An Investigation of the ca. 2.7 Ga Late Archean Magmatic Event (LAME) in the Superior Province using 1-D Thermal Modelling

by

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A thesis submitted in conformity with the requirements for the degree of Doctor of Philosophy
Graduate department of Geology
University of Toronto

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Seema Ahmad, Doctor of Philosophy, Department of Geology, University of Toronto, 2009

Abstract

The Late Archean Magmatic Event (LAME), ca. 2.7 Ga, was the greatest crustal addition event in Earth history. My focus is the Superior Province of Canada, where LAME occurred ca. 2.75 – 2.65 Ga. Mantle plumes impinged on the Abitibi subprovince, where ~ 16 km regional thickness of tonalite-trondhjemite-granodiorite (TTG) melt was produced. Granites (sensu stricto) were the last magmatic phase of LAME, with a Superior-wide regional thickness of ~ 1 – 3 km.

Assuming a crustal source for both TTG and granites, I use 1-D thermal models to investigate the origin of TTG in the Abitibi subprovince and that of late granites in the Superior Province. Melting curves appropriate to the source of TTG and granites are used to determine the thickness of melt produced in the models.

I show that the incorporation of upward melt transfer into a standard model of lower crustal melting may increase the amount of predicted melt by \( \frac{1}{1-f} \), where \( f \) denotes the fraction of melt that is on average being extracted from the source rocks. Partitioning of heat producing elements between melt and restite reduces the amount of melt produced, but the effect is secondary compared to the increase in melt production through upward melt transfer.

For the Abitibi subprovince, I show that the emplacement of a single plume coupled with the emplacement of a 12-km-thick greenstone cover can generate a maximum of ~ 9-km-thickness of TTG melt. However, the emplacement of a series of plumes, each coupled with the
emplacement of a 3-km-thick greenstone cover and a 10-km-thick sill results in ~ 20-km-thickness of TTG melt. My model incorporates delamination of restitic eclogite.

Finally, I show that late granites in the Superior Province may have resulted from thickening of a crust that had been “pre-heated” during earlier arc activity and that prolonged granitic magmatism observed in some areas of the Superior Province may be explained by late underthrusting of fertile source rocks into deeper and hotter regions of the crust.
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Chapter 1: Introduction

Chapter 1

Introduction

The geological processes responsible for the formation of Archean protocrust and its evolution into differentiated continental crust remain elusive despite vigorous investigative efforts on the part of geochronologists, geochemists, structural geologists, geophysicists, etc. Not only does “much work remain to be done” but first order questions remain to be answered, such as the manner in which tonalite-trondhjemite-granodiorite (TTG), the characteristic and dominant lithological component of Archean crust, formed. Central to the evolution of Archean (or more modern) crust is the melting process that allows crustal rocks to be extracted from the mantle in a single or multi-stage process. And fundamental to the melting process is heat (or heat source) though fluids and decompression also play an important role. This thesis is a contribution to the effort to determine how two magma types, TTG and granite, may have been generated in the Archean. The focus of the investigation is the Late Archean Magmatic Event (LAME) in the Superior Province, ca. 2.75 – 2.65 Ga. The investigative tool used is 1-D thermal modelling.

1. Background information

The Superior Province is the largest preserved Archean craton (Fig. 1). Spanning an area of $1.57 \times 10^6$ km$^2$ it constitutes almost one quarter of all exposed Archean crust (Goodwin 1996). It is primarily Mesoarchean to Neoarchean in age with only minor Eoarchean to Paleoarchean components (Henry et al. 2000; Percival et al. 2006).

The prevailing view of its origin is that it was assembled through a series of collision-accretion events that resulted in the amalgamation of continental and oceanic fragments (Langford and Morin 1976; Card 1990; Williams et al. 1992; Stott 1997; Card and Poulsen 1998). In the western Superior Province five such collisions took place between 2720 and 2680 Ma (Percival et al. 2006). The underlying assumption is that plate tectonics was fully operational at the time. Within this overarching model the details of the construction of the Superior Province are far from understood.

The accretion events consisted of a remarkably similar series of steps. They were predated by $\sim 40 – 50$ My of primarily mafic volcanism and TTG plutonism. While much of this magmatic
activity has been interpreted as arc magmatism there is strong evidence for plume activity in the Abitibi and, to a lesser extent, Wawa subprovince. This was followed by the “cessation of arc magmatism, early deformation, synorogenic sedimentation, sanukitoid magmatism, bulk shortening, regional metamorphism, late transpression, orogenic gold localization, emplacement of crust-derived granites, and postorogenic cooling” (Percival et al. 2006).

The accretion events were diachronous, starting in the north and younging southward at intervals of ~ 10 Ma. This feature is ascribed to a series of predominantly northward dipping but southward stepping subduction zones. The end result is a series of east-west trending volcano-plutonic, metasedimentary, plutonic and high-grade gneiss belts (Fig. 1) that make up the Superior Province. The repetitive nature of the accretion events is encouraging and suggests that an understanding of each step may allow the eventual piecing together of the manner in which plate tectonics operated in general in the Neoarchean.

Voluminous magmatism, between 2.75 – 2.65 Ga, accompanied the assembly of the Superior Province. I refer to this as the Late Archean Magmatic Event (LAME). LAME-aged rocks utterly dominate the present surface of the Superior Province (Fig. 2). In terms of juvenile additions to the crust 48 wt% of the Superior crust is estimated to have been added between 2.76 – 2.69 Ga (Henry et al. 2000). This suggests that a regional thickness of 17 – 20 km of melt was generated during LAME, and more if one considers reworking of pre-LAME crust.

LAME appears, in fact, to have been a “worldwide” event, with rock ages clustering around 2.7 Ga found in Finland, North China, Russia, India, Australia (Yilgarn craton and East Pilbara Terrane), Canada (Slave, Rae, and Hearne Provinces) and South Africa (Zimbabwe and Kaapval cratons) (Blenkinsop et al. 1997; Lobach-Zhuchenko et al. 2000; Ernst and Buchan 2001; Ketchum et al. 2004; Polat et al. 2006; Slabunov et al. 2006; Condie 2007; Naqvi and Rana Prathap 2007; Pirajno 2007; van Breemen et al. 2007; Kositcin et al. 2008; Martel et al. 2008). Consideration of the area spanned by juvenile LAME-aged rocks identifies it as the greatest crustal addition event in earth history (Condie 2001).

Thermal modelling of LAME in general, and modelling that is specific to the Superior Province in particular, is rare. This is a serious omission given that LAME appears to have been primarily a thermal event. Furthermore, there do not appear to be any models in which the thickness of melt produced is quantified (an important exercise where voluminous quantities of melt are involved).
Rey et al. (2003) have suggested that the thermal pulse from a mantle plume coupled with the “thermal blanket effect” of plume-generated greenstones was responsible for the generation of granitoids (primarily TTG and to a lesser extent granite) during LAME. However, as I will show, their 1-D thermal models do not appear to produce enough melt (Chapter 6). Mareschal and Jaupart (2006) have suggested that heating due to crustal thickening (again the thermal blanket effect) was responsible for the generation of the late granites during LAME. However, the 1-D thermal models they present are steady state so that the timing of granitic magmatism in their models relative to deformation (thickening) cannot be checked against geochronological data for the Superior Province. Percival and Pysklywec (2007) have suggested that the delamination of restitic eclogite from the base of the crust may have triggered an overturn of the lithospheric mantle that resulted in sudden, rapid heating of the lower crust and the generation of the late granites. To date there is no evidence for such a mechanism operating elsewhere. And the spatial extent of lower crustal heating in their 2-D thermal models is not significantly wider than the width of the eclogitic raft. Hence a series of such rafts would be required to delaminate in succession to account for granite production throughout the Superior Province.

2. Objectives

The principal objectives of this study are to account for, in a quantitative fashion using 1-D thermal models, the following two aspects of LAME:

1. The generation of TTG in the Abitibi subprovince where there is convincing evidence for plume activity.
2. The generation of late granites in the Superior Province.

3. Thesis outline

In Chapter 2, I describe the details of my thermal modelling.

In Chapter 3, I outline constraints on crustal thickness and mantle heat flow at the time of LAME. I calculate average heat production in the Superior crust at 2.7 Ga.

In Chapter 4, I calculate melting curves for amphibolite (source rock for TTG) and tonalite (a source rock for granite).

In Chapter 5, I investigate the effect that upward transfer of melt in a LAME-aged Superior-type crust may have on the amount of melt produced.

In Chapter 6, I analyze the plume with greenstone cover model of Rey et al. (2003) in detail. I show that while increasing the crustal heat production in their model to reflect that of the
Superior crust at 2.7 Ga significantly increases the thickness of TTG melt produced, it still falls short of that observed in the Abitibi subprovince. I then investigate the thickness of TTG melt that may be produced in a LAME-aged Abitibi protocrust that is bombarded by a series of mantle plumes. I compare the thickness of melt produced, as well as the timing of melt generation, to that observed in the Abitibi.

In Chapter 7, I investigate the role that “pre-heating” of the Superior crust during early arc activity (prior to thickening) and late underthrusting may have played in the generation of the late granites during LAME.

In Chapter 8, I summarize my main findings and suggest ways in which this work can be extended and improved. I also discuss related problems in need of thermal and other forms of investigation.
References


Lobach-Zhuchenko, S.B., Chekulaev, V.P., Ivanikov, V.V., Kovalenko, A.V., and Bogomolov, E.S. 2000. Late Archean high-Mg and subalkaline granitoids and lamprophyres as indicators of gold mineralization in Karelia (Baltic Shield), Russia. In Ore-bearing granites of Russia and adjacent countries. Edited by A.A. Kremenetsky, B. Lehmann and R. Seltmann. Institute of Mineralogy, Geochemistry and Crystal Chemistry of Rare Elements, Moscow, Russian Federation, pp. 193-211.


Fig. 1. Map of the Superior Province. Modified from Stott (2009).
Fig. 2. U-Pb crystallization and estimate of crystallization ages of 606 rock samples from the Superior Province. Data from Skulski and Villeneuve (1999).
Chapter 2

Modelling Details

1. Introduction

I have constructed 1-D thermal models of the lithosphere during LAME. In Chapter 5, I have modelled only the crust.

2. Heat Equation

The basic equation that governs heat transfer and hence transient geotherms (temperature-depth profiles) in my models is the heat equation with an advective term:

$$\rho C \frac{\partial T}{\partial t} = \nabla \cdot (K \nabla T) + A - \rho \bar{u} \cdot \nabla T$$  \hspace{1cm} (1)

- $T$ is temperature.
- $\nabla T$ is the gradient vector of the temperature field, $\nabla T = \frac{\partial T}{\partial x} i + \frac{\partial T}{\partial y} j + \frac{\partial T}{\partial z} k$.
- $\nabla \cdot$ denotes the divergence of a vector, e.g. $\nabla \cdot \mathbf{v} = \frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} + \frac{\partial v_z}{\partial z}$.
- $\rho$ is density.
- $C$ is heat capacity.
- $K$ is thermal conductivity.
- $A$ is rate of heat production.
- $\bar{u}$ is a velocity field with respect to a reference coordinate system.

A more familiar form of this equation is obtained by moving the thermal conductivity, $K$, outside the scope of the spatial derivatives. This can only be done, however, if $K$ is constant, which it is not in my modelling.

2.1. Heat equation in one dimension

My 1-D models are governed by the 1-D form of (1) given by:

$$\rho C \frac{\partial T}{\partial t} = \frac{\partial}{\partial x} \left( K \frac{\partial T}{\partial x} \right) + A - \rho \bar{u} \frac{\partial T}{\partial x}$$
or

\[\rho C \frac{\partial T}{\partial t} = K \frac{\partial^2 T}{\partial x^2} + \frac{\partial K}{\partial x} \frac{\partial T}{\partial x} + A - \rho C u \frac{\partial T}{\partial x}.\]  

(2)

In my models, \(A\) represents crustal heat production from the decay of radioactive elements U and Th and of \(^{40}\)K, an unstable isotope of potassium. \(x\) is tied to the erosion surface (which corresponds to \(x = 0\)) and not to material markers in the rock. \(u\) represents the erosion rate which may be constant or time-dependent. Where erosion is not involved \(u = 0\) and (2) reduces to:

\[\rho C \frac{\partial T}{\partial t} = K \frac{\partial^2 T}{\partial x^2} + \frac{\partial K}{\partial x} \frac{\partial T}{\partial x} + A.\]  

(3)

Where steady state (equilibrium) conditions hold \(\frac{\partial T}{\partial t} = 0\) and (3) reduces to:

\[K \frac{\partial^2 T}{\partial x^2} + \frac{\partial K}{\partial x} \frac{\partial T}{\partial x} + A = 0.\]  

(4)

Initial conditions in my models generally consist of a steady state solution or a variant of it.

3. Boundary Conditions

In all of my models the boundary condition at the top of the crust is \(T = 0\), where \(T\) denotes temperature in °C. The boundary condition at the base of the lithosphere (or crust, where only the crust is modelled) is constant heat flow.

4. Thermal Parameters

I use crustal and mantle densities of 2800 and 3250 kg/m\(^3\), respectively, and a heat capacity of 1000 JK\(^{-1}\)kg\(^{-1}\). Other parameters, such as mantle heat flow and crustal heat production are discussed in Chapter 3.

4.1. Thermal Conductivity

I have employed a temperature-dependent crustal and mantle conductivity, \(K\), similar to that used by Mareschal and Jaupart (2006) in their thermal modelling of 2.55 Ga Eastern Abitibi and Slave Province crust (see Appendix A):

\[K = 2.26 - \frac{616.858}{T} + K_0 \times \left(\frac{355.574}{T} - 0.30247\right), \text{ (Wm}^{-1}\text{K}^{-1})\]
where $T$ is the absolute temperature, and $K_0$ denotes the conductivity at the surface. It is based on an empirical formula for thermal diffusivity determined by Durham et al. (1987). Like Mareschal and Jaupart (2006), I chose a surface conductivity of $K_0 = 2.8 \cdot 10^{-9}$.

For mantle conductivity a radiative component was added:

$$K_r = 0.37 \times 10^{-9} T^3, \quad \text{(Wm}^{-1}\text{K}^{-1})$$

(Scharmeli 1979; Mareschal and Jaupart 2006). In my models I obtain the absolute temperature by adding 273.15 to model temperatures.

As pointed out by Mareschal and Jaupart (2006) use of a temperature-dependent conductivity can result in a significant increase in Moho temperature (> 150°C) relative to that resulting from use of a constant conductivity. The latter is commonly employed in thermal modelling of the crust.

4.1.1. Fitting Vicker (1997) mantle xenolith data

As a test of whether the temperatures that result from use of the above thermal parameters (in particular my choice of conductivity) are realistic or not, I use these parameters to determine the mantle heat flow value that provides a best fit geotherm to a P-T array derived from thermobarometry of Late Jurassic peridotite xenoliths from the Abitibi subprovince. As discussed in Chapter 3, section 4.1, the assumption is that such a P-T array represents not only Late Jurassic, but also present day conditions, at least in the upper lithospheric mantle. Therefore, if my choice of thermal parameters is reasonable the determined mantle heat flow value should fall within the estimated range of 10 – 14 mWm$^{-2}$ for current mantle heat flow in the Abitibi subprovince (Guillou et al. 1994).

The peridotite xenoliths are found in kimberlite diatremes that occur near Kirkland Lake in northern Ontario, in the south-central portion of the Abitibi Subprovince. They have Late Jurassic emplacement ages (155-159 Ma) (Schulze 1996). Xenolith textures range from undeformed (coarse) to deformed (porphyroclastic) with most xenoliths retaining evidence of mantle equilibrium conditions (Vicker 1997). I have used the pressure and temperature data determined by Vicker (1997) who used the Brey-Kohler-Nickel (BKN) geothermometer (TBKN) and geobarometer (PBKN) (Brey and Kohler 1990; Brey et al. 1990). TBKN is based on Ca/Mg partitioning between clinopyroxene and orthopyroxene. Compositional variables used are Fe, Mg, Ca and Na contents of clinopyroxene and orthopyroxene. PBKN is based on the Al content of orthopyroxene in equilibrium with garnet. Compositional variables used in PBKN
include Ca, Mg, Fe, Cr, Al and Mn of garnet, and Ca, Mg, Fe, Cr, Al, Mn and Na of orthopyroxene (Vicker 1997).

Interestingly, the temperatures associated with the deepest derived mantle xenoliths (from depths of ~ 200 km) (Fig. 1) approach remarkably closely the adiabatic temperatures associated with a mantle potential temperature of 1300°C, the estimated present day mantle potential temperature. (Mantle potential temperature is the temperature the mantle would have if it rose adiabatically to the surface without melting). This suggests that these xenoliths are indeed derived from near the base of the Abitibi lithosphere.

Mareschal and Jaupart (2006) estimate a crustal thickness of 40 km for the Western Abitibi subprovince and crustal heat production of 1.2 µWm⁻³ for the upper 20 km and 0.4 µWm⁻³ for the lower 20 km.

I calculated geotherms using the above-mentioned data for the Western Abitibi for a range of mantle heat flow values from 10 – 16 mWm⁻² (Fig. 1). I assumed that heat production in the lithospheric mantle was negligible (0 µWm⁻³). Geotherms associated with mantle heat flow of 12 or 13 mWm⁻² appear to provide the best fit to the peridotite xenolith P-T data (Fig. 1). Hence use of the thermal parameters that I have employed in my modelling yield mantle heat flow values that are consistent with current estimates.

5. Femlab 3.1

My models are constructed using Femlab 3.1, a commercially available software package that uses the finite-element method to solve PDEs. It can be accessed as a stand alone product through a GUI but also interfaces with Matlab. (Femlab models can be saved and run as Matlab m-files). Femlab 3.1 contains a variety of ‘applications modes’. These consist of predefined templates set up with equations and variables for specific areas of physics, such as heat transfer, fluid dynamics, electromagnetics, etc. The user specifies material properties, boundary conditions, etc.

I used the ‘Heat Transfer’ mode, specifically the Heat Transfer ‘Conduction’ mode for those models that do not incorporate erosion and the Heat Transfer ‘Conduction and Convection’ mode for those that do. My main use of Matlab was to export solutions from Femlab to Matlab and use Matlab functions to analyze the solutions. This was done for most of my models to determine aspects of the model such as the thickness of melt produced over time.

A Femlab model can consist of an ‘unlimited’ sequence of sub-models (called ‘geometries’ in Femlab). All of my models were constructed using multiple geometries. For each geometry
the user specifies a spatial domain. The domain may be subdivided into ‘subdomains’ which differ in at least one property. For example, a thermal model of a lithosphere may consist of two subdomains, a crust and lithospheric mantle, that differ in their density, thermal conductivity, etc. Alternatively, the crust may consist of an upper and lower crust that differ in their heat production so that a total of three subdomains (upper crust, lower crust and lithospheric mantle) would be specified.

The boundaries of subdomains may be external (no adjacent subdomain) or internal. For example, the top boundary of an upper crustal subdomain is external while its bottom boundary is internal. Boundary conditions are specified only on external boundaries. In a 1-D model this is at the top of the crust and the base of the lithospheric mantle (or crust if only the crust is modelled).

Femlab 3.1 provides a default mesh (‘grid’) for all subdomains but allows this to be modified or overridden in various ways. In my 1-D models I restricted the maximum size of the mesh elements for crustal subdomains to 20 m and for lithospheric mantle subdomains to 250 m.

Femlab 3.1 does automatic smoothing of specified initial conditions (initial geotherms). Where excessive smoothing at the top and bottom of magmatic intrusions into the crust and plume intrusions into the lithospheric mantle appeared to be occurring I adjusted (decreased) the maximum element size at the top and bottom of such intrusions to limit the region over which such smoothing occurred.

An extremely useful feature of Femlab 3.1 which I made use of in all of my models is that the solution geotherm from one geometry can be saved and then used to specify initial conditions (an initial geotherm) for another geometry. An entire solution geotherm need not be used. Rather, segments of it, specified on subdomains, may be used and transformations may also be applied to the geotherm on these segments. I discuss some examples of the manner in which this feature was utilized in my modelling.

In those models in which the initial geotherm was a steady state geotherm I calculated the steady state geotherm in one geometry and used it as an initial geotherm in the succeeding geometry. To model homogeneous crustal thickening (Chapter 7) I stretched the crustal geotherm obtained for one geometry and used it as the initial geotherm for the geometry that modelled the thickened crust. To model the intrusion of a plume (Chapter 6) I first obtained the geotherm prior to intrusion. I then performed a ‘cut and paste’ on the geotherm by moving the portion of the geotherm corresponding to depths greater than that at which the plume was intruded downward by the thickness of the plume. I then assigned an appropriate temperature to
that portion of the geotherm representing the plume to obtain the initial geotherm after plume intrusion. To model upward transfer of melt from the lower crust (Chapter 5) I obtained a geotherm prior to the upward transfer. I then ‘cut out’ the portion of the lower crustal geotherm that represented melt and inserted it into the mid- to upper crust portion of the geotherm to obtain the initial geotherm after upward transfer of melt.

Model parameters were input into Femlab 3.1 in SI units. Femlab 3.1 has an automatic scaling feature and I allowed this to handle the scaling. I specified maximum time steps of 1 My.

6. Basic Checking of Modelling Results

For each application mode there is a default choice of solver to solve the associated PDEs, which I used. My understanding is that Femlab 3.1 is a well tested piece of software in which heat transfer models that are considerably more sophisticated than mine can be developed. Hence I did not undertake extensive independent testing of the accuracy of my modelling results. However, I did do some basic checking.

I reran a published model by Rey et al. (2003) of LANE that incorporates a plume intrusion and thickening by a greenstone cover and obtained good agreement with their results (Fig. 6 in Chapter 6).

I also solved a simple heat flow problem for which an analytic solution is known using parameters that are the same as or similar to those employed in my models (except for thermal conductivity which is constant in the problem) and compared the resulting geotherms with the analytic solution (Appendix B). The two solutions agree to within 0.5°C.
Chapter 2: Modelling Details

References


Fig. 1. Geotherms calculated using crustal thickness and heat production data for the Western Abitibi of Mareschal and Jaupart (2006) and mantle heat flow ranging from 10 – 16 mWm$^{-2}$. The associated mantle heat flow value appears below each geotherm. P-T data for peridotite xenoliths from the Abitibi subprovince calculated using BKN methods (Vicker 1997).
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Chapter 3

The Superior Crust

Abstract

In order to model the thermal evolution of the Superior crust during the Late Archean, it is necessary to estimate crustal thickness, heat production and mantle heat flow at that time. And because the aim of my models is to predict the amounts of TTG and of granites produced during the Late Archean Magmatic Event (LAME), what is known about the actual amounts produced must also be reviewed. The estimates are as follows:

Pre-LAME crustal thickness: between 20 and 46 km.
Pre-LAME mantle heat flow: at least that of present day, 10 – 18 mWm\(^{-2}\).
Bulk average heat production of LAME crust: 1.3 – 1.5 µWm\(^{-3}\).
Average thickness of TTG emplaced during LAME: 5 – 8 km.
Average thickness of late granites: 1 – 3.1 km.

1. Introduction

The aim of my thesis is to test possible mechanisms that might have been responsible for TTG and granite production during LAME by using 1-D thermal modelling of a Superior-like crust during LAME. In order to construct such a model I begin with either a pre-LAME crust (for the Abitibi subprovince) or a late LAME crust (for the late granites) to which I apply the mechanisms that I want to test, for example crustal thickening (for the late granites).

Certain basic quantities must be specified to set up the model, such as crustal thickness, crustal heat production and mantle heat flow. I limit myself to estimating bulk average crustal heat production here, letting the choice of vertical distribution of heat producing elements be dictated by the model. For example, in a model implementing upward melt transfer (Chapter 4), the initial crust is undifferentiated, with uniform average heat production. As melt is produced, heat producing elements (HPEs) are concentrated into it. When melt is transferred upward, the mid-crust becomes enriched in HPEs while the lower crust is depleted.
I attempt to arrive at a range of plausible values for crustal thickness, crustal heat production and mantle heat flow, while recognizing that given the significant reworking of the Superior crust that occurred during LAME (as evinced by the paucity of ages of Superior rocks older than LAME – see Fig. 2 in Chapter 1) significant uncertainty is associated with the results, particularly those applying to a pre-LAME crust.

A test of whether a mechanism being modelled may have played a significant role during LAME is to compare the amount of TTG and granite melt produced in the associated model with that produced during LAME. The manner in which I determine the quantity of melt produced in my models will be discussed in Chapter 4. Here I will estimate the regional thickness of TTG and of late granites produced during LAME. Indeed, the latter quantity has already been estimated by Cruden (2006).

2. Thickness of pre-LAME crust

Jaupart and Mareschal (1999) and Perry et al. (2002, 2006) have estimated a present day average crustal thickness for the Superior Province of 42 and 41 km respectively. The estimated average thickness of all present day continental crust is 41 km (Christensen and Mooney 1995). Considering the uncertainty in these estimates, I conclude that the average thickness of the present day Superior crust is the same as that of all continental crust.

2.1. Minimum thickness

A simple constraint on the minimum thickness of a pre-LAME crust is the thickness of Neoarchean oceanic crust. Neoarchean mantle potential temperatures are estimated to have been ~100 – 200°C hotter than present day (Abbott et al. 1994) resulting in greater decompression melting at mid-ocean ridges and oceanic crustal thicknesses of ~15 – 20 km (McKenzie 1984). (Mantle potential temperature is the temperature the mantle would have if it rose adiabatically to the surface without melting). I take 20 km as a lower limit for the thickness of the pre-LAME crust.

2.2. Variations of crustal thickness in the Archean

Many modellers tend to assign an initial thickness of 35 km to the crust (Archean or otherwise). Indeed, I will do the same in some models for the late granites (Chapter 7). However, the work of Galer and Mezger (1998) suggests that a range of crustal thicknesses might have existed in the Archean, and can be considered for a pre-LAME crust.
2.2.1. Maximum thickness

Galer and Mezger (1998) suggest that stable continental crust in the Archean was likely thicker than stable continental crust today. They have used the principle of isostatic balance between oceanic and continental crustal columns to estimate average crustal thicknesses in the Archean. They point out that the lack of variation of modern continental crustal thickness with basement age is evidence that such a balance is maintained at present (Wise 1974; Durrheim and Mooney 1991; Christensen and Mooney 1995).

Galer and Mezger (1998) observe that continental surfaces have remained close to ambient sea level throughout geologic time (Galer 1991; Buick et al. 1995) and that this is paradoxical in that as the area of the continental land mass increases, continental freeboard (the mean height of continental land above ambient sea level) should decrease. Galer (1991) claims that the main parameter governing continental freeboard is the thickness of the oceanic crust, which is strongly dependent on the mantle potential temperature. As previously discussed, oceanic crust formed at Archean plate spreading centers (if such existed) should have been thicker than present day (Sleep and Windley 1982; Bickle 1986; McKenzie and Bickle 1988; Galer 1991).

Galer and Mezger’s (1998) estimate of the thickness of a post-LAME stable crust (one that stabilized at ~ 2.6 Ga) is 5 ± 2 km thicker than that crust today. Coupled with the assumption of constant freeboard this places limits on the average thickness of oceanic crust during the Archean. Assuming that the area covered by continents in the Archean was comparable to present day, oceanic crust would have been 7 ± 2 km thicker during the Archean (Galer and Mezger 1998). On the other hand, if continental area was minimal (< 5%) in the Archean, oceanic crust would have been 14 ± 2 km thicker (Galer and Mezger 1998). This suggests an oceanic crustal thickness of ~ 15 – 20 km in the Archean (assuming a present day mean thickness of ~ 7 km [White et al. 1992]). This is in good agreement with the estimate for Neoarchean oceanic crust stated earlier, based on Abbot et al. (1994) and McKenzie (1984).

Hence Archean continental crust in isostatic balance with oceanic crust would not be in isostatic balance with secularly thinning oceanic crust and would undergo gradual uplift and erosion over time (Galer and Mezger 1998). The authors claim that it is the mechanism of continental uplift that is responsible for the maintenance of what appears to be an almost constant continental freeboard over time (Galer and Mezger 1992; Harrison 1994).

Based on Galer and Mezger’s (1998) argument, the thickness of a pre-LAME *stable* crust should not have been significantly different from a post-LAME stable crust, namely 46 ± 2 km.
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Galer and Mezger’s (1998) analysis yields a maximum thickness for stable Neoarchean crust, namely with its surface close to sea level. If a characteristic of stable crust is that its surface stabilizes close to sea level it appears to be a truism (though a somewhat paradoxical one) that stable Archean crust was thicker than stable crust today. I suggest that pre-LAME crustal thickness is unlikely to have exceeded 46 km, Galer and Mezger’s (1998) estimate of a stable Neoarchean crust.

3. Mantle heat flow

Pre-LAME mantle heat flow is estimated to have been at least that of present day, namely 10 – 18 mWm\(^{-2}\). Jaupart and Mareschal (1999) estimate that mantle heat flow throughout the Canadian Shield falls in the range 11 – 15 mWm\(^{-2}\). Perry et al. (2006) estimate mantle heat flow in the Superior Province to be 12 – 18 mWm\(^{-2}\). Guillou et al.’s (1994) estimate of mantle heat flow in the Abitibi subprovince is 10 – 14 mWm\(^{-2}\).

4. Crustal heat production

I estimate heat production in the crust during LAME by first estimating current heat production and then extrapolating such heat production back to 2.7 Ga.

4.1. Heat production through geotherms

Rudnick and Nyblade (1999) have compared pressure-temperature arrays derived from thermobarometry of peridotite xenoliths from four Archean cratons (Kalahari, Slave, Superior and Siberia). They assumed that the P-T arrays represent present day conditions in the (upper) lithospheric mantle, noting that the xenoliths range from Mesozoic to Permian and that unless a sub-lithospheric heat flow anomaly is significant, it would take several hundred million years to work its way through a 200-km-thick lithosphere (Nyblade 1999).

Using data for the Kalahari, for which over 100 P-T data points for mantle xenoliths exist, with associated depths of 70 – 200 km, Rudnick and Nyblade (1999) have calculated a range of plausible present day geotherms for the Kalahari craton, along with a best fit geotherm (Fig. 1). They did so by generating approximately 18,000 geotherms corresponding to incremental changes in values of surface heat flow, crustal heat production, crustal conductivity, mantle heat production and crustal thickness, placing prescribed ‘reasonable’ limits on such values. From these they selected those geotherms (referred to as “permissible”) that lay within the 95% confidence limits (two standard deviations) of a linear least-squares fit to the Kalahari P-T data.
The values of the thermal and structural parameters associated with the permissible models served to define a more limited range for each such parameter, a range consistent with the Kalahari data. For example, the crustal heat production used for all models for the Kalahari ranged from $0.3 – 0.9 \, \mu\text{Wm}^{-3}$, but permissible models had a more limited range of crustal heat production, from $0.5 – 0.8 \, \mu\text{Wm}^{-3}$. The crustal heat production for the best fit Kalahari geotherm was $0.7 \, \mu\text{Wm}^{-3}$, with lithospheric mantle heat production of $0 \, \mu\text{Wm}^{-3}$ and mantle heat flow of $18 \, \text{mWm}^{-2}$. The latter value lies within the range proposed by Perry et al. (2006) for the Superior, namely $12 – 18 \, \text{mWm}^{-2}$. The Kalahari geotherm appears to be a good fit to the Superior mantle xenolith data as well (Fig. 1) but, as Rudnick and Nyblade (1999) point out, a best fit geotherm for the Superior (which they compute but do not provide the details of) results in a lower value for crustal heat production due to differences in surface heat flow and crustal thickness between the Kalahari and Superior. This approach towards determining crustal heat production is highly indirect, however, as no measurements of actual heat production of crustal rocks were used. I next examine the results of a more direct approach.

4.2. Heat production through sampling

A direct approach towards determining the vertical distribution of heat producing elements in the crust would be to measure the concentrations of Th, U and K in samples taken along a crustal profile. While no such vertical profiles are available for examination in the Superior, there are two areas that are interpreted as representing oblique cross-sections through the Superior crust.

4.2.1. Michipicoten (Wawa) greenstone belt to Kapuskasing Structural Zone

The first of these (in the central Superior) extends from the Michipicoten (Wawa) greenstone belt near Wawa, Ontario, through an amphibolite grade domal gneiss terrane, to the granulite belt of the Kapuskasing Structural Zone, near Foleyet. It spans a distance of $\sim 100 \, \text{km}$ and is interpreted to correspond to a 25-km vertical profile through a greenstone belt that has been uplifted along a major thrust fault (Percival and Card 1983).

Ashwal et al. (1987) have measured the concentrations of Th, U and K in 58 samples that they consider to be representative of the main lithologies along the 100 km transect. In addition they used the results of major and trace element analyses of 34 samples of metavolcanic rocks from the Michipicoten greenstone belt (Sylvester et al. 1987). They have calculated the mean heat production within the 3 main metamorphic grades present, namely within the Michipicoten greenstone belt, amphibolite terrane, and granulite terrane to be $0.72$, $1.37$ and $0.44 \, \mu\text{Wm}^{-3}$,
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respectively. The mean is based on estimates of the percentage of each rock type present within a given lithologic unit, with the heat production of a particular rock type determined through the sampling.

A model heat production profile appears in Fig. 2. The mean crustal heat production for the section (extending from the surface to 25 km depth) is 0.90 µWm⁻³. We can extrapolate this profile to a 45-km-thick crust, namely one that might have existed during LAME, by considering a 10-km-thick layer on top with mean heat production the same as the Michipicoten greenstone belt and a 10-km-thick layer on the bottom with mean heat production the same as the Kapuskasing granulites. Such a profile yields an average crustal heat production of 0.76 µWm⁻³. If we consider the same profile but with a heat production within the lowermost 10 km of only 0.2 µWm⁻³ (within the lower range of heat production values for granulites [Rolandone et al. 2002]), the average crustal heat production decreases to 0.70 µWm⁻³. This value is the same as that obtained by Rudnick and Nyblade (1999) for the Kalahari craton.

Ashwal et al. (1987) note that “for the Kapuskasing crustal profile, the heat production data are consistent with the granitoid intrusives in the Wawa domal gneiss terrane having been derived from the sampled Kapuskasing granulite terrane, thereby enriching the middle parts of the section and slightly depleting the lower parts in heat producing elements”. Percival and Card (1983) note the presence of both garnet and tonalitic leucosomes within paragneiss and mafic gneiss in the KSZ (Kapuskasing Structural Zone). Therefore conditions existed in the KSZ for the production of tonalite with the fractionated REE pattern typically found in Archean TTG (Chapter 4). This supports the implementation of lower crustal melting as a source of TTG in my models.

4.2.2. Pikwitonei and Sachigo subprovinces

Another area that is interpreted as representing a cross-section through the Superior crust occurs in the Pikwitonei and Sachigo subprovinces (northwestern Superior) in central Manitoba (Fountain and Salisbury 1981). The Sachigo subprovince contains greenschist to amphibolite facies greenstone belts which are surrounded by tonalitic gneisses and granitic plutons. Rocks of the Sachigo (in particular silicic gneisses and some greenstone belts) grade into the Pikwitonei granulite domain to the north and northwest (Fountain et al. 1987).

Fountain et al. (1987) have measured Th, U and K concentrations and the thermal conductivity of 60 samples that span most of the major rock units in the Pikwitonei and Sachigo
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They have constructed a schematic cross-section of the Pikwitonei-Sachigo crust and have assigned a heat production value to each lithologic unit based on their data.

They have determined a representative vertical heat production profile by calculating the weighted mean heat production for their cross section at increments of 2.5 km depth extending down to 30 km (Fig. 2). Peak pressure conditions recorded by the granulites in the Pikwitonei subprovince correspond to 24 – 36 km depth (Arima and Barnett 1984; Mezger et al. 1986; Paktunc and Baer 1986).

I joined data points along their heat production profile with straight line segments and calculated the mean heat production for the resulting profile (extending from the surface to 30 km depth) to be 0.68 µWm⁻³. We can extrapolate this profile to a 50 km-thick crust (one that might have existed during LAME) by considering a 10-km-thick layer on top with heat production the same as the surface heat production of Fountain et al.’s (1987) profile and a 10-km-thick layer on the bottom with heat production the same as their basal heat production. Such a profile leaves the average crustal heat production unchanged at 0.68 µWm⁻³. If we consider the same profile but with a heat production within the lowermost 10 km of only 0.2 µWm⁻³ the average crustal heat production decreases to 0.65 µWm⁻³. These values remain close to that of Rudnick and Nyblade (1999) for the Kalahari craton, namely 0.7 µWm⁻³.

4.3. Heat production through heat flow

Perry et al. (2006) have estimated an average crustal heat production of 0.64 µWm⁻³ for the Superior, based on 64 surface heat flow measurements from the Uchi, English River, Wabigoon, Quetico, Wawa and Abitibi subprovinces.

4.4. Factor increase in heat production between present and 2.7 Ga

The various estimates that I have presented of mean crustal heat production apply to present-day heat production within the Superior crust. But radioactive decay implies that this crust would have been far more radioactive in the Archean. Mareschal and Jaupart (2006) estimate crustal heat production at the time of LAME (~ 2.7 Ga) to be approximately 2.1 times that at present. Mean mantle heat production is estimated to have been twice that at present (Turcotte and Schubert 2002, equation 4–8).

To determine the extent to which such increases in heat production apply to the Superior crust I have used the Th, U and K concentrations of samples measured by Ashwal et al. (1987) and Fountain et al. (1987) to calculate the heat production in these samples at 2.7 Ga (Appendix C). The relative increase in heat production between the present and 2.7 Ga for each such
sample is shown in Fig. 3. The mean increase in heat production for the Kapuskasing samples is 2.3 times. The mean increase in heat production for the Pikwitonei and Sachigo samples combined is 2.4 times. This is somewhat higher than the estimates quoted earlier.

Applying a conservative factor increase in heat production of 2.0 between the present and 2.7 Ga to mean crustal heat production values inferred from Rudnick and Nyblade (1999), Ashwal et al. (1987), Fountain et al. (1987) and Perry et al. (2006) yields mean crustal heat production at 2.7 Ga of 1.3 – 1.5 µWm⁻³. Applying a larger factor increase, of 2.3 (Kapuskasing samples) or 2.4 (Pikwitonei and Sachigo samples), yields mean crustal heat production of 1.5 – 1.8 µWm⁻³ at 2.7 Ga.

5. Regional thickness of TTG and granite produced during LAME

An important constraint on my thermal models is the (regional) thickness of TTG and late granites that were produced during LAME. “Regional thickness” is defined as the total volume produced divided by the area of the Superior Province. In order to test whether a mechanism being modelled can possibly account for LAME I will compare the amount of TTG and granite melt produced in the model with that produced during LAME.

5.1. Late granites

McCaffrey and Petford (1997) have proposed an empirical relationship between the area exposed at the surface of a granitoid pluton and its (subsurface) volume. Cruden (2006) has used this relationship to estimate the regional thickness of late granites produced during LAME to be 1 to 3.1 km; the former value results from the assumption that all granitic plutons are wedge-shaped, the latter that they are tabular. Cruden (2006) points out that published gravity and seismic reflection studies favour the assumption that Superior granitic plutons are wedge-shaped (Jackson et al. 1995; Everitt et al. 1998). Furthermore, as a single contiguous area of granites (treated as a single pluton in the approach used by Cruden [2006]) may in fact correspond to a collection of individual plutons, the above estimates should be regarded as upper bounds. Counterbalancing this is the fact that some granitic rocks (both intrusive and extrusive) have been lost through erosion while others may occur in the subsurface without any portion being exposed at the surface.
5.2. TTG

Unfortunately the above approach cannot be used to estimate the regional thickness of TTG produced during LAME, as TTG occurs in both batholithic and gneissic form within the Superior crust. Such TTG includes both juvenile TTG and TTG with a pre-LAME model age.

A preliminary step to estimating the regional thickness of TTG emplaced during LAME is to estimate the total regional thickness of TTG in the Superior. In the southern portion of the western Superior Province, where granitoid rocks have been mapped on a scale of at least 1:63,360 to 1:50,000, Beakhouse (2007) has estimated the areal percentage of the exposed lithologies (Table 1). The area is dominated by granitoid rocks (> 70%), with TTG comprising 37% of the total area. That the dominance of granitoid rocks extends to depth is supported by other studies, some of which are discussed below.

As previously mentioned, the area extending from the Michipicoten greenstone belt to the granulite belt of the Kapuskasing Structural Zone (KSZ) is thought to represent an oblique cross section through the Superior crust (Percival and Card 1983). The generalized vertical cross-section that is inferred from the oblique section consists of 3 main layers. The upper 5 – 10 km consist of the Michipicoten greenstone belt, intruded by granitoid plutons and surrounded by gneissic migmatitic haloes. The middle 10 – 15 km consist of tabular batholiths of gneissic and xenolithic TTG. The bottom 5 – 10 km consist of upper amphibolite to granulite-facies heterogeneous gneiss and anorthosite; the base of this layer is not exposed.

The vertical cross-section inferred from the Pikwitonei-Sachigo region is dominated by quartzo-feldspathic gneisses (primarily tonalitic to granodioritic gneisses and migmatites) and silicic to intermediate gneisses (Fountain and Salisbury 1981).

Velocity models for seismic refraction profiles across the Abitibi and Pontiac subprovinces are consistent with: (1) an upper crust (~ 0 – 12 km) containing mafic metavolcanics, metasediments and granitic plutons, (2) a relatively uniform middle crust (~ 12 – 30 km) containing quartzo-feldspathic lithologies and (3) a lower crust (~ 30 – 40 km) dominated by mafic lithologies (Grandjean et al. 1995).

Assuming a crustal thickness of 40 km for the Superior crust and that Beakhouse’s (2007) estimate of the areal percentage of TTG reflects the percentage of TTG occurring at depth, the estimated total thickness of TTG is ~ 15 km. This is in good agreement with the thickness of TTG inferred from the Kapuskasing transect. I assume that this thickness is broadly reflective of that across the Superior.
Next, we must arrive at an estimate of how much of this total TTG, consisting of both juvenile TTG and TTG with a pre-LAME model age, was emplaced during LAME.

Based on 210 whole rock Sm-Nd analyses of mafic to felsic metavolcanic, plutonic, gneissic and metasedimentary rocks from the North Cariboo terrane and the Uchi, English River, Winnipeg River, Wabigoon, Quetico and Wawa subprovinces (Noble 1989; Stevenson 1995; Henry et al. 1998; Larbi et al. 1999; Henry et al. 2000), Henry et al. (2000) have identified three main periods of crustal growth in the Western Superior Province at ca. 2.7, 3.0 and 3.4 Ga in which juvenile additions from the mantle dominated over crustal recycling. Using the composition of detrital metasedimentary rocks to estimate the average composition of the crust, they calculate that the Western Superior Province is composed of 48 wt%, 44 wt% and 7 wt% of crust created (derived from the mantle) between 2.69 – 2.76 Ga (LAME), 2.92 – 3.02 Ga and 3.4 ± 0.1 Ga respectively. An alternative calculation yields 33 wt%, 48 wt% and 19 wt% of crust created between 2.69 – 2.76 (LAME), 2.92 – 3.02 and 3.4 ± 0.1 Ga respectively (Henry et al. 2000). The authors suggest that these results are broadly applicable to much of the Superior Province.

While the logic of the following argument is admittedly flawed, one could argue that relative to the total thickness of TTG, the percentage that was emplaced during LAME (derived from both juvenile and older sources) is unlikely to be less than 48 wt% (Henry et al.’s [2000] estimate of the percentage of juvenile Superior crust created during LAME). I choose 48 wt% rather than Henry et al.’s (2000) lower estimate of 33 wt% because of the dominance of LAME-aged rocks in the Superior Province (Fig. 2 in Chapter 1). This yields an estimated 7.2-km-thickness of TTG emplaced during LAME, with the possibility that the thickness was even greater, say 10 km. The flaw in the argument, of course, is that Henry et al. (2000) do not specify the percentage of juvenile component of each major rock type, such as TTG.

Tectonic thickening during the later stages of LAME post-dated the emplacement of most TTG (Percival et al. 2006). The thickening factor is unlikely to have exceeded 1.5 (Chapter 7). This suggests an estimated 5 – 7 km thickness of TTG was emplaced during LAME. Assumption of a less severe thickening factor of 1.25 (Chapter 6) yields an estimated 6 – 8 km thickness of TTG emplaced during LAME. I estimate the regional thickness of TTG emplaced during LAME to be between 5 – 8 km, and likely more, as my analysis essentially focuses on only juvenile TTG.
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5.3. Other Archean TTG events

My estimate of the regional thickness of TTG emplaced during LAME is comparable to that emplaced during other Archean TTG events. Zegers and van Keken (2001) estimate the emplacement of a 5 – 10 km regional thickness of TTG between 3.49 and 3.42 Ga in the Pilbara craton. Similarly, Wells (1980) constructs a thermal model of a TTG event that occurred between 2.9 and 2.8 Ga in West Greenland in which he emplaces a total thickness of ~ 50 km of TTG into his model crust. This is likely excessive but does reflect the fact that a significant thickness of TTG was emplaced during the event.

6. Conclusion

Due to the significant reworking of the Superior crust during LAME and lack of knowledge of the precise source of the TTG that was emplaced during LAME, there is great uncertainty associated with any attempt to characterize the pre-LAME crust. The thermal models that I will construct of the Superior crust during LAME, however, require just such a characterization.

I estimated pre-LAME crustal thickness to have been between 20 – 46 km. I estimated the average crustal heat production of the LAME crust to have been 1.3 – 1.5 µWm$^{-3}$. Mantle heat flow is estimated to have been no less than present day mantle heat flow, so 10 – 18 mWm$^{-2}$ (Guillou et al. 1994; Jaupart and Mareschal 1999; Perry et al. 2006). I estimated the regional thickness of TTG emplaced during LAME to have been 5 – 8 km, while the regional thickness of late granites emplaced has been estimated by Cruden (2006) as 1 – 3.1 km. Comparing the thickness of TTG and granite produced in my models with these estimates is an important test of whether a model can possibly account for LAME.
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References


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Table 1. Areal percentage of lithologies occurring in the southern part of the western Superior Province. See Fig. 1 in Beakhouse (2007) for a map of the relevant area. Only Archean rocks are considered. Proterozoic rocks form a small fraction of the total area (1.97%) and are ignored. Modified from Beakhouse (2007).

<table>
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<th>Lithology</th>
<th>Total Area %</th>
<th>Area %</th>
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Fig. 1. “Present day” geotherm that is a best fit to mantle xenolith P-T data from the Kalahari craton superimposed on mantle xenolith temperature-depth data from the Superior Province (from Rudnick and Nyblade 1999).
Fig. 2. Heat production profiles for the Superior crust inferred from sampling along the Kapuskasing (Ashwal et al. 1987) and Pikwitonei-Sachigo (Fountain et al. 1987) transects.
Fig. 3. Factor increase in heat production between present day and 2.7 Ga in samples obtained along (a) the Kapuskasing transect, and (b) the Pikwitonei-Sachigo transect.
Abstract

Any model that is proposed for LAME must be able to account for the amount and timing of the voluminous quantities of TTG and late granites produced during LAME. I review the results of several experimental studies on the partial melting of amphibolites which support the idea that TTG results from 10 – 40 vol% batch melting of metabasalt under “vapour absent” conditions in the garnet-amphibolite or eclogite stability field. Based on the experimental data, I construct a melting curve for metabasalt that gives vol% melt as a function of temperature. The associated solidus temperature is 850°C with 35 vol% melting at 1025°C. I intend to use this melting curve to determine the amount of TTG melt that is produced through lower crustal melting in my thermal models. There is a difference of up to 150°C in the experimental temperatures that produce a given melt fraction. This is likely due to differences in the bulk composition of the starting materials but adds considerable uncertainty to results inferred from the use of the melting curve.

The source rock for the late granites is thought to be pre-existing TTG or felsic metasedimentary rocks. I use the results of a single experimental study on partial melting of tonalite to construct a melting curve that I will use to determine the amount of granite produced in my thermal models. The associated solidus is 827°C with 35 vol% melting at 980°C.

1. Introduction

Voluminous quantities of tonalite-trondhjemite-granodiorite (TTG) and late granites were produced during LAME in the Superior Province, estimated at regional thicknesses of 5 – 8 km and 1 – 3.1 km respectively (Chapter 3).

A key test of whether a given thermal model can account for LAME is to compare the amount and timing of TTG and granite produced during LAME with that indicated in the model. An example of a model that has been proposed for LAME but which fails this somewhat exacting test is that of Rey et al. (2003) in which crustal temperatures remain too low for significant TTG melt production (Chapter 6).
To determine the amount of melt produced in my models I take a simplistic approach and use melting curves that give the volume % melt that would be produced in metabasalt (source for TTG) or tonalite (a source for granite) at a given temperature. The aim here is to review the melting experiments that provide the data needed to construct such curves.

2. Partial melting origin of TTG

At present there appears to be a general consensus amongst petrologists that the source rock for TTG is metabasalt and that TTG results primarily from dehydration melting of this source. This conclusion is supported by the results of amphibolite melting experiments.

Whether TTG in the Archean resulted primarily from lower crustal melting or melting of the oceanic crust of a subducting slab is a matter of huge debate. Strong arguments and counterarguments exist on both sides (Martin 1986; Martin 1994; Smithies 2000; Zegers and van Keken 2001; Kamber et al. 2002; Martin and Moyen 2002; Condie 2005; Martin et al. 2005; Valley et al. 2005; Bédard 2006; Condie 2008; Foley 2008; Nair and Chacko 2008), a few of which I will outline here.

2.1. Slab melting origin for TTG

To produce 5 – 8 km thickness of TTG, namely the regional thickness of TTG produced in the Superior Province during LAME, requires the melting of a minimum thickness of 12.5 – 20 km of metabasalt. Once such a thickness of basalt has cooled to crustal temperatures extraordinary thermal conditions (mantle plume, delamination, etc.) appear to be required to reheat the entire basaltic packet to the point that TTG (associated with 10 – 40 vol % melting) is produced throughout the entire thickness. Furthermore, evidence is lacking for a widespread high-velocity mafic lower crustal restitic layer (though such a layer may well be present but being interpreted as part of the mantle, or may have delaminated from the crust).

The above observations appear to support a slab melting model for TTG, in which TTG is produced in an incremental fashion that only requires reheating of a much thinner basaltic source that is constantly replenished through continued subduction and automatically disposed of deep in the mantle. However, it is questionable if the rate of TTG melt production under such conditions would have been sufficient to meet Archean constraints, such as in the Minto Block, Superior Province (Bédard 2006).
Chapter 4: Melting curves for the production of TTG and granite

2.2. Lower crustal melting origin for TTG

The presence of amphibolitic migmatites containing tonalitic leucosomes in regions such as the Kapuskasing Structural Zone indicate that some TTG was indeed derived from mid- to lower crustal melting of amphibolite (Percival and Card 1983). However, such migmatites have not been observed on a scale large enough to account for the voluminous quantities of TTG that were produced during the Archean.

In the central Wabigoon and other areas of the Superior Province TTG with Nd model age significantly older than its crystallization ages is found (Whalen et al. 2002; David et al. 2009). This is evidence that some Archean TTG was indeed derived from (or at least incorporated) older (non-juvenile) crust.

The relatively low Mg\# of pre-3.0 Ga TTG and of the majority of post-3.0 Ga TTG suggests that little interaction with the mantle wedge occurred, making a slab melt origin for such TTG unlikely (Smithies 2000).

As well, peridotite xenoliths or metamorphosed equivalents are not commonly found in TTG. This is an original, and at this stage, tentative argument against a slab melt origin for TTG, but one that I would like to explore following the completion of this thesis. Mantle xenoliths are common in lavas that have a known origin within the mantle, such as basalt and komatiites. It does not appear plausible that a silica-rich, and hence viscous, magma such as TTG could pass through the mantle wedge without incorporating mantle fragments. A definite possibility is that the longer residence time at high temperature in an intrusive magma compared to that in an extrusive magma may render peridotite xenoliths unrecognizable. The lack of peridotite xenoliths may also be explained by the absence of a mantle wedge due to an extremely shallow angle for subducting slabs in the Archean. This is implausible, however, as a convecting mantle wedge is the primary source of heat in subduction zones. Without it, voluminous quantities of magma are unlikely to be produced (Gutscher et al. 2000), particularly over a span of tens of millions of years, as during LAME and other Archean TTG events.

For the purposes of this work I therefore adopt the view that the primary mechanism for TTG generation in the Archean was via melting of the lower crust.

3. Metabasalt melting experiments

I will be reviewing the results of experimental studies of “vapour-absent” melting of metabasalt. In the experimental studies I will be reviewing, (organized by authors), experimental
melt compositions are compared against ‘average’ or ‘representative’ major and trace element compositions of Archean TTG and conclusions drawn as to whether Archean TTG could have been produced from the experimental source rock under the experimental conditions being tested.

My aim is to construct a melting curve for metabasalt that gives the volume % melt as a function of temperature. As will be seen, however, there is a wide range of experimentally determined melt fractions that correspond to similar temperatures and pressures. This is a point of significant uncertainty, whether the amount of TTG melt produced in my models (as determined by use of my melting curve) is a reasonable estimate of the amount of melt that would be produced in nature under the conditions being modelled.


Winther and Newton (1991) conducted melting experiments on two low-K tholeiites under a range of temperatures, pressures and degrees of H$_2$O-undersaturation, one a high-Al (18.2 wt% Al$_2$O$_3$) natural basalt from Hat Creek, California, the other a synthetic low-Al (14.8 wt% Al$_2$O$_3$) average Archean tholeiite (Condie and Hunter 1976). Average Archean tholeiite (AAT) is similar to average modern mid-ocean ridge basalt (MORB) in all major elements except iron, in which AAT is enriched compared to MORB. Runs spanned a pressure range of 0.5 – 3 GPa and a temperature range of 750 – 1100°C. Water was added in the form of Al(OH)$_3$ and liquid water, totalling approximately 1%, 2%, 5% and 15% by weight. The high-Al basalt did not produce partial melts of natural TTG composition, their Al$_2$O$_3$ content being too high. The low-Al AAT yielded partial melts with compositions similar to those of typical Archean tonalitic and trondhjemitic “grey gneisses” over a range of conditions.

The precise melt percentages produced in the AAT are not given, but other results emerge from their experiments that constrain the conditions under which TTG may be produced and hence are useful for my modelling work.

All runs on the average Archean tholeiite (AAT) produced either tonalitic or trondhjemitic quenched melt compositions, according to the classification of O’Connor (1965) (Fig. 1). This classification considers only feldspar components, however. The major and minor oxide compositions of quenched melt from all runs producing tonalite (22 runs) and trondhjemite (27 runs) were averaged, respectively, by Winther and Newton (1991). These averages were compared to the composition of the average Amitsoq tonalite (McGregor 1979) and the type locality Trondheim (Follstad) trondhjemite, respectively (Table 1).
In general, higher H$_2$O contents tended to produce tonalitic (rather than trondhjemitic) melts, as did higher temperatures and lower pressures for a given H$_2$O content.

Garnet phenocrysts were ubiquitous in runs above 1 GPa except for those with excess H$_2$O (15%). Residual garnet appears necessary to account for the HREE depletion found in most Archean TTG.

3.1.1. Discrepancy between experimental results and natural TTG

Among tonalites, K$_2$O is low in the average quenched melt composition (0.74%) compared to that for the average Amitsq tonalite (1.53%), reflecting the low-K content of the source. Among trondhjemites, Al$_2$O$_3$ is low in the average quenched melt composition (15.4%) compared to that for the Follstad trondhjemite (16.6%). Other oxide components of the natural rocks fall within one standard deviation of those of the average quenched melt compositions or are very close to them.

There are individual runs however in which K$_2$O and Al$_2$O$_3$ contents of the melt do approach those of the natural rocks. As amphibole enters the melt its K$_2$O content increases. This is favoured by lower H$_2$O content and higher pressures (approaching the stability limit of amphibole). (Amphibole phenocrysts were abundant in most runs below 2 GPa and temperatures up to 1100°C, the maximum temperature tested).

3.1.2. Effect of H$_2$O saturation

Runs with excess H$_2$O (15%) resulted in tonalitic partial melts; however, except for a few runs below 850°C and above 1.5 GPa, Al$_2$O$_3$ and CaO contents were higher than those found in typical tonalitic grey gneisses. On the basis of this and of the experimental work of Beard and Lofgren (1989), Winther and Newton (1991) concluded that water-saturated partial melting of basalt was not a primary mechanism for the production of TTG during the Archean. Wolf and Wyllie (1994) concur with this conclusion.

3.1.3. Ideal conditions to produce tonalite and trondhjemite

Instead, Winther and Newton (1991) found that temperatures of 850 – 1000°C and pressures around 1.5 GPa would be appropriate for the production of tonalite and trondhjemite from average Archean tholeiite, and suggested that such conditions could exist either in a shallow subduction-zone setting or in a deep crustal metamorphic setting. Their results for an AAT with approximately 2 wt% H$_2$O are summarized in the form of a semi-schematic P-T stability diagram (Fig. 2).

Rapp (1990), Rapp et al. (1991) and Rapp and Watson (1995) conducted melting experiments on four natural amphibolite rock powders. Each sample was subjected to a range of pressures from among 0.8, 1.6, 2.2 and 3.2 GPa and temperatures resulting in 10 to 40% melt. Three of the samples (#2, #3 and #4) were low-K, MORB-like olivine tholeiites while the fourth (#1) was an alkali-rich olivine basalt. Sample #1 was taken from the uppermost pillow lava flow of the Josephine Ophiolite Complex, Klamath Mountains, California (Harper 1984). Sample #2 was from the Post Pond Volcanics member of the Ammonoosuc Volcanics, Vermont and New Hampshire (Spear 1982). Mineral assemblages of the Post Pond Volcanics consist of hornblende + andesine +/- quartz +/- Fe-Ti oxide +/- carbonate. Sample #3 was from Nunatak Fjord, Alaska, with a mineral assemblage of aluminous hornblende, plagioclase, epidote, sphene and quartz (Barker et al. 1985). Sample #4 was from pillowed greenstone belt sequences in the Wind River Range, Wyoming. It has been interpreted as a dismembered Late Archean ophiolite (Harper 1985). The samples spanned a range of major element chemistry and contained ~ 1 – 2 wt% water in the form of amphibole. Only in one experiment (at 0.8 GPa and 900°C) was 1 – 2 wt% water added to three of the samples (#1, #2 and #3).

The feldspar compositions of all resulting partial melts that contained more than 10% CIPW normative quartz (29 runs) are shown in Fig. 3 on a ternary feldspar diagram, based on the classification of Barker (1979). The effect of source composition on the type of melt produced is apparent: samples #1 and #2 produced only trondhjemitic melt, except for two low temperature runs for sample #2 that resulted in granitic melt. Melt from sample #3 spanned both the trondhjemite and tonalite fields, whereas melt from sample #4 was entirely tonalitic. It is interesting to note that none of the melts were granodioritic; this is in agreement with the results of Winther and Newton (1991) but leaves open the question of the petrogenesis of the granodioritic component of Archean TTG.

Rapp et al. (1991) compared the major element chemistry of the experimental melts with those of chosen representative Archean tonalite and trondhjemite compositions. As found by Winther and Newton (1991), the K₂O content of most of the experimental melts is less than 1.0 wt% while, on average, Archean TTG has 2.2 wt% K₂O. The K₂O content of some melts does exceed the Archean average (ranging up to 3.8 wt%): such melts are either the result of low degrees of melting or are derived from samples with higher initial K₂O content.
The MgO content of the experimental melts (0.09 – 2.84 wt%) is also lower than that of typical Archean TTG (0.93 – 3.4 wt%). Rapp et al. (1991) suggest that this may be due to more Mg-rich tholeiites during the Archean (Condie 1985).

The Na₂O content of the experimental melts (3.5 – 7.55 wt%) is generally higher than that of Archean TTG (4.38 – 4.87 wt%). Rapp et al. (1991) propose that fractionation of sodic plagioclase prior to emplacement of Archean TTG might explain this discrepancy.

The Al₂O₃ content of the experimental melts (12.46 – 19.71 wt%) is also generally higher than that of Archean TTG (14.52 – 16.09 wt%). Winther and Newton (1991) found no such discrepancy, but their estimate of the Al₂O₃ content of ‘representative’ Archean tonalite and trondhjemite is 16.3 wt% and 16.6 wt% respectively, higher than that of Rapp et al. (1991).

The REE patterns of Archean TTG tend to be highly fractionated (namely high La/Yb ratios) and HREE-depleted (Jahn and Zhang 1984; Martin 1987). Quantitative estimates are La/Yb = 5 – 70 and Ybₙ = 0.3 – 8.5 (Martin 1986). Rapp et al. (1991) calculated the REE patterns of some of their experimental melts. Based on these patterns one can rule out a low pressure (0.8 GPa) origin for Archean TTG, in which the residue is amphibolite. A residue of garnet-amphibolite (1.6 GPa) or eclogite (2.2, 3.2 GPa) appears necessary to achieve REE patterns comparable to those of Archean TTG.

Rapp et al. (1991) found that the temperature range over which tonalitic and trondhjemitic melt is produced at 0.8, 1.6 and 2.2 GPa is quite narrow (approximately 50°C), increasing somewhat at 3.2 GPa. Strictly speaking, experiments conducted at 3.2 GPa are no longer “dehydration melting” experiments as amphibole is unstable at pressures above 2.7 GPa (Rapp and Watson 1995). An alternate source of water, such as small amounts of pore water or unsegregated “wet” silicic melt, is required. Amphibole can still be the source of such fluids, however, if it originates at pressures within its stability field and is then carried to depth.

Rapp et al. (1991) presented their results in the form of a schematic in which P-T conditions likely to result in the production of tonalitic and trondhjemitic melt are shown (Fig. 4). It is interesting to note that there is a difference of more than 100°C between their inferred vapour-absent solidus, and that of Winther and Newton (1991) for a basalt with ~ 2 wt% H₂O. The precise reason for this difference is unclear, although it is likely due to compositional differences in the starting materials. The water content of the Rapp samples is less than 2 wt% (~ 0.70 – 1.65 wt%). The difference in the solidi demonstrates the uncertainty associated with any melting curve employed to determine the amount of melt produced in my thermal models.
3.3. Sen and Dunn (1994)

Sen and Dunn (1994) conducted dehydration melting experiments on an amphibolite from British Columbia, Canada, at pressures of 1.5 and 2 GPa and temperatures ranging from 800 – 1150°C. The mineral assemblage of the starting amphibolite consisted of hornblende (76.3 wt%), An$_{55}$ plagioclase (20.5 wt%), quartz (2.3 wt%), titanite (0.9 wt%) and garnet (< 0.1 wt%) with a water content of ~ 1.5 wt% in the form of amphibole. Though their aim was to test the hypothesis that adakites are produced by melting of a subducting slab their experimental results can be applied equally well to the production of TTG, especially as they compared their results to those of Rapp et al. (1991). Adakites are arc rocks that are interpreted to be partial melts of young (< 25 Ma), relatively hot, subducted oceanic crust (Defant and Drummond 1990). Martin and Moyen (2003) have subdivided adakites into a high-SiO$_2$ (HSA; SiO$_2$ > 60 wt%) and low-SiO$_2$ (LSA; SiO$_2$ < 60 wt%) group and suggest that it is only HSA that are slab melts and modern analogues of Archean TTG. HSA also have lower MgO (0.5 – 4 wt%) than LSA (MgO = 4 – 9 wt%) (Martin et al. 2005).

Trace amounts of melt were found at all temperatures tested, but significant melting (1 – 2 wt%) didn’t occur until 900°C at 1.5 GPa and between 800 and 850°C at 2 GPa. The feldspar compositions of the melts are shown in Fig. 5, where they are compared to those of other dehydration melting experiments (Rapp et al. 1991; Rushmer 1991; Wolf 1992). At both 1.5 and 2 GPa, with increasing temperature, the composition of the melt progressively changes from granitic to trondhjemitic/granodioritic to tonalitic. Rapp et al. (1991) noted that feldspar compositions of average Archean TTG tend to plot near the trondhjemite-tonalite-granodiorite triple point (Martin et al. 1983). This is the case for Sen and Dunn’s experimental melts if one ignores the low fraction granitic melts (Fig. 5).

Sen and Dunn (1994) compared the major element compositions of their experimental melts with those of adakites and with the experimental melts of Rapp et al. (1991). I am primarily interested in the latter comparison. The Na$_2$O and K$_2$O content of Sen and Dunn’s (1994) melts are lower and higher, respectively, than those of Rapp et al. (1991). The amphibolite used by Sen and Dunn (1994) had a lower Na$_2$O/K$_2$O ratio than the amphibolites used by Rapp et al. (1991). With regard to the other major oxides (Al$_2$O$_3$, FeO, CaO and MgO) there is significant overlap between their results and those of Rapp et al. (1991).
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3.4. Rushmer (1991)

Rushmer (1991) conducted dehydration melting experiments at 0.8 GPa on two contrasting natural amphibolites, a meta-alkali basalt (ABA) with an H$_2$O content of ~ 1 wt% contained in hornblende, and a meta-island-arc tholeiite (IAT) with an H$_2$O content of ~ 1 – 1.3 wt% contained in hornblende, cummingtonite and biotite. The ABA was from the amphibolite facies section in the Ivrea Zone, northern Italy, with a mineral assemblage of Mg-hornblende (54.4 wt%), An$_{40-42}$ plagioclase (36.4 wt%), quartz (8.4 wt%) and Fe-Ti oxides (0.8 wt%). The IAT was from the Hidaka metamorphic belt in northern Japan with a mineral assemblage of actinolitic-hornblende (44.0 wt%), An$_{40-45}$ plagioclase (32.0 wt%), quartz (17.0 wt%), cummingtonite (5.0 wt%), biotite (1.0 wt%) and Fe-Ti oxides (1.0 wt%). Rushmer’s work clearly demonstrates the significant effect of composition on both the solidus and the temperature span over which melting occurs.

The “solidus” of the ABA at 0.8 GPa was found to be 925°C, at which 3 – 5 vol% melt was present. The melting interval was between 925 and 1000°C, with melt reaching 45 – 48 vol% at 1000°C. The solidus of the IAT at 0.8 GPa was 800°C. The melting interval was 800 to 950°C with melt reaching 35 – 40 vol% at 950°C. Rushmer attributed the lower melting temperature of the IAT primarily to the breakdown of accessory biotite and cummingtonite but pointed out that the presence of a significant tremolite component in the starting hornblende and of significant quartz (17 wt%) also played a role in the lower melting temperatures of this amphibolite.

Rushmer (1991) claimed that the quenched melt compositions plot in the tonalite field in the Ab-An-Or projection. Technically this may be true but the anorthite component of her quenched melts (or at least those for which she was able to obtain detailed compositions, namely ABA at 950 and 1000°C and IAT at 950°C) is much higher (see numbered field 6 in Fig. 5) than what is currently recognized for natural tonalites. Furthermore, garnet does not appear as a phase in Rushmer’s melting experiments, hence the melt produced is unlikely to possess the highly fractionated and HREE-depleted signature of Archean TTG.

3.5. Wolf and Wyllie (1994)

The final experimental work that I discuss is that of Wolf and Wyllie (1994). They conducted dehydration melting experiments on a low-K, high-Mg, calcic amphibolite (67.4% hornblende, 32.5% anorthite) at 1 GPa with temperatures ranging from 750 to 1000°C. The amphibolite sample was from the Owens Mountain area of the western Foothills Metamorphic Belt of the Sierra Nevada, California. Its solidus temperature was found to be less than 750°C. Wolf and
Wyllie (1994) did note that it was difficult to achieve equilibrium at temperatures below 900°. Hence experimental results for temperatures below 900°C may or may not be reliable. Hornblende persisted till just above 975°C, hence the melting interval was significant, from 750 to 975°C. Melt volume reached 47% at 1000°C. Garnet appeared at 850°C. As garnet was not a phase in any of Rushmer’s (1991) experiments (conducted at 0.8 GPa) the garnet-in boundary must lie between 0.8 and 1 GPa, though its precise position is of course composition-dependent. The experimental melts produced were tonalitic until hornblende was fully consumed (just above 975°C) at which point the melts became quartz dioritic. The tonalitic melts were however significantly more calcic (plotting within numbered field 5 in Fig. 5) and aluminous (18.95 – 22.39 wt% Al₂O₃) than natural tonalites, likely due to the plagioclase composition (An⁹⁰).


Moyen and Stevens (2006) conducted a more comprehensive review and analysis of the published data on experimental melting of amphibolites than I have above. They concluded that major and (calculated) trace element compositions of the experimental melts suggest that TTG melts formed at pressures greater than 1.5 GPa and temperatures between 900°C and 1100°C. They suggest that the low geothermal gradients associated with such P-T conditions are likely to occur only in subduction zones. In fact crustal temperatures of 900 – 1100°C at depths of ~ 50 km (corresponding to ~ 1.5 GPa) are not low, even for the Archean!

Based on the range of compositions of natural TTG however, they concluded that TTG was generated over a large range of depths, from 1.0 – 2.5 GPa (corresponding to ~ 33 to 83 km).

3.7. Implications for melt segregation models

All of the experimental data cited above supports a petrogenetic model for TTG that involves batch melting of metabasalt. However, at least 10 – 15 vol% (but no more than 40 vol%) melting is required to produce a melt of TTG composition. If this petrogenetic model is correct it implies that smaller melt percentages formed in the initial stages of melting must remain within the metabasaltic source. This is contrary to the contention of some that the percentage of melt in a source region is unlikely to exceed a few percent at any given time (Philpotts 1990).

Alternatively, Wickham (1987) suggests that melt fractions of 25 – 30 vol% are necessary for efficient segregation of a granitic (sensu lato) melt from its source. This view appears to be more consistent with the batch melting model for TTG and also with the voluminous quantities of TTG emplaced during LAME (Chapter 3) and other Archean TTG events (Wells 1980;
Zegers and van Keken 2001). It may be that the relatively high solidus temperature of metabasalt leads to more ductile conditions in the source region, allowing a higher melt fraction to be retained before fracturing occurs.

An enigmatic aspect of this petrogenetic model is the nature of the controls that led primarily to the production of TTG melt in the Archean, rather than granitic or dioritic melt, associated with lesser and greater degrees of melting, respectively.

4. Melting curves for metabasalt

None of the experimental studies discussed above have reproduced all the geochemical characteristics currently associated with natural tonalites and trondhjemites. Neither is it the case that all the discrepancies have been satisfactorily accounted for. However, the similarities between natural TTG and the experimental melts are sufficient that the currently favoured model for TTG petrogenesis is that it represents a batch partial melt of metabasalt produced under “vapour absent” conditions in the garnet-amphibolite or eclogite stability field.

An important constraint on my thermal models is the volume of TTG that was produced during LAME. To that end I require a melting curve that gives the volume % of TTG melt produced from a metabasaltic source as a function of temperature. Petford and Gallagher (2001) constructed such a curve using data from all of the previously mentioned experimental studies (except Winther and Newton [1991]) as well as one by Beard and Lofgren (1991) (Fig. 6).

There are two problems with Petford and Gallagher’s (2001) curve. The first is that they used a mix of vol% melt and wt% melt data without converting the wt% data (of Rapp and Watson [1995], and Sen and Dunn [1994]) to vol% data. The second is that the parameterization of the curve given in their paper (equation 2 in Petford and Gallagher [2001]) yields a curve that is different from that which appears in their graphical representation (Fig. 1 in Petford and Gallagher [2001]).

I constructed a melting curve (Fig. 6) based on the same data employed by Petford and Gallagher (2001), with two exceptions. I omitted the data of Beard and Lofgren (1991) for experiments conducted at 0.69 GPa (corresponding to too shallow a depth) and added additional data from Sen and Dunn (1994) for experiments conducted at 2 GPa. All such data, except that of Sen and Dunn (1994), is given in vol% form in Wolf and Wyllie (1994), which I used, as well as the data of Sen and Dunn (1994) converted to vol%.
Chapter 4: Melting curves for the production of TTG and granite

The data did not appear to warrant anything more complicated than a linear approximation. I first calculated a “best fit” line ($r^2 = 0.48$) but that did not appear to reflect the trend of the data. I then chose a line that seemed a good visual fit to the trend. Its equation is given by:

$$m_{\text{basalt}} = 0.2T - 170,$$

where $m_{\text{basalt}}$ denotes volume % melt and $T$ denotes temperature in °C. The corresponding solidus temperature is 850°C, somewhat higher than that associated with Petford and Gallagher’s (2001) melting curve, namely 822°C. The calculated $r^2$ value is 0.21. It’s interesting to note that Petford and Gallagher (2001) deal with data with a similar spread and fit a curve to it for which they claim $r^2$ equals 0.92. I am sceptical of this result.

In visually fitting a line to the trend of the data I essentially took the approach of choosing a line that passed through the middle of the range of temperatures associated with each melt fraction. In terms of regression this approach is better reflected by allowing melt fraction, rather than temperature, to be the independent variable. Fig. 7 shows the results of using such an approach, in which the best fit line to the modified data set differs by less than 5°C from my melting curve in the solidus. My melting curve is skewed towards higher temperature data but this is desirable as such data are associated with melting at higher pressures as occurs in my models (Chapter 6).

In my thermal models I assume that up to 35 vol% melt gives melt of TTG composition. This is consistent with the experimental data and with Wickham’s (1987) estimate of the melt fraction necessary for efficient segregation of a granitic melt from its source.

5. Late granites

5.1. Source rock

I am not only interested in determining the volume of TTG produced in my thermal models but also the volume of late granites. Due to the high K-content of the granites (4 – 6 wt% K₂O), it is unlikely that they are derived directly from metabasalt. Winther and Newton (1991) test this hypothesis in their experimental study, in which they consider Late Archean granites in general, and unequivocally rule out this possibility. In the other experimental studies discussed above only a handful of the partial melts produced fall into the granite field (at extremely low degrees of melting) but none possess the high K-content associated with the late granites. Given that up to 90% of felsic rocks in early Archean granite-greenstone and high-grade gneiss terrains may consist of tonalites and trondhjemites, with lesser granodiorites (Glikson and Sheraton 1972;
Windley and Smith 1976; Hunter et al. 1978; Martin et al. 1983; Nutman and Bridgewater 1986; Martin 1987), these are a compelling candidate for the source of late granites, along with felsic metasedimentary rocks. The melting temperature of tonalite exceeds that of felsic metasedimentary rocks, hence the volume of granitic melt obtained in our models through use of a tonalite melting curve will be a minimum bound on the amount of melt that can be expected.

Rutter and Wyllie (1988) conducted vapour-absent melting experiments at 1 GPa and at temperatures between 775 and 1000°C on a tonalite sample obtained from the Central Sierra Nevada Batholith, California, with modal composition 13.1% quartz, 4.4% K-feldspar, 58.9% plagioclase and 23.6% mafic minerals (Piwinskii 1968). The sample contained 0.8 wt% H₂O stored in biotite (12.5 modal %) and hornblende (9 modal %). Between 825 and 900°C biotite melting produced 20% melt and by 1000°C hornblende had melted producing 35% melt. Pelitic gneisses with higher biotite content would be expected to yield even greater melt at the above temperatures. Rutter and Wyllie (1988) do not discuss melt composition but do make a reference to a “granodiorite magma”.

Melt composition is discussed in detail by Patiño Douce (2005) who conducted vapour-absent melting experiments on a tonalite at pressures of 1.5 to 3.2 GPa and temperatures of 900 to 1150°C. I do not use the data from his study because 1.5 GPa is greater than the pressures I consider for the source of the late granites in my thermal models. However, all partial melts produced in the study (ranging from trace amounts to 50 wt% partial melt) are granitic, not granodioritic, in composition. In my thermal models I only consider up to 35 vol% melt as representing granitic melt as segregation from the source is likely to occur at higher melt fractions.

5.2. Melting curve for tonalite

My melting curve for tonalite is the “best fit” line (r² = 0.93) to the data of Rutter and Wyllie (1988) (Fig. 8). Its equation is given by:

\[ m_{\text{tonalite}} = 0.2282 T - 188.68, \]

where \( m_{\text{tonalite}} \) denotes volume % melt and \( T \) denotes temperature in °C. The corresponding solidus temperature is ~ 827°C.

5.3. H₂O release from crystallization of tonalite

If tonalite is indeed the result of say up to 40 vol% melting of amphibolite then we would expect the liquidus for tonalite to not exceed the temperature at which 40% melt is produced
from amphibolite, namely \(~ 1050^\circ\)C. However, the trend of the Rutter and Wyllie (1988) data for the partial melting of a tonalite indicates that the experimental liquidus would in fact be much higher. This discrepancy may be due to the loss of fluid during crystallization of the tonalite.

It is an interesting exercise to compare the percentage of H\(_2\)O in the Rutter and Wyllie (1988) tonalite sample, namely 0.8 wt\%, to that in melt produced by 40 vol\% melting of amphibolite. The difference, considered over the thickness of TTG produced during LAME, allows an estimate of the amount of water that might have become available (in a crustal column) during the crystallization (or metamorphism) of TTG during LAME. Such water might have played a significant role in generating further melt, or as an agent of metamorphism and mineralization.

Assuming that the amphibolite source contained 1 – 2 wt\% H\(_2\)O and that all H\(_2\)O entered a TTG melt resulting from 40 vol\% melting, retention of 0.8 wt\% H\(_2\)O in the crystallized TTG suggests that 2 – 4 wt\% H\(_2\)O would have escaped, which is significant!

6. Conclusion

Numerous experimental studies (Rushmer 1991; Winther and Newton 1991; Rapp et al. 1991; Rapp and Watson 1995; Sen and Dunn 1994; Wolf and Wyllie 1994) support the idea that TTG represents batch partial melt of metabasalt under “vapour absent” conditions in the garnet-amphibolite or eclogite stability field. Melt compositions that fall within the TTG field are achieved at 10 – 40 vol\% melting of metabasalt at temperatures ranging from 850 – 1050\(^\circ\)C.

The temperatures at which a given experimental melt fraction is achieved differ by up to 150\(^\circ\)C between studies (Fig. 6). This is likely due primarily to differences in the bulk compositions of the starting materials. As a result there is significant uncertainty associated with use of the melting curve for metabasalt that I have constructed, which is based on the experimental data, and gives volume % melt as a function of temperature. The curve will be used to determine the amount of TTG melt produced in my thermal models.

The high-K content of the late granites precludes their direct derivation from a basaltic source (Winther and Newton 1991). Rather they are thought to be derived from partial melting of TTG or felsic metasedimentary rocks. The melting curve for TTG that I have constructed, based on the experimental data of Rutter and Wyllie (1988), represents the higher end of the temperature range at which granitic melt is produced. Hence application of this curve to my
thermal models would yield the minimum amount of granite that can be expected under the conditions being modelled.
Chapter 4: Melting curves for the production of TTG and granite

References


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Table 1. Average quenched melt compositions from runs on AAT (Average Archean tholeiite) and comparison with natural tonalite and trondhjemite. Modified from Winther and Newton (1991).

<table>
<thead>
<tr>
<th></th>
<th>Tonalites</th>
<th>Trondhjemites</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>melt avg.(22 runs)</td>
<td>melt avg.(27 runs)</td>
</tr>
<tr>
<td>T (°C)</td>
<td>750 – 1100</td>
<td>800 – 1100</td>
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<tr>
<td>P (kb)</td>
<td>5 – 30</td>
<td>5 – 30</td>
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<tr>
<td>H$_2$O %</td>
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<td>0.8 – 5.1</td>
</tr>
<tr>
<td></td>
<td>wt%, wt%</td>
<td>wt%, wt%, wt%</td>
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<td>SiO$_2$</td>
<td>68.15, 3.20, 67.8</td>
<td>71.84, 1.61, 71.4</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>0.68, 0.37, 0.48</td>
<td>0.47, 0.24, 0.23</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>16.3, 1.12, 16.3</td>
<td>15.4, 0.75, 16.6</td>
</tr>
<tr>
<td>“FeO”</td>
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<td>2.17, 0.74, 1.41</td>
</tr>
<tr>
<td>MnO</td>
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<td>0.07, 0.05, 0.02</td>
</tr>
<tr>
<td>MgO</td>
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<td>0.52, 0.23, 0.52</td>
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<tr>
<td>CaO</td>
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<td>2.36, 0.62, 3.07</td>
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<tr>
<td>Na$_2$O</td>
<td>4.32, 1.25, 4.68</td>
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<tr>
<td>K$_2$O</td>
<td>0.74, 0.26, 1.53</td>
<td>1.22, 0.46, 1.31</td>
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<tr>
<td>Total</td>
<td>100.1, 100.0</td>
<td>100.0, 100.0</td>
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Fig. 2. Semi-schematic P-T stability diagram of AAT with ~ 2 wt% H₂O added (Winther and Newton 1991). Solidus is bold line. Curve A represents a geotherm that produces REE-fractionated magmas of trondhjemitic to tonalitic composition near the amphibole-out line and is considered feasible for the origin of many Archean gneiss compositions. am=amphibole, gt=garnet, pl=plagioclase, grano=granodiorite, trondh=trondhjemite, px=pyroxene. Modified from Winther and Newton (1991).
Fig. 3. CIPW-normative feldspar plot of all partial melts containing > 10% CIPW normative quartz in experimental runs of Rapp et al. (1991). Classification is after Barker (1979). Or=orthoclase, Ab=albite, An=anorthite. Filled symbols: partial melts of sample #1; open symbols: partial melts of sample #2; half-filled symbols: partial melts of sample #3; open symbols with dots in center: partial melts of sample #4. Circles with horizontal bars are partial melts at 0.8 GPa and 900°C in which 1 – 2 wt.% water was added. Modified from Rapp et al. (1991).
Fig. 4. P-T conditions suitable for the generation of trondhjemite and tonalite melts by vapour-absent partial melting of amphibolite/eclogite, as well as inferred vapour-absent solidus (Rapp et al. 1991). Shaded field represents generalized region over which amphibole and garnet stability fields overlap. Also shown are conditions for trondhjemite-tonalite melt produced by vapour-absent melting of amphibolite at 0.8 GPa, 900°C with 1 – 2 wt.% extraneous water (Rapp et al. 1991), and solidus for basalt with 5 wt.% water added (from Wyllie [1983]). Curve A represents a geotherm under which tonalite-trondhjemite melt would be produced. Modified from Rapp et al. (1991).
Fig. 6. Plot of volume % melt versus temperature from experimental studies of dehydration melting of amphibolites. Data from Rushmer (1991), Wolf and Wyllie (1994), Rapp et al. (1991) and Sen and Dunn (1994) at given pressures. Also shown are a melting curve from Petford and Gallagher (2001), best linear least-squares fit to the data (actual best fit line) and a visual fit (my “best fit” line). The latter will be used to determine the amount of TTG melt produced in my thermal models.
Fig. 7. Plot of temperature versus volume % melt from all experimental studies listed in Fig. 6. Also shown are a best linear least-squares fit to the data (actual best fit line) and my melting curve (my “best fit” line) plotted in Fig. 6.
Fig. 8. Plot of volume % melt versus temperature from an experimental study of dehydration melting of tonalite at 1.0 GPa (Rutter and Wyllie 1988). Also shown is the best linear least-squares fit to the data which is the melting curve I use to determine the amount of granitic melt produced in my thermal models.
Chapter 5: Effect of Upward Magma Transfer on Total Melt Produced during Crustal Melting

Chapter 5

Effect of Upward Magma Transfer on Total Melt Produced during Crustal Melting

Abstract

Under natural conditions of crustal melting melt segregates from its source rock and is transferred upwards to higher levels of the crust. As a volume of melt is extracted and transferred upward, the still fertile roof above the source region moves downward and can in turn undergo partial melting. I demonstrate the effect that the upward transfer of melt may have on the total amount of melt predicted by a thermal model. One-dimensional (1-D) models show that, under reasonable Neoarchean thermal conditions, more than 40% more melt can be generated than if such melt transfer did not occur. I model a 40-km-thick crust with a uniform distribution of heat producing elements that experiences increased mantle heat flow for 40 My, resulting in crustal anatexis. This is the approximate time span over which basaltic volcanism and TTG plutonism occurred in the Superior Province during the Late Archean Magmatic Event (LAME), ca. 2.7 Ga. More generally, such an increase in mantle heat flow may be a result of arc or plume activity or extension. I use a simplified set of criteria under which melt is transferred upwards to the mid-crust in my model. I compare the amount of melt produced in this model to one in which no such upward transfer of melt is implemented. Without upward transfer of melt 3.48 km of melt is produced. With upward transfer of melt 4.91 km of melt is produced, an increase of 41% which is significant. I show that the incorporation of upward melt transfer into a standard model of lower crustal melting may increase the amount of melt predicted by the model by \( \frac{1}{1-f} \) where \( f \) denotes the fraction of melt that is on average being extracted from the source rocks.

It is well known that melts concentrate heat producing elements (HPEs) such as U, Th and \(^{40}\)K and that the upward transfer of such melts results in the redistribution of HPEs in the crust. This results in crustal cooling and may be of primary importance in the long-term stabilization of continental crust. Assuming reasonable partitions of HPEs between the melt and the restite, I show that their upward transfer leads to more rapid cooling of the crust, and therefore reduces
somewhat the ultimate amount of melt produced (4.54 km). However, this is still 30% more melting than if no upward migration occurred.

1. Introduction

While the details of the crustal melting process are enormously complex the basic sequence of steps involved is not, namely melt generation, segregation, ascent and emplacement (Brown 1994). The thermal effect of magma emplacement at various levels of the crust (generally in the form of sills) is a classic thermal modelling exercise (Wells 1980).

However, to my knowledge, little modelling has been done to determine the effect that the transfer of such melt out of a crustal source region might have on the total amount of melt produced. Upward transfer of melt is rarely implemented in thermal models of crust undergoing melting. In such models it is either demonstrated that crustal temperatures have exceeded the solidus of the assumed source rock or, where the actual volume of melt produced is important it is calculated by applying a melting curve that gives the fraction of melt expected at a given temperature (and possibly also pressure) to the final geotherm.

That there should be some effect is clear. During crustal melting a source region may yield up to 50 vol% melt. As such melt is expelled from the source region, overlying crust will move downwards, into hotter regions of the crust. If the downward moving crust is fertile, then melting of this (new) source is expected, as long as heating of the crust is sustained. I conduct a quantitative investigation of the resulting increase in the amount of melt produced as a result of these transfers.

A secondary effect of the generation and upward transfer of melt is the concentration of HPEs in melt, their depletion in restite and their subsequent transfer in melt to higher crustal levels. Such a redistribution of HPEs results in a cooler crust and should decrease the amount of melt produced. Indeed, this ‘cooling mechanism’ is considered of primary importance in the differentiation and stabilization of continental crust (Morgan 1985). My models show that this partitioning also has an impact on the amount of melt produced.

2. Modelling parameters and initial conditions

While my aim is to present a result that I believe has applicability beyond LAME my models broadly reflect conditions existing in the Superior crust during LAME. I am concerned with melting and heat transfer in the crust (the presumed source of TTG during LAME; see Chapter 3) and not in the mantle. I assume a crustal thickness of 40 km and uniform heat production of
I construct 1-D transient heat conduction models. I assume an ambient mantle heat flow of 13 mWm\(^{-2}\). My choice of thermal parameters results in a Moho temperature of 795°C for the initial crust. I increase the mantle heat flow to 34 mWm\(^{-2}\) ‘instantaneously’ at the start time of the models (0 My) and keep it at this increased level for 40 My. This represents a period of heating from a mantle heat source, possibly as a result of arc or plume activity or extension. This time period is also coincident with the ~ 40 My of basaltic volcanism and TTG magmatism that occurred during LAME. I run the models for a total time span of 100 My.

I simplify the models by imposing a somewhat idealized (artificial) set of conditions under which melt is transferred (upwards) out of the source region. In order for melt to leave the source region I require that \(\frac{1}{3}\) of the source (by volume) must comprise melt (Wickham 1987). Dehydration melting experiments at 1 GPa by Wolf and Wyllie (1994) on an amphibolite sample from the western Foothills Metamorphic Belt of the Sierra Nevada yielded 30 vol% tonalitic melt at 925°C. I assume that a melt fraction of \(\frac{1}{3}\) is reached at 925°C. Ideally a transfer of melt upwards should take place each time some minimum thickness of melt is produced under such conditions. However, this would result in a large number of transfers making the models somewhat cumbersome. Therefore to keep the number of transfers reasonable I have elected to implement such transfers at 2 My intervals. This results in models containing a sequence of 20 submodels or ‘geometries’ (Chapter 2). Because crustal heating is rapid in my models, a significant portion of the source region reaches temperatures higher than 925°C during the somewhat long interval (2 My) between melt transfers. However, I assume that the melt fraction remains unchanged, at \(\frac{1}{3}\), just as it would if I had incorporated more frequent melt transfers.

Melt is transferred from the lower crust to the mid-crust (17 km depth). Crust between this depth and the top of the melt source region is moved downwards to accommodate the newly emplaced melt. The choice of 17 km results in a final crustal configuration (following all upward transfer of melt) in which the base of the transferred melt region is approximately at the mid-crustal level (20 km).

Finally, I assume that once the melt fraction of \(\frac{1}{3}\) has been transferred out of a source region the remaining \(\frac{2}{3}\) of the source is completely infertile, incapable of producing any further melt regardless of the temperature it may reach.
Chapter 5: Effect of Upward Magma Transfer on Total Melt Produced during Crustal Melting

3. Models

3.1. Model 1: Upward Transfer of Melt

I begin by modelling the upward transfer of melt at 2 My intervals with no associated change in the (uniform) distribution of HPEs in the crust (Model 1). Every 2 My I determine the depth to the 925°C isotherm. Crust between this depth and the top of the restitic layer from previous melt extractions is regarded as crust that contains a melt fraction of $\frac{1}{3}$ which is ready to be expelled. The manner in which such melt is transferred upwards at 2 My intervals is shown in Fig. 1. The period of enhanced mantle heat flow is between 0 and 40 My. The first melt is ready to be expelled at 6 My and the last at 40 My. (A return to ambient mantle heat flow conditions at 40 My results in a rapid end to further melt generation).

The cumulative melt transferred over the course of the model run is shown in Fig. 2. The final transfer of melt upwards at 40 My is shown in Fig. 3a. A total of $\sim 4.91$ km of melt is transferred upwards with $\sim 9.83$ km of restite remaining in the lower crust, while the final distribution of HPEs remains uniform (Fig. 4a).

3.2. Model 2: No Upward Transfer of Melt

For comparison purposes I run a model in which no upward transfer of melt is implemented. Every 2 My I determine the depth to the 925°C isotherm. I regard crust below this depth as containing a melt fraction of $\frac{1}{3}$. The idea behind this approach is that under natural conditions melt would have been transferred out of this source region once a critical melt fraction (assumed to be $\frac{1}{3}$ in my models) was achieved, leaving behind an anhydrous restitic layer that would have resisted further melting. The first melt is generated at 6 My and the last at 40 My.

The cumulative melt produced over the course of the model run is shown in Fig. 2. The total thickness of melt produced is $\sim 3.48$ km with an associated restitic layer of thickness $\sim 6.96$ km; the final distribution of HPEs remains uniform (Fig. 4b). The geotherm at 40 My and the final melt and restite layer are shown in Fig. 3b.

3.3. Model 3: Upward Transfer of Melt with Redistribution of HPEs

This model is similar to Model 1 in which melt is transferred upwards at 2 My intervals. However, in addition, I implement a redistribution of HPEs with concentration of HPEs in the melt and a concomitant depletion of HPEs in the restite in a manner that leaves the total heat production of the model crust unchanged. I assign a heat production of 2.6 $\mu$Wm$^{-3}$ to the melt and 0.8 $\mu$Wm$^{-3}$ to the restite. This is similar to the estimated heat production of the Western
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Abitibi upper and lower crust, respectively, extrapolated to 2.7 Ga. Current heat production in the Western Abitibi upper and lower crust is estimated at 1.2 and 0.4 µWm⁻³ respectively (Mareschal and Jaupart 2006). At 2.7 Ga heat production would have been approximately double, hence 2.4 and 0.8 µWm⁻³. As in Model 1 the first melt is ready to be expelled at 6 My and the last at 40 My. The manner in which melt is transferred upwards and the associated redistribution of HPEs is shown in Fig. 5.

The cumulative melt transferred over the course of the model run is shown in Fig. 2. The final transfer of melt upwards at 40 My is shown in Fig. 3c. A total of ~ 4.54 km of melt is transferred upwards with ~ 9.08 km of restite remaining in the lower crust. The final distribution of HPEs is shown in Fig. 4c.

3.4. Results

My modelling shows that under the particular thermal conditions assumed in my models significantly greater thicknesses of melt are predicted by models in which upward transfer of melt is implemented (4.91 and 4.54 km in Models 1 and 3, respectively) than in a model in which no upward transfer of melt takes place (3.48 km in Model 2). In the case where the distribution of HPEs remains unchanged (Model 1) there is a 41% increase in the amount of melt generated. In the case where the HPEs are redistributed during the generation and upward transfer of melt (Model 3) there is a 30% increase in melt.

While the redistribution of HPEs (Model 3) does result in a decrease in the amount of melt generated its effect is minor compared to the increase resulting from the mechanism of upward transfer of melt. The difference between the amount of melt generated in the two upward transfer models is 375 m. The difference between the amount of melt generated in the two upward transfer models compared to that in the model with no upward transfer (Model 2) is 1435 m (Model 1) and 1060 m (Model 3).

However, the redistribution of HPEs in Model 3 does have a definite cooling effect on the crust which can be seen if we examine the Moho temperatures of the three models (Fig. 6a). The difference in Moho temperature between Model 3 and Model 2 in which there is no upward transfer of melt is established at ~ 50°C by 40 My, the end of the period of increased mantle heat flow (Fig. 6b). The difference in Moho temperature between Model 3 and Model 1 in which there is upward transfer of melt but no redistribution of HPEs is > 30°C by 40 My and approaches 50°C by 100 My (Fig. 6b).
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4. Discussion

4.1. Generalization to different thermal histories

While the general result that upward transfer of melt during lower crustal melting is likely to lead to increased melt production appears indisputable the manner in which such an increase might be quantified remains unclear. I investigate this problem using my model results and derive a formula for such an increase that may have wider applicability.

4.1.1. Using Model 2-type geotherms to estimate Model 1-type geotherms

The vast majority of published thermal models of lower crustal melting resemble Model 2 in that they do not incorporate the upward transfer of melt. The difference in Moho temperature between Model 2 and Model 1 in which there is upward transfer of melt but no redistribution of HPEs is a maximum of 19°C at 21 My (Fig. 6b) suggesting that the advection of heat via the upward transfer of ~ 5 km of melt in Model 1 has only a minor effect on lower crustal temperatures. Further evidence of the similarity of lower crustal temperatures in Models 1 and 2 is that the top of the last melt layer (at 40 My), which coincides with the 925°C isotherm, occurs at a depth of ~ 30.03 km in Model 1 and at ~ 29.57 km in Model 2, a difference in the level of the 925°C isotherm of 464 m. This suggests that lower crustal temperatures of relatively simple models such as Model 2 can be used to approximate lower crustal temperatures of more sophisticated models such as Model 1.

I caution that the precise set of conditions under which this can be done cannot be ascertained from a single set of models such as mine. Rather, a wide range of thermal conditions must be modelled (a possible future project). One requirement would certainly be that the upward transfer of melt should not have a significant cooling effect on the lower crust. One can imagine this to be the case when the heat loss in the lower crust through upward melt transfer is adequately compensated by heat flow from the mantle. I would expect this to be the case when heat flow from the mantle was significant, hence during a significant crustal heating event, such as LAME.

Furthermore, there is a requirement that is implicit in all my models, and that is that the crust that moves downward into the zone of partial melting is fertile. One can envision such fertile crust as being more readily available in the initial stages of formation of continental crust rather than in mature crust.

The assumption that the lower crustal temperatures of relatively simple models such as Model 2 can be used to approximate lower crustal temperatures of more sophisticated models
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such as Model 1 provides us with a means of estimating how much more melt would be generated in a model such as Model 2 if upward transfer of melt had been implemented.

4.1.2. Estimate of increased melt production resulting from upward melt transfer

Suppose that in a thermal model such as Model 2 the eventual thickness of the lower crustal partially melted layer is \( d \). The top of this layer would correspond to an isotherm for a characteristic temperature \( T \) such as 925°C in my models. Suppose that the melt fraction being extracted from the partial melt region is \( f \), corresponding to the fraction of \( \frac{1}{3} \) employed in my models. With no upward transfer of melt the total thickness of melt generated would be \( d f \).

If upward transfer of melt is implemented then a temperature of \( T \) would be reached at approximately the same depth, as shown by the similarity of lower crustal temperatures in Models 1 and 2, again giving a lower crustal partial melt layer of thickness \( d \). Assuming that all crust that entered this region was fertile, melt of cumulative thickness \( f \times d \) would certainly be extracted (though not necessarily as a single partial melt layer). This would allow fertile crust of thickness \( f \times d \) to move downward (into the region vacated by the melt) from which melt of thickness \( f \times (f \times d) \) would be extracted. This would in turn allow fertile crust of thickness \( f \times (f \times d) \) to move downward (again into the region vacated by the melt) from which melt of thickness \( f \times (f \times (f \times d)) \) would be extracted. Hence melt of total thickness \( f \times d + f \times (f \times d) + f \times (f \times (f \times d)) + \cdots \) would ultimately be extracted. The sum of this geometric series is given by

\[
d(f + f^2 + f^3 + \cdots) = d \frac{f}{1-f}
\]

with the term on the right hand side giving the total thickness of melt that would ideally be transferred upward.

4.1.3. Application of Equation (1) to Model 2

We can apply equation (1) to Model 2 in which there is no upward transfer of melt. The highest crustal level at which the 925°C isotherm occurs is 29,576 m (at 40 My) giving a partially melted layer of thickness \( d \) equalling 10,434 m. Assuming \( f \) equals \( \frac{1}{3} \) equation (1) implies that the amount of melt that would be produced in a model similar to Model 2 but in which upward transfer of melt is implemented is 5,217 m. In Model 1 which is similar to Model
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2 but which incorporates upward transfer of melt 4,913 m of melt is produced, a minor difference of 304 m or 6% from the amount predicted by equation (1).

The difference can be attributed to the fact that my analysis is based on the assumption that the 925°C isotherm at 40 My occurs at the same depth in Model 2 as in its counterpart model in which upward transfer of melt is implemented (Model 1). This is not the case: in Model 1 the 925°C isotherm occurs at a depth lower by 464 m than the level at which this isotherm occurs in Model 2. Hence melt in Model 1 is being extracted from a thinner partial melt region than that assumed in the derivation of equation (1). Another source of discrepancy may be that my analysis is based on the assumption that melt is fully extracted from the region below the 925°C isotherm at 40 My. Such is not the case in Model 1. At 40 My the 925°C isotherm occurs at a depth of 30,030 m while the top of the restite layer occurs at a depth of 30,174 m, leaving a layer of thickness 144 m from which melt is not extracted. By 42 My (at which time further melt would be extracted if necessary) the crust has cooled (Fig. 6a) and the 925°C isotherm has moved downward, below this depth.

4.1.4. Alternate derivation of Equation (1)

We can in fact arrive at the estimate appearing in equation (1) of the amount of melt that would be produced if upward transfer of melt is incorporated into a standard model of crustal melting in a much simpler manner than that outlined above. What we are in fact doing in the analysis that leads to equation (1) is assuming that no further melt remains in the region of thickness \(d\) from which melt is extracted. This implies that the entire region must consist of restite. For a melt fraction \(f\), the ratio of melt to restite is \(\frac{f}{1-f}\) implying a total melt thickness of \(d \frac{f}{1-f}\) for the case of restite of thickness \(d\). This is the same result as appears in equation (1).

4.1.5. General application of Equation (1)

The percentage increase in the amount of melt produced that would result from the incorporation of upward melt transfer into a standard thermal model of crustal melting is shown in Fig. 7 for varying melt fractions, \(f\). For higher melt fractions, ranging from \(\frac{1}{3}\) to \(\frac{1}{2}\), the increase is significant, from 50 – 100%.

If one is modelling a melting event in which melt on the order of 1 or even 10 km was produced, such as the major TTG event that occurred in the Pilbara Craton between 3.48 – 3.42 Ga (Zegers and van Keken 2001) then my result may be useful. In a model in which ~ 5 km of
melt is produced, given a melt fraction of $\frac{1}{3}$ application of my result would increase this amount to 7.5 km.

I apply equation (1) to a thermal model of LAME in Chapter 6.

4.1.5.1. Rey et al. (2003) model

A final point I wish to make is that where one is attempting to demonstrate that a particular model of lower crustal melting can account for a large-scale melting event such as LAME (Percival et al. 2006) it is necessary to provide an estimate of the amount of melt predicted by the proposed thermal model. It is not sufficient to simply show that the model Moho temperature exceeds the solidus or even the liquidus of the assumed source rock.

An example of this is the model of Rey et al. (2003). Rey et al. (2003) propose that crustal thickening coupled with a mantle plume may have been responsible for LAME. In their model, in which a 10-km-thick greenstone cover results from plume activity (Chapter 6) initial heat flow into the base of the crust is 25 mWm$^{-2}$ and reaches a maximum 44 mWm$^{-2}$ after plume emplacement. The increased mantle heat flow in my models of 34 mWm$^{-2}$ is intermediate to that in their models.

Their emphasis is on the production of granitoids (TTG and granite). They assume that at 900°C a melt fraction of 0.2 – 0.4 of TTG composition is produced from melting of metabasalt. This is similar to my assumption that a melt fraction of $\frac{1}{3}$ (~ 0.33) is produced at 925°C in my models.

However, the Moho temperature in their 1-D thermal model never exceeds 950°C. In Model 2 in which no upward transfer of melt is implemented melt of thickness 3.48 km is produced but with the Moho temperature reaching 1168°C which is more than 200°C hotter than the temperature reached in the Rey et al. model (2003). In Model 1, in which upward transfer of melt is implemented, 4.91 km of melt is produced with the Moho temperature reaching 1154°C.

This highlights the necessity of quantifying the amount of melt produced in some thermal models of lower crustal melting and suggests an area in which equation (1) may be useful.

In addition, note that while the thickness of melt produced in my models may be comparable to the thickness of TTG produced in some areas of the Superior Province during LAME it still falls short of the (at least) 10 – 15 km vertical thickness of juvenile TTG comprising the mid- to lower Abitibi-Wawa crust inferred from exposures along the Kapuskasing uplift (Percival and Card 1983; Grandjean et al. 1995). A melt thickness of 10 km could not be produced, even in my models incorporating upward melt transfer, without temperatures in the lowermost 20 km of
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crust exceeding 925°C. Such high temperatures are not recorded at such shallow depths in the Superior Province supporting the idea that some form of restite recycling must have been involved in the generation of such melt (Beakhouse 2007).

5. Conclusion

Using 1-D crustal thermal models that broadly reflect thermal conditions during LAME in the Superior Province I have shown that during periods of lower crustal melting the upward transfer of melt out of the source region and the associated downward movement of fertile source material can have a significant effect on the total thickness of melt produced.

I have derived a formula that allows a quantitative determination of the increase in melt production.

I have also shown that the partitioning and upward transfer of HPEs in the melt exerts a cooling effect on the crust but that melt production is still significantly enhanced over conditions in which no upward transfer of melt occurs.
References


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(a) 

(b)
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Fig. 1. In Model 1, crustal temperatures first exceed 925°C between 4 and 6 My. Crust below the 925°C isotherm is assumed to contain a melt fraction of $\frac{1}{3}$ which is ready to be expelled. (a) Geotherm at 6 My, just prior to the first upward transfer of melt; $\frac{2}{3}$ of the partial melt region remains in place as restite. (b) Geotherm at 6 My, immediately following melt transfer. The melt is emplaced in the mid-crust, with the top of the melt layer at 17 km depth. (c) Geotherm at 8 My, just prior to the 2nd upward transfer of melt. The crust has continued to heat up with an additional thickness of crust now below the 925°C isotherm. (d) Geotherm at 8 My, immediately following melt transfer.
Fig. 2. Cumulative thickness of melt generated in the various models over time. In Model 1, in which melt is transferred upwards but the HPE distribution remains unchanged, a total of 4.91 km of melt is generated. In Model 2, in which the melt remains in place, a total of 3.48 km of melt is generated (assuming that \( \frac{1}{3} \) of the partial melt region below the 925°C isotherm consists of melt). In Model 3, in which melt is transferred upwards and HPEs are concentrated in the melt, a total of 4.54 km of melt is generated.
Fig. 3. Final location of all generated melt and restite at 40 My, immediately following the last upward transfer of melt in Models 1 and 3. (a) Model 1, in which upward transfer of melt occurs but there is no associated redistribution of HPEs. (b) Model 2, in which no upward transfer of melt occurs, and (c) Model 3, in which upward transfer of melt occurs and the HPEs are redistributed (concentrated in the melt and depleted in the restite).
Fig. 4. Final distribution of HPEs in (a) Model 1, (b) Model 2 and (c) Model 3. Final location of all generated melt and restite is also shown.
Fig. 5. In Model 3, crustal temperatures first exceed 925°C between 4 and 6 My. Crust below the 925°C isotherm is assumed to contain a melt fraction of $\frac{1}{3}$ which is ready to be expelled. (a) Distribution of HPEs at 6 My, just prior to the first upward transfer of melt. (b) Geotherm at 6 My, just prior to the first upward transfer of melt; $\frac{2}{3}$ of the partial melt region remains in place as restite. (c) Geotherm at 6 My, immediately following melt transfer. The melt is emplaced in the mid-crust with the top of the melt layer at 17 km depth. (d) Distribution of HPEs at 6 My, immediately following melt transfer.
Fig. 6. (a) Moho temperatures of Models 1, 2 and 3 over time. The rapid decline in Moho temperature in all the models after 40 My coincides with the end of the period of increased mantle heat flow. (b) Difference in Moho temperatures between the models.
Fig. 7. The percentage increase in the thickness of melt produced in a thermal model of crustal melting in which upward transfer of melt is incorporated compared to that produced in a standard model in which there is no such melt transfer.
Chapter 6: The Role of Mantle Plumes in the Production of TTG during LAME

The Role of Mantle Plumes in the Production of TTG during LAME

Abstract

Catastrophic slab avalanche events resulting in the generation of a superplume or multiple major plumes have been proposed to account for periods of high crustal growth such as that during the Late Archean Magmatic Event (LAME) ca. 2.7 Ga. However, convincing evidence for mantle plumes in the Superior Province during LAME is limited to the Abitibi subprovince and to a lesser extent the Wawa subprovince. Rey et al. (2003) have suggested that the thermal pulse from a mantle plume coupled with the thermal blanket effect of plume-generated greenstones was responsible for the generation of granitoids (primarily TTG and to a lesser extent granite) during LAME. Application of an appropriate melting curve for a metabasaltic source for TTG to the “Rey model” results in thicknesses of granitoid melts that are well below the regional thicknesses of ~ 5 – 8 km of TTG and ~ 1 km of granite estimated to have been emplaced in the Superior Province during LAME. Modifications of the Rey model through use of a more appropriate value for crustal heat generation in the Superior Province at ca. 2.7 Ga and of a temperature-dependent thermal conductivity result in a significantly greater thickness of TTG melt (~ 7 – 9 km). However, this still falls short of the estimated 16 km of TTG melt emplaced in the Abitibi subprovince during LAME. I show that comparable thicknesses of melt can be produced in models that incorporate several plumes (or plume pulses) (the timing of which is based on plume activity in the Abitibi subprovince) along with repeated delamination of lower crustal restite. However, such models still fail to explain the manner in which hydration of lower crustal TTG source rocks (a requirement based on experimental evidence) occurs. Furthermore the timing of TTG generation in the models is not fully consistent with that observed in the Abitibi.

1. Introduction

The aim of this investigation is to determine in a quantitative fashion, using 1-D thermal models coupled with the application of melting curves that give the fraction of melt expected at a given temperature, the viability of mantle plumes as a mechanism for the generation of the
voluminous quantities of TTG melt emplaced in the Superior Province during LAME (Chapter 3). A focus of my investigation will be the Abitibi subprovince where the most convincing evidence for the involvement of mantle plumes during LAME exists.

1.1. The 2.7 Ga LAME

The Late Archean Magmatic Event (LAME) that occurred in the Superior Province ca. 2.7 Ga appears to have been a “worldwide” event (Chapter 1). Though the “global” nature of LAME is well recognized it has not been accounted for to any great degree, perhaps because the challenges of accounting for LAME even locally have proved formidable. Several authors have, however, suggested that mantle plumes may have played a major role during LAME. Their reasoning and the evidence for mantle plume activity during LAME in the Superior Province is discussed below.

1.2. Mantle plumes and LAME

1.2.1. Slab avalanches and superplumes

Numerical models of mantle convection suggest that the effect of the endothermic spinel to perovskite and magnesiowüstite phase transition at ~ 660 km depth may result in temporary inhibition and eventual avalanching of downgoing slabs today and that such an effect may have been much greater in the past due to higher mantle temperatures (Machetel and Weber 1991; Schubert and Tackley 1995; Tackley 2002). In particular, in the Archean, slabs may not have penetrated the 660-km discontinuity as easily as they do today.

Stein and Hofmann (1994) have suggested that periods of high crustal formation rates such as LAME may have been the result of the intermittent breakdown of predominantly two-layer mantle convection into single-layer convection. During such periods cold subducted slabs that had accumulated at the 660-km discontinuity may have sunk into the deep mantle (“slab avalanche”) leading to the rise of multiple major plumes from the core-mantle boundary, a mantle overturn event. In the MOMO (mantle overturn and major orogenies) model of Stein and Hofman (1994), such plume activity would have led to high rates of crust formation through (1) the generation of oceanic plateaus that may eventually have accreted to continental crust, (2) continental flood basalts and (3) magmatic underplating of existing continental crust. During intervening periods of two-layer mantle convection, lower rates of crust formation resulting from arc processes would have prevailed. Such a model can explain several features of the isotope and trace-element geochemistry of major orogenic mafic and granitoid rocks that suggest derivation from a source that is intermediate between depleted and primitive mantle
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(Stein and Hofmann 1994). A mixing of primitive and depleted mantle can be achieved by the emplacement of plume heads that have entrained large amounts of lower mantle material, into the depleted upper mantle.

Condie (1998, 2000, 2004, 2007) has suggested that a mantle “superplume event” may have been responsible for the peak in juvenile crust production during LAME. A “superplume” is defined by Condie (2001, 2004) as a large single plume that originates primarily from the D” layer with the lateral extent of the plume head reaching 1500 to 3000 km in diameter after spreading at the base of the lithosphere. A “superplume event” is a short-lived (~ 100 My) mantle event during which superplumes along with smaller plumes reach the base of the lithosphere. During a superplume event, plume activity may be focused in one or more mantle upwellings (Condie 2004). Evidence cited for a superplume event at 2.7 Ga is a high production rate of juvenile crust (based on the distribution of U/Pb zircon ages coupled with Nd isotopic data), a peak in BIF (banded iron formation) deposition and increased CO2 (indicated by a minor peak in the deposition of black shales) (Condie 2004).

As in the MOMO model (Stein and Hofmann 1994), such plumes may have been triggered by the catastrophic sinking of slabs that had piled up at the 660-km seismic discontinuity at different locations into the lower mantle. When the slabs reached the D” layer at the base of the lower mantle, plumes would have been produced. Such plumes would not only have produced juvenile crust through decompression melting but they may also have heated the upper mantle resulting in enhanced magmatic activity at arcs and ocean ridges (Condie 1998). Arc activity may have been coupled to such avalanches as subduction may have been initiated above such regions of downwelling (Peltier et al. 1997).

1.2.2. Criteria for identification of ancient mantle plumes

While a mantle overturn event cannot be ruled out as a cause of LAME the evidence cited by Stein and Hofmann (1994), Condie (1998, 2000, 2004, 2007) and others (Breuer and Spohn 1995) for the involvement of mantle plumes in an event such as LAME is fairly indirect. Here I review more direct evidence for mantle plumes. Campbell (2001) suggests the following criteria be used to identify ancient mantle plumes, namely (1) uplift prior to volcanism, (2) a radial or modified radial pattern of dikes that feed the volcanism, (3) massive flows that can be correlated over large distances, (4) chains of plume tail-related volcanoes that show a systematic age progression due to the movement of the overlying plate relative to the plume, and (5) high temperature magmas such as komatiites and picrites or high-pressure magmas such as alkali
basalts, nephelinites, and type-1 (high-Mg, low-K) kimberlites. In contrast to (3) the high volatile content of island arc basalts generally results in pyroclastic deposits that cannot be correlated over large distances due to the high viscosity of the melt (Campbell 2001).

1.2.2.1 Komatiites: anhydrous (high-T) or hydrous (moderate-T) melts?

Amongst the criteria outlined above perhaps the clearest evidence for an ancient mantle plume is the presence of komatiites. Komatiites are defined as ultramafic volcanic rocks that contain at least 18 wt% MgO (Arndt and Nisbet 1982) while many contain as much as 33 wt% MgO. They occur with tholeiitic basalts and generally constitute no more than 10% of the volcanic rocks (Philpotts 1990).

The liquidus of a komatiite increases with MgO content. All komatiites have suffered geochemical alteration to varying degrees but nearly fresh komatiites have been found in the 2.7 Ga Belingwe greenstone belt in Zimbabwe. These contain original olivine crystals and melt in the form of glassy inclusions (Nisbet et al. 1987; McDonough and Ireland 1993). Based on an inferred MgO content of 25.6 wt% for the parental melt of a Belingwe komatiite sample Nisbet et al. (1993) calculate an eruptive temperature of 1520°C. For a relatively fresh 2.7 Ga komatiite sample from Alexo Township, Ontario, a melt MgO content of 28 wt% implies an eruptive temperature of 1560°C (Nisbet et al. 1993). Such high eruptive temperatures require mantle potential temperatures of 1800 – 1900°C (Nisbet et al. 1993). Mantle potential temperature is the temperature the mantle would have if it rose adiabatically to the surface without melting. Herzberg et al. (2007) estimate present-day ambient mantle potential temperature in the range 1280°C – 1400°C. The estimate of Lee et al. (2006) is somewhat higher at 1370 ± 50°C. Ambient mantle potential temperature at 2.7 Ga is estimated to have been only 100 – 200°C higher than today (Abbott et al. 1994), implying a mantle plume source for komatiites (Campbell et al. 1989; Bickle 1993; McDonough and Ireland 1993).

However, the analysis by Nisbet et al. (1993) of the mantle potential temperature of a komatiitic source is based on the assumption that Archean komatiitic melts were anhydrous. It has been suggested that komatiitic melts may in fact have been hydrous melts generated in a subduction zone setting or in a wetter (undegassed) part of the Archean mantle (Grove et al. 1997; Parman et al. 1997; Wilson et al. 2003; Parman and Grove 2005; Barr et al. 2006; Grove and Parman 2006). Komatiites from the Barberton Mountain Land, South Africa, of age 3.49 Ga, contain augite as almost the only pyroxene, with high CaO contents (0.4 – 0.43 mole fraction of wollastonite [Wo] component in the augite cores) (Parman and Grove 2005). Melting
experiments by Parman et al. (1997) indicate that an anhydrous magma at 1 bar would first crystallize pigeonite, then augite, and the Wo components of the augites (0.3 – 0.35) would be lower than those of the Barberton komatiites. Alternatively, at a minimum pressure of 0.19 GPa (~ 6 km depth) a relatively low temperature (1370 – 1400°C), hydrous magma (4 – 6 wt% H₂O), would tend to crystallize augite, with higher Wo components (0.38 – 0.42) (Parman et al. 1997). However, Bouquain et al. (2006) claim that high Wo components (0.26 – 0.46) have been found in augites in demonstrably anhydrous, extrusive komatiites in Alexa and Belingwe. This debate has yet to be resolved and indeed it may be found that komatiites may be produced under both hydrous and anhydrous conditions. For the purposes of this investigation I will assume the more conventional origin of komatiites, namely in a very hot mantle plume.

1.2.3. LAME-age mantle plumes in the Superior Province

Ernst and Buchan (2003) outline a similar set of criteria to those of Campbell (2001) for identifying ancient mantle plumes, in particular the presence of a large igneous province (LIP) in the form of an oceanic plateau, oceanic flood basalts or continental flood basalts. LIPs are interpreted to result from decompression melting of a mantle plume head. They suggest that in the Archean rock record a plume is most likely indicated by greenstone belts that contain komatiites. Ernst and Buchan (2004) have compiled a database of LIPs in Canada and adjacent regions dating from 3 Ga to the present. Eighty potential LIPs and LIP remnants have been identified with ages ranging from 3100 to 17 Ma. LAME-aged plume activity is recognized in four regions in Canada, within the Rae, Slave and Superior Provinces. In the Superior Province plume-generated components have been identified in the Wawa and Abitibi subprovinces. They range in age from 2.75 – 2.70 Ga in the Abitibi belt and 2.75 – 2.74 Ga in the Wawa belt (Ernst and Buchan 2004).

A compilation of Archean greenstone belts that are most likely to have been derived from mantle plume sources appears in Tomlinson and Condie (2001). Komatiites are generally found in the two main and one minor Archean greenstone associations that may represent plume-generated sequences. These associations are (i) the mafic plain association which is interpreted as accreted oceanic plateaus, (ii) the platform association which is interpreted as plume-generated magmas erupted through rifted continental crust, and (iii) the rifted arc association which is interpreted as arcs that rifted in response to mantle plumes (Thurston and Chivers 1990; Condie 2001; Tomlinson and Condie 2001; Ernst and Buchan 2003). In the Superior
Province LAME-aged plume-generated sequences have only been identified in the Abitibi and Wawa greenstone belts (Tomlinson and Condie 2001).

The Abitibi and Wawa belts comprise only a fraction of the Superior Province (Fig. 1) in which LAME was widespread (Skulski and Villeneuve 1999; Percival et al. 2006). The findings of Ernst and Buchan (2004) and Tomlinson and Condie (2001) call into question the superplume model that has been proposed to account for periods of high crustal formation rates such as LAME (Stein and Hoffman 1994; Condie 1998, 2000, 2004, 2007). It is possible that in the future more subtle indicators of plume involvement during LAME may be discovered than are recognized at this time. At present, however, it appears that while mantle plumes may have been a contributor to LAME they are unlikely to have been the primary cause of LAME.

1.2.4. Plumes and the Abitibi subprovince

The geochemistry of volcanic rocks of the Abitibi subprovince (Fig. 1) suggests that it was constructed through a combination of plume and arc activity (Ayer et al. 2002; Scott et al. 2002; Sproule et al. 2002; Wyman et al. 2002). Condie (2000, pg. 157) and others have pointed out that plume-derived lavas contaminated by continental crust may acquire a pseudo-subduction zone geochemical signature (Ta-Nb depletion and Th enrichment) and that some Archean greenstone belts that have been described as arc-type based on geochemistry may represent crustally contaminated plume-type magmas. However, this should not be applicable to the Abitibi crust as it is predominantly juvenile (Carignan et al. 1993; Vervoort et al. 1994; Ludden and Hynes 2000; Ayer et al. 2002). The manner in which a plume (or plumes) and a subducting slab may have interacted is the subject of several (qualitative) models in which the Abitibi belt is constructed in an autochthonous fashion (Scott et al. 2002; Wyman et al. 2002; Daigneault et al. 2004; Benn and Moyen 2008).

Despite the evidence for subduction in the tectonomagmatic evolution of the Abitibi subprovince I believe that the construction of a plume-only model based on geological data from the Abitibi belt is still a valuable exercise. Comparison of model predictions with geological data can offer insight into the role that mantle plumes may (or alternatively may not) have played in the generation of TTG in the Abitibi. After all, it is unlikely to be a coincidence that the greatest thickness of TTG melt seen during LAME in the Superior Province is in a region where there is the strongest evidence for the operation of a mantle plume. To that end we begin with a brief review of salient aspects of the geology of the Abitibi subprovince.
1.2.4.1 Abitibi greenstone belt

The Abitibi greenstone belt is located in the southeastern Superior Province (Fig. 1). It straddles the provinces of Ontario and Quebec (Fig. 2) and is one of the largest known contiguous greenstone belts, 700 km long by 300 km wide (Card 1990; van Breemen et al. 2006). It is dominated by primarily metavolcanic supracrustal rocks with granitoid plutons and batholiths occupying approximately one-third of the surface area of the belt (Fig. 2; Chown et al. 1992). The regional metamorphic grade is primarily greenschist to subgreenschist with amphibolite grade rocks occurring adjacent to large magmatic intrusions (Easton 2000).

The debate over the allochthonous vs. autochthonous origin of greenstone belts (de Wit 1998; Hamilton 1998; Bleeker 2002; Thurston 2002) is one that has been conducted for the Abitibi. Traditionally an allochthonous origin has been ascribed to the Abitibi: it has been interpreted as the product of accretionary tectonics, involving the amalgamation and thrust stacking of magmatic arcs, backarcs and oceanic plateau fragments (Jackson and Fyon 1991; Desrochers et al. 1993; Kimura et al. 1993; Jackson et al. 1994; Jackson and Cruden 1995; Polat et al. 1998; Polat and Kerrich 2001). However, based on detailed geochronological and field data a radically different picture of the Abitibi has gradually emerged, namely that the belt is of primarily autochthonous or paraautochthonous origin (Heather and van Breemen 1994; Heather et al. 1995; Ayer et al. 2002; Bleeker 2002; Ayer et al. 2005; Peschler et al. 2006; Benn and Moyen 2008; Thurston et al. 2008).

1.2.4.2 Abitibi supracrustal assemblages

The metavolcanic rocks of the Abitibi subprovince represent nearly 50 My of semi-continuous volcanic activity starting at ca. 2750 Ma (Ayer et al. 2002). Volcanic rocks include komatiites, komatiitic basalts, abundant tholeiitic basalts and calc-alkaline mafic to felsic lavas, with intermingling of rocks of different compositions. Volcanic activity was followed by ~ 20 My of siliciclastic sedimentation. Based on 450 precise U-Pb ages, stratigraphy and geochemistry the supracrustal rocks of the Abitibi have been subdivided into nine lithostratigraphic assemblages, 7 volcanic assemblages formed between > 2750 and 2696 Ma and two sediment dominated assemblages formed between 2690 and 2670 Ma (Fig. 3) (Ayer et al. 2002; Ayer et al. 2005; Thurston et al. 2008). Stratigraphy has been shown to be upward facing with ~ 20 % of samples containing zircons with the same ages as of the underlying assemblages (van Breemen et al. 2006).
The Abitibi belt has been divided into a northern and southern part based on interpreted stratigraphic and structural differences (Fig. 2) (Dimroth et al. 1982, Ludden et al. 1986, Chown et al. 1992). Recent work suggests that the division may be artificial. The above-mentioned nine assemblages encompass the entire Abitibi belt (Thurston et al. 2008). However, komatiites (indicators of mantle plume activity in the Archean) are more prominent in the southern Abitibi (Dimroth et al. 1982; Ludden et al. 1986). For that reason I will focus on the southern Abitibi belt, and on that part of it in Ontario (Fig. 2).

As I am investigating the role that mantle plumes may have played in the generation of TTG in the Abitibi during LAME, the volcanic assemblages, particularly those containing komatiites, are of interest:

The oldest Abitibi assemblage consists of pre-2750 Ma supracrustal fragments. These are rare, however (only 3 have been found), and occur at the eastern and western margins of the Abitibi belt (Ketchum et al. 2008; Thurston et al. 2008).

The Pacaud assemblage (2750 – 2735 Ma) consists primarily of tholeiitic volcanic rocks with calc-alkaline intermediate to felsic volcanic rocks and minor komatiites (Ayer et al. 2005). The Pacaud assemblage has an ocean basin, plume and arc volcanic signature (Ayer et al. 2002).


The Kidd-Munro assemblage (2719 – 2711) has been divided into a lower and an upper part. The lower Kidd-Munro assemblage (2719 – 2717) consists primarily of intermediate to felsic calc-alkaline volcanic rocks (Ayer et al. 2005). The upper Kidd-Munro assemblage (2717 – 2711) consists primarily of tholeiitic mafic and komatiitic rocks with local occurrences of tholeiitic felsic volcanic rocks and graphitic sedimentary units (Ayer et al. 2005). The Kidd-Munro assemblage has a mixed plume and island arc volcanic signature (Ayer et al. 2002).

The Tisdale assemblage (2710 – 2704 Ma) has been divided into a lower and an upper part. The lower Tisdale assemblage (2710 – 2706 Ma) consists primarily of tholeiitic mafic volcanic rocks with local occurrences of komatiites and intermediate to felsic calc-alkaline volcanic rocks and iron formation (Ayer et al. 2005). The upper Tisdale assemblage (2706 – 2704 Ma)
consists primarily of calc-alkaline felsic to intermediate volcanic rocks (Ayer et al. 2005). The Tisdale assemblage has a mixed plume and island arc volcanic signature (Ayer et al. 2002).

The Blake River assemblage (2704 – 2696 Ma) has been divided into a lower and an upper part. The lower Blake River assemblage (2704 – 2701 Ma) consists primarily of tholeiitic mafic volcanic rocks with isolated tholeiitic felsic volcanic rocks and turbiditic sedimentary rocks (Ayer et al. 2005). The upper Blake River assemblage (2701 – 2696 Ma) consists primarily of calc-alkaline basalt and andesite with local occurrences of bimodal tholeiitic basalt and rhyolite (Ayer et al. 2005). The Blake River assemblage has a rifted island arc volcanic signature (Ayer et al. 2002).

The Porcupine and Timiskaming assemblages are dominantly sedimentary assemblages. The Porcupine assemblage (2690 – 2685 Ma) consists primarily of wacke, siltstone and mudstone along with calc-alkaline felsic volcanic rocks, conglomerates and iron formation (Ayer et al. 2005). The Timiskaming assemblage (2676 – 2670 Ma) consists primarily of subaerial conglomerate and fluvial sandstone along with alkaline and calc-alkaline volcanic rocks (Ayer et al. 2005).

1.2.4.3. Abitibi komatiites

Komatiites are found within four of the seven recognized volcanic assemblages attesting to the importance of mantle plumes in the construction of the Abitibi greenstone belt. These assemblages are the Pacaud assemblage (2750 – 2735 Ma), the Stoughton-Roquemaure assemblage (2723 – 2720 Ma), the upper Kidd-Munro assemblage (2717 – 2711 Ma) and the lower Tisdale assemblage (2710 – 2706 Ma). The volume % of komatiites in the above assemblages is low. Komatiites represent << 1% of the Pacaud assemblage, ~ 2% of the Stoughton-Roquemaure assemblage and ~5% of the Kidd-Munro and Tisdale assemblages (Sproule et al. 2002).

Abitibi komatiites are derived from LILE-depleted mantle, unlike most Phanerozoic plumes that are derived from enriched mantle (Sproule et al. 2002). Komatiites of the Abitibi belt can be subdivided into three groups on the basis of Al$_2$O$_3$/TiO$_2$ ratios, namely Ti-depleted, Al-depleted+Ti-enriched and Al-undepleted. The Pacaud assemblage only contains Ti-depleted komatiites, the Stoughton-Roquemaure assemblage contains predominantly Al-depleted+Ti-enriched and lesser Al-undepleted komatiites, the Kidd-Munro assemblage contains predominantly Al-undepleted and lesser Al-depleted+Ti-enriched komatiites, and the Tisdale assemblage contains predominantly Al-undepleted komatiites with extremely rare Al-
undepleted+Ti-enriched komatiites (Sproule et al. 2002). The temporal trend in komatiite composition suggests a decreasing involvement of garnet in the source region and hence shallower depths of melt extraction (Sproule et al. 2002). Abitibi komatiitic melts are estimated to have been extracted at depths ranging from ~70 – 270 km (2 – 9 GPa) (Sproule et al. 2005). The depth of komatiite melt extraction is not necessarily equivalent to the final depth to which the plume source (or plume restite) rises. Rather, it is a maximum bound on this depth. All the komatiites may have originated from a single, long-lived plume but the prolonged timespan of komatiitic magmatism (~ 44 My) makes multiple plumes (up to four separate ones, one for each of the plume-generated assemblages) more likely (Sproule et al. 2002).

1.2.4.4. Abitibi plutonic rocks

The plutonic rocks of the Abitibi subprovince have been broadly divided into synvolcanic (coeval with the Pacaud to Blake River assemblages), syntectonic (coeval with the Porcupine and Timiskaming assemblages) and post-tectonic (post-dating the Timiskaming assemblage) suites (Chown et al. 1992, 2002; Sutcliffe et al. 1993; Ayer et al. 2005). My focus is on the synvolcanic granitoid rocks (2747 – 2696 Ma) consisting primarily of tonalites and granodiorites. Compositinally these rocks are similar to calc-alkaline members of coeval volcanic assemblages (Ayer et al. 2005).

Plutonic rocks of the Abitibi subprovince are not nearly as well studied as the economically important volcanic ones. Consequently there are far fewer plutonic rock ages available. Ages for plutonic rocks are listed in Table 1 and are shown relative to the timing of volcanic activity in Fig. 3. Age-dating of batholiths such as the Kenogamiissi, Ramsey-Algoma and Round Lake batholith (Fig. 2) reveals a protracted magmatic history spanning ~ 50 – 80 My (Table 1).

1.3. Aim and general methodology of the models

As previously mentioned, the aim of this investigation is to determine in a quantitative fashion, using 1-D thermal models, the thickness of TTG melt that would be produced in a Superior-type LAME-age crust during plume activity. In all my models, the presumed source of TTG melt is the lower crust (as opposed to a subducting slab). The volume of melt produced is determined by applying a melting curve for metabasalt (source of TTG – see Chapter 4), that gives the vol% melt generated at a given temperature, to the model geotherms. I construct two types of models:

1. Models in which a single plume and lava layer are emplaced.
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2. Models in which the emplacement of a single plume, lava layer and lower crustal ‘sill’ is repeated 5 times to mimic a series of plumes or plume pulses (for which there is evidence in the Abitibi). In these models delamination of lower crustal restite also occurs.

Models of type (1) for LAME have already been constructed by Rey et al. (2003). Indeed, the starting point (prior to presenting my own models) will be an in-depth examination of the models of Rey et al. (2003) and the thickness of TTG melt that they generate.

1.3.1. Zegers and van Keken (2001) model for 3.4 Ga in Pilbara and Kaapval

My models of type (2) with delamination of lower crustal restite have some similarity with the model of Zegers and van Keken (2001) and Zegers (2004) for the generation of TTG in the Archean though the treatment in my models is significantly more quantitative.

Zegers and van Keken (2001) and Zegers (2004) suggest that the earliest continental crust, such as that of the Pilbara and Kaapval cratons, may have formed through a process of delamination of lower eclogitic portions of oceanic plateau-like protocrust leading to uplift, extension and the production of TTG. In particular, they propose this model for the voluminous 3.48 – 3.42 Ga TTG event in the Eastern Pilbara craton, in which an estimated 5 – 10 km thickness of TTG was produced. In their model delaminated lower crustal eclogite is envisaged as sinking rapidly through the underlying (harzburgitic) mantle root of the protocrust with little interaction. Zegers and van Keken (2001) suggest that the heat to melt the remaining lower crust to produce TTG was provided by the hot, depleted mantle that rose to replace the eclogite and by the melt produced by decompression melting of this mantle. Plumes do not appear as a feature of their model.

However, based on the geochemistry of ultramafic (komatiitic) and mafic magmas that were emplaced in the East Pilbara between 3.515 – 3.24 Ga (overlapping the above-mentioned TTG event) Smithies et al. (2005) conclude that the Pilbara protocrust and subcontinental lithospheric mantle (SCLM) were the result of a series of mantle plume events, with each plume heating and remelting portions of previous plumes that had accreted to the SCLM (see also Van Kranendonk et al. 2002; Smithies et al. 2007; Van Kranendonk et al. 2007). They propose this as a general model for the formation of Paleoarchean protocrust and Archean SCLM. Geochemical data from mantle xenoliths support the idea that the Pilbara SCLM formed through the accumulation of plume-related melting residues (Griffin and O’Reilly 2007).

Given the evidence for the operation of mantle plumes in the East Pilbara, plumes appear to be a more logical source of heat (as in my models and those of Rey et al. [2003]) than relatively
cool upwelling lithospheric mantle and the minimal volume of mantle melt that would be
generated during such upwelling as in the Zegers and van Keken (2001) model. However, the
model of Zegers and van Keken (2001) cannot be ruled out without a more detailed quantitative
analysis.

2. Examination of the Rey Models

As previously mentioned, Rey et al. (2003) address the problem of what caused LAME, and
they do so using the same tool that I am using, 1-D thermal modelling. Some aspects of the
“Rey models” are problematical, however. I begin with a detailed examination of the Rey
models.

2.1. Mantle plumes and crustal thickening as a cause of LAME

Like Condie (2004), Rey et al. (2003) propose the existence of a widespread (“global”) region of mantle plumes at 2.75 Ga that impinged on the Superior and Yilgarn lithospheres.
Decompression melting of these plumes led to the formation of a thick crustal greenstone cover ranging from 6- to 12-km-thick. The greenstone cover formed a thermal blanket that led to heating of the crust “from above”. At the same time hot plume material penetrated the lower portion of the lithospheric mantle, heating the crust from below. Crustal temperatures increased rapidly, exceeding the melting temperature of tholeiite (in the presence of excess water) resulting in the production of TTG melt and the (dehydration) melting temperature of TTG resulting in granitic melt. Based on data from Nelson (1998) for the Abitibi and Wawa subprovinces and the Eastern Goldfields Province of the Yilgarn Craton Rey et al. (2003) conclude that granitoid production (encompassing both TTG and granites) peaked approximately 40 My after the first appearance of greenstones.

2.2. Set-up of the Rey Models

Rey et al. (2003) present three “plume with greenstone cover” models (the “Rey models”).
The models consist of an initial 42-km-thick crust on which greenstone covers of 6-, 10- and 12-
km thicknesses have been deposited and below which a 50-km-thick layer of hot (1700°C)
plume material has been emplaced. The plume penetrates the lithospheric mantle and is emplaced at the base of the mechanically strong portion of the lithosphere, marked by the 900°C isotherm (Houseman and Molnar 1997) (Fig. 4).

In their modelling, Rey et al. (2003) employ a thermal conductivity of 2.75 Wm⁻¹K⁻¹, a
density of 2750 kg/m³, and a heat capacity of 1000 Jkg⁻¹K⁻¹ for the crust and mantle. They
assume a uniform distribution of heat-producing elements in the crust with crustal heat production of 0.98 μWm⁻³ and negligible heat production (0 μWm⁻³) in the lithospheric mantle. They associate such a distribution with an initial “undifferentiated” crust. While such an assumption may be acceptable for the Abitibi pre-LAME crust as isotopic studies indicate that the Abitibi crust is predominantly juvenile (Carignan et al. 1993; Vervoort et al. 1994; Ludden and Hynes 2000; Ayer et al. 2002), it would not be applicable to the rest of the Superior Province, in which there is ample evidence of older, pre-LAME granitoids (Skulski and Villeneuve 1999; Percival et al. 2006), making it difficult to argue that the pre-LAME crust was truly undifferentiated. This is a minor point however, and I do not consider this as one of the shortcomings of the Rey models.

The value of crustal heat production used in the Rey models is derived from Taylor and McLennan’s (1985, 1995) estimate of the average composition of present-day Archean crust, extrapolated to 2.75 Ga. The average crustal composition determined by Taylor and McLennan is very general compared to crustal heat production estimates that are more specific to the Superior crust. The Rey models use an average crustal heat production of 0.98 μWm⁻³, whereas, based on the work of Perry et al. (2006), Rudnick and Nyblade (1999), Ashwal et al. (1987) and Fountain et al. (1987), a more accurate estimate for the Superior crust is 1.3 – 1.5 μWm⁻³ (Chapter 3). Had the choice of heat production employed by Rey et al. (2003) allowed their model crust to heat up sufficiently to produce adequate amounts of TTG and granite melt one could argue that increasing the heat production would simply result in greater melt. However, I will show that the Rey models do not in fact produce adequate melt.

In the Rey models, the choice of the thickness of the pre-LAME crust and the associated mantle heat flow is made as follows. Present-day thicknesses of Archean crust are estimated at 35 to 50 km and erosion of 5 km is assumed after crustal stabilization at 2.6 Ga (Galer and Mezger 1998). Hence crustal thicknesses at 2.6 Ga are estimated at 40 to 55 km; pre-LAME crustal thicknesses are assumed to have been the same. Pre-LAME mantle heat flow is estimated to be higher than present-day, so at least 15 mWm⁻². The Moho temperature is assumed to have been no higher than 800°C, else gravitational collapse is expected to have thinned (and cooled) the crust. Coupled with a (uniform) crustal heat production of 0.98 μWm⁻³ the above constraints allow a range of possible pre-LAME crustal thickness and mantle heat flow values (Fig. 3 in Rey et al. 2003) from which “average” respective values of 42 km and 25 mWm⁻² are chosen.
2.3. Determining how much melt is produced in the Rey models

In the Rey models it appears that TTG melt is assumed to be produced when lower crustal temperatures exceed the solidus for wet tholeiite and granite melt produced when lower crustal temperatures exceed the solidus for dehydration melting of TTG. In Fig. 4, which presents the transient geotherms of the Rey models, two solidi are shown: one for wet tholeiite (“wts”) and another for dehydration melting of TTG (“tnls”). No justification is provided for why the wet tholeiite solidus is applicable, namely why excess water should be present at lower crustal depths right across the Superior.

In personal correspondence with Rey regarding the use of the wet tholeiite solidus I was instead directed to the text of the Rey et al. (2003) paper in which it is stated that “melting experiments at 500-1000 MPa show that a temperature of ca. 900°C is required to produce 20-40% volume liquid from amphibolite and tonalite dehydration melting”.

However, even the above amphibolite dehydration melting data does not appear to me to be applicable to the Rey models. The final model favoured by Rey et al. (2003) to account for LAME consists of a 10-km-thick greenstone cover deposited on top of a 42-km-thick crust (Fig. 4b). Hence the total crustal thickness is 52 km. Pressure at the Moho is then ~ 1600 MPa. In such a crust melting at 1000 MPa would only occur after approximately 19 km of lower crust had already undergone partial melting.

Ignoring the data for water-saturated melting, there are three recognized experimental studies in which significant melting of metabasalt occurs in the 900 to 950°C range: Wolf and Wyllie (1994) at 1.0 GPa, Rushmer (1991), samples IAT and ABA, at 0.8 GPa and Beard and Lofgren (1991), sample 478 at 0.69 GPa (Chapter 4). Rapp (1997) has analyzed the major element composition of the melt produced in the Wolf and Wyllie (1994) study at 1.0 GPa and has found that it does not fall within the TTG field. Hence the results of the Wolf and Wyllie (1994) study should be ignored. The Rushmer (1991) and Beard and Lofgren (1991) studies have been conducted at pressures corresponding to approximate depths of 26 and 23 km respectively. These depths are significantly less than those at which melting is expected to occur in the Rey models in which the Moho is at 52 km depth. At the same time Rey et al. (2003) have not considered the results of the Rapp et al. (1991) experimental study, conducted at 0.8 and 1.6 GPa (corresponding to approximate depths of 26 and 53 km respectively) in which significantly higher temperatures are required for significant amounts of TTG melt to be produced. I believe the Rapp et al. (1991) study is pertinent to the Rey models.
The melting curve for metabasalt that I have derived (Chapter 4) is based on all of the above studies (except for Beard and Lofgren, 1991) as well as one by Sen and Dunn (1994) conducted at 1.5 and 2.0 GPa (corresponding to approximate depths of 50 and 66 km respectively). Hence I believe that it is more representative of the actual temperatures at which dehydration melting of amphibolite would occur in the Rey models than the studies on which Rey et al. (2003) have focused.

My melting curve is given by:

\[ m_{\text{basalt}} = 0.2 \, T - 170, \]

where \( m_{\text{basalt}} \) denotes volume % melt and \( T \) denotes temperature in °C. This results in a solidus and liquidus of 850°C and 1350°C respectively. Because melt fractions in excess of 40 vol% have a mafic composition (Rapp and Watson 1995) I assumed a maximum degree of melting of 35 vol% (which occurs at 1025°C) in all the models. In reality, melt may segregate from the source at melt fractions even lower than this leaving an anhydrous restite that is unlikely to produce further melt unless significantly higher temperatures are achieved.

2.4. Thickness and timing of melt production in the Rey models

In order to apply my melting curve to two of the Rey models (in which 10- and 12-km-thick greenstone covers have been deposited) I set up and ran the Rey models. The models were only run for a time span of 100 My, reflecting the approximate time span of LAME across the Superior Province. To determine the maximum thickness of TTG melt that could be produced by the models I assumed that the lower crust consisted entirely of metabasalt so that only TTG (and no granite) was produced.

The results of my running of the Rey models appear in Figs. 5 and 6. The Moho temperatures for my run of the “plume plus 10 km thick greenstone” Rey model correspond closely to those determined by Rey et al. (2003) for the same model (Fig. 6). Applying my melting curve to the results of my run of the Rey models gives maximum cumulative TTG melt thicknesses of approximately 713 m and 1.25 km for 10- and 12-km-thick greenstone covers respectively (Fig. 7). The thicknesses of TTG melt produced by the Rey models are too low!

Given that the amphibolite and TTG dehydration melting curves are similar (at least up to 20 – 40 vol% melting – see Chapter 4) the above thicknesses could also be interpreted as the approximate total thickness of TTG and granite (\textit{sensu strico}) melt produced. Clearly they are insufficient when compared to the estimates of the regional thickness of granite melt (~ 1 km) (Cruden 2006) and TTG melt (~ 5 – 8 km) produced during LAME in the Superior Province.
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This is particularly true if one considers the estimated thickness of 16 km of TTG melt emplaced in the Abitibi subprovince (section 4.3) where there is the clearest evidence of mantle plume involvement during LAME. Hence a key problem with the Rey models is that there is not enough melt generated to account for even the TTG melt produced during LAME, let alone the additional granite melt.

Rey et al. (2003) note about their final models that “the peak of the thermal anomaly (> 900°C) is compatible with the production of 20 – 40% volume liquid, and it is reached 40 My after the emplacement of the plume and the deposition of the greenstone sequence”. (In other words maximal crustal temperatures are attained at ~ 40 My in their models). There is an implication in Rey et al. (2003) that a peak in crustal temperatures would correspond to a peak in granitoid production. This is not the case. A peak in crustal temperatures actually corresponds to the tail end of a crustal melting event. Using my own model reruns of the Rey models I have calculated the thickness of TTG melt produced over 1 My intervals in the models. The results show that granitoid production in the Rey models peaks at 23 – 24 My (Fig. 8) and not at 40 My. This is a minor point, however, and not one that I consider to be a problematical aspect of the Rey models.

2.5. Correction of the Rey Models

The manner in which temperatures have been assigned immediately following plume emplacement in the Rey models is also questionable. I will only discuss the Rey model with a 12-km-thick greenstone cover in detail (Fig. 4c). (The model with a 10-km-thick greenstone cover has been set up similarly). Rey et al. (2003) state the thickness of the initial crust as 42 km and of the initial lithosphere as 112 km. If one calculates a steady state temperature profile for such a lithosphere using their model parameters one obtains a temperature at the base of the lithosphere of 1333°C, which is eminently reasonable.

The expectation then is that in order to simulate the emplacement of a 12-km-thick greenstone cover on top of the initial crust the entire initial temperature profile should be shifted downwards 12 km. And in order to simulate the additional emplacement of a 50-km-thick plume layer at the base of the mechanically strong portion of the lithosphere (occurring 22 km below the Moho) the initial temperature profile from 22 km below the Moho to the base of the lithosphere should be shifted downwards an additional 50 km. The initial temperature for the plume layer can then be set at 1700°C.
Rather than taking the above approach, Rey et al. (2003) begin by calculating a steady state temperature profile for a 162-km-thick “lithosphere” (Fig. 4c). The thickness of 162 km is obtained by adding the thickness of a 50-km-thick plume layer to that of a lithosphere of thickness 112 km. To simulate the emplacement of a 12-km-thick greenstone cover they shift the calculated temperature profile downwards by 12 km but to simulate the emplacement of a 50-km-thick plume layer they replace the temperature profile from 22 km below the Moho to 72 km below the Moho (equivalent to the thickness of the plume layer) by 1700°C (Fig. 4c). As a result, at the time of plume emplacement (prior to any heat flowing from the plume layer to the surrounding mantle) the temperature immediately above the plume layer is 896°C and immediately below the plume layer is 1351°C, a difference of approximately 450°C. I feel that this is not reasonable, to simply have the 50-km-thick layer of initial mantle of temperature between 896°C and 1351°C vanish! Clearly the approach they have taken serves to keep temperatures in their model domain high, though still not high enough to result in sufficient crustal melt production.

I tested the effect on crustal melting in the Rey models of the alternative (and I believe more plausible) approach to determining temperatures immediately following plume emplacement that I outlined previously, in which the mechanically weak portion of the lithospheric mantle (with its associated initial temperatures) is moved downwards to accommodate the plume layer. I ran a “corrected Rey model” with a 12-km-thick greenstone cover which only differed from the Rey models in shifting mantle at depth 22 km below the Moho downward by 50 km to accommodate the plume layer (Fig. 9). The cumulative TTG melt thickness produced was only 375 m (Fig. 7). The melting result using this alternative approach contrasts significantly with that for the Rey model with a 12-km-thick greenstone cover, in which a cumulative TTG melt thickness of ~ 1.25 km was produced (Fig. 7).

2.6. Assessment of the Rey Models

The thermal evidence for the plume and crustal thickening model proposed by Rey et al. (2003) to account for LAME does not appear very strong, particularly under the more stringent (and more plausible) numerical implementation that I have proposed in my corrected Rey model. I will now determine if the model can be salvaged (from a thermal viewpoint) by using a higher value for crustal heat production (for which there is supporting evidence in the Superior Province) and by introducing additional variations in the implementation.
3. Modelling plume emplacement alone

I have pointed out two major shortcomings of the Rey models, namely that the models do not produce adequate thicknesses of melt (thicknesses that are comparable to those produced during LAME) and that the manner in which temperatures immediately following plume emplacement are set in the models does not appear reasonable. Here I explore the possibility of modifying the Rey models in a manner that allows the above two problems to be addressed. I naturally choose modifications that promote increased melting. I consider the model that I have constructed to be an end-member model, namely one in which melt production is maximized through a choice of thermal parameters that are within reasonable bounds but are unlikely to have actually been met during LAME. The virtue of such a model is that if the thickness of melt produced is still inadequate then either a fundamental modification of the Rey models is required or (if no such modification can be found) a plume and crustal thickening model can be ruled out as a cause of LAME.

3.1. Model 1: Set-up

In Model 1 the temperature profile immediately following plume emplacement is set as in the corrected Rey model, namely the mechanically weak portion of the lithospheric mantle is shifted downwards by the thickness of the plume layer emplaced above it, carrying its initial temperature downwards with it.

The following features distinguish Model 1 from the corrected Rey model:

1. Temperature-dependent crustal and mantle conductivities were used. When I first examined the Rey models one item that struck me was the value of crustal conductivity used, namely 2.75 Wm$^{-1}$K$^{-1}$, which is higher than one normally finds in the literature when a crust undergoing heating is being modelled. The reasoning is that while such a conductivity (or even higher) might be applicable at low temperatures, as the crust heats up, the associated conductivity goes down. A lower value of conductivity increases heating. Hence I used the temperature-dependent crustal and mantle conductivities described in Chapter 2.

2. Higher crustal heat production was assumed. Like Rey et al. (2003) I assumed a uniform distribution of heat-producing elements in the crust. As mentioned previously, the average crustal heat production value used in the Rey models (0.98 $\mu$Wm$^{-3}$) is too low for the Superior Province. It is derived from data about Archean crust in general. Data that is specific to the Superior Province indicates that average crustal heat production in the Superior Province at 2.7
Ga would likely have been closer to 1.4 μWm⁻³ (Chapter 3). Hence I have used the latter value of crustal heat production.

While my choice is a better estimate of the average heat production of the Superior crust during LAME the assumption of a uniform distribution is decidedly unrealistic. It was chosen because such a distribution serves to maximize crustal temperatures through the “thermal blanket effect”.

3. A lower mantle heat flow was assumed. The thickness of the initial crust is 42 km as in the Rey models. With my choice of thermal parameters and crustal thickness, had I maintained an ambient mantle heat flow of 25 mWm⁻² as in the Rey models, the initial temperature at the Moho would have been 1171°C, which is clearly too high! Hence I decreased the mantle heat flow to 10 mWm⁻², which is at the low end of the range of estimates for current mantle heat flow in the Superior Province (10 – 18 mWm⁻²; Chapter 3). My choice of thermal parameters resulted in an initial Moho temperature of 803°C, which is hot, and should therefore maximize crustal melting, but still below the metabasalt solidus of 850°C associated with my melting curve.

In the Rey models the base of the mechanically strong portion of the lithosphere, where the plume layer is emplaced, is assumed to occur at the 900°C isotherm. I maintain this convention in my model (while recognizing that there is nothing to preclude insertion of the plume layer at a slightly different level). Hence the plume layer is emplaced 22 km below the Moho (the same depth as in the Rey models) where the initial temperature is 900°C. The initial lithospheric thickness is 180 km with a temperature of 1329°C at the base of the lithosphere. The thickness of the greenstone cover is 12 km.

The model was run over a time span of 100 My.

3.2. Model 1: Results

The results for Model 1 appear in Figs. 10 and 7. The cumulative TTG melt thickness is 6.7 km (Fig. 7), a significant increase over that in the Rey models (713 m and 1.25 km for a 10- and 12-km-thick greenstone cover, respectively).

3.3. Model 1: Discussion

The results of Model 1 certainly appear more promising than those of Rey et al. (2003). An even greater thickness of generated melt is predicted for this model if upward transfer of TTG melt from lower to mid- to upper crustal levels is implemented.
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At the end of the model run (at 100 My) partial melt is present in the lower 23.9 km of the crust. The average melt fraction over this region is 0.28. In Chapter 5 it was shown that the incorporation of upward melt transfer into a standard model of lower crustal melting may increase the amount of melt predicted by the model by $\sim 1 / (1 - f)$, where $f$ denotes the fraction of melt that is on average being extracted from the source rocks. Application of this result to Model 1 suggests that $\sim 9.3$ km thickness of melt may be generated in this model.

Though the distribution of heat producing elements in Model 1 is decidedly unrealistic such a result suggests that a plume and crustal thickening model may well be able to account for TTG melt thicknesses of $\sim 5 – 8$ km generated in the Superior Province during LAME.

However, as was pointed out earlier, there simply isn’t convincing evidence at present for plume involvement during LAME in the Superior Province except in the Abitibi and Wawa subprovinces, though such evidence may well be found in the future. And for the Abitibi (where there is evidence for more pervasive plume activity during LAME than in the adjoining Wawa subprovince [Ernst and Buchan 2001]), the above thickness of melt still falls short of the $\sim 16$ km that I estimate to have been emplaced during LAME (section 4.3). If we consider the fact that Model 1 is an end-member model constructed in a manner that essentially gives a maximum bound on the amount of melt that can be generated in a Rey-type model it doesn’t appear that such models can account for the thickness of TTG melt generated in the Abitibi subprovince.

4. Modelling plume emplacement with added ‘sill’, and delamination of restite

I explore an alternative plume-based model in which the thickness of TTG melt generated is even greater than that in Model 1. I am reluctant to present this as a model that explains “all” TTG produced in the Abitibi during LAME, for as mentioned earlier, the geochemistry of Abitibi volcanic rocks suggests that it was constructed through a combination of plume and arc activity (Ayer et al. 2002; Scott et al. 2002; Sproule et al. 2002; Wyman et al. 2002).

4.1. The Restite Problem

While Model 1 has demonstrated that $\sim 7 – 9$ km thickness of TTG melt could possibly have been produced in the Superior Province under the type of mantle plume conditions envisioned by Rey et al. (2003), a model in which restitic material is presumed to persist in the lower crust (as in Model 1 and in the Rey models) is simply inconsistent with the $\sim 15 – 20$ km thickness of predominantly juvenile TTG that constitutes the Abitibi mid- to lower crust. The estimated TTG thickness is based on the assumption that the crustal cross section constructed from observations...
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along the Kapuskasing uplift, where mid- to lower crustal levels (0.33 – 1.1 GPa) of the Abitibi-Wawa belt are exposed, is representative of the Abitibi crust (Chapter 3; Percival and Card 1983; Mäder et al. 1994).

Experimental studies indicate that melt of TTG composition results from ~ 10 – 40 vol% dehydration melting of metabasalt (Chapter 4). Higher degrees of melting result in mafic liquids (Rapp and Watson 1995). Therefore the minimum thickness of a metabasaltic source required to produce 20 km thickness of TTG melt is ~ 50 km, with a minimum thickness of ~ 30 km of mafic restite remaining after extraction of all such melt.

As outlined in Chapter 4, whether TTG in the Archean resulted primarily from lower crustal melting or melting of the oceanic crust of a subducting slab is a matter of huge debate (see Chapter 4 and references therein). I favour the lower crustal melting model for TTG genesis in the Archean and adopt it in this investigation.

The exposure of middle to lower Abitibi-Wawa crust in the Kapuskasing uplift suggests that no more than the lowermost 10 – 15 km of Abitibi crust is of mafic (and hence possibly restitic) composition (Chapter 3; Percival and Card 1983; Percival and West 1994). Hence significant thicknesses of restite must either reside below the seismic Moho or have been recycled into the mantle during LAME.

Geobarometry of mantle xenoliths in Jurassic-aged kimberlite diatremes in the Kirkland Lake area in the south-central Abitibi subprovince indicates pressures of equilibration of ~ 68 – 198 km (Vicker 1997). Compositionally these xenoliths range from lherzolites and harzburgites to dunites. If large-scale (> 30-km-thick) restitic eclogite existed below the seismic Moho (at ~ 40 km depth) one would expect some of the more shalllowly derived xenoliths to be eclogitic but no such xenoliths have been found. Hence, in my models, I assume that excess restite is disposed of by delamination and subsequent recycling into the mantle.

4.2. Implementing Plumes and Restite Disposal

The four Abitibi supracrustal assemblages that contain komatiites and hence are interpreted to have a plume origin are the Pacaud (2750 – 2735 Ma), Stoughton-Roquemaure (2723 – 2720 Ma), Upper Kidd-Munro (2717 – 2711 Ma) and Lower Tisdale (2710 – 2706 Ma) assemblages.

The start time of my models (0 My) corresponds to 2750 Ma, the time at which plume-generated rocks were first emplaced in the Abitibi. Hence in terms of model time the Pacaud assemblage was emplaced between 0 and 15 My (15 My span), the Stoughton-Roquemaure
between 27 and 30 My (3 My span), the Upper Kidd-Munro between 33 and 39 My (6 My span) and the Lower Tisdale between 40 and 44 My (4 My span).

In my models I invoke a series of 5 plumes (or plume pulses) at 0, 8, 27, 33 and 40 My. Four of these (at 0, 27, 33 and 40 My) correspond to the initial time at which rocks in each of the four plume-related assemblages were emplaced. An extra plume pulse was introduced at 8 My to reflect the prolonged plume activity of the Pacaud assemblage (15 My) compared to that of the other assemblages (3 – 6 My).

In Model 1 and in the Rey models a hot slab (1700°C) representing a plume is ‘instantaneously’ emplaced in the mantle while a cool slab (0°C) is instantaneously emplaced at the top of the crust representing plume-generated mafic lavas erupted in an underwater setting. In my “plume with restite disposal” models I introduce an additional slab (or sill) emplaced in the mid- to lower crust representing intra- and under-plated mafic magmas, which along with pre-existing crust constitute the source rocks for TTG. For each of the three type of “slabs” (lava, sill and plume) crustal or mantle material is moved downward to accommodate it. The supra- and intra-crustal additions to the crust incrementally thicken it. Whenever the crustal thickness exceeds ~ 50 km, lower crust below this depth is instantaneously removed. This represents the removal of restitic (and likely eclogitized) portions of the lower crust. Mantle material is moved upward to replace it.

4.3. Thickness of melts emplaced in Abitibi during LAME

Due to the juvenile nature of the Abitibi crust (Corfu and Noble 1992; Kerrich et al. 1999; Wyman et al. 2002) and the pre-LAME xenocrystic zircons that have been found at its margins (Ketchum et al. 2008; Thurston et al. 2008) both an oceanic crust (Ayer et al. 2002) and a highly rifted continental crust (Bleeker 2002) have been suggested as a precursor (pre-LAME) Abitibi crust. Late Archean mantle potential temperatures are estimated to have been ~ 100 – 200°C higher than present day (Abbott et al. 1994) resulting in greater decompression melting at mid-ocean ridges and oceanic crustal thicknesses of ~ 15 – 20 km (McKenzie, 1984). In my models I assume an initial crustal thickness of 25 km, which is near the lower end of the estimated range of pre-LAME crustal thicknesses for the Superior Province (Chapter 3).

I calculate the thickness of the plumes and plume-generated lavas and magmas emplaced in the models as follows. The present Abitibi crust is ~ 40 km thick. I assume an idealized Abitibi crust in which the upper and lower 10 km of the crust consist of mafic rocks and the middle 20 km of TTG. My idealized crust is not dissimilar to that proposed by Grandjean et al. (1995)
based on velocity models for seismic refraction profiles across the Abitibi and Pontiac subprovinces: the crust consists of three layers: (1) an upper crust (~ 0 – 12 km) with variable velocities (5.6 – 6.4 km/s) with higher velocities associated with mafic metavolcanics and lower velocities with metasediments and granitic plutons, (2) a relatively uniform middle crust (~ 12 – 30 km) with velocities of 6.4 – 6.6 km/s reflecting quartzo-feldspathic lithologies and (3) a lower crust (~ 30 – 40 km) with velocities increasing from 6.9 km/s at the top to 7.3 km/s at the base. The higher velocities of the lower crust are interpreted to result from more mafic lithologies (Grandjean et al. 1995). Metamorphic pressure estimates indicate that in much of the Abitibi some 10 km of the crust has been eroded (Easton 2000). I assume that most of the eroded rocks comprised mafic lavas so that prior to erosion the idealized Abitibi crust consisted of a 20-km-thick mafic upper layer, a 20-km-thick TTG middle layer and a 10-km-thick mafic lower layer.

The Abitibi crust experienced significant tectonic thickening coeval with the Porcupine and Timiskaming sedimentary assemblages (Ayer et al. 2005). Such thickening post-dated the emplacement of basalts and most TTGs (Fig. 3). I am not aware of any good estimates of the degree of such thickening. In the analogue models of Benn and Peschler (2005) which attempt to explain late tectonic features of the Abitibi crust, crustal shortening of 20% is implemented. The structures produced, when scaled, compare well to those found in the Abitibi. The degree of shortening corresponds to vertical thickening by a factor of 1.25.

If this degree of thickening produced the above described idealized 50-km-thick Abitibi crust then prior to thickening the crust must have consisted of a 16-km-thick mafic upper layer, a 16-km-thick middle TTG layer and an 8-km-thick mafic lower layer.

4.4. Thickness of plume-generated magmas and lavas

In my models the mafic upper layer is constructed by instantaneously emplacing a cool (0°C) 3-km-thick mafic (basalt) layer at the top of the crust with each plume (or plume pulse). Given that there are 5 such plumes (or plume pulses) this results in an upper mafic layer of total thickness 15 km, comparable to but slightly less than 16 km.

If we assume that the 16-km-thick TTG layer resulted from 20 – 25 vol% melting of a mafic source then the thickness of the source (which includes the thickness of TTG) must have been in the range of 64 – 80 km. Given an initial 25-km-thick model crust, which may be considered as a partial source for TTG, additional source material of 39 – 55 km thickness is required. In my models, with each plume, I also introduce a mafic “sill” of thickness 10 km, giving a total
addition over time of 50 km. This is done via the instantaneous emplacement of a 10-km-thick mafic underplate at 0 My (when the crust is 25 km thick) and emplacement of 10-km-thick mafic sills at a depth of 30 km at subsequent times (8, 27, 33 and 40 My).

The ratio of intrusive to extrusive plume-generated rocks of 10:3 in my models is not unreasonable. The Ontong-Java plateau in the southwestern Pacific is the largest of all oceanic plateaus. Its total volume is estimated to be $44.4 \times 10^6$ km$^3$ (Coffin and Eldholm 2001) while the volume of extruded lavas is estimated at $6 \times 10^6$ km$^3$ (Courtillot and Renne 2003). Given an estimated crustal thickness of 36 km (Courtillot and Renne 2003) this implies a 5 km thickness of extrusive rocks. If an initial thickness of 7 km is subtracted for pre-existing oceanic crust this implies a ratio of ~ 10:2 for intrusive to extrusive rocks generated during Ontong-Java plume activity which is comparable to but slightly greater than the corresponding ratio in my models of 10:3.

4.5. **Thickness of plumes (or plume pulses)**

I run two “plume with restite disposal” models. I assume a plume temperature of 1700°C (as in the Rey models and in Model 1) in Model 2 and a cooler plume temperature of 1500°C in Model 3. The rationale for employing the latter cooler temperature is discussed in the next section. In the Rey models the plume is emplaced at 64 km depth. I round off this value to an emplacement depth of 60 km. The requirement that with each plume (or plume pulse) 13-km-thickness of plume-generated melt be produced (to account for the 3-km-thick basalt layer and 10-km-thick mafic sill) imposes a constraint on the thickness of the associated plume.

McKenzie (1984) has calculated the thickness of partial melt that would be produced by decompression melting of adiabatically rising mantle of varying initial temperatures. (Latent heat of melting is considered). His results accurately predict the thickness of present day oceanic crust and can account for the volume of melt that has been generated by the Hawaiian plume. An examination of Fig. 13 in McKenzie (1984) shows that a 40-km-thick mantle layer of potential temperature ~ 1700°C would contain ~ 13 km of partial melt if the top of such a layer rose to a depth of 60 km. The restitic portion of such a layer would then have a thickness of ~ 27 km. In my models the emplaced plumes represent this restitic portion. For simplicity I assume a plume thickness of 30 km rather than 27 km.

4.6. **Rationale for a cooler (1500°C) plume**

McKenzie’s (1984) work on the thickness of partial melt produced by decompression melting of adiabatically rising mantle suggests that if melt is retained within the mantle source
(facilitating further melting) then significant cooling of such mantle would occur due to absorption of heat (latent heat of melting) during melting. Mantle of potential temperature 1700°C that retains its melt during upwelling would cool to ~ 1525°C at a depth of 90 km and to ~ 1460°C at a depth of 60 km (McKenzie 1984). These depths correspond to the bottom and top of my emplaced model plumes. Further evidence to support a cooler plume head (whether it is the result of McKenzie style cooling or due to the entrainment of ambient mantle into the plume head during its rise) is the relative paucity of high-temperature mafic lavas found in Abitibi plume-related assemblages. Komatiites comprise no more than ~ 5% of such lavas (Sproule et al. 2002) providing a strong argument against uniformly high temperature plumes existing at relatively shallow depths such as in the Rey models and my own.

4.7. Crustal heat production

I assume uniform crustal heat production of 0.6 µWm⁻³ in Model 3 (with plume temperature 1500°C) and heat production of 1.0 µWm⁻³ in Model 2 (with plume temperature 1700°C). (Heat production of 0.98 µWm⁻³ was used in the Rey models).

Both the crustal heat production values I have used are less than the estimated average heat production of the present day Superior crust extrapolated to 2.7 Ga (1.3 – 1.5 µWm⁻³) (Chapter 3). I have not employed heat production as high as this as it would only have been achieved during the later stages of LAME as the Abitibi crust matured and approached its present intermediate bulk composition. As well, the distribution of heat producing elements would not have been uniform but decreasing downward.

During the initial stages of LAME the crust would have been primarily mafic with lower associated crustal heat production. When extrapolated to 2.7 Ga heat production of samples of metagabbro collected in the Sachigo subprovince from the Pikwitonei-Sachigo region (which is interpreted as an oblique cross-section of the Superior crust) ranges from 0.3 – 0.6 µWm⁻³ (Table 3 in Appendix C) (Fountain et al. 1987). Hence the heat production values I have employed are intermediate between those of an immature and mature Abitibi crust, with one value (0.6 µWm⁻³) skewed towards that of an immature crust and the other (1.0 µWm⁻³) towards that of a mature crust.

The coupling of lower heat production (0.6 µWm⁻³) with a cooler plume (1500°C) and higher heat production (1.0 µWm⁻³) with a hotter plume (1700°C) serves to bracket the thickness of TTG that might be produced under the conditions being modelled. Heat production in the
lithospheric mantle is assumed to be negligible (0 µWm⁻³). The heating effect of crustal thickening is implicit in the models.

4.8. Models 2 and 3 Set-up

The models are run for a time span representing 100 My. The initial crust is 25 km thick. The 3-km-thick plume-generated basalt layers emplaced at the top of the crust have a temperature of 0°C. The 10-km-thick plume-generated mafic sills emplaced in the lower crust have a temperature of 1300°C.

The 3-km-thick basalt layers are emplaced on top of previous layers. The 10-km-thick sills are emplaced above previous sills with varying interleaved thicknesses of initial crust. The 30-km-thick plumes (or plume pulses) are emplaced on top of previous ones with some intervening initial lithospheric mantle.

The details of the emplacement of the plumes (or plume pulses) and the associated plume-generated lavas and magmas in the models are as follows:

1. At 0 My, a 3-km-thick plume-generated basalt layer is emplaced at the top of the crust. A 10-km-thick plume-generated mafic sill is emplaced at the Moho. Magmatic thickening by 13 km of the initial 25-km-thick crust results in a crust of thickness 38 km. A 30-km-thick plume (or plume restite) is emplaced 35 km below the Moho. The reasoning is this: the initial plume thickness, which includes the 30-km-thick plume restite, the 3-km-thick basalt layer and the 10-km-thick mafic sill, is 43 km. As previously mentioned, this plume is emplaced at 60 km depth. As the initial thickness of the crust is 25 km, it is emplaced 35 km below the Moho. The total thickness of melt in the plume is 13 km. Of this, 3 km is emplaced at the top of the crust and 10 km at the Moho, leaving 30 km of plume restite to be emplaced in the mantle.

2. At 8 My, a 3-km-thick basalt layer is emplaced at the top of the crust. A 10-km-thick mafic sill is emplaced at 30 km depth. This results in magmatic thickening of the crust from 38 km to 51 km. A 30-km-thick plume (or plume restite) is emplaced 22 km below the Moho (the latter depth below the Moho is the same as in the Rey models). The initial plume thickness is 43 km, with emplacement at 60 km depth. As the initial thickness of the crust is 38 km, it is emplaced 22 km below the Moho.

3. At 27 My, a 3-km-thick basalt layer is emplaced at the top of the crust. A 10-km-thick mafic sill is emplaced at 30 km depth. This results in magmatic thickening of the crust
from 51 km to 64 km. The lowermost 11 km of crust is removed ‘instantaneously’, leaving a crust of thickness 53 km.

When removing lower crustal restite, for simplicity, I leave all previously emplaced 10-km-thick mafic sills intact, i.e. I either remove an entire sill or I don’t remove any portion of it. This determines the precise thickness of lower crustal material that is removed. Following delamination, the 10-km-thick sill that was emplaced at 30 km depth at 8 My now sits above the Moho, between 43 – 53 km depth.

A 30-km-thick plume (or plume restite) is emplaced 9 km below the Moho.

The initial plume thickness is 43 km, with emplacement at 60 km depth. As the initial thickness of the crust is 51 km, it is emplaced 9 km below the Moho.

4. At 33 My, a 3-km-thick basalt layer is emplaced at the top of the crust. A 10-km-thick mafic sill is emplaced at 30 km depth. This results in magmatic thickening of the crust from 53 km to 66 km. The lowermost 13 km of crust is removed in an instantaneous fashion, leaving a crust of thickness 53 km. A 30-km-thick plume (or plume restite) is emplaced 7 km below the Moho.

The initial plume thickness is 43 km, with emplacement at 60 km depth. As the initial thickness of the crust is 53 km, it is emplaced 7 km below the Moho.

5. At 40 My, a 3-km-thick basalt layer is emplaced at the top of the crust. A 10-km-thick mafic sill is emplaced at 30 km depth. This results in magmatic thickening of the crust from 53 km to 66 km. The lowermost 13 km of crust is removed in an instantaneous fashion, leaving a crust of thickness 53 km. A 30-km-thick plume (or plume restite) is emplaced 7 km below the Moho.

The initial plume thickness is 43 km, with emplacement at 60 km depth. As the initial thickness of the crust is 53 km, it is emplaced 7 km below the Moho.

4.9. Determining amount of TTG melt produced

To determine the amount of TTG melt produced in the models I used the melting curve given by equation 1, the same as that used to determine the amount of melt produced in the Rey models. As for the Rey models I assumed a maximum degree of melting of 35 vol%.

The general approach that I take to determine the thickness of melt produced in my models is to subdivide the initial crust into 1-metre-thick sections. I then monitor the location of these sections over time as well as the maximum temperature attained in each section over time. To determine the thickness of melt produced up to a given time I use the maximum temperature
that has been attained within each 1-metre-thick section up to that time to calculate the volume/thickness of melt that has been produced within the section. I then sum the melt thicknesses associated with each such section to determine the total thickness of melt produced in the model crust up to the specified time.

If the initial temperature of each of the 1-metre-thick sections is below the solidus the above approach is relatively straightforward to implement. However, such is not the case in the two “plume with restite disposal” models due to the introduction of the 10-km-thick sills with an initial temperature of 1300°C, well above the metabasalt solidus.

A different approach must be taken to determine the thickness of TTG melt produced in these sills which cool initially and then heat up again due to the arrival of the next plume or the thermal blanket effect. I again subdivide these sills into 1-metre-thick sections but to determine the amount of TTG melt produced in each section I ignore its temperature during initial cooling and only begin to monitor it once the section begins to heat up again.

Ideally the sills should cool to below the solidus but they do not in all cases. The reason they don’t is primarily an artefact of the simple manner in which I have implemented the intrusion of plume-generated magmas into the lower crust. In reality these magmas would be dispersed both in time and space as sporadic thin dikes or sills but in my models they intrude as a single 10-km-thick sill. The time taken for a sill to cool is directly proportional to the square of its thickness (Turcotte and Schubert [2002], eqn. 4-150). Hence a 10-km-thick sill (such as in my models) takes 100 times as long to cool as a 1-km-thick sill. A more realistic implementation involving thinner and staggered sills would have resulted in more rapid cooling, prior to reheating.

If any of the 1-metre-thick sections into which I subdivide the hot sills don’t cool to below the solidus prior to heating up again I use the lowest temperature that they attain during the initial cooling period to determine the amount of initial melt that they contain. I refer to this initial melt as “incipient melt”.

4.10. Model 2: 1700°C plumes, moderate heat production

In Model 2 the temperature of the plumes is 1700°C. Crustal heat production is 1.0 µWm⁻³ and ambient mantle heat flow is 40 mWm⁻². The choice of thermal parameters results in an initial Moho temperature of 613°C (at 25 km depth). The initial 1350°C isotherm occurs at a depth of ~ 71.4 km while the initial 1450°C isotherm occurs at a depth of ~ 79.8 km. This corresponds to a relatively thin lithosphere which is consistent with an initial mafic (oceanic) crust or an initial highly rifted continental crust.
Geotherms immediately prior to and immediately following plume emplacement at 0, 8, 27, 33 and 40 My are shown in Fig. 11, as well as transient geotherms spanning 50 – 100 My.

The cumulative thickness of incipient melt present in the model sills along with the cumulative thickness of TTG melt produced in Model 2 over time is shown in Fig. 12. Over a 100-My time span, corresponding to 2750 – 2650 Ma, TTG melt of total thickness ~ 20.3 km was produced, which includes incipient melt of 6.9 km. Such thicknesses of melt were produced by 56 My (corresponding to 2694 Ma).

In the southern Abitibi tonalite melt production ended by ~ 2695 Ma though granodiorite continued to be produced until ~ 2685 Ma (Fig. 3).

4.11. Model 3: 1500 °C plumes, low heat production

In Model 3 the temperature of the plumes is 1500°C, lower than that in Model 2 (1700°C). Crustal heat production is 0.6 µWm⁻³, also lower than that in Model 2 (1.0 µWm⁻³). Ambient mantle heat flow is 45 mWm⁻², slightly higher than that in Model 2 (40 mWm⁻²). The choice of thermal parameters results in an initial Moho temperature of 613°C (at 25 km depth), the same as that in Model 2. The initial 1350°C isotherm occurs at a depth of ~ 66.2 km while the initial 1450°C isotherm occurs at a depth of ~ 73.7 km, suggesting a relatively thin lithosphere.

The cumulative thickness of incipient melt present in the model sills along with the cumulative thickness of TTG melt produced in Model 3 over time is shown in Fig. 13. Over a 100-My time span, corresponding to 2750 – 2650 Ma, TTG melt of total thickness ~ 16.2 km was produced, which includes incipient melt of 1.3 km. Such thicknesses of melt were produced by 42 My (corresponding to 2708 Ma).

5. Discussion

Some features of Models 2 and 3 are consistent with geological data for the Abitibi while others are not.

5.1. Evaluation of the Plume with Restite Delamination Models: Pros

5.1.1. Thickness of TTG melt

Models 2 and 3 are able to account for the estimated 16 km thickness of TTG melt that was produced in the Abitibi subprovince during LAME. Indeed, this was my prime motivation in devising the models and I have not encountered any other numerical models in the literature that are based on geologically reasonable assumptions in which a similar quantity of TTG melt is produced.
While the models only represent the geological conditions that would prevail during actual plume activity in an extremely crude manner, Model 3 with its combination of a relatively cool plume, hot (but not very hot) sills and low crustal heat production does support the idea that the voluminous quantity of TTG melt that was emplaced in the Abitibi during LAME could have resulted from plume activity coupled with restite delamination.

5.1.2. Thick, depleted Abitibi SCLM

The models are also able to account for the thick, depleted SCLM that is inferred for the Abitibi. The final structure of the lithospheric mantle in my models consists of five plume restite layers, each of thickness 30 km, interleaved with minor layers of initial mantle, with the base of the plume layers at 238 km depth (Fig. 14). This depth is reasonably close to that inferred for the base of the Abitibi lithosphere from geothermobarometry of mantle xenoliths from the Kirkland Lake region, Abitibi subprovince (Vicker 1997; Rudnick and Nyblade 1999) (see Fig. 1 in Chapter 2). The precise depth assigned depends on the manner in which the base of the lithosphere is defined – thermally, petrologically or seismically. Based on global S-wave tomographic models the Abitibi lithosphere has an estimated thickness in excess of 280 km (Abbott et al. 2000).

Based on my model one would expect a depleted mantle root beneath the Abitibi extending down to 238 km depth (or less, allowing for thermal or mechanical erosion). Scully et al. (2004) have used the major and trace element chemistry of garnet xenocrysts from 12 kimberlite and alkaline rock bodies in the Superior Province to determine the protoliths of the garnets and hence reconstruct the lithology of the Superior SCLM (at the time the kimberlites were emplaced). The latter is an important qualifier as tectonothermal events or the introduction of fluids or melts into the SCLM after LAME (or even during the late stages of LAME following plume emplacement) can be expected to have modified it (Griffin et al. 2003b, 2003a).

Emplacement ages of kimberlite samples from the Abitibi subprovince and adjoining Pontiac subprovince range from 135 – 165 Ma. Kimberlite samples from the adjoining Wawa subprovince are ca. 1140 Ma in age. The garnet xenocryst data for the Abitibi, Pontiac and Wawa subprovinces suggest a depleted lherzolite source (up to at least 140 km depth) though there is some evidence for minor depleted harzburgite components in the Wawa subprovince (Scully et al. 2004). Such a source is consistent with a plume origin (Griffin et al. 2009) though other mechanisms cannot be ruled out (Canil 2004). (The debate over the origin of Archean SCLM is similar to but perhaps less heated than that over the origin of Archean TTG).
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In further support of a plume origin for Archean SCLM Griffin et al. (2009) point out that most Archean SCLM has likely undergone some degree of metasomatism (Stachel et al. 1998; Griffin et al. 1999; McCammon et al. 2001; Griffin et al. 2003b; Simon et al. 2003) which increases with depth so that Archean SCLM grades into refertilized lherzolite towards its base. The present Wawa SCLM appears to be composed primarily of garnet lherzolite (Scully et al. 2004), however garnet inclusions in diamonds emplaced at ca. 2.67 Ga in the Wawa subprovince indicate a harzburgitic paragenesis (Stachel et al. 2006). Based on seismic and gravity data for the Kalahari craton Griffin et al. (2009) suggest that most Archean SCLM consists of highly depleted dunite/harzburgite of plume-related origin which is generally not sampled by mantle xenoliths. Rather it is the low velocity regions peripheral to such high velocity cores that are sampled instead.

If the Abitibi pre-LAME crust was oceanic or highly rifted continental crust then the associated thin lithospheric mantle would have to be reconstructed, both in thickness and composition, into the current SCLM or possibly even a more depleted version of it. One possible mechanism is via the accretion of restitic plume material as in my models (Wyman and Kerrich 2002).

5.1.3. Delamination of restite

In my models I assume that lower crust in excess of ~ 50 km depth delaminates. Rapp (1990) has conducted melting experiments on natural amphibolites over a pressure range of 0.8 – 3.2 GPa and temperatures of 1000 – 1150°C. The composition of the partial melts ranges from trondhjemitic-tonalitic to dioritic-gabbroic. The calculated density of the restite for three experiments at 1100°C (and one at 1050°C) at a pressure of 1.6 GPa (~ 53 km depth) ranges from 3.5 – 3.7 g/cm³ (Fig. 15) suggesting that it would be gravitationally unstable with respect to underlying depleted lithospheric mantle of density < 3.4 g/cm³ (Griffin et al. 2003a).

Jull and Kelemen (2001) have calculated the instability times for a dense, lower crustal layer to sink into the mantle. They show that high Moho temperatures (> 700°C) are required for this process to occur in ~ 10 My, but that if Moho temperatures approach 1200°C then the process may be ‘instantaneous’. In Model 3 the Moho temperature just prior to the three delamination events at 27 My, 33 My and 40 My is 1035°C, 1151°C and 1209°C, respectively. Hence even in the cooler of the two ‘plume with restite delamination’ models (with plume temperature 1500°C), model conditions are consistent with rapid, if not instantaneous, delamination.
5.2. Evaluation of the Plume with Restite Delamination Models: Cons

5.2.1. Hydrated source for TTG

Experimental petrology and major element geochemistry of Archean TTG, ranging from the Eoarchean onwards, appear to require a hydrated metabasaltic source (Moyen and Stevens 2006). Trace element geochemistry of most Archean TTG appears to require that (a) garnet be present in the restite (implying pressures > 0.8 – 1.2 GPa) to account for low concentrations of HREEs and (2) that if the source does not possess a “subduction signature” then some amphibole remain in the restite (to account for negative Nb, Ta and Ti anomalies) (Rapp et al. 2003; Foley 2008).

The melting curve that I have used to calculate the amount of TTG melt generated in my models is based on the results of experimental studies of dehydration melting of metabasalt (Chapter 3). Experimental petrologists conducting these studies have generally concluded that Archean TTG could have resulted from dehydration melting at the base of overthickened crust or dehydration melting of a subducting slab. The amounts of water involved are not large. The water contents in the starting materials of the experimental studies on which my melting curve (equation 1) is based are: 0.5 – 1.9 wt% (Wolf and Wyllie 1994), 0.8 – 1.4 wt% (Rushmer 1991), 0.5 – 1.5 wt% (Rapp et al. 1991) and 1.5 wt% (Sen and Dunn 1994).

The requirement of a hydrated source is a problematic aspect of my models as the primary source for TTG in my models is the mafic sills along with some initial crust. One possibility is that the small amounts of water required for the generation of TTG melt may have been present in the plume-generated magmas (Rapp, pers. comm.).

Alternatively, hydrated basalts may have been transported downwards from the surface to lower crustal depths through a process of vertical tectonics (Mareschal and West 1980; Robin and Bailey 2008). Such a process is not and cannot be implemented in my 1-D models. If 40% melting of such a source is assumed (which is high) 40 km thickness of hydrated basalt must be cycled into the lower crust to produce 16 km thickness of TTG. If 20% melting is assumed then 80 km thickness of hydrated basalt is needed. The latter requirement may not be unreasonable if TTG generation is staggered over multiple crust building events but appears excessive for a single crust building event such as LAME in the Abitibi unless a mechanism can be suggested that results in deep hydration of the crust.

If lower crustal melting is indeed the primary mechanism by which TTG was generated in the Abitibi (and not through melting of the oceanic crust of a subducting slab) then the water needed may have come from the mantle in the form of hydrous plume-generated or arc basalts.
In the late Miocene Cordillera Blanca batholith in Peru Na-rich trondhjemitic type magmas are inferred to be partial melts of newly underplated Miocene crust in an arc setting (Petford and Atherton 1996). Source rocks of this type would possess a “subduction signature” which would be inherited by derived TTG. Such a model may be applicable to the Abitibi. But regardless of whether TTG source rocks arrive “from above” as in my model or “from below” as postulated for the source of the Cordillera Blanca batholith my model shows that the thermal pulse from a plume (if shallowly emplaced as in my models) would be sufficient to melt such rocks.

Indeed, the requirement for a voluminous hydrated source for TTG is a problem that plagues most models that propose lower crustal melting as a source of Archean TTG (Zegers and van Keken 2001; Bedard 2006). One reason that such a model has not been abandoned, however, is that geochemical evidence such as the lower Sr, Mg, Ni, Cr and Nb/Ta content of TTGs compared to that of adakites argues for lower crustal melting rather than slab melting as a source of most Archean TTGs (Chapter 4; Smithies 2000; Condie 2005).

5.2.2. Timing of TTG production in the Abitibi

In the Abitibi subprovince the vast majority of rock ages that have been determined are for volcanic and not for plutonic rocks. Hence precise intervals of tonalite episodes have yet to be determined for the Abitibi (Ayer, pers. comm.). With this important proviso in mind I compare the timing of TTG production in my models with that in the Abitibi (Fig. 16). The match between the timing of major pulses of TTG production in my models and TTG ages in the Abitibi is not very good. However, the discrepancies may offer some insight into the factors responsible for this.

I reiterate what was stated earlier that, as there is evidence for the operation of subduction during LAME in the Abitibi, the plume models I have presented can only be considered a partial investigation of the mechanisms that may have been responsible for the production of TTG in the Abitibi during LAME. Fig. 16 shows that in the Abitibi TTG was produced within a few million years of the emplacement of the oldest volcanic rocks in the Pacaud assemblage (2750 – 2735 Ma). In both Models 2 and 3 significant thicknesses of TTG are not produced until 2742 Ma (though minor quantities of TTG begin to be produced as early as 2749 Ma). In contrast to this, in Model 1, TTG is produced within a few million years of the introduction of a plume (Fig. 7). The difference is due to the significantly cooler initial Moho temperature in Models 2 and 3 (613°C) compared to that in Model 1 (803°C). The pre-LAME Abitibi crust may have
been hotter than the initial crust in my models or melting may have occurred at shallower depths at which lower melting temperatures would be applicable.

TTG production in my models is waning during the emplacement of the Deloro assemblage (2730 – 2724 Ma in the southern Abitibi). Because a subduction model is favoured for the generation of the Deloro assemblage I do not expect my models to be able to account for TTG produced during this period.

In the Abitibi little TTG appears to be coeval with the Stoughton-Roquemaure or Tisdale assemblages while significant TTG is produced during the associated periods in my models. Periods of TTG production preceded the emplacement of both of these assemblages and it is possible that suitable TTG sources may have been exhausted during these former periods. With regard to the Tisdale assemblage there is only a gap of 1 My separating the upper Kidd-Munro assemblage (2717 – 2711 Ma) and the lower Tisdale assemblage (2710 – 2706 Ma) suggesting that a single plume may have been responsible for the generation of these two assemblages. The main thermal pulse from this plume may have waned by the time of the Tisdale assemblage (2710 – 2706 Ma) or, as mentioned earlier, the generation of TTG coeval with the (earlier) upper Kidd-Munro assemblage (2717 – 2711 Ma) (Fig. 16) may have exhausted the TTG source.

In both Models 2 and 3 the TTG episode associated with the upper Kidd-Munro assemblage is telescoped to the earlier part of the span of this assemblage (Fig. 16). This is inconsistent with the later appearance of TTG coeval with this assemblage (Fig. 16). This suggests that the input of heat in my models is too rapid (plume is too shallowly emplaced?) or that the emplacement of plume-generated intrusive magmas as single 10-km-thick sills results in a hotter source than would be present if such magmas were emplaced as thinner sills distributed over the time period associated with the given assemblages.

5.2.3. Model mid-crust

The final configuration of my model crust contains initial crust between 15 and 30 km depth (Fig. 14). Such crust is nowhere to be found in the Abitibi. Rather this region is occupied by TTG which is mantled by the upward younging supracrustal assemblages (Ayer et al. 2002). An improvement over my implementation would be to incorporate the upward transfer of TTG melt from the source region. The effect would be to force all initial crust downwards into the region of melting requiring a lesser amount of plume-generated magma to be introduced as a source of TTG.
6. Conclusion

A comprehensive treatment of the thermal conditions existing in the Abitibi during LAME appears to require a somewhat complex model that incorporates several plumes (or plume pulses), an active subduction zone and the intermittent opening of slab windows (Benn and Moyen 2008). I view the purely plume models that I have presented as preliminary ones which extend the work of Rey et al. (2003) but unlike their models demonstrate how significant thicknesses of TTG (up to 20 km) can be generated as a result of plume activity.

Finally, I point out that based on lithologic and geochemical data Tomlinson and Condie (2001) suggest that ~ 80% of Early Archean (> 3.0 Ga) greenstones worldwide may have plume affinity while their estimate for Late Archean (3.0 – 2.5 Ga) greenstones is ~ 35%. Hence my models may have greater applicability to the Early Archean than to the Late Archean.
References


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Table 1. Geochronology of plutonic rocks of Abitibi subprovince.

<table>
<thead>
<tr>
<th>Date (Ma)</th>
<th>Rock type</th>
<th>Geological unit</th>
<th>Reference</th>
</tr>
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<tbody>
<tr>
<td>2747.3 ± 2.6</td>
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<td>Ketchum et al. (2008)</td>
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<td>Round Lake batholith</td>
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Table 1. Geochronology of plutonic rocks of Abitibi subprovince (continued).

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Note: Hornblende (Hb), Biotite (Bt), Muscovite (Mu), Quartz (Qz), Feldspar (Fs).
Fig. 1. Location (and size) of Abitibi and Wawa subprovinces within the Superior Province. Modified from Thurston et al. 2008.
Fig. 2. Stratigraphic map of the Abitibi greenstone belt. Modified from Thurston et al. 2008.
Fig. 3. Schematic timeline for the emplacement of the volcanic and sedimentary assemblages and of plutonism in the southern Abitibi. (Modified from Ayer et al. 2005). In the southern Abitibi Deloro-aged rocks span 2730 – 2724 Ma. Superimposed are ages of plutonic rocks appearing in Table 1 for the entire Abitibi subprovince.
Fig. 4. Transient geotherms of the Rey models following deposition of 6-, 10- and 12-km-thick greenstone covers and the emplacement of a 50-km-thick, 1700 °C plume head at the base of the strong portion of the lithosphere, located at the ~ 900 °C isotherm (22 km below the Moho). wts: wet tholeiite solidus; tnls: biotite dehydration (tonalitic) solidus. From Rey et al. 2003.
Fig. 5. Transient geotherms (at 0, 1, 5, 10, 20, 40, 60, 80 and 100 My) of the Rey models I have rerun with (a) a 10-km-thick cover and (b) a 12-km-thick cover. The numbers on the geotherms denote the corresponding time in My.
Fig. 6. Moho temperatures in (a) the Rey model with a 10-km-thick greenstone cover (modified from Rey et al. 2003) and (b) the Rey models I have rerun with a 10- and 12-km-thick greenstone cover.
Fig. 7. Cumulative thickness of TTG melt produced in my rerun of the Rey models with a 10- and 12-km-thick greenstone cover, in my “corrected” Rey model with a 12-km-thick greenstone cover and in Model 1 (modified Rey model) with a 12-km-thick greenstone cover. In Model 1 temperatures exceeded those associated with 35 vol. % melting. However, in calculating the amount of TTG melt that was produced in the model I assumed that this was the maximum degree of melting that had occurred.
Fig. 8. Incremental thickness of TTG melt produced over 1 My intervals in my rerun of the Rey models with a 10- and 12-km-thick greenstone cover.
Fig. 9. Transient geotherms (at 0, 1, 5, 10, 20, 40, 60, 80 and 100 My) of the “corrected” Rey model with a 12-km-thick cover. The initial mantle below the depth at which the plume was emplaced (22 km below the Moho) was shifted downwards by 50 km to accommodate the plume. The numbers on the geotherms denote the corresponding time in My.
Fig. 10. Transient geotherms (at 0, 1, 5, 10, 20, 40, 60, 80 and 100 My) of Model 1. The numbers on the geotherms denote the corresponding time in My.
Fig. 11. Geotherms immediately prior to and following plume (and basalt and mafic sill) emplacement in Model 2 at (a) 0 My, (b) 8 My, (c) 27 My, (d) 33 My and (e) 40 My. (f) Transient geotherms at 10 My intervals from 50 – 100 My (50 and 100 My geotherms are labelled). Broken lines delineate adiabatic temperatures in the mantle (assuming a gradient of 0.6 °C/km) for mantle potential temperatures ranging from 1200 – 1600 °C.
Fig. 11. (continued).
Fig. 11. (continued).
Fig. 12. Cumulative thickness of TTG melt produced in Model 2 (1700°C plumes) over time (0 My in the model corresponds to 2750 Ma). The arrows denote the point at which the 5 plumes are emplaced. Incipient melt thickness is also shown.
Fig. 13. Cumulative thickness of TTG melt produced in Model 3 (1500°C plumes) over time (0 My in the model corresponds to 2750 Ma). The arrows denote the point at which the 5 plumes are emplaced. Incipient melt thickness is also shown.
Fig. 14. Lithospheric column showing the final location of emplaced basalt layers, mafic sills (s1, s2 and s3 have been removed by delamination) and plumes (or plume restite) in Models 2 and 3.
Fig. 15. Densities of melts and residues calculated from melting experiments on natural amphibolites at 1100 °C and pressures ranging from 16 to 32 kbar. Modified from Rapp (1990).
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ASSEMBLAGES

VOLCANISM

SEDIMENTATION
iron formation
turbidites
alluvial-fluvial
plutonic rock ages

INTRUSIONS
tonalite
granodiorite

Unconformity
Unconformity

SYNVO CANIC
SYNTE CTIC
LATE-TECTONIC

0 500 1000 1500 2000 2500 3000 3500
thickness (m)

2750 2740 2730 2720 2710 2700 2690 2680 2670 2660 2650
time (Ma)

incipient melt □ non-incipient melt

other granodiorite tonalite
Chapter 6: The Role of Mantle Plumes in the Production of TTG during LAME

ASSEMBLAGES

VOLCANISM

SEDIMENTATION
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(b)
Fig. 16. Thickness of TTG melt produced in (a) Model 2 and (b) Model 3 over 1 My intervals. A schematic timeline for the emplacement of the volcanic and sedimentary assemblages and of TTG plutonism in the southern Abitibi is also shown (modified from Ayer 2005). Superimposed are ages of plutonic rocks for the entire Abitibi subprovince.
Chapter 7: The Origin of Late Granites in the Superior Province

Chapter 7

The Origin of Late Granites in the Superior Province

Abstract

Late granites comprise the final phase of magmatic activity during LAME (ca. 2.75 – 2.65 Ga) in the Superior Province. Relative to accretion and major orogenic thickening (spanning approximately 2720 – 2680 Ma) some of these granites were produced early (within 5 – 10 My) and some late (over a span of ~ 60 My). This range of timing is difficult to explain using classical crustal thickening models in which (relative to the timing of crustal thickening) melting would tend to begin later and end earlier than what is observed in some areas of the Superior Province. Furthermore, in classical crustal thickening models, melting occurs at the base of a crust that typically exceeds 50 km in thickness. The geochemistry of the late granites of the Superior Province suggests a shallower source.

Using 1-D thermal modelling, I show that the earliest of the late granites may have resulted from thickening of a crust that had been “pre-heated” during preceding arc or plume activity while the latest of the granites may have been the result of continued underthrusting of fertile source rocks to deeper and hotter levels of the crust. Evidence for such underthrusting, which appears to have extended up to 100 My following accretion, is present in the Kapuskasing Structural Zone in the Wawa Subprovince. Pre-heating of the crust during earlier arc or plume activity would have resulted in an anhydrous, infertile, lower crust, forcing the production of late granites from a shallower source.

1. Introduction

As outlined in Chapter 1 the prevailing view of the origin of the Superior Province, tentatively accepted in this investigation, is that it was assembled through a series of collision-accretion events that resulted in the amalgamation of continental and oceanic fragments (Langford and Morin 1976; Card 1990; Williams et al. 1992; Stott 1997; Card and Poulsen 1998). In the western Superior Province five such collisions took place between 2720 and 2680 Ma (Percival et al. 2006). The underlying assumption is that plate tectonics was fully operational at the time.
The accretion events consisted of a remarkably similar series of steps. They were predated by ~ 40 – 50 My of predominantly mafic volcanism and TTG plutonism, ascribed primarily to arc activity. This was followed by the “cessation of arc magmatism, early deformation, synorogenic sedimentation, sanukitoid magmatism, bulk shortening, regional metamorphism, late transpression, … emplacement of crust-derived granites, and postorogenic cooling” (Percival et al. 2006).

In the Superior Province, as in many Archean cratons, a late “granite bloom” comprised the final stage of magmatic activity prior to stabilization. Though such granites are generally classified as “late syntectonic” to “post-tectonic” granite experts argue that such a classification may be misleading as most (all?) granites show evidence of some deformation (Cruden, pers. comm.).

The heat source responsible for the production of the late granites during LAME remains enigmatic, as it does for much of more recent granitic magmatism. Here I investigate it using 1-D thermal models of the Superior lithosphere during LAME. My aim is to construct models that can explain the timing of granitic melt production relative to crustal thickening as well as the estimated thickness of granitic melt produced during LAME.

I begin with a review of late granitic magmatism in the Superior Province.

1.1. Late granitic magmatism

1.1.1. Composition and occurrence

The late granites of the Superior Province range from quartz monzonite to granite, in many cases occurring as part of granite-granodiorite suites (Sylvester 1994; Cruden 2006). Their composition, coupled with the results of experimental melting studies, suggests that they are crustally derived (Patiño Douce and Johnston 1991; Skjerlie and Johnston 1993; Sylvester 1994). Sm-Nd isotopic data indicate a range of crustal residence times for the source rocks, from brief (juvenile sources) to protracted (Beakhouse and McNutt 1991; Ducharme et al. 1997).

They span both I-type and S-type granites with inferred sources being predominantly TTG or metasediments, respectively. They tend to be leucocratic with SiO₂ > 70% (Sylvester 1994). Negative Eu anomalies are common, suggesting partial melting within the plagioclase stability field. I-type granites tend to be mildly peraluminous whereas S-type granites are strongly peraluminous (Beakhouse 2007).

Compositionaly, many of the I-type granites can be categorized as CA1 granites (Sylvester 1994; Percival and Pysklywec 2007). The geochemistry of CA1 granites is consistent with
melting outside the garnet stability field, suggesting melting depths of less than 30 – 35 km (Skjerlie and Johnston 1993; Skjerlie et al. 1993; Sylvester 1994).

I-type granites appear to be restricted to regions that are dominated by granitoid rocks that have been subjected to medium- to high-grade metamorphism and where older TTG (> 2.85 Ga) is present, such as the Winnipeg River and central Wabigoon subprovinces (Beakhouse 2007).

S-type granites tend to occur at boundary zones between subprovinces within (or proximal to) highly metamorphosed, migmatitic metasedimentary rocks. Examples are the Lacorne Block at the southern margin of the Abitibi subprovince, adjacent to the Pontiac subprovince (Feng and Kerrich 1990; Mulja et al. 1995a, 1995b; Ducharme et al. 1997) and the boundary zones between the Wabigoon, Winnipeg River and English River subprovinces (Breaks and Moore 1992; Larbi et al. 1999).

1.1.2. Timing

We examine the ages of granites, monzogranites and quartz monzonites in the Superior Province, along with ages of felsic pegmatites that often accompany granitic magmatism (Fig. 1). While isolated ages for granites older than 2700 Ma appear, the data indicate that major late granitic activity began between 2700 – 2695 Ma and most such activity ended by ~ 2655 Ma, with isolated granitic activity continuing to ~ 2630 Ma. This is similar to the age range of 2695 – 2655 Ma for late granitic-granodioritic magmatism estimated by Cruden (2006). Hence, Superior-wide, late granitic magmatism appears to have lasted ~ 40 My, with isolated activity extending this span to ~ 65 My.

The initial appearance of pegmatites is essentially coeval with that of the late granites (Fig. 1) but, as discovered by Krogh (1993), pegmatites continued to be generated for ~ 50 My after granitic magmatism ended, until ~ 2580 Ma. Due to the diachronous nature of the accretionary events that shaped the Superior Province (Card and Poulsen 1998; Percival et al. 2006) the timing of granitic magmatism relative to orogenic deformation (which is an important constraint on my thermal models) cannot be ascertained from Superior-wide ages of granites. Hence we next turn to the details of late granitic magmatism at a local scale within selected regions of the Superior Province.

1.1.3. Late Granites of the Winnipeg River, Wabigoon and Abitibi subprovinces

As pointed out by Percival and Pysklywec (2007), the emplacement of late granites in the Superior Province followed regional tectonism by 5 – 20 My. The following review presents some examples of the range in timing of granitic magmatism relative to orogenic thickening in
the Superior Province. Some granites were produced early, within 5 – 10 My of orogenic thickening (Fig. 2), and in some regions granitic magmatism extended over a span of 50 My.

1.1.3.1. Winnipeg River and Wabigoon subprovinces

The ages of volcanic and plutonic rocks in the primarily LAME-aged western Wabigoon and older Winnipeg River subprovinces suggest that the two subprovinces developed independently as juvenile magmatic arcs with collision between the two subprovinces occurring at ~ 2707 ± 8 Ma (Beakhouse and McNutt 1991; Davis 1996; Beakhouse 2007). In the Winnipeg River subprovince, crustal I-type magmatism began at 2702 Ma (within 5 – 13 My of collision) and continued to 2660 Ma (Gower et al. 1983; Beakhouse and McNutt 1991; Beakhouse 2007). The timing of I-type granitic (and granodioritic) magmatism overlapped that of high-grade metamorphism associated with tectonic crustal thickening (Beakhouse and McNutt 1991; Beakhouse 2007).

The Sioux Lookout Terrane (SLT) forms the boundary zone between the Winnipeg River and Wabigoon subprovinces (Larbi et al. 1999). Within the SLT the S-type leucogranites of the Ghost Lake batholith (2685 Ma) (Breaks and Moore 1992; Larbi et al. 1999) and pegmatite dykes (2650 ± 3 Ma) (Larbi et al. 1999) that cross-cut the Gullwing Lake batholith reflect both an earlier and later episode of granitic activity, with the latter occurring 35 – 50 My following collision.

A gap of only 5 – 10 My exists between collision of the western and central Wabigoon regions at ca. 2700 Ma and high-K magmatism in the central Wabigoon at 2693 – 2685 Ma (Whalen et al. 2004).

1.1.3.2. Abitibi subprovince

Leucogranites are rare in the Abitibi subprovince. However, granitic, leucocratic, S-type plutons (Preissac, Lamotte and Lacorne plutons) do occur within the fault-bounded Lacorne Block (Feng and Kerrich 1992) located at the southern margin of the Abitibi subprovince, adjacent to the Pontiac subprovince.

Granitoids of the Lacorne Block consist of an early diorite-granodiorite suite (2690 – 2670 Ma) and a later leucocratic monzogranite-granite suite with associated pegmatites (Rive et al. 1990; Ducharme et al. 1997). Examples of the latter suite are the Lamotte pluton and leucogranitic phases of the Preissac and Lacorne plutons (Ducharme et al. 1997). The ages of the Preissac leucogranites range from 2681 to 2660 Ma while those of the Lamotte and Lacorne leucogranites from 2647 to 2630 Ma (Ducharme et al. 1997 and references therein). Syn-
deformational deposition of greywacke in the northern Pontiac subprovince and within the southern Abitibi subprovince is bracketed at 2685 ± 3 Ma (Davis 2002) suggesting the proximity of the two subprovinces at that time. Field observations and isotopic data suggest that in the Pontiac subprovince D1 deformation began no earlier than 2694 Ma (Benn et al. 1994), an estimated age for collision between the Pontiac and southern Abitibi subprovinces.

Crustal thickening resulting from collision of the Pontiac subprovince with the southern Abitibi volcanic arc has been suggested as a cause of crustal heating leading to the generation of granitic melts (Feng and Kerrich 1992; Feng et al. 1992, 1993).

Hence initial granitic magmatism post-dated the start of crustal thickening by 13 My, while granite production spanned 51 My.

1.1.4. Regional thickness of late granites

An important constraint on my thermal models is the volume of late granites produced during LAME. As discussed in Chapter 4, Cruden (2006) estimated the regional thickness of late granites emplaced during LAME as 1 – 3.1 km, based on an empirical relationship between the area of an exposed pluton and its volume. The former value results from the assumption that all granitic plutons are wedge-shaped, the latter that they are tabular. Cruden (2006) points out that published gravity and seismic reflection studies favour the assumption that Superior granitic plutons are wedge-shaped (Jackson et al. 1995; Everitt et al., 1998), hence a thickness of 1 km is likely more accurate.

Furthermore, as a single contiguous area of granites (treated as a single pluton in the approach used by Cruden [2006]) may in fact correspond to a collection of individual plutons, both of the above estimates should be regarded as upper bounds (under the differing assumptions about the geometry of the granitic plutons). Counterbalancing this is the fact that some granitic rocks have been lost through erosion of volcanic edifices and plutons, while others may occur in the subsurface without any portion being exposed at the surface.

It is worth noting that compared to the earlier part of LAME the late granitic component is relatively minor. In Chapter 3, I estimated the regional thickness of TTG melt produced during LAME as 5 – 8 km. The actual thickness is undoubtedly greater as I focused on juvenile TTG in deriving my estimate. An estimated regional thickness of 1 km of granitic melt indicates that the late granites were really a minor component of the cumulative magmatic activity that made up LAME. Therefore, significant heating of the Superior crust may not have been required for their generation.
1.2. Previous Models for Late Granites in the Neoarchean

Several models have been proposed to account for the late granites. I discuss a few.

1.2.1. Classical crustal thickening

A classical crustal thickening model, in which an ambient crust is tectonically thickened and heats up due to the “thermal blanket effect” is perhaps the most commonly proposed model for the late granites (Sylvester 1994; Mareschal and Jaupart 2006). However, as I will show, the rate of heating of such a crust would likely have been too slow to meet the time constraints imposed by the earliest of the late granites, which in some regions appeared within 5 – 10 million years of orogenic thickening (Fig. 2).

Such models have also been rejected by some authors on the grounds that they cannot explain the prolonged nature of granitic activity (up to 60 My in some areas of the Superior Province) (Percival and Pysklywec 2007).

1.2.2. Delamination

Percival and Pysklywec (2007) suggest that a 20-km-thick and 150-km-wide raft of restitic eclogite may have delaminated from the base of the crust, triggering an overturn of the lithospheric mantle that resulted in sudden, rapid heating of the lower crust. A close examination of their model (see Figs. 2 and 3a in Percival and Pysklywec [2007]) reveals the extent of melting to be extremely narrow, both in space and time. In addition, melting occurs at too great a depth (~ 60 km) for granitic melt to be in equilibrium with plagioclase. Furthermore, a series of such eclogitic rafts would be required to delaminate in succession to account for granite production throughout the Superior Province.

In the Archean Slave Province a similar late “granite bloom” occurred between 2610 and 2585 Ma (Davis et al. 2003). Davis et al. (1994) suggested that partial lithospheric delamination following collision gave rise to mantle-derived magmas and the late granites. In their model delamination occurs at ~ 120 km depth. Its occurrence at such a late stage of assembly, however, poses a difficulty in accounting for the thick, buoyant Slave lithosphere (190 km or more; Jones et al. 2005).

1.2.3. Crustal thinning followed by thickening

Thompson (1989) has proposed an alternate model for the above-mentioned late granites in the Slave Province. He suggests that granitic magmatism and the formation of low pressure-high temperature metamorphic terranes in the Slave can be accounted for by initial thinning of the crust accompanied by sediment deposition, followed by homogeneous shortening and
Chapter 7: The Origin of Late Granites in the Superior Province

thickening of the crust and overlying sedimentary basins. Schematic illustrations of the P-T paths followed by a rock (1) initially at the top of the crust and subsequently at the bottom of a sedimentary basin, and (2) at the base of the crust, are shown (see Fig. 3 in Thompson [1989]). They are based on geological data.

A problem with the model is that at least some of the temperatures associated with the P-T paths shown (which again, are based on geological data) are unlikely to be achieved in the model that Thompson (1989) is proposing. For example, during the extensional phase, a rock initially at the surface (0°C) which eventually forms the base of a sedimentary basin reaches a temperature of 250°C at depth 5.5 km within 14 My (Fig. 3 in Thompson [1989]). No numerical modelling is done in support of the model. Modelling by Jackson (1987) is cited for the extensional phase, but in that model (in which an initial 30-km-thick crust is thinned to 15 km) a rock initially at the surface that forms the base of a sedimentary basin only reaches 23°C at 14 My, and does not reach 100°C even after 200 My (see Fig. 5b in Jackson, 1987). I would expect that a numerical implementation of Thompson’s (1989) model would support crustal melting but not the achievement of the given low P-high T conditions.

1.3. Essential features of my model for late granites

I discuss below the two main ideas on which my variant of the classical crustal thickening model for the late granites is based. I propose a crust that has been pre-heated by earlier arc (or plume) activity to account for the earliest of the late granites, and prolonged ductile underplating of mid-crustal rocks à la Krogh (1993) (section 1.3.2) to account for the latest of the granites.

1.3.1. Crustal heating from arc activity

As previously mentioned, the prevailing interpretation of the majority of basalts and TTGs that preceded the late granites during LAME (Fig. 2) is that they resulted from arc activity (Percival et al. 2006). The first of the two ideas on which my model for the late granites is based is that arc activity (or as an alternate, plume activity, as in Chapter 6) produces anomalously hot crust. This idea is important for the generation of the earliest late-stage granites as it suggests that at the end of arc (or plume) activity in the Superior Province source rocks with elevated but subsolidus temperatures may have been present in the deep crust that could well have produced granitic melt within a few million years following orogenic thickening.

That a crust heated during basaltic volcanism and TTG magmatism may have played a role in later granitic magmatism and metamorphism has been suggested by some authors. Whalen et al. (2004) point out that only a narrow temporal gap of 5 – 10 My exists between collision of the
western and central Wabigoon regions at ca. 2700 Ma and high-K magmatism in the central Wabigoon at 2693 – 2685 Ma and that this gap is insufficient for significant crustal heating to have occurred as a result of collisional thickening unless the central Wabigoon crust was already hot as a result of arc activity. The North Cariboo superterrane experienced widespread calc-alkaline and late sanukitoid magmatism between 2745 and 2704 Ma (Corfu and Stone 1998; Percival 2007). Percival (2007) suggests that this prolonged magmatic activity produced a hot North Cariboo upper plate that at the time of collision and overthrusting of the lower plate English River synorogenic sediments and Winnipeg River terrane may have had a “hot iron effect” on these rocks resulting in low P-high T metamorphism.

Thermal models of subduction zones show a significant increase in mantle heat flow under the arc and back arc regions relative to that prior to subduction (Conder et al. 2002; van Keken et al. 2002; Kelemen et al. 2003). This effect is particularly dramatic in more recent thermal models which employ a temperature-dependent viscosity for the mantle (Conder et al. 2002; van Keken et al. 2002; Kelemen et al. 2003). Details of lower crustal and upper mantle temperatures reached in subduction zones appear in section 7.2, and will be compared with temperatures reached in my models.

1.3.2. Krogh’s (1993) tectonic ductile underplating model

The second cause of heating in my model, that I believe was responsible for the prolonged granitic magmatism in some regions of the Superior Province, is that a process of tectonic “ductile underplating” (Krogh 1993) may have been operating at mid- to lower crustal levels up to ~ 100 My after termination of major volcanic and plutonic activity recorded in the upper crust. Such a mechanism may have resulted in underthrusting of mid-crustal source rocks into hot, lower crustal regions leading to heating and the production of granitic melt much later than that predicted by classical crustal thickening models.

I believe that Krogh’s (1993) model is an extremely important one that has implications for the final structure of Archean crust and associated mantle root that extend far beyond its application to granitic magmatism. I discuss it in some detail here.

The Kapuskasing uplift in the Wawa subprovince is interpreted as an oblique cross section of the Superior crust (Percival and Card 1983) (Chapter 2). The lowermost exposure of crust occurs in the Kapuskasing Structural Zone (KSZ) which consists of mafic granulite, tonalite, metasediments and anorthosite. Granulites are subhorizontally laminated with E-W lineations.
The maximum recorded metamorphic temperatures and pressures in the KSZ range from 750 – 790°C and 10 – 11 kbar, respectively (Mäder et al. 1994).

U-Pb ages of various components of an upper mafic granulite gneiss, a lower mafic granulite gneiss, a tonalite underlying the lower mafic granulite gneiss and a conglomerate of amphibolite grade that extends into granulite grade were determined (Krogh 1993). Krogh’s (1993) results indicate a pattern of successive emplacement and metamorphism younging downward.

In the upper mafic granulite (prograde) metamorphism occurred at 2660 ± 2 Ma and 2647 ± 2 Ma and rehydration and rupturing (with infilling by pegmatites) at 2640 ± 2 Ma. The latter event is coeval with prograde metamorphism in the lower mafic granulite. This unit was rehydrated and disrupted at 2630 ± 2 Ma which coincides with the age of a crosscutting pegmatite in the upper mafic granulite. Mafic granulite xenoliths in the underlying tonalite were pulled apart and infilled by pegmatites at 2601 ± 2 and 2586 ± 2 Ma. The latter age is similar to that of a pegmatite dyke (2582 ± 2 Ma) in the lower mafic granulite.

The youngest U-Pb protolith age determined for the conglomerate is 2667 ± 2 Ma which gives a maximum age of deposition. Hence a late tectonic mechanism appears to have been operating that was 1) carrying supracrustal rocks downwards tens of kilometres into granulite grade and 2) successively underplating granulite grade rocks with those of lower grade. Krogh (1993) suggests that given the young ages of metamorphism in the Superior Province the operation of this mechanism may have been widespread. Krogh’s late tectonic ductile underplating model is consistent with seismic reflection profiles across the Western Superior and Opatica Abitibi terranes (see Foldouts 1a, 1b, 2a and 2b in van der Velden et al. 2006).

Krogh (1993) does not propose a heat source responsible for the granulitization of the successive ductile underplates. Moser et al. (1996) suggests that a series of four lithospheric mantle (and lower crustal) delamination events occurred at 2660, 2640, 2630 and 2585 Ma. However, as even he points out, supporting evidence in the form of coeval basaltic activity (Kay and Kay 1993) is lacking. A further problem with this model, and indeed any model that invokes the delamination of mantle lithosphere at a late stage of assembly of an Archean craton, is in accounting for the thick, depleted mantle roots that are a characteristic feature of Archean cratons (Abbott et al. 2000).

Using 1-D thermal modelling I will show that remanent heat from earlier arc activity coupled with in situ heating from crustal thickening would have sufficed to maintain high crustal temperatures and that the tectonic emplacement of fertile rocks into even deeper (and hotter) crustal levels than those exposed in the KSZ (or into locally hotter regions of the crust) could
have prolonged granitic magmatism beyond that predicted by classical crustal thickening models.

2. Modelling parameters and initial conditions

I construct 1-D transient heat conduction models of the Superior lithosphere during LAME. The thermal models presented here are divided into ‘non-arc models’, in which the crust is considered to be at normal temperature before being thickened and ‘arc models’ in which a “pre-heated” crust is thickened. In this investigation, I focus on pre-heating as a result of arc activity, as the majority of basalts and TTGs that were generated during LAME in the Superior Province have been interpreted as arc magmatism (Percival et al. 2006). However, such preheating could also have been the result of plume activity, as modelled in Chapter 6.

Table 1 summarizes the main features that distinguish the eight models that make up this investigation.

2.1. Crustal thickness

We begin with an initial crust of 35 or 40 km thickness in steady state. In the majority of my models the initial crust is subjected to heating through increased mantle heat flow over a 40 My period, prior to thickening. This is meant to model crustal heating during arc activity, with the 40 My time span approximately that over which primary magmatic activity (involving basalts and TTGs) occurred during LAME. The crust is then thickened in an ‘instantaneous’ and homogeneous fashion to 55 km.

In their thermal modelling of the Superior crust, Mareschal and Jaupart (2006) assume crustal thickening to 50 km. I suggest that a somewhat greater thickness is justified on the following grounds.

The present thickness of the Superior crust is ~ 40 km (Jaupart and Mareschal 1999; Perry et al. 2002, 2006). Metamorphic pressures estimates at present day erosion levels indicate erosion of 8 – 18 km (Easton 2000; Stone 2000; Bédard 2003; Valli et al. 2004; Hynes and Song 2006; Percival and Pysklywec 2007). Hence, at the time that peak metamorphic conditions were attained, crustal thicknesses would have ranged from 48 – 58 km.

However, England and Thompson (1984) have shown that maximum metamorphic temperatures and maximum pressures may not be synchronous. Estimates of the latter should be regarded as a minimum bound on actual pressures reached.
Furthermore, there is evidence for mid-crustal orogen-parallel (east-west) ductile extension in the Wawa Subprovince from 2660 – 2640 Ma, that resulted from both pure shear (flattening) and simple shear (Moser et al. 1996). This was accompanied by deep crustal granulite-facies metamorphism. Such metamorphism appears to have occurred across the Superior Province at this time (Percival 1990; Krogh 1993), which suggests that the same mechanism of ductile extension may have operated elsewhere.

Cruden (pers. comm.) suggests that the mid-crustal extensional fabric in the Wawa is the result of syn-convergent, orogen-parallel ductile flow (Cruden et al. 2006). This is consistent with Krogh’s (1993) tectonic ductile underplating model. Regardless of the cause, however, the findings of Moser et al. (1996) indicate that erosion was not the only mechanism involved in thinning of the crust, again supporting the view that surface pressure estimates may not be indicative of maximum thickening. Therefore, the crust may in fact have been thickened by significantly more than 8 – 18 km during LAME. It should be noted, however, that such thickening was far from uniform, with depth of burial < 8 km indicated within some regions of the Superior Province (Easton 2000).

The majority of my models incorporate erosion, either in constant or exponentially decreasing form. The models are run for a 100 My time span following crustal thickening.

2.2. **Crustal heat production**

I assign a vertical distribution of heat producing elements so that following thickening the heat production profile of the lowermost 40 km of the thickened crust matches that inferred for the present day (~ 40-km-thick) Western Abitibi crust extrapolated to 2.7 Ga. Mareschal and Jaupart (2006) estimate present-day heat production in the upper 20 km of the Western Abitibi crust to be 1.2 µWm$^{-3}$ and in the lowermost 20 km to be 0.4 µWm$^{-3}$. Assuming that heat production at 2.7 Ga was approximately twice present-day, corresponding values of heat production in my models are 2.4 µWm$^{-3}$ and 0.8 µWm$^{-3}$. I chose this distribution because it is simple and because the average crustal heat production of 1.6 µWm$^{-3}$ is only slightly higher than the range of 1.3 – 1.5 µWm$^{-3}$ that I have estimated for LAME crust (Chapter 3). I assign a heat production of 0 µWm$^{-3}$ to the now eroded cap that overlay this 40 km thickness on the assumption that the majority of such rocks were mafic. Heat production in the lithospheric mantle is assumed to be negligible (0 µWm$^{-3}$).
2.3. Mantle heat flow

I assume ambient mantle heat flow values of 12 – 15 mWm$^{-2}$, which are within the range of estimates for current mantle heat flow (11 – 15 mWm$^{-2}$) throughout the Canadian Shield (Jaupart and Mareschal, 1999). Contrary to popular usage, in the discussion of my models “mantle heat flow” refers to heat flow into the base of the lithosphere and not necessarily into the base of the crust. During periods of changing crustal temperatures heat flow into the base of the crust can differ significantly from that into the base of the lithosphere.

A constraint on heat flow during the late stages of LAME (following crustal thickening) is the prevailing view that much of the thick cratonic root of the Superior Province (> 200 km; Jaupart and Mareschal 1999) was in place by the end of LAME. Cool mantle conditions during the late stages of LAME are also suggested by diamonds that occur in ~ 2.67 Ga heterolithic breccias and lamprophyre dikes in the Wawa and Abitibi subprovinces (Wyman et al. 2006). In my models I choose mantle heat flow values such that the base of the thermal boundary layer is in the diamond stability field both prior to arc activity and following crustal thickening, though the former is not strictly necessary.

2.4. Thickness of granitic melt produced

To determine the thickness of granitic melt produced in my models, I apply the melting curve for tonalite that I derived in Chapter 3, based on dehydration melting experiments of Rutter and Wyllie (1988). (The main hydrous minerals present in the source tonalite were biotite and amphibole).

The melting curve is given by:

$$m_{\text{tonalite}} = 0.2282 T - 188.68,$$

where $m_{\text{tonalite}}$ denotes volume % melt and $T$ denotes temperature in °C. This results in a solidus and liquidus of 827°C and 1265°C, respectively. I assumed a maximum degree of melting of 35 vol% (which occurs at 980°C) in the models as melt is likely to segregate from the source at melt fractions greater than this leaving an anhydrous restite that is unlikely to produce further melt unless significantly higher temperatures are achieved. The approach that I have used to determine the amount of granitic melt produced is conservative in that melting temperatures for sources of granites other than tonalite, such as pelites, are lower than those for tonalite (Clemens 2006). As well, there is evidence that the segregation of granitic melt from source rocks may be incomplete (Beakhouse 2007) suggesting that granitic “melt” fractions may be much greater than 35%.
3. **Non-arc Models: Models 1 and 2**

In my “non-arc” models no period of crustal heating via increased mantle heat flow (reflecting arc activity) precedes crustal thickening. These models can be viewed as classical crustal thickening models.

3.1. **Model 1: homogeneous crustal thickening from 35 to 55 km**

In Model 1, a 35-km-thick crust is homogeneously thickened to 55 km. Mantle heat flow is 15 mWm$^{-2}$. I have not incorporated erosion. The initial Moho temperature is 609°C and increases to 921°C after 100 My (Fig. 3). Granites appear 54 My after thickening (Fig. 4). The total thickness of granitic melt produced after 100 My is ~ 2.7 km.

3.2. **Model 2: homogeneous crustal thickening from 40 to 55 km**

In Model 2, a 40-km-thick crust is homogeneously thickened to 55 km. Mantle heat flow is 13 mWm$^{-2}$. I have not incorporated erosion. The initial Moho temperature is 736°C and increases to 972°C after 100 My (Fig. 3). Granites appear 26 My after thickening (Fig. 4). The total thickness of granitic melt produced after 100 My is ~ 4.8 km.

3.3. **Evaluation of the Non-arc models**

In Models 1 and 2, granitic melt is produced 54 My and 26 My, respectively, after crustal thickening. Given the lack of a significant gap between initial granite production and orogenic deformation in some areas of the Superior Province, a 54 My gap is too long and Model 1 can be ruled out. The 26 My gap in Model 2 is also somewhat long and can be expected to be even longer if erosion is incorporated into the model. Furthermore, there is a more subtle problem associated with both of these models and that is that initial melt is generated at 55 km depth. (If a higher level granite source is assumed then initial granites would be produced even later). Granite geochemistry suggests a shallower source for the granites (located at less than ~ 30 – 35 km depth) (Sylvester 1994). My arc models do not completely overcome this problem but melting does take place at a shallower depth in those models than in Models 1 and 2 as the arc models include a 15-km-thick layer of “anhydrous” lowermost crust, the result of crustal heating during arc activity. No granites are presumed to be generated within this layer.
4. Arc Models without Underthrusting: Models 3, 4, 5 and 6

4.1. Crustal heating

In my arc models a 40-My period of crustal heating precedes crustal thickening. Such heating could also have been the result of plume activity, as in the Abitibi subprovince (Chapter 6). There are various ways in which such heating can be implemented. In Chapter 6, crustal heating resulted from the emplacement of mantle plumes, mafic sills and greenstone covers.

Here, I heat the crust by instantaneously and homogeneously thinning the lithospheric mantle portion only, which results in increased mantle heat flow. The lithospheric mantle remains thinned for 40 My. To decrease mantle heat flow at the end of arc activity I reverse the process, namely I instantaneously and homogeneously thicken the lithospheric mantle.

This device for heating the crust is admittedly artificial, and as previously mentioned, is just one of several others that can be employed to achieve crustal heating. One reason I did not use the device of imposing a heat flow of choice at the base of the crust during arc activity and following crustal thickening is that I wished to continue to model the entire lithosphere and not simply the crust. As well, it would represent less realistic conditions as during a period in which crustal temperatures are changing heat flow from the mantle into the base of the crust does not remain constant.

I do not mean to suggest that all crustal heating during arc activity was the result of thinning of the lithospheric mantle only. Magmatic intraplate and underplating as well as in situ heating from magmatic thickening would have contributed to such heating. My approach is simply a convenient means of forcing significant heating of the crust to occur during “arc activity” in my models.

4.2. Anhydrous lower crust

I force a 15-km-thick layer of “anhydrous” lowermost crust to be present in the thickened crust by heating the crust sufficiently during the 40 My prior to thickening that immediately after thickening (to 55 km) the temperature at 40-km depth is ~ 827°C (the tonalite solidus associated with my melting curve, $m_{\text{tonalite}}$). Again, such an anhydrous layer may also have been the result of lower crustal melting during plume activity, as in Chapter 6. I associate supersolidus temperatures, at crustal depths greater than 40 km in the thickened crust, to dehydration of this lowermost 15 km of crust during arc activity, rendering it relatively infertile. Hence, in determining the amount of granitic melt produced in my models I ignore the lowermost 15 km of crust. I recognize that the choice of a single temperature to separate fertile
from infertile crust is highly artificial but I do it to avoid the complexities of having to implement varying degrees of fertility of the crust in my models. While my implementation of a lower crustal anhydrous layer may be artificial the idea behind it is not. Some mechanism must be invoked to explain why a thickened and heated Superior crust that produced granitic magmas appears to have only produced it from depths less than ~ 30 – 35 km.

The above approach forces Moho temperatures at the time of thickening to be between 1034 and 1083°C in the various arc models. These are consistent with petrological data from modern arc rocks as well as more recent thermal models of subduction zones (Kelemen et al. 2003), as discussed in section 7.2.

As the base of the “fertile” granite-producing layer is already at the solidus at the time of crustal thickening granite production in the arc models begins immediately upon thickening.

4.3. Models without erosion

4.3.1. Model 3: homogeneous crustal thickening from 35 to 55 km; no erosion

In Model 3, an initial 35-km-thick crust in steady state with mantle heat flow 15 mWm$^{-2}$ is subjected to a 40-My period of heating by homogeneously thinning the lithospheric mantle only to ~ 21% of its original thickness. The degree of thinning of the lithospheric mantle is dictated in this, and all other models, by the requirement that following instantaneous thickening the temperature in the lowermost 15 km of crust should lie above the tonalite solidus. During the period of arc activity the heat flow into the base of the crust varies between 41 and 45 mWm$^{-2}$ (Fig. 5).

At the end of this period of heating the crust is instantaneously and homogeneously thickened to 55 km and the lithospheric mantle is homogeneously thickened so that mantle heat flow is ~ 13 mWm$^{-2}$ after thickening.

Fig. 6 shows the transient geotherms for Model 3. The Moho temperature immediately after thickening is 1083°C and (because I have not incorporated erosion into this model) increases to 1122°C after 100 My (Fig. 3). The dip in Moho temperature immediately following crustal thickening (Fig. 3) results from the instantaneous manner in which heat flow into the base of the crust is decreased at the time of thickening. Such a decrease in temperature is limited to the bottom 7 km of the crust.

Granites appear immediately after thickening (Fig. 4). The total thickness of granitic melt produced after 100 My (in the fertile layer 15 km or more above the Moho) is ~ 3.0 km. The
maximum temperature reached at the base of the fertile layer (at 15 km above the Moho) is 1018°C at 100 My (Fig. 7).

4.3.2. Model 4: homogeneous crustal thickening from 40 to 55 km; no erosion

In Model 4, an initial 40-km-thick crust in steady state with mantle heat flow 13 mWm\(^{-2}\) is heated for 40 My by homogeneously thinning the lithospheric mantle only to \(\sim 26\%\) of its original thickness. The crust is then instantaneously and homogeneously thickened to 55 km and the lithospheric mantle is homogeneously thickened so that mantle heat flow is \(\sim 12\) mWm\(^{-2}\) after thickening. The Moho temperature immediately after thickening is 1039°C and (because I have not incorporated erosion into this model) increases to 1093°C after 100 My (Fig. 3). The dip in Moho temperature immediately following crustal thickening (Fig. 3) is limited to the bottom 6 km of the crust. Granites appear immediately after thickening (Fig. 4). The total thickness of granitic melt produced (in the fertile layer) after 100 My is \(\sim 2.6\) km. The maximum temperature reached at the base of the fertile layer (at 15 km above the Moho) is 997°C at 100 My (Fig. 7).

4.4. Models with erosion

Based on burial pressures from fluid inclusion studies from various Archean cratons (including the Superior Province) Galer and Mezger (1998) estimate that since the time of stabilization Archean cratons have undergone approximately 5 ± 2 km of erosion. Indications are that the Superior crust stabilized at \(\sim 2.6\) Ga; given a present day thickness of \(\sim 40\) km Galer and Mezger’s (1998) estimate suggests that the crust was \(\sim 45\)-km-thick at that time. If I assign an approximate age of 2.7 Ga for orogenic thickening of the Superior crust, then a 55 km thickened crust (as in my models) should experience 10 km of erosion over 100 My. I employ erosion rates that meet this constraint.

4.4.1. Model 5: homogeneous crustal thickening from 35 to 55 km; constant erosion

In Model 5, an initial 35-km-thick crust in steady state with mantle heat flow 15 mWm\(^{-2}\) is heated for 40 My by homogeneously thinning the lithospheric mantle only to \(\sim 21\%\) of its original thickness. The crust is then instantaneously and homogeneously thickened to 55 km and the lithospheric mantle is homogeneously thickened so that mantle heat flow is \(\sim 13\) mWm\(^{-2}\) after thickening. Constant erosion of 0.1 mm/yr begins immediately upon thickening. The Moho temperature immediately after thickening is 1083°C and decreases to 1038°C after 100 My (Fig. 3). The dip in Moho temperature immediately following crustal thickening (Fig. 3) is limited to the bottom 7 km of the crust. Granites appear immediately after thickening (Fig. 4) with granite
production ending after 39 My. The total thickness of granitic melt produced (in the fertile layer) is \(\sim 1.3 \text{ km}\). The maximum temperature reached at the base of the fertile layer (at 15 km above the Moho) is 947°C at 40 My (Fig. 7).

4.4.2. Model 6: homogeneous crustal thickening from 35 to 55 km; exponentially decreasing erosion

Model 6 is set up exactly like Model 5 above, except that the erosion rate is exponentially decreasing rather than constant. I assumed an erosion rate of \(c_0 \exp(-c_1t)\) mm/yr where \(c_0 = 0.2157\), \(c_1 = 1.801 \times 10^{-8}\) and \(t\) is measured in years. Crustal thickness decreases more rapidly in this model than in Model 5, with erosion of 5 km of crust 30 My after thickening, contrasted with a similar amount of erosion 50 My after thickening in Model 5. The Moho temperature immediately after thickening is 1083°C and decreases to 1012°C after 100 My (Fig. 3). The dip in Moho temperature immediately following crustal thickening (Fig. 3) is limited to the bottom 7 km of the crust. Granites appear immediately after thickening (Fig. 4) with granite production ending after 30 My. The total thickness of granitic melt produced (in the fertile layer) is \(\sim 0.9 \text{ km}\). The maximum temperature reached at the base of the fertile layer (at 15 km above the Moho) is 929°C at 31 My (Fig. 7).

4.5. Evaluation of the Arc models (without Underthrusting)

In all the arc models granitic melt is produced early (immediately) after thickening. This is not a perfect fit to the timing of initial granitic magmatism across the Superior Province but it is a better fit than that in the classical crustal thickening models (54 and 26 My after thickening in Models 1 and 2, respectively). The thickness of granitic melt in the models that incorporate erosion (Models 5 and 6) is 1.3 and 0.9 km, respectively. This is consistent with Cruden’s (2006) estimate of \(\sim 1 \text{ km}\) regional thickness of late granitic-granodioritic melt in the Superior Province. As well, in Models 5 and 6, the timespan of granitic activity (39 and 30 My, respectively) is similar to the span over which the majority of granites were produced in the Superior Province (\(\sim 40 \text{ My}\)). I next examine the effect of underthrusting on prolonging granitic magmatism.

5. Arc Model with Underthrusting: Model 7

Krogh’s (1993) work in the Kapuskasing suggests the workings of a mechanism that was carrying higher level rocks into the deep crust and was continuing to operate well after the main phase of late granitic magmatism ended at \(\sim 2655 \text{ Ma}\). I implement this “tectonic ductile
underplating” somewhat crudely in my model as a brittle underthrust. I investigate if such a mechanism operating after the end of the main period of granitic magmatism can result in renewed granitic activity.

5.1. Model 7: homogeneous crustal thickening from 35 to 55 km; exponentially decreasing erosion; late underthrusting

Ideally a 2-D model should be used to model late underthrusting (namely underthrusting occurring after the main phase of granitic magmatism). I have modelled it using a 1-D model (Model 7) and leave 2-D modelling of it for future work. I use Model 6, involving homogeneous thickening and exponentially decreasing erosion, as a base model. In Model 6, granitic activity ends 30 My after thickening, followed by crustal cooling.

I model the occurrence of underthrusting 10 My after the end of granitic activity as follows. During the crustal heating that occurs following crustal thickening in Model 6, temperatures in the lowermost ~ 22.5 km of crust reached or exceeded the tonalite solidus rendering this portion of the crust infertile (at least according to my convention). I take the lowermost 23.5 km of crust, hence the depleted layer along with an overlying 1 km of fertile crust, and move it downward 10 km relative to adjoining crust. I invoke Krogh’s (1993) late ductile underplating model to justify this.

I demonstrate that associated crustal thickening is not required for renewed granitic activity by simultaneously (homogeneously and instantaneously) thinning the resulting crust by 10 km so that crustal thickness remains unchanged. Such a demonstration is necessary since evidence for late thickening of the Superior crust (namely thickening after the main period of granitic magmatism) is lacking. However, such thinning is also consistent with late extension documented in the KSZ (Krogh 1993) and in the Wawa Gneiss Domain (WGD) (Moser et al. 1996). The WGD is an amphibolite facies region exposed in the Kapuskasing uplift that is at a structurally higher level than the KSZ. As in the KSZ kinematic indicators show that vertical shortening was achieved primarily through E-W, orogen parallel elongation (Moser et al. 1996). Extension in the WGD is estimated to have occurred between 2660 and 2645 Ma (Moser et al. 1996) hence was less protracted than in the KSZ.

Fig. 8 shows the transient geotherms for Model 7 at the time of underthrusting and 10 My afterwards. The Moho temperature 100 My after initial thickening is similar to that in Model 6, 1014°C (compared to 1012°C in Model 6). An additional 121 m of granitic melt is produced within 2 My after underthrusting (Fig. 4). While this is not a significant thickness of melt Model 7 does demonstrate the ability of the crust to continue to produce granites through tectonic
Chapter 7: The Origin of Late Granites in the Superior Province

means, even during a cooling period. This suggests a causal relationship may exist between Krogh’s (1993) ductile underplating model and the production of the latest of the late-stage granites.

6. Arc Model without crustal thickening: Model 8

For completeness I run a final model, Model 8 (based on Model 3) in which there is increased mantle heat flow during arc activity but no associated crustal thickening. In Model 8, an initial 35-km-thick crust in steady state with mantle heat flow 15 mWm\(^{-2}\) is subjected to a 40 My period of heating by homogeneously thinning the lithospheric mantle only to ~ 21% of its original thickness. At the end of this period of heating the thickness of the crust remains unchanged but the lithospheric mantle is homogeneously thickened so that mantle heat flow is ~ 13 mWm\(^{-2}\).

Moho temperatures for this model are shown in Fig. 3 and show a rapid decline after the end of arc activity. At the end of arc activity the Moho temperature is 1083°C. 100 My after the end of arc activity the Moho temperature has decreased to 751°C. This model does not yield any granites and demonstrates what may be an important role for crustal thickening during LAME, and that is in sustaining crustal temperatures well after arc activity ends, allowing production of the later late-stage granites to take place.

7. Discussion

7.1. Arc Models and the Late Granites

My modelling suggests that the heat source for the late granites may have been remanent heat from ~ 40 My of arc activity (or plume activity, as in the Abitibi subprovince) coupled with in situ heating from orogenic thickening. The results of Models 5 and 6 (arc models with erosion) are not an unreasonable fit to the late granites in the Superior Province, in terms of timing and thickness of melt produced, though initial granites are perhaps produced too early. Model 7 demonstrates how granitic activity may have been prolonged by underthrusting occurring after main granitic activity had ended. If lithospheric mantle conditions became cool enough to support diamond formation at the end of arc activity then crustal thickening would have been both necessary (Model 8) and sufficient to sustain high crustal temperatures over the entire period of granite formation.

An interesting pattern emerges from the ‘arc models’ collectively and one that is quite different from that of many models that have been proposed to account for the late granites.
Most such models invoke a separate crustal heating event immediately prior to the appearance of the granites, such as lithospheric mantle overturn (Percival and Pysklywec 2007) or in the Slave Province, lithospheric delamination (Davis et al. 1994).

My arc models suggest that the main crustal heating event associated with LAME was arc (or plume) activity (Fig. 2). Fertile lower crust susceptible to melting at the high temperatures reached during arc activity would have melted, leaving anhydrous restite that would have required even higher temperatures to produce melt. And in a crust that was already hot, but in which lower crustal fertile source rocks had already been exhausted of melt during arc activity, the existence of higher level fertile crust along with the movement of fertile mid- and upper crust downwards into not simply hot, but very hot regions may have played a critical role in the generation of the granites.

7.2. Crustal heating during arc activity

A fundamental point that remains to be addressed is if the temperatures in my arc models are within reasonable bounds or are too high.

Maximum Moho temperatures reached in my models during arc activity range from 1034 – 1083°C. Based on P-T estimates of arc mantle-melt equilibration and metamorphism of lower crustal arc rocks Kelemen et al. (2003) estimate that temperatures in modern arcs are > 1000°C at ~ 30 km depth and approach 1300°C at depths > 45 km. Tatsumi (2003) suggests that the high temperatures (~ 1300°C) found beneath arc crust/mantle boundaries are likely localized though temperatures > 1350°C in the mantle wedge are still required for the generation of primitive arc magmas.

In the majority of thermal models of subduction zones mantle wedge temperatures are several hundred degrees too low to give rise to primitive arc magmas (Peacock 1990; Davies and Stevenson 1992; Iwamori 1998). Van Keken et al. (2002), however, used a stress and temperature-dependent mantle viscosity in their subduction model which predicts temperatures of ~ 950 – 1000°C at 45 km depth beneath NE Japan and the Cascades; the 1300°C isotherm occurs at 65 km depth. It is not unreasonable to assume that temperatures reached during arc activity in the Neoarchean may well have been higher. Hence crustal temperatures attained in my models during arc activity appear to be consistent with petrological data from modern arc rocks as well as more recent thermal models of subduction zones.

I also determine if the temperatures reached in my arc models following crustal thickening are within reasonable bounds of those recorded in the Superior crust. To do this I compare peak
metamorphic temperatures and pressures attained in Model 6, my arc model with exponentially decreasing erosion, with those of high-grade rocks of the Superior crust.

Starting with a variety of initial crustal depths in Model 6, I determined the maximum temperature reached by a rock from each such level and the corresponding depth at which the maximum temperature was attained (Fig. 9). As Fig. 9 shows, the peak metamorphic temperatures and (corresponding) pressures determined for Model 6 are actually intermediate to those recorded in the Superior. My model cannot explain the highest of the low P-high T metamorphic temperatures found in the Superior, however no upward transfer of melt has been implemented in my models. This would tend to increase temperatures attained in the mid- to upper crust. However, my main objective in doing this analysis was to determine if the assumption of significant crustal heating during arc activity would result in crustal temperatures in my models that were too high and Fig. 9 suggests that it does not.

7.3. Wawa diamonds and late cool lithospheric mantle

The overall lack of constraints on mantle heat flow during the Archean remains the main stumbling block in devising representative thermal models of Archean lithosphere. Unfortunately there are no good constraints on mantle heat flow prior to LAME and the extent to which such heat flow might have increased during the main period of arc activity.

A constraint on heat flow during the late stages of LAME is the prevailing view that much of the thick cratonic root of the Superior Province (> 200 km; Jaupart and Mareschal 1999) was in place by the end of LAME. Mantle potential temperatures during the Late Archean are estimated to have been ~ 100 – 200°C hotter than today (Abbott et al. 1994). Figs. 6 and 8 show that high lower crustal temperatures during the late stages of LAME are not inconsistent with a thick thermal boundary layer if the lithospheric mantle is thickened following arc activity, possibly as a result of collision or (the model that I favour) accretion of the subducting plate (Bostock 1999). Such cool mantle conditions during the late stages of LAME are also suggested by diamonds that occur in ~ 2.67 Ga heterolithic breccias and lamprophyre dikes in the Wawa and Abitibi subprovinces (Wyman et al. 2006). Note that the base of the thermal boundary layer in the arc models (Models 3 –8) is in the diamond stability field (prior to and) following arc activity.

7.4. Implications of the Arc models

In my model for the late granites both arc activity and horizontal tectonics (resulting in crustal thickening and prolonged tectonic underplating) play a primary role. Because both are
associated with modern plate tectonics the first appearance of late granites in the Archean may be an indication of the onset of a modern plate tectonic regime. Interestingly, most such granites appeared after ~ 3.1 Ga (Sylvester 1994) which coincides with the minimum age for the onset of modern style subduction proposed by Smithies et al. (2007), based on evidence that the 3.12 Ga Whundo Group, Pilbara Craton, was a modern style oceanic arc.

Archean granites share features with more recent granites such as those generated in the Pan-African orogeny (~ 600 Ma), the British Caledonides and the Lachlan Fold Belt (~ 400 Ma) and the Hercynides of Europe (~ 300 Ma) (Sylvester 1994). My model (or at least the two main ideas on which it is based) may have possible application outside of the Archean.

7.5. An aside regarding choice of melting curve

Finally, I would like to point out an observation that emerged from this study that might be useful to thermal modellers interested in using the same approach that I have to determine the thickness of melt produced in a model, namely application of a melting curve. My choice of melting curve was based on the results of partial melting experiments on a tonalite, one of the inferred sources for the late granites. In hindsight, however, I believe that greater consideration should be given to the choice of melting curve as follows.

Suppose that rather than the late granites my focus was on reproducing thermal conditions in the KSZ, particularly those recorded in migmatitic, granulite-grade mafic gneisses. These make up a significant part of the KSZ and contain up to 25 vol% trondhjemitic leucosomes (Hartel and Pattison 1996). Pressure-temperature estimates for these gneisses are 9 kbar and 685 – 735°C, however, based on dehydration melting experiments on amphibolites, higher pressures and temperatures (11 kbar and 850°C) are favoured by Hartel and Pattison (1996). In fact there is a significant range in melting temperatures for amphibolites determined by the various experimental studies.

Had I constructed a thermal model for the KSZ and chosen a melting curve for amphibolite for which solidus temperatures were significantly greater than 735°C (such as one based on an experimental study of Rapp et al. [1991] in which ~ 13 vol% melt is produced at 1000°C) then regardless of the manner in which my model was constructed, it could not be a successful one. The choice of melting curve would simply not allow migmatites to be produced at temperatures similar to those recorded in the KSZ. Arguments could be offered of the type that “peak” temperatures may in fact be “cooling” temperatures but given that prior metamorphic history is generally not preserved in granulites it is difficult to quantify the extent to which the two
temperatures may differ. Because metamorphic temperatures are an important constraint on a thermal model I suggest that melting conditions recorded in actual rocks, if available, should be carefully considered when choosing a melting curve.
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References


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### Table 1. Summary of the main features of my models.

<table>
<thead>
<tr>
<th>Model</th>
<th>Lithosphere</th>
<th>Thickening</th>
<th>Erosion</th>
<th>‘Underthrusting’</th>
<th>Dimension</th>
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<td>1</td>
<td>‘non arc’ (cool)</td>
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<td>no</td>
<td>no</td>
<td>1-D</td>
</tr>
<tr>
<td>2</td>
<td>‘non arc’ (cool)</td>
<td>40 to 55 km</td>
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<td>1-D</td>
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<tr>
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<td>no</td>
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<tr>
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<td>no</td>
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<tr>
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<td>‘arc’ (hot)</td>
<td>No</td>
<td>No</td>
<td>No</td>
<td>1-D</td>
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Fig. 1. U-Pb crystallization and estimate of crystallization ages for pegmatites and late granites of the Superior Province. Data from Skulski and Villeneuve (1999).
Fig. 2. Timing of major phases of magmatic activity and deformation and sedimentation during the assembly of the western Superior Province in the Neoarchean. My focus is on the syn- to post-orogenic granites. BL–RL–BI: Bee Lake – Rice Lake – Black Island; v: volcanic rocks; p: plutonic rocks. (Modified from Percival et al., 2006).
Fig. 3. Evolution of Moho temperatures following crustal thickening in the non-arc models (Models 1 and 2) and arc models without underthrusting (Models 3 – 6). Models 5 and 6 incorporate constant and exponentially decreasing erosion, respectively. Also shown is the evolution of Moho temperatures after arc activity ends in Model 8, in which no crustal thickening is implemented.
Fig. 4. Thickness of granitic melt produced following thickening in all the models (except Model 8).
Fig. 5. Heat flow into the base of the crust during arc activity in all arc models with a 35-km-thick initial crust (Models 3, 5, 6, 7 and 8).
Fig. 6. Transient geotherms for Model 3. The red geotherm represents steady-state temperatures prior to arc activity under ambient mantle heat flow conditions (15 mW/m²). The grey geotherms represent temperatures during arc activity, spanning 40 My, at 10 My intervals. Significant heating of the crust during arc activity is implemented by thinning the mantle lithosphere to ~ 21% of its original thickness. We point out that this is simply a convenient means of implementing crustal heating during arc activity. In actuality such heating would have been accomplished through a combination of thinning of the mantle lithosphere, magmatic intra- and underplating and in situ heating from magmatic thickening. The black geotherms represent temperatures after crustal thickening, for a span of 100 My, at 10 My intervals. During this period mantle heat flow is ~ 13 mW/m². Green lines delineate adiabatic temperatures in the mantle (assuming a gradient of 0.6 °C/km) for mantle potential temperatures ranging from 1200 – 1500 °C. The graphite-diamond transition of Kennedy and Kennedy (1976) is also shown.
Fig. 7. Evolution of temperatures at the base of the fertile layer (15 km above Moho) in the arc models without underthrusting (Models 3 – 6). Models 5 and 6 incorporate constant and exponentially decreasing erosion, respectively.
Initial granitic magma production ceases 30 My after initial thickening. 10 My after this we introduce underthrusting of predominantly anhydrous lower crust that contains a topmost 1-km-thick layer of fertile crust. This results in renewed, minor granitic activity. Green lines delineate adiabatic temperatures in the mantle (assuming a gradient of 0.6 °C/km) for mantle potential temperatures ranging from 1200 – 1600 °C. The graphite-diamond transition of Kennedy and Kennedy (1976) is also shown.
Fig. 9. Filled diamonds indicate peak metamorphic temperatures reached by rocks at varying initial depths in Model 6 (arc model with exponentially decreasing erosion) and the corresponding depths at which these temperatures were reached. Black rectangles show estimated metamorphic conditions for Archean granulites. The latter portion of the figure is modified from Percival (1994). A: Ashuanipi (Percival 1991a, 1991b); ER: English River (Perkins and Chipera 1985; Chipera and Perkins 1988); K: Kapuskasing uplift (Mäder et al. 1994); M: Minto block (Bégin and Pattison 1994); MRV: Minnesota River Valley (Perkins and Chipera 1985; Moecher et al. 1986); P: Pikwitonei (Mezger et al. 1990); WR: Winnipeg River (Beakhouse 1991).
Chapter 8

Conclusions

1. Summary

I set out to investigate the origin of TTG in the Abitibi subprovince and the origin of late granites in the Superior Province during LAME. The primary investigative tool used was 1-D thermal modeling. I have shown:

1. Incorporating the upward transfer of melt in a thermal model of lower crustal melting of a LAME-type crust in which the base of the crust is subjected to high mantle heat flow (as might be expected during arc or plume activity) can significantly enhance the thickness of melt produced.

2. A thickness of TTG comparable to that occurring in the predominantly juvenile Abitibi subprovince can be produced through a series of plume pulses with timing comparable to that of plume activity in the Abitibi. However, the timing of TTG melt produced in my models is not fully consistent with that of TTG in the Abitibi. Furthermore, my models cannot account for the hydration of the TTG source.

3. A crust that is pre-heated through earlier arc activity can produce granitic melt upon thickening that is comparable in initial timing and estimated regional thickness to late granitic melt produced in the Superior Province during LAME. Late underthrusting can prolong the time span of such granitic activity.

2. Discussion and future work

There are discrepancies between the two models outlined in 2 and 3 above and geological data for the Superior Province so the first, obvious area of future work would be to address such discrepancies:

2.1. Plume models

In Chapter 6, I pointed out three main problems with my plume with restite delamination model for the Abitibi.
Chapter 8: Conclusions

One problem was that the final configuration of my model crust contained initial crust between 15 and 30 km depth (see Fig. 14 in Chapter 6). In the Abitibi, this region is occupied by TTG. Hence an improvement over my implementation would be to incorporate the upward transfer of TTG melt from the source region into this region, essentially flushing out (or down) all initial crust.

A second problem was that my model could not account for the hydration of lower crustal source rocks, which appears to be required to produce melt of TTG composition. This is not really a problem if hydrated surficial volcanics and volcanogenic sediments sink down into the mid- to lower crust through a process of vertical tectonics, such as modelled in 2-D by Mareschal and West (1980) and Robin and Bailey (2008). Hence an interesting extension of my model would be to introduce plumes and restite delamination into the above models.

A third problem is that the timing of TTG melt generation in the Abitibi differs from that in my models. It suggests that TTG in the Abitibi was not purely the result of plume activity. Of the two competing models for the production of TTG in the Archean, melting of a subducting slab vs. lower crustal melting, the former is favoured by some for the Abitibi because it provides a means of introducing and disposing of large quantities of a suitable source (hydrated oceanic crust) (Beakhouse 2007) and because the geochemistry of some Abitibi TTG suggests low degrees of melting (Benn and Moyen 2008).

Benn and Moyen (2008) suggest that the interstratification of plume-type and subduction-type lavas in the Abitibi, and the concomitant emplacement of TTG plutons with a possible slab melt origin, might be explained by the initiation of subduction beneath an oceanic plateau that is experiencing plume activity, followed by the opening of a slab window in the subducting slab. Ridge subduction may have provided such a window. However, Arndt (2008) suggests that a major plume that encountered a subducting plate on its ascent would “snuff out the subduction”. He does not specify precisely how. One possibility is that it may punch its way through the subducting slab. Hence the slab window proposed by Benn and Moyen (2008) may not simply have been a fortuitous occurrence but may have been created by the Abitibi plume.

There are many interesting thermal models that can be devised to test various aspects of the model of Benn and Moyen (2008) but I suggest that an interesting initial one may be to establish a Neoarchean subduction zone (in a 2-D model) and allow a plume that is hot enough to produce komatiites to impinge on the base of the subducting slab. Two possible effects of the plume could be tested. The first is the likelihood that the plume could generate its own slab window through thermal and mechanical erosion of the subducting slab. The second is if the plume (if
sufficiently large) might possess the ability to push up on and flatten out the slab (at least locally), leading to flat subduction and slab melting (TTG) under conditions that a mantle wedge (and hence mantle contamination of the TTG) is minimized.

2.2. Late Granites

In my model for the late granites I have assumed a very simple post-collisional tectonic scenario in which a single crustal thickening event occurs. However, the record of post-collisional events throughout the Superior is far more complex. For example, in the English River subprovince, Nitescu et al. (2006) have made a case for crustal extension rather than crustal thickening as the cause of late granite production. This model should be investigated, preferably through 2-D modelling of possible slab break off or slab rollback occurring following collision.

2.3. Upward melt transfer

In Chapter 5, I derived a result that I suggested may be generally applicable, regarding the increase in melt that may be expected through implementation of upward melt transfer within a standard model of lower crustal melting. That result is, however, based on a single model, hence on a very specific set of thermal conditions. A case for the general applicability of the result can only be made by testing the result under a variety of thermal conditions. This is an obvious area of future work.
References


The following expression for thermal conductivity of the crust (based on Durham et al. 1987) appears in Mareschal and Jaupart (2006):

\[
K = 2.26 - \frac{618.241}{T} + K_0 \left( \frac{255.576}{T} - 0.30247 \right)
\]

(1)

where \( K \) is thermal conductivity (in Wm\(^{-1}\)K\(^{-1}\)), \( T \) is the absolute temperature and \( K_0 \) is the thermal conductivity at the surface (for \( T = 273 \) K). Mareschal and Jaupart (2006) use it in modelling crust of the Eastern Abitibi subprovince and Slave Province at 2.55 Ga. It contains an error, however, which we correct here, prior to using it in our modelling.

(1) is of the form

\[
K = a - \frac{b}{T} + K_0 \left( \frac{c}{T} - d \right)
\]

(2)

where

\[
a = 2.26, \quad b = 618.241, \quad c = 255.576, \quad d = 0.30247.
\]

Substituting \( T = 273 \) K in (2) yields:

\[
K = K_0 = a - \frac{b}{273} + K_0 \left( \frac{c}{273} - d \right).
\]

(4) can only hold for all values of \( K_0 \) if
\( a - \frac{b}{273} = 0 \) and \( \frac{c}{273} - d = 1 \). \hspace{1cm} (5)

However, values appearing in (3) yield:

\( a - \frac{b}{273} = -0.005 \) and \( \frac{c}{273} - d = 0.634 \).

In comparison to (5) the latter value of 0.634 is obviously in error suggesting that there is an error in \( c \) or \( d \) (or both).

We rederive the expression for thermal conductivity that appears in (1). We begin with the empirical formula for thermal diffusivity \( \alpha \) (in mm\(^2\)/s) determined by Durham et al. (1987):

\[
\alpha = 0.47 + \frac{78}{T} - 1.31Q + 1540 \frac{Q}{T} \tag{6}
\]

where \( Q \) is the volume fraction of quartz and \( T \) is absolute temperature. If we let \( \alpha_s \) denote the thermal diffusivity at the surface (for \( T = 273 \text{K} \)) then substitution into (6) yields:

\[
\alpha_s = 0.47 + \frac{78}{273} - 1.31Q + 1540 \frac{Q}{273} \tag{7}
\]

Letting

\[
a_s = 0.47 + \frac{78}{273}, \quad b_s = -1.31 + \frac{1540}{273} \tag{8}
\]

in (7) yields \( \alpha_s = a_s + b_sQ \). This implies:

\[
Q = \frac{\alpha_s - a_s}{b_s}, \tag{9}
\]

Substituting (9) into (6) yields:
\[
\alpha = 0.47 + \frac{78}{T} - 1.31 \left( \frac{\alpha_0 - a_0}{b_0} \right) + 1540 \left( \frac{1}{T} \right) \left( \frac{\alpha_0 - a_0}{b_0} \right) \Rightarrow
\]
\[
\alpha = 0.47 + \frac{78}{T} - 1.31 \frac{\alpha_0}{b_0} + 1.31 \frac{a_0}{b_0} + \frac{\alpha_0}{b_0} \frac{1540}{T} - \frac{a_0}{b_0} \frac{1540}{T} \Rightarrow
\]
\[
\alpha = 0.47 + 1.31 \frac{a_0}{b_0} + \frac{78}{T} \frac{a_0}{b_0} \frac{1540}{T} - \frac{1.31 \alpha_0}{b_0} + \frac{\alpha_0}{b_0} \frac{1540}{T} \Rightarrow
\]
\[
\alpha = \left( 0.47 + 1.31 \frac{a_0}{b_0} \right) - \left( 1540 \frac{a_0}{b_0} - 78 \right) \frac{1}{T} + \alpha_0 \left( \frac{1540 \frac{1}{T} - 1.31}{b_0} \right). \tag{10}
\]

Multiplication of (10) by 10^{-6} yields \( \alpha \) in \( \text{m}^2/\text{s} \). Further multiplication by \( \rho \) (density) and \( C \) (heat capacity) yields:

\[
K = \left( 0.47 + 1.31 \frac{a_0}{b_0} \right) \rho C 10^{-6} - \left( 1540 \frac{a_0}{b_0} - 78 \right) \rho C 10^{-6} \frac{1}{T} + K_0 \left( \frac{1540 \frac{1}{T} - 1.31}{b_0} \right). \tag{11}
\]

The last two terms are independent of the choice of \( \rho \) and \( C \). A choice of \( \rho \approx 3235 \text{ kg/m}^3 \) (actually \( \rho = 3234.495 \text{ kg/m}^3 \)) and \( C = 1000 \text{ JK}^{-1} \text{kg}^{-1} \) yields a first term that is the same as in Mareschal and Jaupart’s (2006) expression for thermal conductivity (1). Substitution of the above values for \( \rho \) and \( C \), and values from (8) for \( a_0 \) and \( b_0 \) into (11) yields the following expression for thermal conductivity:

\[
K = 2.26 - \frac{616.858}{T} + K_0 \left( \frac{355.574}{T} - 0.30247 \right) \tag{12}
\]

which primarily differs from (1) in the third term in which there is likely a typographical error in Mareschal and Jaupart (2006).

A choice of \( K_0 = 2.8 \text{ Wm}^{-1} \text{K}^{-1} \) as in Mareschal and Jaupart (2006) yields:

\[
K = 2.26 - \frac{616.858}{T} + 2.8 \left( \frac{355.574}{T} - 0.30247 \right). \tag{13}
\]
This is the thermal conductivity we use for the crust in our modelling. For all the theoretical considerations a better test of the appropriateness of any expression for thermal conductivity is how well it compares against experimental data. A graph of (13) appears in Fig. 1. The values fall within the experimentally determined range of thermal conductivities for crustal rocks (Vosteen and Schellschmidt 2003).
References


Fig. 1. Crustal thermal conductivity used in my modelling.
Appendix B

Comparison of Femlab 3.1 and Analytical Solution

The following 1-D heat flow problem appears in Carslaw and Jaeger (1965) (pg. 113) along with an analytic solution.

Problem 1: \( 0 < x < l \). Zero initial temperature. Constant flux \( F_0 \) into the region at \( x = l \). \( x = 0 \) kept at zero temperature.

Solution: \( T(x,t) = \frac{2F_0}{K} (\kappa t)^{1/2} \sum_{n=0}^{\infty} \left[ (-1)^n \left( \text{ierfc} \left( \frac{(2n+1)l-x}{2(\kappa t)^{1/2}} \right) - \text{ierfc} \left( \frac{(2n+1)l+x}{2(\kappa t)^{1/2}} \right) \right) \right] \) (1)

where \( T \) denotes temperature, \( t \) time, \( \kappa \) thermal diffusivity and \( K \) thermal conductivity.

As a crude test of the accuracy of the modelling solutions I have obtained using Femlab 3.1 I solve Problem 1 in Femlab 3.1 using parameters that are the same as or similar to those employed in my models (except for thermal conductivity which is temperature-dependent in my models) and compare the resulting geotherms against the analytic solution (1).

I first determine an approximation to the analytic solution. \( \text{ierfc}, \text{erfc} \) and \( \text{erf} \) (the error function) are defined as follows (Carslaw and Jaeger 1965, pg. 51):

\[
\text{ierfc}(u) = \int_u^\infty \text{erfc}(\xi) \, d\xi
\]

\[
\text{erfc}(u) = 1 - \text{erf}(u)
\]

\[
\text{erf}(u) = \frac{2}{\sqrt{\pi}} \int_0^u e^{-\xi^2} \, d\xi.
\]

In addition, the following identity holds (Carslaw and Jaeger 1965, pg. 484):

\[
\text{ierfc}(u) = \frac{1}{\sqrt{\pi}} e^{-u^2} - u \text{erfc}(u)
\]

(2)

Letting:

\[
c(t) = \frac{2F_0}{K} (\kappa t)^{1/2}, \quad s_i(x,t) = \frac{(2n+1)l-x}{2(\kappa t)^{1/2}}, \quad s_s(x,t) = \frac{(2n+1)l+x}{2(\kappa t)^{1/2}}
\]

in (1) yields:
Appendix B: Comparison of Femlab 3.1 and Analytical Solution

\[ T(x,t) = c(t) \sum_{n=0}^{\infty} (-1)^n \left\{ \text{erfc}(s_1(x,t)) - \text{erfc}(s_2(x,t)) \right\}. \] \hspace{1cm} (3)

Substituting for \text{erfc} from (2) into (3) yields:

\[ T(x,t) = c(t) \sum_{n=0}^{\infty} (-1)^n \left\{ \frac{1}{\sqrt{\pi}} e^{-s_1^2(x,t)} - s_1(x,t) \text{erfc}(s_1(x,t)) - \frac{1}{\sqrt{\pi}} e^{-s_2^2(x,t)} + s_2(x,t) \text{erfc}(s_2(x,t)) \right\}, \]

or

\[ T(x,t) = c(t) \times \sum_{n=0}^{\infty} (-1)^n \left\{ \frac{1}{\sqrt{\pi}} \left( e^{-s_1^2(x,t)} - e^{-s_2^2(x,t)} \right) - (s_1(x,t) \text{erfc}(s_1(x,t)) - s_2(x,t) \text{erfc}(s_2(x,t))) \right\}. \] \hspace{1cm} (4)

I compute an approximation to the analytic solution (1) in Matlab using (4) and the built-in function \text{erfc}. I choose parameters as follows:

- \( x \) represents crustal thickness of 35 km.
- \( F_0 \) represents mantle heat flow of 40 mWm\(^{-2}\).
- \( K \) represents thermal conductivity of 2.5 Wm\(^{-1}\)K\(^{-1}\).
- \( \kappa \) represents thermal diffusivity of \( 8.9286 \times 10^{-7} \) m\(^2\)/s (calculated using crustal density of 2800 kg/m\(^3\) and heat capacity of 1000 Jkg\(^{-1}\)K\(^{-1}\)).

For time \( t \) of 10, 50 and 100 My I obtain an approximation to the analytic solution (1) by summing the series in (4) from \( n = 0 \) to \( n = 2^{15} \).

I also solve Problem 1 in Femlab 3.1 using a maximum time step of 1 My and a maximum element (grid) size of 20 m.

The results are shown in Fig. 1, with temperatures obtained using Femlab 3.1 differing by less than 0.5 °C from those given by the approximation to the analytic solution (1) calculated in Matlab.
References

Fig. 1. Solutions to (heat flow) Problem 1 obtained at 10, 50 and 100 My using Femlab 3.1 and an approximation to the analytic solution (1). Parameter values used are listed in the text.
Appendix C

Increased Heat Production of Superior Rocks at 2.7 Ga

Using the measured K₂O, Th and U content of rock samples from the Kapuskasing Uplift and the Pikwitonei-Sachigo region I have calculated the heat production of these samples at the present time and at 2.7 Ga, along with the factor increase in heat production from the present to 2.7 Ga (Tables 1 – 3). The calculations were done using constants listed in Table 4.
References


Table 1. Heat production in rock samples from the Kapuskasing uplift. Data in columns A – F is from Ashwal et al. (1987). My calculations appear in columns G – I. The density of the samples is not given in Ashwal et al. (1987), hence a density of 2.7 g/cm³ was assumed.

<table>
<thead>
<tr>
<th>A Sample no.</th>
<th>B Description</th>
<th>C Th, ppm</th>
<th>D U, ppm</th>
<th>E K₂O, wt%</th>
<th>F Present Heat production, µWm⁻³</th>
<th>G Present heat production, µWm⁻³</th>
<th>H 2.7 Ga heat production, µWm⁻³</th>
<th>I Factor Increase, H ÷ G</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Leucogranite, Oswald L. Pluton, E of ILCZ</td>
<td>63.45</td>
<td>4.76</td>
<td>3.86</td>
<td>6.09 ± 0.14</td>
<td>6.28</td>
<td>10.28</td>
<td>1.64</td>
</tr>
<tr>
<td></td>
<td>average =</td>
<td>61.70</td>
<td>4.82</td>
<td>3.93</td>
<td>5.99 ± 0.15</td>
<td>6.19</td>
<td>10.22</td>
<td>1.65</td>
</tr>
<tr>
<td>2</td>
<td>Flaser tonalite, Abitibi B.</td>
<td>3.65</td>
<td>0.58</td>
<td>1.66</td>
<td>0.57 ± 0.06</td>
<td>0.66</td>
<td>1.73</td>
<td>2.62</td>
</tr>
<tr>
<td>3</td>
<td>Paragneiss, graphite-bearing</td>
<td>4.94</td>
<td>0.55</td>
<td>1.71</td>
<td>0.66 ± 0.05</td>
<td>0.75</td>
<td>1.85</td>
<td>2.47</td>
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<td>4</td>
<td>Mafic volcanic, Abitibi B., E of ILCZ</td>
<td>0.31</td>
<td>0.21</td>
<td>0.38</td>
<td>0.12 ± 0.04</td>
<td>0.13</td>
<td>0.40</td>
<td>2.94</td>
</tr>
<tr>
<td>5</td>
<td>Mafic gneiss</td>
<td>2.47</td>
<td>0</td>
<td>2.08</td>
<td>0.36 ± 0.06</td>
<td>0.49</td>
<td>1.61</td>
<td>3.29</td>
</tr>
<tr>
<td>6A</td>
<td>Mafic tonalite gneiss, garnet-rich</td>
<td>0.41</td>
<td>0.45</td>
<td>0.97</td>
<td>0.24 ± 0.05</td>
<td>0.29</td>
<td>0.94</td>
<td>3.17</td>
</tr>
<tr>
<td>6B</td>
<td>Mafic tonalite gneiss, garnet-free</td>
<td>5.27</td>
<td>0.48</td>
<td>0.99</td>
<td>0.59 ± 0.04</td>
<td>0.65</td>
<td>1.35</td>
<td>2.10</td>
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<td>Paragneiss</td>
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<td>0.99</td>
<td>1.74</td>
<td>1.02 ± 0.05</td>
<td>1.12</td>
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<td>2.14</td>
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<td>7A</td>
<td>Mafic gneiss</td>
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<td>0.20</td>
<td>0.23</td>
<td>0.11 ± 0.03</td>
<td>0.12</td>
<td>0.30</td>
<td>2.51</td>
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<td>8</td>
<td>Mafic gneiss with leucosome</td>
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<td>0.56</td>
<td>0.62</td>
<td>0.48 ± 0.04</td>
<td>0.52</td>
<td>1.04</td>
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<td>9</td>
<td>Banded gneiss</td>
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<td>Mafic granulite gneiss, garnet bearing</td>
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<td>0.10</td>
<td>0.24</td>
<td>2.37</td>
</tr>
<tr>
<td>A Sample no.</td>
<td>B Description</td>
<td>C Th, ppm</td>
<td>D U, ppm</td>
<td>E K₂O, wt%</td>
<td>F Present Heat production, µWm⁻³</td>
<td>G Present heat production, µWm⁻³</td>
<td>H 2.7 Ga heat production, µWm⁻³</td>
<td>I Factor Increase, H ÷ G</td>
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<td>13</td>
<td>Xenolithic tonalite</td>
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<td>0.23</td>
<td>1.00</td>
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<td>1.22</td>
<td>2.11</td>
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<td></td>
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<td>average =</td>
<td>5.20</td>
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<td>0.57</td>
<td>1.20</td>
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<td>14</td>
<td>Tonalite gneiss</td>
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<td>0.92</td>
<td>1.70 ± 0.07</td>
<td>1.73</td>
<td>2.56</td>
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<td></td>
<td></td>
<td>21.38</td>
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<td>0.80</td>
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<td>1.77</td>
<td>2.57</td>
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<td>average =</td>
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<td>Granite</td>
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<td>8.46</td>
<td>1.79</td>
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<td>Paragneiss (minus Leucosome)</td>
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<td>Leucosome of paragneiss</td>
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<td>2.30</td>
<td>3.85</td>
<td>1.67</td>
</tr>
<tr>
<td></td>
<td></td>
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<td>1.17</td>
<td>2.07</td>
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<td>2.30</td>
<td>3.95</td>
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<td>0.74 ± 0.05</td>
<td>0.80</td>
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<td>21</td>
<td>Mafic xenolith from xenolithic tonalite</td>
<td>1.26</td>
<td>0.27</td>
<td>0.56</td>
<td>0.21 ± 0.03</td>
<td>0.24</td>
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Appendix C: Increased heat production of superior rocks at 2.7 Ga
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<tr>
<th>A Sample no.</th>
<th>B Description</th>
<th>C Th, ppm</th>
<th>D U, ppm</th>
<th>E K₂O, wt%</th>
<th>F Present Heat production, µW m⁻³</th>
<th>G Present heat production, µW m⁻³</th>
<th>H 2.7 Ga heat production, µW m⁻³</th>
<th>I Factor Increase, H ÷ G</th>
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Appendix C: Increased heat production of superior rocks at 2.7 Ga
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<th>F</th>
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<td>D U, ppm</td>
<td>E K2O, wt%</td>
<td>F Present Heat production, µWm⁻³</td>
<td>G Present heat production, µWm⁻³</td>
<td>H 2.7 Ga heat production, µWm⁻³</td>
<td>I Factor Increase, H ÷ G</td>
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Table 2. Heat production in rock samples from the Pikwitonei region. Data in columns A – G is from Fountain et al. (1987). My calculations appear in columns H – J. bi, biotite; gar, garnet; mag, magnetite; pyx, pyroxene; trond., trondhjemite.

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<th>E U, ppm</th>
<th>F K₂O, wt.%</th>
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<th>H Present heat production, µWm⁻³</th>
<th>I 2.7 Ga heat production, µWm⁻³</th>
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Table 3. Heat production in rock samples from the Sachigo region. Data in columns A – G is from Fountain et al. (1987). My calculations appear in columns H – J. bi, biotite; hb, hornblende; plag, plagioclase; sill, sillimanite.

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SILICIC INTRUSIVE ROCKS

|    |                    |     |     |     |                   |                  |                  |                   |                       |
|    |                    |     |     |     |                   |                  |                  |                   |                       |
| P16 | Bear Lake granite  | 2.628 | 5.75 | 4.31 | 1.87 | 1.67 | 1.78 | 3.96 | 2.23 |
| P21 | Bear Lake granite  | 2.680 | 34.60 | 1.69 | 4.49 | 3.33 | 3.52 | 6.68 | 1.90 |
| P41 | Leucogranite        | 2.659 | 10.40 | 1.72 | 2.87 | 1.45 | 1.59 | 3.66 | 2.29 |
| P43 | Granodiorite        | 2.744 | 6.04 | 1.03 | 1.63 | 0.87 | 0.96 | 2.18 | 2.28 |
| P54 | Quartz diorite      | 2.640 | 6.25 | 1.37 | 1.79 | 0.96 | 1.05 | 2.40 | 2.29 |
| P57 | Granodiorite        | 2.695 | 2.65 | 0.59 | 1.53 | 0.49 | 0.57 | 1.57 | 2.74 |

METAVOLCANIC AND INTRUSIVE ROCKS

<p>| | | | | | | | | | |
|    |                    |     |     |     |                   |                  |                  |                   |                       |
|    |                    |     |     |     |                   |                  |                  |                   |                       |
| P19 | Metabasalt         | 3.076 | 0.24 | 0.03 | 0.21 | 0.05 | 0.06 | 0.20 | 3.14 |
| P20 | Metabasalt         | 3.069 | 1.44 | 0.28 | 0.25 | 0.23 | 0.24 | 0.50 | 2.05 |
| P22 | Metatuff           | 2.902 | 37.60 | 6.19 | 3.16 | 4.96 | 5.10 | 9.15 | 1.79 |
| P28 | Metabasalt         | 3.019 | 0.13 | 0.02 | 0.09 | 0.02 | 0.03 | 0.09 | 2.93 |
| P29 | Metagabbro         | 3.025 | 0.32 | 0.17 | 0.22 | 0.10 | 0.11 | 0.30 | 2.65 |
| P38 | Metagabbro         | 3.113 | 1.70 | 0.33 | 0.14 | 0.25 | 0.26 | 0.47 | 1.80 |
| P39 | Metabasalt         | 2.996 | 0.61 | 0.21 | 0.10 | 0.12 | 0.13 | 0.26 | 2.04 |
| P40 | Metagabbro         | 3.023 | 1.55 | 0.37 | 0.10 | 0.24 | 0.25 | 0.44 | 1.77 |</p>
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<td>U, ppm</td>
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Appendix C: Increased heat production of superior rocks at 2.7 Ga
Appendix C: Increased heat production of superior rocks at 2.7 Ga

Table 4. Constants used in my calculations of heat production in Tables 1 – 3.

**Avogadro’s number:** $6.0221415 \times 10^{23}$

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<td>$^{238}\text{U}$</td>
<td>$1.55125 \times 10^{-10}$</td>
<td>47.4</td>
</tr>
</tbody>
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