMOCALC$^d$</th>
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<td></td>
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<td></td>
<td></td>
<td>T (°C)</td>
<td>T (°C)</td>
<td>T (°C)</td>
<td>P (GPa)$^e$</td>
<td>T (°C)</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>St-free</td>
<td>244</td>
<td>Grt + Bt +</td>
<td></td>
<td></td>
<td>585 ± 70</td>
<td>0.71 ± 0.08</td>
<td>Insufficient</td>
<td>Insufficient</td>
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<tr>
<td>St-free</td>
<td>262b</td>
<td>Grt + Bt +</td>
<td>Stable</td>
<td></td>
<td>610 ± 70</td>
<td>0.56 ± 0.12</td>
<td>Insufficient</td>
<td>Insufficient</td>
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<tr>
<td></td>
<td></td>
<td>Pl</td>
<td></td>
<td></td>
<td>[n=5]</td>
<td>[n=6]</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>St-bearing</td>
<td>251</td>
<td>Grt + Bt +</td>
<td>PI + Si +</td>
<td></td>
<td>625 ± 50</td>
<td>0.59 ± 0.11</td>
<td>628 ± 35</td>
<td>0.64 ± 0.16</td>
<td>625 ± 50</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pl</td>
<td>St</td>
<td></td>
<td>[n=6]</td>
<td>[n=3]</td>
<td></td>
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<tr>
<td>Amp-rich</td>
<td>252a</td>
<td>Amp + Pl</td>
<td>700 ± 70</td>
<td></td>
<td></td>
<td></td>
<td>Insufficient</td>
<td>Insufficient</td>
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<td>[n=5]</td>
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<tr>
<td>Amp-rich</td>
<td>268a</td>
<td>Amp + Pl</td>
<td>600 ± 70</td>
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<td>Insufficient</td>
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<td>[n=5]</td>
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<tr>
<td>Best estimates of the peak metamorphism$^y$</td>
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<td></td>
<td></td>
<td>Insufficient</td>
<td>Insufficient</td>
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</tbody>
</table>

(b) Thermobarometric estimates of the LP-HT metamorphism of the Jean Lake area

<table>
<thead>
<tr>
<th>Unit</th>
<th>Sample</th>
<th>Assemblage</th>
<th>Tr-Ed</th>
<th>P-T Grid$^a$</th>
<th>GS$^b$</th>
<th>GAR$^b$</th>
<th>GASP$^b$</th>
<th>GPI$^c$</th>
<th>THERMOCALC$^d$</th>
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<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>T (°C)</td>
<td>T (°C)</td>
<td>T (°C)</td>
<td>P (GPa)$^e$</td>
<td>T (°C)</td>
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</tr>
<tr>
<td>St-free</td>
<td>244</td>
<td>Grt + Bt +</td>
<td></td>
<td></td>
<td>540 ± 80</td>
<td>0.25 ± 0.11</td>
<td>Insufficient</td>
<td>Insufficient</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pl</td>
<td></td>
<td></td>
<td>[n=4]</td>
<td>[n=3]</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

$^a$ the number of mineral pairs used for the calculations is in parentheses.
$^b$ the grid by Spear and Cheney (1989).
$^c$ Garnet–staurolite thermometer (Perchuk, 1989).
$^d$ Uncertainty considers the discrepancy of the calculated results among different pairs and the uncertainties (at 2σ-level) related to the equation given by the authors of the original papers.
$^e$ Uncertainties in pressure estimates correspond to the standard deviation (2σ) with the average previously calculated temperature.
$^h$ THERMOCALC computer program (Powell et al., 1998).
$^i$ pressure is an average of all estimates and the temperature is that of the sample 252a using the Tr-Ed thermometer.

Abbreviations: Amp = amphibole, Bt = biotite, Cpx = clinopyroxene, Grt = garnet, Kfs = K-feldspar, Ky = kyanite, Pl = plagioclase, Sil = sillimanite, St = staurolite, Tr = tremolite.
Table 4
Timing of sedimentation, deformation (D) and metamorphism in the Quetico metasedimentary belt

<table>
<thead>
<tr>
<th>Events</th>
<th>Previous studies</th>
<th>This study</th>
<th>Proposed timing</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sedimentation</td>
<td>2698 to &lt;2690 Ma (1,2)</td>
<td>2698 to &lt;2690 Ma</td>
<td></td>
</tr>
<tr>
<td>D1</td>
<td>2698–2689 Ma</td>
<td></td>
<td>2698–2689 Ma</td>
</tr>
<tr>
<td>D2</td>
<td>2689–2684 Ma (3), ? to &lt; 2666 Ma (4)</td>
<td>2689–2684 Ma</td>
<td></td>
</tr>
<tr>
<td>D3</td>
<td>2684–2671 Ma</td>
<td></td>
<td>2671–2667 Ma</td>
</tr>
<tr>
<td>D4</td>
<td>2667 ± 20 Ma</td>
<td></td>
<td>2671–2667 Ma</td>
</tr>
<tr>
<td>MP–MT metamorphism (M1)</td>
<td></td>
<td>Synchronous with D1</td>
<td></td>
</tr>
<tr>
<td>Peak of metamorphism (M3)</td>
<td></td>
<td>Synchronous with D3</td>
<td></td>
</tr>
<tr>
<td>LP–MT metamorphism (M4)</td>
<td>2671–2667 Ma (5), synchronous with D2 (4)</td>
<td>2671–2667 Ma</td>
<td></td>
</tr>
</tbody>
</table>

References: (1) Davis et al. (1990); (2) Zaleski et al. (1999); (3) Percival (1989); (4) Pan et al. (1998); (5) Percival and Sullivan (1998).

In 1989), Greenschist facies retrograde metamorphism is recorded at the end of D4 deformation in the Jean Lake area. Similar retrograde metamorphism is observed in many other places in the Quetico belt (Percival, 1989; Pan et al., 1994; Pan and Fleet, 1999). The D3 and D4 are described in Sawyer (1983) and Williams (1991).

The timing of these deformation events and their relationships with regional metamorphism are in debate. Percival (1989 and references therein) suggested that regional D2 deformation is bracketed between 2689 and 2684 Ma and the LP–MT event between 2671 and 2667 Ma based on ages of intrusive rocks and detailed structural study in Shebandowan and adjacent Quetico belt (Fig. 2). Pan et al. (1998) proposed a much later date for the regional D2 based on 2666 ± 1 Ma pegmatitic rocks concordant to the penetrative S2 foliation. This would imply that the D2 deformation occurred during the regional metamorphism dated at 2670–2650 Ma (Percival, 1989).

This is not consistent with our data from the Jean Lake area, where the LP–MT metamorphism was synchronous with the D4 deformation. To confirm this time relationship, we dated monazite that crystallized during the LP–MT metamorphic stage (D4).

7. Dating of monazite

Thirty-six grains of monazite were examined in sample 251. Monazite shows no compositional zoning, contains high U (up to 0.8 wt.%), Th (up to 5 wt.%), and light rare earth elements (REE). Chondrite-normalized REE patterns of all grains are similar, suggesting a single generation of monazite.

8. Discussion

8.1. Pressure–temperature–time path

The age of 2667 ± 20 Ma for the D4 and LP–MT metamorphism is in good agreement with the time
Fig. 13. Weighted histogram of the age of monazite. The age \( t \) is calculated from the U, Th, and Pb concentrations of the sample, assuming that Pb is a radiogenic decay product.

\[
Pb = \frac{\frac{Th}{232} \exp(\lambda_{232} t) - 1}{208} + \frac{U_{238}}{238} \cdot 0.040992 \exp(\lambda_{238} t) - 1} + \frac{U_{238}}{238} \cdot 0.040992 \exp(\lambda_{238} t) - 1}
\]

where Pb, U, Th are their concentration, and \( \lambda_{232}, \lambda_{235}, \lambda_{238} \) are the decay constants of \(^{232}\)Th, \(^{235}\)U, \(^{238}\)U, respectively. The mean age was estimated from a least-squares approach considering errors of the measured concentrations of U, Th, and Pb at the 95% level. Each measurement is represented by its probability density function. Note that there is no unit for the vertical scale because the integration of these factors on time yields a constant (Montel et al., 1996). The dash curve is computed with the program Final4 (Montel et al., 1996) and the dash-dot line represents the sum of all curves. The U, Th, and Pb contents were determined using a Camebax electron microprobe, with an accelerating voltage of 15 kV and beam current of 100–150 nA. The counting times for peak/background were 300 s/100 s for Pb, 50 s/20 s for U, and 40 s/20 s for Th. The standards are uraninite (U), Th oxide (Th), and a synthetic glass (Pb).

8.2. Tectonic setting of the sedimentation of the Quetico belt

A variety of tectonic settings have been proposed for the deposition of the sediments in the Quetico belt. We briefly present the evidence for and against several possible settings.

8.2.1. Ensialic basin

Given the elongate shape of the Quetico metasedimentary belt, an aborted rift basin in sialic crust was proposed by Percival (1989). However, as discussed by Percival (1989) and Williams (1990), the turbiditic character of the Quetico sedimentary rocks is markedly different from alluvial conglomeratic sequences that commonly develop in a continental extensional basin.

8.2.2. Back-arc basin or intra-arc basin

The evidence against this model is essentially the same as for the ensialic rift environment described above. In addition, the Quetico belt does not contain rift-related mafic igneous rocks that are common in a back-arc or intra-arc basin.

8.2.3. Forearc setting

This was first suggested by Devaney and Williams (1988) based on a sedimentological-structural study in the Beardmore-Geraldton area (Fig. 2) in the southern margin of the Wabigoon Subprovince. They outlined north-dipping thrust slices of volcanoclastic rocks and interpreted the area as a forearc basin in an accretion model. Sediments in modern forearc basins are thin (Lallemand, 1999), which is not consistent with the observed 20 km of burial corresponding to MP-MT metamorphism shortly (<9 My) after sedimentation.

8.2.4. Accretionary prism

This model, proposed by Percival (1989) and Williams (1990), suggests that submarine fans and abyssal turbidites were first deformed during their
Fig. 14. Estimated $P$–$T$–$t$ path (solid curve) for the rocks in the Jean Lake area compared to the paths of modern accretionary prisms. Queyras (Schwartz et al., 2000), Sambagawa (Ernst, 1988), and Franciscan (Ernst, 1988). Solid circles with error bars are $P$–$T$ values estimated using thermobarometeric calculations. The metamorphic facies shown by thick light gray curves are from Spear (1993). Dot-double-dash lines represent two geothermal gradients; 10 and 30 °C/km. Shaded areas are the intersections between $P$–$T$ stability area derived from mineral assemblages (see Fig. 11), and $P$–$T$ estimated from thermobarometric calculations (solid circles).

accretion onto the active Wabigoon arc (D1) and later during the docking of the Wawa arc from the south (D2–D3).

This model is compatible with the observed deformation and MP–MT metamorphism shortly after sedimentation. Burial of sediments up to 0.6 ± 0.1 GPa in less than 9 My is observed in modern subduction zones, such as Sambagawa belt in Japan, where sediments were buried to similar depths in less than 1 My (Enami, 1998; Inui and Toriumi, 2002; Aoya et al., 2003). Furthermore, the younging age of sedimentation toward the south in the Quetico belt (Zaleski et al., 1999) is compatible with a southward propagation of the active front in a north dipping subduction as shown by Platt (1986). An oblique component of the convergence in the Quetico belt likely contributed to the transpressional structures, dextral wrench zones, and transcurrent faults along subprovince boundaries. The late LP–MT metamorphism is explained by thermal relaxation after cessation of subduction.

Seismic reflection profile from the Superior Province illustrates north-dipping crust and the remnant of the subducted lithosphere down to 80–100 km in the mantle (Ludden et al., 1993; Calvert et al., 1995). Island arcs were continuously accreted to a continent to the north through subduction of intervening oceanic crust and collision. The development of the Superior Province is very similar to modern tectonic processes. This is further supported by petrological and structural studies in the Superior Province (Desrochers et al., 1993; Tomlinson et al., 1996; Davis, 1998).

The proposed interpretation, Phanerozoic style plate tectonics in Archean time, has been challenged by
Hamilton (1998) based on the lack of blueschist, ophiolite complexes, and tectonic mélanges of accretionary prisms. The absence of blueschist is explained by higher temperature gradients in accretionary prisms in Archean. Upper sections of ophiolitic rocks have been identified in Archean terranes including those in Canadian Shield (Card, 1990; Tomlinson et al., 1996). It is likely that Archean oceanic lithosphere was thick due to hotter mantle temperatures and that only upper sections of oceanic plates could have been delamination and obducted (Hoffman and Ranalli, 1988).

Modern accretionary prisms commonly contain highly deformed tectonic mélanges. For example, a mélangé unit of the Shimanto accretionary prism in Japan includes slivers of basaltic pillow lavas, shales, and volcanic ash layers, in association with highly sheared argillaceous sedimentary rocks of turbidites (Taira, 1981; Taira et al., 1988, 1992, 1997). Similar assemblages of highly deformed rocks are reported in the Quetico belt, although they are not identified as tectonic mélangé units. For example, Williams (1989) reported the occurrence of thin sheets of gabbro, tonalite, granodiorite, and ultramafic to anorthositic rocks interbedded with tectonized turbiditic sediments. On the southern side of the Quetico belt, blocks (gabbro, rhyolite, tonalite, and volcanic rocks) in a mylonitic sedimentary matrix were also reported and identified as part of a mélangé unit by Polat et al. (1998) and Polat and Kerrich (1999).

Thus, we suggest that an accretionary prism setting best explains the sedimentologic, metamorphic, tectonic and geophysics characteristics of the Quetico belt.

8.3. Source and evolution of the Quetico accretionary prism

The sedimentary rocks in the Jean Lake area are contemporaneous with the sedimentation of the Quetico sediments. They include a suite of sanukitoids of ~2690 Ma in the Wahibgoon Subprovince (Stevenson et al., 1999), and voluminous mafic-ultramafic rocks ~2692 Ma in the southern Wahibgoon belt along the boundary with the Quetico belt (Sutcliffe et al., 1989; Blackburn et al., 1991; Pettigrew and Hattori, 2002).

High contents of MgO and CaO suggest a proximal source, since these elements are easily leached during weathering and transportation (Nesbitt and Young, 1989). Immature sediments may be formed during sudden uplift and erosion of a source terrane, or by discharge of pyroclastic material from a volcano, but the Jean Lake is located more than 100 km from the southern Wahibgoon Subprovince. The mafic sedimentary rocks are too immature to be transported for such a long distance from the source. This problem may be explained by lateral transport of sediments during the development of accretionary prisms.

In modern accretionary prisms, sediments are thickened through repeated underplating and near-horizontal thrusting (Cloos, 1986; Platt, 1986; Cloos and Shreve, 1988; Lallemand, 1999; Hashimoto and Kimura, 1999). Underplating at the base of the wedge is compensated by extension in the shallow part of the wedge, resulting in lateral transport of sediments away from the wedge toward the prism front (Platt, 1986; Lallemand, 1999; Fig. 15).

In the Quetico belt, sediments would have been moved southward, and buried during the development of the accretionary prism (Fig. 15). We suggest that the rocks in the Jean Lake area were originally deposited close to the Quetico–Wahibgoon boundary, displaced towards the south, and buried up to ~20 km by rapid underthrusting along a MP–MT metamorphic gradient to amphibolite facies conditions during the D1 deformation.

Subsequent dextral transpressive deformation (D2–D3) displaced the sedimentary rocks to the west from the original depositional site. The dextral component during D2–D3 was likely caused by strain partitioning in an oblique convergent system as described in many modern subduction zones, such as the Ryukyu, Higurangi, Aleutian, and western North American subduction zones (Lallemand, 1999). The rocks were then uplifted during the D4 deformation. This exhumation was likely accompanied by the de-
vertical and horizontal scales

underplating of sediment

Wabigoon granite greenstone belt

eroded komatiitic basalts

Fig. 15. Schematic model showing the evolution of the Quetico accretionary prism based on the general model presented by Platt (1986). Thick arrows show the direction of movement.

8.4. Comparison with modern subduction zones

Our data suggest the burial of sedimentary rocks to a depth of about 20 km and their metamorphism up to amphibolite facies conditions in less than 9 My, and their exhumation up to LP–MT conditions within about 15 My. The burial and exhumation rates are comparable with those of Phanerozoic terranes. The typical burial rate of sediments in Phanerozoic prisms is about 3–8 cm per year (Demets et al., 1990) and their exhumation rate is about 0.2–1 mm per year (Duchêne et al., 1997; Schwartz et al., 2000; DeSigoyer et al., 2000). Using these burial and exhumation rates, our \(P-T\) path suggests that the Quetico rocks would have been buried in less than \(\sim 1\) My and exhumed within 12–60 My. Comparable burial and exhumation rates of the Quetico belt with those from modern subduction zones suggest that the Neoproterozoic tectonic regime was essentially similar to the modern analogs.

Modern subduction zones commonly contain HP–LT metamorphic rocks, such as blueschists and eclogites. They formed under low geothermal gradients, \(~10^\circ C/km\) (Fig. 14; Ernst, 1988; Ernst and Lion, 1999; Schwartz et al., 2000). Like many other Archean sedimentary terranes, the Quetico belt does not contain HP–LT rocks. The lack of such metamorphic rocks in Archean terranes has been in debate (e.g. De Wit, 1998). This may be attributed to high geothermal gradients \(\sim 30^\circ C/km\) during Archean time possibly caused by local effects such as magma emplacement and ridge subduction, as proposed to account for high geothermal gradients of several Phanerozoic accretionary prisms (Sakaguchi, 1999).

Upwelling of magmas may be supported by the occurrence of small, yet numerous intrusions in the belt including 2680 ± 1 Ma alkaline igneous rocks (Hattori and Percival, 1999).

9. Conclusion

Our study in the Jean Lake area in the central Quetico belt shows the mafic rocks formed by sedimentary process with a contribution of komatitic basalts, that were likely exposed on the southern Wabigoon Subprovince in late Archean time. The tectono-metamorphic evolution recorded by these rocks suggests that the Quetico sedimentary rocks formed in an accretionary prism above a north-dipping subduction slab beneath the Wabigoon Subprovince. Our proposed interpretation is suggested by a rapid burial of sediments (less than 9 My) down to amphibolite facies conditions \(0.6 \pm 0.1\) GPa and \(700 \pm 70^\circ C\), followed by exhumation under regional compressive deformation in about 15 My. This \(P-T-t\) evolution is
comparable to those of modern accretionary prisms. During their burial, the temperature gradient recorded was around 30 °C/km which is significantly higher than modern counterparts (10 °C/km). The lack of HP-LT rocks, such as blueschists, in the Quetico belt is attributed to such high geothermal gradients.

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References


