Insight into the cooling history of the Valhalla complex, British Columbia

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A B S T R A C T

Migmatitic paragneiss and amphibolite gneiss from the Valhalla metamorphic core complex, southeastern B. C. record a complex retrograde history that includes melt-involving net transfer reactions and partial re-equilibration at lower pressure conditions. Pelitic paragneiss from ~1.5 km above the Gwillim Creek shear zone contains diffusion zoned garnet porphyroblasts that indicate a slow then fast cooling history that is different from fast cooling from peak recorded in samples from within the shear zone. Fast cooling is consistent with transport of this migmatitic crustal root zone of the Canadian Cordillera up a cold footwall ramp of the thrust sense Gwillim Creek shear zone, followed by exhumation via the Valkyr Shear zone-Slocan Lake detachment after melt crystallization. Forward modeling of the interdiffusivity of Fe+Mg in garnet, governed by a retrograde net transfer reaction, indicates distinctly different cooling histories for rocks above the Gwillim Creek shear zone, suggesting these rocks may have been at similar peak temperatures at different times. Thermobarometry and thermodynamic modeling of these gneisses constrain a cooling and decompression P–T path segment consistent with retrograde amphibolite facies re-equilibration. 2-D thermal modeling suggests that slow followed by fast, then slower cooling can be accommodated by movement up a shallow thrust ramp, then rapid thrusting up a steeper ramp, and finally normal-sense shearing with erosional denudation. Ductile flow onto a cool footwall, coupled with erosion, is a viable mechanism of cooling and partial exhumation of lower crustal rocks.

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1. Introduction

Rapid exhumation of metamorphic core complexes of the Shuswap terrane in the Omineca Belt of British Columbia is well documented (Laberge and Pattison, 2007; Matthews, 1981; Norlander et al., 2002; Parrish et al., 1988; Spear, 2004; Spear and Parrish, 1996; Vanderhaeghe et al., 2003). However, the cooling history of rocks across strike appears to be variable. For example, migmatites of the Valhalla complex show evidence of rapid early retrograde cooling from peak metamorphism (~800 to ~650 °C in 1–2 m.y., Spear, 2004), whereas rocks ~60 km away in the Grand Forks complex may have cooled much more slowly (~11 °C/m.y.) during/following decompression that led to formation of cordierite-bearing assemblages in pelitic rocks (Fig. 1a; Laberge and Pattison, 2007).

Variable retrograde cooling/decompression histories at mountain range scale elucidate the tectonic processes responsible for exhumation of different parts of the root of the Canadian Cordillera. At finer scales, variations in cooling history provide further details on the thermal structure of the lower–mid crust during cooling and the timescales and mechanisms of exhumation of rocks in metamorphic core complexes.

 Petrologic cooling rates determined by forward modeling of diffusion of major cations in garnet were used to constrain the early retrograde cooling history of the Valhalla complex (Spear, 2004; Spear and Parrish, 1996). The present contribution examines, in detail, differences between petrologically determined cooling rates from different structural levels within the Valhalla complex. In addition, thermobarometry results from a partially re-equilibrated amphibolite gneiss indicate ~2.0±0.5 kbar of decompression occurred, possibly accompanied by melt crystallization during cooling.

Previous work has focused on the paragenetic, structural, tectonic and exhumation history of the Valhalla complex as a whole (Carr et al., 1987; Gordon et al., 2008, 2009a,b; Schuabs et al., 2002; Spear, 2004; Spear and Parrish, 1996). This study reports detailed results from diffusion modeling and thermobarometry suggesting a more complex early cooling history than previously reported. Two-dimensional thermal modeling indicates that the recorded cooling/exhumation histories can be produced by overthrusting onto a cold footwall, where late stage cooling and decompression is an erosional response to thrusting and the initiation of low angle normal sense shearing along the Valkyr shear zone–Slocan Lake fault detachment system. An episode of rapid ductile flow of partially melted rocks up a cool basement ramp is an important and distinct phase in the
transport and exhumation of a portion of allochthonous crust in a mature orogenic belt such as the Canadian Cordillera.

2. Geologic setting

The Valhalla metamorphic core complex is part of the regionally metamorphosed Shuswap complex of the southeastern British Columbian Omineca Belt (Fig. 1a). The area represents the exposed root of the Canadian Cordillera (Armstrong, 1982; Brown and Read, 1983; Coney and Harms, 1984). The Shuswap terrane experienced magmatic and regional metamorphism in the Jurassic–Cretaceous that persisted until the early Cenozoic exhumation and formation of several dome-shaped metamorphic core complexes (Parrish et al., 1985a,b). Tertiary regional exhumation was coeval across the Shuswap terrane (Carr et al., 1987; Parrish, 1995), but comparison of retrograde histories suggests that differences in the exhumation/cooling mechanism may exist for different gneiss domes within the Shuswap complex (Norlander, et al., 2002; Spear and Parrish, 1996).

Metamorphism in the Valhalla complex reached granulite facies conditions in the Early Tertiary (Parrish, 1995; Parrish et al., 1988; Spear, 2004; Spear and Parrish, 1996). Paragneisses in the cores of the Valhalla and Passmore domes (Fig. 1b) contain the peak assemblage garnet + biotite + quartz + sillimanite + plagioclase + K-feldspar + ilmenite + zircon + monazite + rutile [ + melt] (Spear and Parrish, 1996). Widely spread migmatisation resulted in the formation of tonalitic leucosomes present throughout the lower plate of the core shear zone. Leucosome formation and emplacement of the Airy quartz monzonite and the Ladybird leucogranite further up-section (see Fig. 1b) is interpreted to have occurred as a result of prograde dehydration melting of muscovite and biotite (Spear and Parrish, 1996).

The high grade gneisses and migmatites of the lower plate of the Valhalla complex lie in the footwall of a major low angle detachment system, the Vulkar shear zone (VSZ), which is characterized by mylonitic fabrics that indicate top to the east shear sense (see Fig. 2). An amphibolite facies penetrative mylonitic fabric occurs over a diffuse zone in the top 2–2.5 km of the footwall of the VSZ (Carr et al., 1987). On the east side of the Valhalla complex, the VSZ is cut by the younger green schist facies ductile–brittle Slocan Lake normal fault. Both of these extensional structures were active in Paleocene–Eocene time as evidenced by U–Pb geochronology of concordant and crosscutting leucogranites and a syn-deformational Rb–Sr muscovite growth age (Carr et al., 1987).

In the cores of the Valhalla and Passmore domes are exposures of the ductile Gwillim Creek shear zone (GCSZ), interpreted as a thrust sense structure (Carr et al., 1987; Parrish et al., 1985a,b; Fig. 1b). A third culmination of the GCSZ, recognized by downwarp intensifying shear fabrics, occurs in the China Creek dome south of the mapped exposures of the GCSZ (Simony and Carr, 1997). Diffuse strain associated with this structure is inherent throughout the lower plate, manifest as shear fabrics that are interpreted to have formed synchronous with or immediately following peak metamorphic conditions of 820 ± 30 °C and 8.0 ± 1.0 kbar (Spear, 2004; Spear and Parrish, 1996). The GCSZ is interpreted to be a projection of a prominent seismic reflector recognized in a seismic lithoprobe traverse (Cook et al., 1988) and interpreted as a thrust to the east thrust sense shear zone (Cook et al., 1988; Parrish, 1995; Parrish et al., 1988). Thrusting on the GCSZ is suggested as a mechanism for Paleocene fast cooling recognized by Spear (2004).

2.1. Geochronologic studies of cooling history of the Valhalla complex

The cooling history of the Valhalla complex has been extensively discussed in the literature (Carr et al., 1987; Duca et al., 2003; Gordon et al., 2008, 2009a,b; Schuabs et al., 2002; Spear, 2004; Spear and Parrish, 1996). U/Th–Pb zircon and monazite geochronology constrain the higher temperature cooling history in the context of crystallization of partial melts manifest as rim growth on zircon and monazite (Gordon et al., 2008; Parrish, 1990; Spear, 2004; Parrish and Parrish, 1996). A depth profiling study of zircon from various structural levels within the Valhalla complex revealed rim growth as young as 51 ± 2 Ma at temperatures >650 °C determined by Ti thermometry (Gordon et al., 2009b). These ages are interpreted as growth due to fluid circulation suggesting a protracted period (5–7 m.y.) at amphibolite facies conditions following leucogranite/migmatite crystallization (Gordon et al., 2009b). The fact that zircon rim depth profiling ages and monazite growth ages (54 ± 4 Ma; Th–Pb, Gordon et al., 2008) overlap with 40Ar/39Ar biotite and muscovite ages suggests that rapid cooling occurred at ~52–48 Ma (Gordon et al., 2008, 2009b). Geothermochronologic studies constrain the absolute timing of cooling of the Valhalla complex in terms of melt crystallization and exhumation from mid-crustal levels. The earliest portion of the cooling history, leading up to melt crystallization (and related zircon/monazite growth), however, is best constrained by diffusion profiles preserved in garnet (e.g. Faryad and Chakraborty, 2005; Olker et al., 2003; Storm and Spear, 2005). Near-peak fast cooling to temperatures of <650 °C is required to explain the garnet diffusion zoning observations of Spear (2004).

3. Valhalla gneisses

Samples discussed in this study are from within the mapped paragneiss unit of Carr et al. (1987). Paragneisses for this study (V10B and V10C) and from Spear (2004; V7C, V6B) are from within the Passmore dome both within the GCSZ (within structural domain P-I of Schuabs et al., 2002; samples V7C and V6B), and approximately 5 km away from the surface expression of the GCSZ in the Passmore dome (within structural domain P-II; Schuabs et al., 2002; samples V10B and V10C are from a single outcrop). A cross section of the Passmore dome (Fig. 2) suggests that the P-II sample location is approximately 1.5 km above the GCSZ. A migmatitic amphibolite gneiss (sample V14) from within the paragneiss unit between the Valhalla and Passmore domes preserves a partial retrograde re-equilibration. Its distance above the GCSZ is not well constrained, but is likely on the order of 2–3 km above the structure. Representative mineral analyses are shown in Table 1.
The paragenesis and reaction history of structural domain P-I samples (V6B, V7C) are discussed in Spear (2004) and Spear and Parrish (1996). The P-II samples (V10C and V10B), discussed in detail below, share a similar petrologic history (equivalent metamorphic assemblage: garnet + biotite + sillimanite + K-feldspar + plagioclase + quartz + ilmenite + rutile), but show a slightly different modal mineralogy (0–5 vol.% more K-feldspar, estimated from X-ray maps) and cooling history, as evidenced by the results of diffusion modeling.

Table 1
Selected mineral analyses. Analyses performed on Cameca SX100 with an accelerating voltage of 15 kV and a current of 10–20 nA. Mineral abbreviations are after Kretz (1983). *a* = not analyzed, *b.d.* = below detection.

<table>
<thead>
<tr>
<th>Sample</th>
<th>V14</th>
<th>V14</th>
<th>V14</th>
<th>V14</th>
<th>V10C</th>
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<th>V10C</th>
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<td>Rim</td>
<td>Core</td>
<td>Rim</td>
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<td>SiO₂</td>
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<td>47.62</td>
<td>51.88</td>
<td>37.80</td>
<td>38.06</td>
<td>36.29</td>
<td>34.57</td>
<td>49.11</td>
<td>53.08</td>
<td>65.76</td>
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<td>TiO₂</td>
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<td>1.03</td>
<td>0.24</td>
<td>0.04</td>
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<td>7.66</td>
<td>1.86</td>
<td>21.63</td>
<td>22.02</td>
<td>20.91</td>
<td>18.41</td>
<td>32.33</td>
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<tr>
<td>MgO</td>
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<td>11.96</td>
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<td>Na₂O</td>
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<td>n.a.</td>
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<td>98.34</td>
<td>99.71</td>
<td>100.28</td>
<td>100.85</td>
<td>99.61</td>
<td>94.91</td>
<td>100.15</td>
<td>100.43</td>
<td>100.17</td>
</tr>
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</table>

* a Fe³⁺ in amphibole was estimated using the method of Schumacher (Appendix 2 of Leake et al. (1997)).

The paragenesis and reaction history of structural domain P-I samples (V6B, V7C) are discussed in Spear (2004) and Spear and Parrish (1996). The P-II samples (V10C and V10B), discussed in detail below, share a similar petrologic history (equivalent metamorphic assemblage: garnet + biotite + sillimanite + K-feldspar + plagioclase + quartz + ilmenite + rutile), but show a slightly different modal mineralogy (0–5 vol.% more K-feldspar, estimated from X-ray maps) and cooling history, as evidenced by the results of diffusion modeling.
discussed below. Peak conditions constrained by metamorphic reactions are consistent with the results of Spear and Parrish (1996). Garnet growth occurred during dehydration melting by the reaction:

\[ \text{sillimanite} + \text{biotite} + \text{plagioclase} + \text{quartz} = \text{garnet} + \text{K-feldspar} + \text{melt} \]  

(1)

The migmatitic amphibolite gneiss (sample V14) contains compositional banding and leucocratic layers that are interpreted to represent what remains of original leucosome–melanosome textures, possibly made diffuse during partial re-equilibration in subsolidus conditions. The amphibolite gneiss sample contains the assemblage garnet + plagioclase + amphibole + biotite + clinopyroxene + quartz + sulfides. Distinct leucosomes are widely spaced in this sample, but do occur as continuous bands of melt. Evidence of partial melting is preserved as growth of plagioclase to form corona textures around garnet. Retrograde amphibole shows a similar texture, enveloping relict clinopyroxene grains (Fig. 3c).

4. Mineral chemistry and reaction history

4.1. Paragneiss

Paragneiss V10C garnet is generally low in calcium with \( X_{\text{Grs}} \) ranging from 0.013 to 0.024. \( X_{\text{Grs}} \) zoning is difficult to recognize, but highest \( X_{\text{Grs}} \) values generally occur near garnet rims. The lowest \( X_{\text{Grs}} \) values occur near quartz inclusions and may represent preserved prograde subsolidus garnet. \( X_{\text{Gps}} \) ranges from ~0.011 to 0.066, with the highest values at the garnet rims (Fig. 4b). All garnet is Fe-rich with Fe/(Fe+Mg) core values between ~0.72 and 0.74. Line traverses from core to rim reveal zoning in Fe/(Fe+Mg) from ~0.73 to 0.90, interpreted to have resulted from diffusion, with low values in the core and relatively steep increases to higher values at the rim. X-ray mapping shows local increases in Fe/(Fe+Mg) in garnet touching biotite (up to 0.90 at rim biotite; up to 0.85 at biotite inclusions), as well as a general “global” increase in Fe/(Fe+Mg) up to 0.86 in areas without biotite along the garnet rim (Fig. 4c). In general, garnet cores appear to be homogeneous with respect to Fe/(Fe+Mg), while increases in Fe/(Fe+Mg) occur in the outer 0.5–0.8 mm of the garnet.

Garnet from paragneiss V10B is extensively resorbed (Fig. 5a) and shows \( X_{\text{Grs}} \) ranging from 0.017 to 0.023. The highest \( X_{\text{Grs}} \) values are along garnet rims (Fig. 5b). Garnet from this sample is Fe-rich with Fe/(Fe+Mg) ranging from 0.785 to 0.880 (Fig. 5c). This range is smaller than that of V10C, and higher core values and a lack of homogeneity among cores in V10B suggest that the garnet cores were affected by diffusion during near-peak retrograde cooling, destroying the peak Fe/
High grade dehydration melting reactions for more mafic bulk compositions grow orthopyroxene in the residue (e.g. Beard and Lofgren, 1991). The amphibole gneiss sample (V14) is particularly noteworthy for the reason that it contains leucosomes yet no orthopyroxene has been found in thin section. However, it is still possible that orthopyroxene did begin to grow in this sample during amphibole melting, by the reaction:

\[
\text{amphibole} + \text{quartz} = \text{orthopyroxene} + \text{clinopyroxene} + \text{plagioclase} + \text{melt}
\]  

(Pattison, 2003). Alternatively, orthopyroxene could have been produced by a biotite melting reaction such as:

\[
\text{biotite} + \text{plagioclase} + \text{quartz} = \text{orthopyroxene} + \text{garnet} + \text{liquid}
\]  

(Patiño Douce and Beard, 1995). The retrograde form of the amphibole “melting” reaction (2) may have destroyed orthopyroxene upon melt crystallization.

Garnet from amphibolite gneiss sample V14 shows no Ca zoning and a narrow range in X$_{Gd}$ from 0.210 to 0.224. Fe/(Fe+Mg) zoning indicates a diffusional increase toward garnet rims (Fig. 6a) with a core–rim range of 0.717–0.867. X$_{Gd}$ zoning shows a uniform core (−0.027) and a near rim increase (to −0.089; Fig. 6b). Garnet is rare and heavily resorbed, with a rim of plagioclase and amphibole (Fig. 6c). Garnet occurs primarily in biotite-rich bands, interpreted as melanosomes.

Clinopyroxene is extensively resorbed and shows various replacement textures; grains are in places completely enveloped in amphibole (Fig. 6d). The heaviest resorption is in proximity to garnet. Clinopyroxene composition ranges from diopside to aluminian augite. Fe/(Fe+Mg) ranges from 0.319 to 0.427 with no recognizable zoning pattern.

Amphibole commonly occurs adjacent to garnet-bearing domains at least partially as a product of the granulite to amphibolite facies transition general reaction:

\[
\text{clinopyroxene} + \text{garnet} + \text{melt} = \text{amphibole} + \text{plagioclase} + \text{quartz} + (\pm \text{biotite})
\]  

which may have taken place during melt crystallization (reaction 3 of Pattison, 2003). Gibbs method (Spear, 1993) thermodynamic modeling of this continuous reaction suggests it was active very near to peak conditions. Amphibole shows somewhat patchy zones of higher Fe/(Fe+Mg), ~0.50, that may be relict paragenetic cores from near peak conditions. Lower Fe/(Fe+Mg) amphibole (~0.41–0.45) surrounds the higher zones and is common along clinopyroxene inclusions (Fig. 6d). This more Mg-rich edenitic amphibole may reflect rim overgrowth occurring during retrograde partial re-equilibration. Amphibole compositions correlate to potassic edenite–postasic ferropargasite (Fig. 6e).

Biotite occurs both in the granoblastic leucosomes and in selvedges. It is interpreted to have crystallized from the melt and grown by retrograde reactions involving mafic phases. The presence of biotite inclusions in garnet indicates that biotite was a prograde phase in this rock and that the rock probably underwent dehydration melting along with the parageneses. Biotite inclusions in garnet contain Fe/(Fe+Mg) of 0.39–0.48 and a slightly higher Ti content (0.28–0.34 cations per 11 oxygens) than matrix biotite. Matrix biotite contains 0.23–0.30 Ti cations per 11 oxygens and Fe/(Fe+Mg) ranges from 0.48 adjacent to garnet to 0.51 in leucosomes. These compositions likely reflect retrograde net transfer of Fe and Mg with garnet, amphibole, and clinopyroxene.

Plagioclase occurs in nearly monomineralic granoblastic bands interpreted as leucosomes, as well as in garnet+biotite domains rimming garnet (Fig. 6c). Plagioclase rims on garnet are generally single, untwinned crystals. X$_{Ab}$ ranges from ~0.59 in plagioclase cores to ~0.79 in rims with the highest values adjacent to clinopyroxene and amphibole. The majority of plagioclase in this sample occurs in a crystallized melt layer and shows annealed granoblastic textures and a narrower range of X$_{Ab}$ (0.59–0.64).

5. Amphibolite gneiss P–T estimates

For thermobarometry, ferric iron was estimated by simultaneous stoichiometric and charge normalization (Spear, 1993) for clinopyroxene and garnet, and by the method of Schumacher (Appendix 2 of Leake et al. (1997)) for amphibole.
Equilibrium between unzoned clinopyroxene and garnet cores is interpreted to give near-peak temperature estimates using Fe–Mg exchange thermometry (calibration of Ravna, 2000b). Garnet cores were paired with clinopyroxene analyses; the results are summarized in Fig. 6. At 7–8 kbars, garnet + clinopyroxene pairs give temperatures of 700–850 °C. While this is a large range, it is consistent with the peak P–T estimates of Spear and Parrish (1996). Peak pressure estimates were made using the garnet + plagioclase + quartz + clinopyroxene calibration of Eckert et al. (1991) and the garnet + plagioclase + quartz + amphibole calibration of Kohn and Spear (1990). The Mg–Tschermakite exchange model of Kohn and Spear (1990) was most appropriate for V14 amphibole, despite a generally low Al composition. The peak pressure estimate of 7.2 ± 0.6 kbars here is a minimum, as it is possible that plagioclase was not part of the peak assemblage, and that plagioclase cores grew upon melt crystallization. Lower grade re-equilibration is interpreted to have followed melt crystallization (reaction 4). We suggest that amphibole may have been present at peak conditions as grains are zoned and relict cores are present. Therefore, amphibole rims in contact with garnet were used in the retrograde thermobarometry.

Thermobarometric estimates of retrograde conditions using garnet + amphibole (calibration of Ravna, 2000a) thermometry and garnet + plagioclase + quartz + amphibole barometry (Kohn and Spear, 1990), compared with peak estimates, suggest a cooling and decompression P–T path segment consistent with retrograde amphibolite facies re-equilibration. Garnet + plagioclase + quartz + amphibole barometry using amphibole, garnet, and plagioclase rims indicate retrograde re-equilibration pressures of 4.6 ± 0.6 kbar at 515–595 °C (determined from garnet + amphibole Fe–Mg exchange temperature pairs). Based on these results, peak and retrograde
assemblages suggest cooling of \(-150\text{--}250^\circ\) and decompression on the order of \(-1.5\text{--}2.5\) kbars (from \(7.2 \pm 0.6\) to \(4.6 \pm 0.6\) kbar). Based on Gibbs modeling of reaction (4; see below), we propose a path, shown in Fig. 7, described by cooling, then decompression based on a lack of high T decompression textures and the absence of cordierite in adjacent pelitic samples (see Section 4.1 above; Spear and Parrish, 1996).

The early portion of this retrograde P–T path was thermodynamically modeled using the Gibbs method (Spear, 1993) and the internally consistent thermodynamic database of Holland and Powell (1998, 2001). A fixed bulk composition (see Fig. 8) was used to establish the equilibrium conditions for the operation of retrograde reaction (4). The proposed P–T trajectory results in (1) modeled plagioclase compositions, (2) Fe/(Fe+Mg) in garnet and biotite, and (3) amphibole growth that are all consistent with observed reaction texture and mineral compositions. This reaction has a similar affect to reaction (1r) in that both garnet and biotite become more Fe-rich. Biotite compositions in sample V14 show little variation in Fe/(Fe+Mg) and are probably significantly affected by retrograde Fe–Mg exchange at lower grade conditions. Plagioclase rims on garnet are higher in \(X_{an}\) than matrix grains, consistent with modeled plagioclase compositions during reaction (4). Garnet profiles (summarized in Fig. 9) suggest zoning related to net transfer reaction (4) that occurred at temperatures high enough for Fe–Mg interdiffusion. The small volumetric proportion of garnet in this rock and the severely embayed nature of the grains suggest that extreme resorption of garnet has taken place.

6. Garnet diffusion modeling

Petrologic cooling rates determined from modeling diffusion in metamorphic minerals have been extensively discussed in the literature (e.g. Dodson, 1973, 1986; Ehlers et al., 1994; Lasaga, 1983; Lasaga et al., 1977; Spear, 1991; Spear, 2004; Spear and Florence, 1992; Spear and Parrish, 1996; Storm and Spear, 2005; Wilson and Smith, 1984, 1985). Garnet diffusion modeling was carried out using a finite difference program developed by Frank Spear (see Spear and Parrish, 1996; Storm and Spear, 2005). Interdiffusion of Fe + Mg was modeled assuming spherical garnet geometry for diffusivity. Suitable
garnet porphyroblasts were chosen for their size and for the evidence of the ReNTR (reaction 1r) preserved at their rims and from zoning patterns.

6.1. Previous results from the Valhalla complex

Garnet diffusion modeling was performed by Spear and Parrish (1996) using garnet and biotite inclusions within garnet. Interdiffusion of Fe + Mg was driven exclusively by retrograde Fe–Mg exchange. Since diffusivities for garnet are generally much lower than for biotite, biotite shows uniform Fe/(Fe + Mg) and the smallest exchange reactions. For the purposes of this study, diffusivities from Chakraborty and Ganguly (1992) are most appropriate for the garnet chemistry in the Valhalla paragneisses (see discussion in Spear and Parrish, 1996).

Spear (2004) recognized that Fe + Mg interdiffusion near garnet rims was governed by the operation of ReNTR (1r). Numerical diffusion modeling constrained cooling rates from near peak temperatures to reach >200 °C/m.y. for a short period (<0.5 m.y.) in samples V6B and V7C directly adjacent to the GCSZ within the Passmore Dome (Figs. 1b, 2). Three of four four-step cooling models include an initial moderate cooling step of 20–50 °C/m.y. prior to the fastest cooling rates, which generally began as the rocks cooled through 805–775 °C. Spear (2004) also modeled diffusion in garnet adjacent to biotite inclusions governed by Fe–Mg exchange, yielding similar fast cooling steps but generally slower late cooling than is recorded along the garnet rims.

6.2. Results from this study

6.2.1. Paragneiss

The results of an electron microprobe line traverse from a V10C garnet are compared with those from sample V7C in Fig. 9. Fe/(Fe + Mg) interdiffusion modeling was performed with the assumption that the ReNTR, reaction (1r) above, was solely responsible for the garnet rim compositions in equilibrium with the matrix during initial cooling. Intrinsic to the interdiffusion modeling technique is the assumption that garnet was homogenized in Fe/(Fe + Mg) at peak conditions, and therefore at the start of the ReNTR (1r). The observation of uniform Fe/(Fe + Mg) in garnet cores (Fig. 4c) suggests preservation of this homogenized garnet composition from the time of the initiation of the ReNTR. The significance of the reaction that determines the boundary conditions for diffusion is critical to the interpretation of the cooling path experienced by a particular rock (see Spear, 2004).

Of particular importance is the behavior of the governing ReNTR reaction in T–X[Fe/(Fe + Mg)] space. A Gibbs method (Spear, 1993) thermodynamic model of this reaction (Fig. 10) for a similar paragneiss bulk composition shows a T–X trajectory with a positive slope that is significantly shallower than that for Fe–Mg exchange (ReER). That is, for a small change in temperature, the equilibrium composition (for both garnet and biotite) changes a significantly larger amount while the ReNTR reaction proceeds than during operation of the ReER. The temperature at which the garnet rim boundary condition was set can be constrained by the thermodynamic model for this reaction. The ReNTR began with the garnet in equilibrium with biotite and the melt near 780 °C. Based on estimates from nearby samples (results of Schuab et al., 2002; Spear, 2004; Spear and Parrish, 1996), it is possible that peak T for this rock was higher, perhaps beyond the stability of biotite preventing the operation of the ReNTR until cooling through 780 °C. Based on the net change in Fe/(Fe + Mg) from garnet core–rim, ~35 (±10)° of cooling took place prior to the “freezing in” of the garnet rim composition, suggesting that the ReNTR operated until the rock reached 744 °C. Alternatively, examining the stability of the reactants for the ReNTR on a petrogenetic grid suggests that the ReNTR could have operated until the solidus (vapor-absent muscovite melting curve) was reached at ~710 °C (see Spear, 2004). Assuming that muscovite was present early in the paragenesis of these paragneisses and considering that no retrograde muscovite is found in any samples, this requires that melt due to muscovite dehydration melting was lost at some point. Garnet rim–adjacent biotite pairs give temperatures of 733 °C (Hodges and Spear, 1982 calibration, assuming all Fe is Fe2+), suggesting that places along the garnet rim may have been affected by retrograde Fe–Mg exchange reactions. For the purposes of this study, diffusion profile traverses were conducted in places where biotite does not occur directly along the garnet rim and our interpretation is that diffusion was exclusively driven by the ReNTR. The subsolidus change in the equilibrium Fe/(Fe + Mg) due to the retrograde exchange reaction has a comparatively smaller effect on the shape
of the diffusion profile due to the decrease in the diffusivities with temperature. Hence a ReER overprint would be minor regarding the penetration distance of the diffusion zoning profile.

The ReNTR reaction (1r) consumes garnet as it proceeds. This effect of this change on the final diffusion profile, as discussed by Storm and Spear (2005), is to increase the slope near the garnet rim. Also, models with slow early (higher T) cooling + resorption generally yield diffusion profiles that are similar to those from models with faster early cooling without resorption. Hence models that include resorption require more time at high temperatures to produce a certain diffusion penetration distance. Estimates of the amount of garnet resorption (in μm) were based on the thicknesses of plagioclase ± sillimanite ± biotite corona that occurs between the garnet and foliation-defining matrix biotite (see Fig. 4c). The location of the electron microprobe traverse on garnet 2 that included preserved garnet core material was close to the edge of sample and did not indicate the extensive resorption recognized in garnet 1.

Our modeling approach follows that of Spear (2004), using the highest and lowest estimates for the garnet rim composition at the end of the ReNTR: 744 °C based on Gibbs modeling, and 710 °C based on petrogenetic grid considerations. Models were run with an estimate of the degree of resorption at each measured profile, as well as with no resorption for comparison. Garnet in sample V10C is irregularly shaped, and deformation was likely occurring during cooling, complicating any estimates of resorption based on linear volume change. For simplicity, garnet radius was assumed to change linearly with temperature during the operation of the ReNTR. A large number of different cooling models were examined, beginning with a single-step cooling model; however, it quickly became evident that 3+ steps were required to fit the measured Fe–Mg profile.

A simple comparison of the diffusion profiles from V7C (from Spear, 2004) and V10C with the same horizontal scale is shown in Fig. 9. This comparison ignores compositional effects on interdiffusivity, which for the respective garnet compositions equates to a 5–6% difference in the interdiffusivity (2–7 μm difference in the characteristic diffusion length scale in 10 m.y.). Note the greater diffusion penetration distance for sample V10C, which requires slower cooling upon initiation of the ReNTR. Modeling results for V10C are shown in Fig. 11. The corresponding T–t histories are shown in Table 2 and plotted in Fig. 12. In all but one model, a three-step cooling model was used to fit the diffusion profiles.

### Table 2

<table>
<thead>
<tr>
<th>Garnet</th>
<th>Solidus (°C)</th>
<th>Δr (mm)</th>
<th>Time (m.y.)</th>
<th>T (°C)</th>
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6.2.2. Amphibolite gneiss
Garnet Fe/(Fe + Mg) profiles for amphibolite gneiss V14 are noticeably narrower than those of migmatitic pelite V7C (see Fig. 9). A possible explanation is that more severe garnet resorption occurred during the activity of reaction (4), which would have extensively steepened the resulting diffusion profile. With no way to assess the degree of resorption (garnet reaction rims are not preserved as in V10C), modeling the retrograde diffusion history recorded by garnet was impossible. The challenge of accurately modeling amphibole compositions (e.g. Dale et al., 2000; Powell and Holland, 1999), as well as the possibility that melt was present below garnet stability in this rock adds to the complexities of choosing boundary conditions for a model of Fe–Mg interdiffusion in garnet due to reaction (4). Sample V14 is located between the Valhallata and Passmore domes (see Fig. 1b), at least 10 km from either mapped exposure of the GCSZ. Therefore assessing the structural distance above this perceived thrust fault is difficult and the interpretation of a T–t path would be speculative. Nevertheless, preservation of the narrow diffusion profile requires a period of relatively fast cooling, consistent with previous results and our interpretation of the other samples.

6.3. Discussion of the diffusion modeling results
The initial slow cooling step in Fig. 12 is required in all models to explain the diffusion penetration distance recorded by V10C. This initial slow cooling step acts to drive Fe enrichment deeper into the cores of garnet porphyroblasts, as seen in the profile comparison (Fig. 9). The diffusion modeling results of Spear (2004) for V7C and V6B do not reflect initial slow cooling, suggesting a different cooling history whereby fast cooling began at or shortly after peak T was obtained. The majority of the models for V10C then require a fast cooling step (cooling rate ≥ 200°/m.y.) on the order of 0.2–0.25 m.y. in duration. We interpret this as the conductive cooling pulse resulting from thrust ramp transport along the GCSZ. However, the duration of the slow cooling step prior to this fast cooling step is much greater than expected for conductive heat flow at such a short distance above the GCSZ (~1.5 km). To better illustrate this, the characteristic thermal diffusion length scale at the GCSZ is given by:

\[ h = \sqrt{\kappa t} \]

For a distance of 1.5 km from the GCSZ (thermal diffusivity \( \kappa = 6.5 \times 10^{-6} \text{ m}^2/\text{s} \)), \( t = 120,000 \) years. So the delay before the onset of fast cooling recognized for sample V10C is almost two orders of magnitude higher than that expected for simple heat conduction due to the presence of a footwall heat sink.

To explain this discrepancy, we propose a model whereby sample V10C began cooling slowly due to transport up an unmapped shallow thrust-sense shear zone up section from the GCSZ, possibly due to heterogeneous strain associated with the formation of the GCSZ. Transport would have to occur for 8–10 m.y., prior to juxtaposition and main phase thrusting. This model infers that the timing of the peak temperatures recorded by these two sets of samples differs by ~8–10 m.y.

Fast cooling due to rapid thrust movement on the GCSZ then was coeval between samples V6B, V7C and V10C. It should be noted that the duration of this slow cooling step depends in part on the starting temperature of the thermodynamic model for the ReNTR. If the ReNTR began to change the garnet rim composition at higher temperatures than suggested from our Gibbs modeling results (Fig. 10), moderately slow cooling rates would produce similar diffusion profiles in less time while diffusivities are highest, yet a slow cooling step is still certainly required to produce the observed profiles. The results of 2-D thermal modeling of this interpretation are discussed below.

7. Thermal modeling
In order to constrain a mechanism for relatively fast cooling with rates that vary with distance from a thrust ramp, a forward modeling scheme to combine heat conduction and advection was used. Similar models were discussed by Spear (2004), suggesting rapid retrograde thrust ramp cooling of Valhalla complex paragneisses. Here, we take the modeling further to constrain the later cooling and decompression history in conjunction with erosion and tectonic denudation to form a core complex by the arrest of thrusting and the initiation of the Valkyr shear zone–Slocan Lake normal fault system (Carr et al., 1987).

7.1. Approach
A 2-D finite difference model of the Valhalla complex was used with the initial conditions of a doubly thickened continental crust (60 km in thickness). Assumptions include: (1) basal heat flow of 27.5 mW/m², (2) thickness of the radiogenic layer of 30 km, (3) uniform heat generation of 2.0 μW/m² within the radiogenic layer, (4) thermal conductivity as a function of T up to a maximum at 2.0 W/m°C, (5) uniform crustal density of 2700 kg/m³, and (6) uniform heat
capacity of 1000 J/kg °C. These initial conditions were tuned to fit estimates of peak conditions for the Valhalla paragneisses. Previous models performed to test the feasibility of thrust ramp cooling (Spear, 2004), used similar parameters to test different thrust geometries and rates and their effect on cooling rates. In contrast, this model includes thermal conductivity as a function of temperature (after Whittington et al., 2009) and a slightly lower heat production and thickness of the radiogenic layer. Also, models for this study include erosion and deposition as a function of the elevation of the surface, as well as the incorporation of the VSZ–Slocan Lake fault system (treated as a single detachment fault). The ramifications of these differences are discussed in the Appendix.

Several models with varying fault geometries (5–30° dip on the GCSZ and VSZ) and horizontal thrust/divergence rates were used to predict retrograde P–T paths. In order to replicate the high cooling rates determined by diffusion modeling, a 25°-dipping thrust sense ramp on the GCSZ was paired with a 10°-dipping normal sense on the VSZ. A horizontal thrust velocity of 50 mm/year was used for the steep ramp portion of the GCSZ, with short duration (0.8 m.y) of transport up this ramp. Movement on the VSZ at 1.5 mm/year resulted in the cooling rates of 20–25°/m.y, constrained by geochronology (Spear and Parrish, 1996 and references therein).

7.2. Erosion

Erosion was incorporated into the thermal model in an attempt to better replicate the likely surface topography during overthrusting of the thickened Cordilleran onto a static cold North American basement. Models do not include subsidence, so predicted erosion rates are maxima. All models require very high erosion rates during thrusting in order to prevent uplift to unfeasibly high elevations. Rapid uplift of very high terrain (–9 km above the original surface elevation) was permitted in order to minimize high temperature decompression, for which no evidence is present in the paragneisses. While rapid overthrusting results in high uplift and erosion rates, our model in which erosion is a function of elevation produces rapid denudation (up to 48 mm/year during short duration 50 mm/year overthrusting).

At the end of overthrusting, a dramatic change in the landscape is expected as high erosion rates crash with arrested uplift and erosion acts to reduce average elevation at rates up to 6 mm/year. While this is an order of magnitude smaller than the maximum erosion rates required during thrusting, the result is a dramatic drop in elevation of up to 3000 m in <1 million years. The effects of climate change on erosion rates are not considered, hence erosion is assumed to be an efficient geomorphic response to tectonic forcing.

7.3. Paragneiss juxtaposition

In order to test the hypothesis that sample V10C underwent fast cooling in tandem with the samples of Spear (2004) following a period of slow cooling, a model was run that includes early movement on a shallow thrust ramp, prior to rapid thrusting on the main GCSZ. This model requires that V10C began to move at higher temperatures and slowly cooled through the start of biotite crystallization (the start of the ReNTR 1r) near 780 °C prior to juxtaposition with the GCSZ rocks (locations V6 and V7). An example of a model that roughly fits the diffusion model results is shown in Fig. 13. This model includes early thrusting up a 10° ramp at a rate of ~4 mm/year followed by transport up the GCSZ at a rate of 50 mm/year for ~0.9 m.y. Cooling rates of ~150°/m.y. were predicted by this model during rapid thrusting and coupled erosion (Fig. 13e).

8. Discussion

The timing of zircon crystallization with respect to our predicted T–t paths is critical to understanding the tectonic history of the Valhalla complex. The results of this study and those of Spear (2004) are consistent with melt crystallization coincident with fast cooling, which is interpreted to have continued below the solidus. The recent zircon depth profiling studies of Gordon et al. (2009a,b) indicate very young zircon rim growth in a migmatic paragneiss from the Valhalla dome. Their interpretation is that melt crystallization occurred 5–7 m.y. earlier and that this 51±2 Ma age represents the growth of thin zircon rims at 650–700 °C by fluid infiltration. Rapid melt crystallization at ~60 Ma is consistent with fast cooling at this time as suggested by Spear (2004). Our results and those of Spear (2004) are, however, not consistent with Ti-in-zircon temperatures as high as 650–700 °C at ~51 Ma as reported by Gordon et al. (2009a,b). These young zircon rim growth ages also seem to conflict with concordant U–Pb titanite cooling ages of 56–61 (Heaman and Parrish, 1991; Spear and Parrish, 1996) for the Mulvey orthogneiss that surrounds the Valhalla dome (Fig. 1b). The interpretation that young, thin ~10° 18O rim growth on zircon occurred due to infiltration of metamorphic fluids with a δ18O composition of ~10‰ suggests these fluids were derived from a local magmatic and/or metamorphic source (Gordon et al., 2009b; Holk and Taylor, 2007). This δ18O isotopic composition is at least 4%, lower than model fluid compositions demonstrably derived from pelite anatexis in the Fall Mountain Nappe of New Hampshire (roughly 14.5%; Kohn et al., 1997), so it seems unlikely that the Valhalla paragneisses were the source of the fluids. This interpretation also infers that fluid composition was sufficiently high in Zr, P, and LREE as well as other elements to allow recrystallization and replacement of major phases in the Valhalla complex. The source of such fluids is undetermined, and a suitable magmatic source of similar age is not exposed in the Valhalla complex. One possibility may involve dehydration and possibly anatexis of rocks in the footwall of the GCSZ leading to upward transport of fluids during exhumation. Alternatively, a systematic error in the zircon depth profile age determinations would suggest that the 650–700 °C Ti-in-zircon temperatures represent melt crystallization temperatures; however, this fails to explain the interior to rim increase in δ18O values (~7‰–10‰; Gordon et al., 2009b).

Gordon et al. (2008) suggest that melting may have continued during decompression. No evidence for this can be found in the migmatic paragneiss samples, and our results are more consistent with an initial phase of nearly isobaric cooling, or initial cooling with only minor decompression (see Fig. 7) such that melt crystallization began near peak conditions and proceeded during progress of the ReNTR 1r as the P–T path crossed garnet isopleths, resulting in garnet resorption. Symplectic corona textures in gedrite–cordierite pods (Marshall and Simandl, 2006) within the GCSZ have been interpreted to suggest that the Valhalla complex experienced isothermal decompression. The observation that these rocks contain gedritic amphibolite, sillimanite or kyanite, and no pyroxene or staurolite (Simandl et al., 2000; Marshall and Simandl, 2006; Gordon et al., 2009a,b, pers. comm.) does suggest a decompression event. However, the temperature of this decompression is poorly constrained by Fe–Mg exchange and we suggest that it may have occurred subsequent to the initial, nearly isobaric cooling step through the solidus as discussed above.

8.1. Cooling related to deformation

These new constraints on the cooling history of rocks from different structural levels indicate that the rocks experienced peak thermal conditions at different times (perhaps ~8 m.y. apart). This conclusion suggests that thrust-sense shearing, arguably related to the GCSZ, began before rocks adjacent to the main GCSZ had obtained peak temperature. Results of the thermal models including an early low angle thrust ramp (see Fig. 13) infer that net heating of deeper rocks (i.e. V7C) would actually be contemporaneous with minor slow cooling of rocks in the hanging wall of such a structure (i.e. V10C, see
mapped GCSZ is discussed by Schuabs et al. (2002) and Carr and Simony (2006). A negative strain gradient immediately upward from the GCSZ is constrained by geochronology and must clearly post-date the intrusion of the metamorphosed China Creek pegmatites at ~90 Ma (Carr and Simony, 2006). Carr and Simony (2006) suggest that GCSZ shear-sense motion along a splay of the GCSZ was at ~67 Ma. Our results and interpretations are remarkably consistent with the coherent thrust sheet model presented by Carr and Simony (2006).

Our thermal modeling results support the interpretation of Carr and Simony (2006) that the GCSZ was a large-scale partially splayed thrust-sense shear zone that acted to transport the entire crustal section above to the east. The rapid cooling documented in this study and others (Schuabs et al., 2002; Spear, 2004; Spear and Parrish, 1996) occurred along a steeper ramp of the GCSZ, prior to deformation associated with the VSZ, which experienced a much smaller displacement than the GCSZ (<10 km vs. 80–100 km, Carr and Simony, 2006). Our results suggest that perhaps the initiation of slow cooling from our thermal model, ~8 Ma prior to fast cooling, represents the formation of the GCSZ as a significant shallowly dipping crustal shear zone. In this model, later fast cooling, which crossed the solidus for Valhalla complex migmatites, represents the transport of the hanging wall up a more steeply dipping (possibly up to 25°) ramp duplex structure. Melt weakening due to the presence of anatectic melt in the crustal section above the GCSZ may have affected this structure (e.g. Hollister and Crawford, 1986). With U/Th–Pb zircon and monazite ages for the crystallization of leucosome material at ~59 Ma (Gordon et al., 2008; Crawford, 1986), Carr et al. (1987) and Schuabs et al. (2002) show an upper splay of the GCSZ resulting in repetition of the Mulvey orthogneiss exposure, possibly in a duplex structure in the core of the Valhalla dome (Carr and Simony, 2006; see Fig. 1b) that may correlate to an unrecognized shear zone near the Passmore dome. Alternatively, heterogeneous strain, possibly accompanied by ductile flow and a varying strain rate, may have influenced the observed variable cooling histories. A negative strain gradient immediately upward from the mapped GCSZ is discussed by Schuabs et al. (2002) and Carr and Simony (2006). As a result of diffuse strain above the GCSZ, rocks farther up section moved a larger distance relative to the GCSZ footwall. Nevertheless, we prefer the interpretation that a discrete high strain zone between the two sample localities (“V7C and V10C) is necessary in order to explain the period of slow cooling observed.

The age of the initiation of movement of the GCSZ is poorly constrained by geochronology and must clearly post-date the intrusion of the metamorphosed China Creek pegmatites at ~90 Ma (Carr and Simony, 2006). Carr and Simony (2006) suggest that GCSZ shearing began near 90 Ma and continued for ~30 Ma. We infer that perhaps the initiation of slow cooling from our thermal model, ~8 Ma prior to fast cooling, represents the formation of the GCSZ as a significant shallowly dipping crustal shear zone. In this model, later fast cooling, which crossed the solidus for Valhalla complex migmatites, represents the transport of the hanging wall up a more steeply dipping (possibly up to 25°) ramp duplex structure. Melt weakening due to the presence of anatectic melt in the crustal section above the GCSZ may have affected this structure (e.g. Hollister and Crawford, 1986). With U/Th–Pb zircon and monazite ages for the crystallization of leucosome material at ~59 Ma (Gordon et al., 2008; Crawford, 1986), our results suggest that the earliest phase of significant shallowly dipping crustal shear zone. In this model, later the initiation of slow cooling from our thermal model, ~8 Ma prior to fast cooling, represents the formation of the GCSZ as a significant shallowly dipping crustal shear zone. In this model, later fast cooling, which crossed the solidus for Valhalla complex migmatites, represents the transport of the hanging wall up a more steeply dipping (possibly up to 25°) ramp duplex structure. Melt weakening due to the presence of anatectic melt in the crustal section above the GCSZ may have affected this structure (e.g. Hollister and Crawford, 1986). With U/Th–Pb zircon and monazite ages for the crystallization of leucosome material at ~59 Ma (Gordon et al., 2008; Crawford, 1986), our results suggest that the earliest phase of significant shallowly dipping crustal shear zone. In this model, later fast cooling, which crossed the solidus for Valhalla complex migmatites, represents the transport of the hanging wall up a more steeply dipping (possibly up to 25°) ramp duplex structure. Melt weakening due to the presence of anatectic melt in the crustal section above the GCSZ may have affected this structure (e.g. Hollister and Crawford, 1986).
Simony, 2006). Our thermal model, which includes the initiation of thrust-sense shearing contemporaneous with the start of slow cooling observed in sample V10C (see Section 7.3), incorporates a horizontal eastward displacement of the hanging wall section of 32 km during shallow thrusting, followed by 45 km during steeper thrusting, resulting in a total shortening of 77 km. This is relatively close to shortening estimates of 80–100 km presented by Carr and Simony (2006) based on their inferred ramp geometry. However, thermal models incorporating erosion require a steeper ramp to produce the cooling observed. Perhaps even more rapid transport on a shallower ramp occurred, or the modeled thermal regime for the cold footwall was not cold enough. A larger degree of late doming in the Valhalla complex may have altered the initial ramp angles further than expected.

The very rapid cooling observed in the Valhalla complex is not recognized in adjacent southern Cordilleran metamorphic core complexes. Evidence for high temperature decompression suggests that much of the Shuswap terrane remained hot during exhumation and cooled less rapidly than the Valhalla (Kruckenberg and Whitney, 2011; Laberge and Pattison, 2007; Norlander et al., 2002). Heterogeneity in the retrograde history for exhumed orogenic roots can be related to the formation of localized décollement ramp geometries, such as demonstrated for the GCSZ. The formation of such décollement ramps may be related to variability in the strength of an underthrust indenter, in this case the cool North American basement (Beaumont et al., 2010; Carr and Simony, 2006). Recent whole-orogen crustal and upper mantle scale modeling has shown that such heterogeneity in structural style is achievable during underthrusting of cool basement beneath allochthonous crust. Ductile flow of migmatic crust up relatively steep ramp segments of the main decollement may be a result of encountering strong, detached blocks of underthrust basement crust attached to the upper mantle (Beaumont et al., 2010). Dome formation above basement ramps is further supported by a recent study in the Frenchman’s Cap dome of the Shuswap terrane (Gervais et al., 2010). In the Frenchman’s Cap dome, minor top-to-the-east thrust sense shearing gives way to minor upright folding and a lack of Cordilleran penetrative fabrics in the dome core, suggesting the dome represents the basal limit of deformation due to drag folding following thrust transport up a basement ramp (Gervais et al., 2010).

These “petrologic” cooling rates based on diffusion modeling help constrain the portion of a P-T-t (Pressure-Temperature-time) path where (1) crystallization and/or diffusional closure of geochronometers does not occur, (2) large uncertainties are the norm, and (3) complexities in interpretation cause radiometric geochronology and thermochronology to fall short of providing a detailed history near the peak of metamorphism. Deciphering cooling histories with this technique allows for resolution of more subtle aspects of the metamorphic and tectonic history of an orogen. Remarkably, our results show how zoning related to diffusion can preserve millions of years of thermal and deformational history not recorded by radiometric geochronology. Diffusion analysis is gaining applicability and with improvements in our ability to determine the boundary conditions for diffusion will become an even more powerful tool.

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Appendix. Thermal modeling considerations

1. Thermal diffusivity

Recent work using laser flash analysis has shown that thermal diffusivity is a relatively strong function of temperature up to ~580 °C (Whittington et al., 2009). This has noted effects on models of a steady state geotherm for significantly thickened crust in a collisional orogen. In general, it causes the lower crust to remain hotter and better insulated from the surface heat sink (Braun, 2009). In light of these results, models were run with a numerical approximation to heat conductivity as a function of temperature. Our initial condition was set up by starting with a steady state (k = constant) geotherm and iteratively recalculating the geotherm and the resultant heat conductivity as given in Whittington et al. (2009). We used a constant thermal conductivity of 2.0 W/m-K at high temperatures (the lower crust). A tolerance at each grid point of temperature change <0.001 °C was imposed, reaching a “steady T” solution for the upper crust. Resultant geotherms for t = 0 are plotted in Fig. A.1. It is important to note that these are transient (not steady state), as allowing any heat conduction will change the T and hence k versus depth profile. In general, the k = f(T) geotherm shows lower temperatures in the upper crust and an inflection from concave to convex occurring where k becomes constant, eventually approximating the steady-state k = constant geotherm in the lower crust. The corresponding geotherm plot results in a smaller ∫dz, hence a smaller bulk heat flow over the full crust that is because of a slightly lower heat production in these models, required to fit the peak conditions recorded in the Valhalla complex.

2. Erosion function

In thermal models, the erosion (and deposition) rate was set as a function of elevation, given by the general equation:

\[
\frac{dz}{dt} = \frac{Mz}{N - z}.
\]

(A.1)

M and N are erosion parameters. M is a scale factor for the erosion curve (Fig. A.2), which requires tuning to determine acceptable values. N corresponds to the maximum elevation allowable, in meters. The function is asymptotic to M at high elevations.

Figure A.1. Geotherms based on thermal conductivity as a constant or as a function of temperature. Black line shows geotherm used in 2-D thermal modeling.

and solved numerically to give the amount of erosion for each model time step. All eroded material is assumed to be transported out from the model grid within each time step and does not affect the thermal evolution of the model. In deposition, a similar N value was used but with a larger M value by a factor of 2, resulting in a faster basin fill response to faulting to avoid elevations that are considerably below initial surface level (far below sea level). Erosion parameters used in the modeling shown in Fig. 13 are M = 0.4 mm/year and N = 9000 m. N was chosen to limit elevation to near maximum elevations on earth today. Deposition parameters used in modeling shown in Fig. 13 are M = 0.8 mm/year and N = 10,000. The larger deposition scale factor M generally prevents the formation of a large negative relief basin. Like erosion, deposition does not balance flow of material into or out of the model.

References


