6 GEODYNAMICS AT THERMAL AND RHEOLOGICAL DISCONTINUITIES IN THE MANTLE LITHOSPHERE

6.1 Introduction

Strong lateral thermal and rheological discontinuities in the upper mantle may influence the deformation patterns, heat flow evolution and surface response of intraplate lithosphere. Many stable continental platforms are bordered by hot and weak regions [e.g. Cloetingh et al., 2004; Hyndman et al., 2005]. These ‘border realms’ of stable continental platforms form long-lived features of deformed, and hot and weak lithosphere. The North American Craton for instance is bordered to its west by a hot mobile belt that connects a long-lived and still active subduction system to a continental interior [e.g. Hyndman et al., 2005]. In Europe, the Eastern European Craton is juxtaposed to areas with thin and hot lithosphere that got thermally perturbed in Tertiary time [Blundell et al., 1992; Gee and Stephenson, 2006 and references therein]. Despite many differences, both cratons contain borders that exhibit stark thermal and rheological contrasts to areas of much warmer and thinner continental lithosphere.

Studies on seismicity and vertical motions in the last decade have shown how continental interiors are subject to active tectonics, despite their large distance to active plate boundaries [Stein and Mazzotti, 2007; Cloetingh et al., 2005]. In addition, ample regional and continent-wide studies have documented the compositional, thermal and rheological structure for the continental lithosphere of North America [e.g. Chulick and Mooney, 2002] and Europe [Cloetingh and Burov, 1996; Artemieva and Mooney, 2001; Artemieva et al., 2006]. These studies elucidate various profound thermal and rheological discontinuities in the upper mantle-lithosphere of cratons.

Ongoing research on continental intraplate lithosphere reveals the interplay of various tectonic processes that operate far from active tectonic margins including buckling and flexure that underlay large-wavelength topographic motions [Stephenson and Lambeck, 1985; Stephenson and Cloetingh, 1991; Cloetingh et al., 1999; Nielsen et al., 2007]. On the other hand, also mantle flow modelling studies have given explanation to particular surface expressions [e.g. Psyklywec and Shahnas, 2003; Shaw and Psyklywec, 2007]. The surface expression from mantle flow dynamics may entail larger wavelengths than the regional isostatic signal from the lithosphere. Mantle flow drives epeirogenic motions that explain large wavelength uplift and subsidence trends within continental interiors [e.g. Mitrovica et al., 1989; Russell and Gurnis, 1994; Psyklywec and Mitrovica, 2000].

First-order discontinuities in the thermal structure of the upper mantle can trigger mantle flow instabilities [Lenardic et al., 2003] and edge-driven convection [King and Anderson, 1995, 1998; Boutilier and Keen, 1999] also reflected in anomalous topographic features [Shahnas and Psyklywec, 2004]. Moreover, the rheological structure of the mantle lithosphere, given its rigid plate’s thermal, compositional and viscosity structure, may exert a top-down influence on convection to which the lithosphere in return again responds [Anderson, 2001].

The base of the mantle lithosphere may be defined as a purely thermal boundary layer or alternatively as compositionally distinct from the underlying mantle lithosphere as well [e.g. Jordan, 1978]. Abrasion of mantle-lithosphere by mantle flow, removes material that is denser than the overlying crust. This increases the overall buoyancy of the lith-
osphere and may lead to surface uplift. Changes to the base of the thermal lithosphere may also impose thermal re-adjustment to the overlying lithosphere and consequently affect its density and rheological structure and heat-flux patterns.

6.1.1 The case of the Canadian Cordiller

This study is inspired by the southern Canadian Cordillera, where a hot upper mantle and thin lithosphere of the Cordillera stand in strong contrast with the continental lithosphere of the stable North American plate. High surface heat flow values (> 80 mW/m²), high electrical conductivities [Gough, 1986; Jones and Gough, 1995] and low seismic velocities in the uppermost mantle are found for the Omineca and Intermontane belts of the Cordillera (see Figure 1.2) [Gough, 1986; Jones and Gough, 1995; Hyndman and Lewis, 1999]. To the east, these proxies change, over a noticeable transition zone, back to values typical for cratonic lithosphere of the platform-craton (e.g. surface heat flow of ~50 mW/m²). As a result, the Cordillera to platform transition also exhibits a strong rheological discontinuity with an approximated integrated strength difference of 0.5-2.0·10³ GPa-m versus 10.0-50.0·10³ GPa-m for the Cordillera and platform lithosphere, respectively [Lowe and Ranalli, 1993; Hyndman and Lewis, 1999] (see Chapter 5).

A similar pattern has recently been observed in high resolution tomography studies [van der Lee and Frederiksen, 2005; Mercier et al., 2009; Sigloch et al., 2008]. These studies outline a first-order transition in seismic velocities from Phanerozoic mantle underneath the Cordillera relative to the mantle of the North American Craton. In western Canada, Frederiksen et al. [2001] and van der Lee and Frederiksen [2005] have documented low seismic velocities underneath the central Cordillera with a significant increase in seismic velocity that roughly follows the deformation front in the southern Cordillera in the south.

The hot and thin lithosphere of the Cordillera is attributed to upwelling of hot asthenosphere, driven by viscous drag along a subducting plate. As mantle material is dragged from beneath the Cordillera down along the plate, it is replaced by hot uprising material central beneath the Cordillera that flows back toward the trench as shallow sub-horizontal flow. Recent geodynamic modelling [Currie et al., 2004, 2008] of this corner flow dynamics has provided new insights in the thinning of Cordillera-style continental lithosphere by mantle upwelling. Hyndman et al. [2005] refers to the Cordilleran lithosphere as a continental backarc, i.e. the situation of a hot and thinned continental plate overriding a subduction system. Initially back-arcs were recognized as high heat flow regions in Pacific ocean-ocean subduction systems, within volcanic-arc, back-arc and island-arc systems. In the Mediterranean for instance, backarcs also refer to continental plates as upper-plate of a subduction and exposed to large extension and oceanization [Jolivet and Faccenna, 2000]. In the North and South American Cordilleran system the term backarc has been used for a zone of thin and hot continental lithosphere above the subducting Pacific plate as synthesized by Hyndman et al., [2005]. Even though the geodynamic implications of the Cordilleran backarcs in comparison to global backarcs [i.e. Hyndman et al., 2005; Currie and Hyndman, 2006] was disputed by Schellart et al. [2007], the use of backarc also for the Canadian Cordillera was not disputed. In lack of any better term, this paper uses backarc purely in the descriptive sense of thin hot continental lithosphere in vicinity of a subduction system.

This study concentrates on the implications of an established thermal and rheological transition between a hot backarc lithosphere (i.e. in sense of Hyndman et al., [2005]) to
a stable continental cratonic platform. The effect of plate boundary forcing is disregarded and the mantle dynamics from active plate margins reduced to representative boundary conditions. The geodynamic response of the transition and stable platform is studied in terms of changes in sub-lithospheric mantle flow, thickness evolution of the thermal and mechanical lithosphere and surface topography and heat flow signal.

This study presents numerical models for a coupled asthenosphere-lithosphere system. First a series of benchmark models are presented in which the ambient evolution of mantle flow and surface heat flow and deflection responses are tested for various simple thermal boundary layer models of different thicknesses. Subsequently, the study deals with the geodynamic response to a thermal perturbation that is imposed in the upper mantle at the onset of the model run. Last, the implications of various rheological structures are discussed (i.e. variant rheologies for the mantle lithosphere) for the overall model response and especially the surface heat flow and surface deflection signal.

6.2 Numerical modelling description

6.2.1 Governing principles

The numerical experiments use the plane strain viscous-plastic finite element code SOPALE (i.e. Simplified OPtimized Arbitrary-Lagrangian-Eulerian code) that solves a system of thermo-mechanical equations with conservation of mass, momentum and energy [Fullsack, 1995]. This system of thermo-mechanical equations are solved for temperature (T), velocity (v) and the stress tensor (σ) [see for further details Pysklywec et al., 2002].

The model assumes a viscoplastic strain from the calculated deviatoric stresses by picking the deformation mechanism that requires the lowest differential stress (i.e., the yield stress σ_y for plastic strain or viscous stress σ_v for creep mechanisms) In generic form, the constitutive relation between stress and the viscoplastic strain is of the form:

\[ \sigma = \eta \dot{\varepsilon} \]  

(6.1)

The model assumes viscous deformation from a power law creep expression in which the effective viscosity as function of strain rate and temperature is given by [see also Pysklywec et al., 2002]:

\[ \eta (\dot{\varepsilon}, T) = \left[ \frac{n+1}{n} \cdot \frac{1-n}{2n} \right] \cdot \left[ A^{-1/n} \cdot \left[ \dot{\varepsilon}^{n+1} \right] \cdot \dot{\varepsilon}^{Q/nRT} \right]. \]  

(6.2)

The variables A, n, Q are the viscosity parameters, powerexponent and activation energy from uniaxial laboratory experiments and R is the ideal gas constant. Note that the first bracketed term on the right-hand side of equation (6.2) is necessary for the conversion of the uniaxial laboratory experimental data to a state of stress that is independent of the choice of coordinate system.

For stresses above the yield stress (σ_y), the deformation shifts from continuous viscous creep to brittle yielding. This break of stress by brittle deformation is described by the Drucker-Prager yield criterion, which is an equivalent of the Coulomb yield criterion for plane strain [Fullsack, 1995] and given by:

\[ \sigma_{\text{shear}} = \sigma_{\text{normal}} \cdot \sin (\Phi) + C_0. \]  

(6.3)

In this expression the material parameters \( \Phi \) and \( C_0 \) represent the internal friction coefficient and the material cohesion, respectively. The internal friction angle sets the slope
of the brittle yield curve and confines the maximum stress level that the lithosphere sustains through viscous deformation before plastic yielding occurs.

The two-dimensional temperature structure is governed by the energy balance equation that contains conductive, advective and heat production terms. For low effective velocities, conduction will dominate, which is the case for the lithosphere, whereas the temperature evolution in the convecting asthenosphere occurs primarily by heat advection. It follows that the temperature gradient in the conductive lithosphere can be represented by simpler heat diffusion equation for which temperature as a function of depth was given in equation 5.2. An idealized case for the mantle vertical temperature structure under vigorous convection is the adiabatic temperature gradient [e.g. Turcotte and Schubert, 2002] that is given by:

\[
\frac{\partial T}{\partial z} = \frac{\alpha g T}{c_p}.
\]  

(6.4)

The adiabatic temperature gradient holds an approximate gradient of \( \sim 0.5 \) K/km at start-up for an thermal expansion coefficient \( (\alpha) \) of \( 3.0 \times 10^{-5} \) K\(^{-1}\) and a heat capacity \( (c_p) \) of 1212 J·kg\(^{-1}\)·K\(^{-1}\).

In SOPALE, the system of equations are solved for \( T, v, \) and \( \sigma \) using a coupled Lagrangian-Eulerian finite element method [Fullsack, 1995]. The Eulerian mesh serves as solver grid and remains essentially undeformed, whereas the Lagrangian mesh acts as tracker grid that advects along with the deforming material. This technique is useful for dealing with high strain materials (e.g., convecting mantle) and for explicitly tracking moving material interfaces, such as the free surface and the base of the crust. Previous numerical experiments [e.g. Fullsack, 1995] with SOPALE have extensively verified the accuracy of the computational code by an extensive series of benchmarking models that also include tests to Rayleigh-Taylor instabilities in agreement with other numerical and analytical studies [e.g. Houseman and Molnar, 1997; van Keken et al., 1997].

### 6.2.2 Model set-up

Figure 6.1 shows the set-up of the model with a solution space of 3000 km width and 660 km depth. A depth of 660 km is chosen as base for the models as it corresponds to a phase-change (i.e., spinel to perovskite plus magnesiowustite phase-change), which provides a positive buoyancy that balances thermal buoyancy and inhibits convection to the deeper mantle [e.g. Korenaga and Jordan, 2004] as well as a significant viscosity increase from the upper to lower mantle [Forte and Mitrovica, 1996].

The computational mesh has 201x101 and 601x201 nodes in the x and y directions for the Eulerian and Lagrangian meshes, respectively. The top 200 km of the mesh is at a higher resolution than the lower part to increase the stability and accuracy of the calculations in the thermal boundary layer (Figure 6.1). All models have a hot upper mantle and thin lithosphere to the left and a much cooler upper mantle and thicker lithosphere to the right of the model domain.

As the study focuses on the geodynamic effects of intraplate thermal discontinuities, the models disregard plate boundary forcing and dynamics from active plate margins. Any geodynamic response observed in the models, results purely from the mantle flow and lithospheric response to the thermal and rheological discontinuity in the upper mantle. We thus exert no horizontal displacements along the side boundaries of the model. However, material along the side boundaries can freely move vertically.
The numerical model set-up showing the mesh dimension and resolution and initial-stage and boundary conditions. The initial stage temperature structure defines the depth to the base of the lithosphere. The crustal thickness is 35 km. The rheological properties of the crust, mantle lithosphere and asthenospheric mantle are given in Table 6.1.

Table 6.1

<table>
<thead>
<tr>
<th>Parameter</th>
<th>crust</th>
<th>mantle lithosphere</th>
<th>mantle</th>
</tr>
</thead>
<tbody>
<tr>
<td>$C_p$ (J·kg$^{-1}$·K$^{-1}$)</td>
<td>1136</td>
<td>1212</td>
<td>1212</td>
</tr>
<tr>
<td>k (W·m$^{-1}$·K$^{-1}$)</td>
<td>2.5</td>
<td>3.5</td>
<td>3.5</td>
</tr>
</tbody>
</table>

Plastic rheology

<table>
<thead>
<tr>
<th>Parameter</th>
<th>value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$C_s$ (MPa)</td>
<td>1.0</td>
</tr>
<tr>
<td>$\Phi_{eff}$</td>
<td>15° to 2°</td>
</tr>
</tbody>
</table>

Viscous rheology

<table>
<thead>
<tr>
<th>Parameter</th>
<th>wet quartzite (*1)</th>
<th>dry olivine (*2)</th>
<th>wet olivine (*2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A (Pa·s$^{-1}$)</td>
<td>1.10e-28</td>
<td>4.85e-17</td>
<td>4.89e-15</td>
</tr>
<tr>
<td>n</td>
<td>4</td>
<td>3.5</td>
<td>3.5</td>
</tr>
<tr>
<td>$Q$ (kJ·mol$^{-1}$)</td>
<td>223</td>
<td>515</td>
<td>498</td>
</tr>
</tbody>
</table>

Material dependent parameter values (*1) [Gleason and Tullis, 1995]; (*2) Hirth and Kohlstedt [1996]

Table 6.2

<table>
<thead>
<tr>
<th>Parameter</th>
<th>value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\rho_{crust}$</td>
<td>crustal reference density 2800 kg/m$^3$</td>
</tr>
<tr>
<td>$\rho_{mantle}$</td>
<td>mantle reference density 3300 kg/m$^3$</td>
</tr>
<tr>
<td>g</td>
<td>gravity acceleration 10 m/s$^2$</td>
</tr>
<tr>
<td>R</td>
<td>gas constant 8.314 J·K$^{-1}$·mol$^{-1}$</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>thermal expandability 3.0·10$^9$ K$^{-1}$</td>
</tr>
<tr>
<td>$\kappa$</td>
<td>Thermal diffusivity 1.0·10$^8$ m$^2$/s</td>
</tr>
<tr>
<td>$T_o$</td>
<td>Reference temperature 293 K</td>
</tr>
</tbody>
</table>

Parameter Values (common to all models)
In the reference model, the crust has a uniform thickness of 35 km at onset of the model and creep parameters according to a wet quartzite composition [Gleason and Tullos, 1995] (Table 6.1). The plastic rheology (\( \Phi_{ef} \) and \( C_0 \)) of the crust has a cohesion of 1.0 MPa and strain weakening controlled by a decrease in friction angle (15° to 2°) with accumulating strain (Table 6.1). The mantle has a normally wet olivine rheology representative for non-dehydrated asthenosphere [Hirth and Kohlstedt, 1996]. For the initial reference model, the base of the lithosphere is defined as a thermal boundary layer, with the mantle lithosphere having the same composition as the rest of the mantle. In addition, models with a dry olivine rheology [Hirth and Kohlstedt, 1996] are performed to account for additional rigidity of the cratonic mantle lithosphere. The plastic rheology of the mantle under high confining pressures is defined by a high material cohesion and no internal friction angle (Table 6.1).

All materials have a temperature dependent density structure with reference densities of 2800 kg/m³ and 3300 kg/m³ for respectively the crust and mantle (Table 6.2). The thermal rock parameters are further summarized in Tables 6.1 and 6.2 for the reference model, and deviations for alternative model runs are shown in Table 6.3. The temperature structure is thus defined for the initial stage and the model starts after which the temperature structure evolves freely, only governed by the heat conduction and convection equations.

### 6.2.3 Characterization of the asthenosphere and lithosphere

The convection in the asthenosphere can be characterized by the behaviour and vigour of mantle flow. The Rayleigh Number describes the tendency of a system to convect and represents the ratio of buoyancy forces to viscous resisting forces. For a bottom heated layer with Newtonian viscosity (\( n=1 \)) [e.g. Turcotte and Schubert, 2002] the dimensionless Rayleigh Number is given by:

\[
Ra = \frac{\alpha \rho_0 g \Delta T L^3}{\kappa \eta},
\]

where \( L \) is the thickness of the layer and \( \Delta T \) is the temperature difference between the top and bottom. The mantle’s Rayleigh number is of the order 10⁶ for typical thermal expansion coefficient of 2-3·10⁻⁵ K⁻¹, viscosities of 10²⁰ - 10²¹ Pa·s and thermal diffusivities in range of 10⁻⁵-10⁻⁷ m²/s. When the Rayleigh number is below the critical value (i.e. \( Ra_c \), in order of 10³), heat transfer primarily occurs by conduction, whereas convection dominates for greater values. A large layer thickness and a large temperature difference favour convection, whereas a high viscosity, high thermal diffusivity or low thermal expansion coefficient hinder convection. For instance, reducing the thermal expansion coefficient (as for the thermal conductivity) makes the model less prone to the development of convection cells.

The lithosphere instead forms a stagnant high viscous lid above a low viscosity convective asthenosphere. The base of the lithosphere forms a mechanical and thermal boundary layer zone [e.g. Schmeling and Marquart, 1993]. The mechanical boundary is marked by the transition from convection by flow in the low viscous asthenosphere to a lithosphere that exhibits stronger resistance to viscous flow. The first-order viscosity structure of the asthenosphere is rather uniform with global averages for the upper mantle around 10²⁰-10²¹ Pa·s. Some lower values have been reported for the western US (10¹⁸ - 10¹⁹ Pa·s) [Dixon et al., 2004]. Instead, the lithosphere contains a more distinct rheological stratification with viscosity ranging between 10²⁰-10²⁵ Pa·s. With strain rates in range of 10⁻¹⁵
This gives the range of earlier found mantle lithosphere rigidities (10-1000 MPa; see Chapter 5).

This also determines the difference in vertical temperature structure between the asthenosphere and lithosphere. The vertical temperature profile in the asthenosphere can be idealized by the mantle adiabat, whereas in the lithosphere it is best represented by a conductive temperature gradient (Figure 6.2a). The mechanical transition implies also a change from advection to conduction-dominated heat transfer that consequently mark the thermal boundary layer. Figure 6.2b outlines the definition of the thermal boundary layer as a zone between the conductive lithosphere and underlying convecting mantle [Parson and McKenzie, 1978; Doin et al., 1997; Jaupart and Mareschal, 1999]. While the BTL in reality forms a transition zone, for plotting purpose the BTL will be represented as the 1600 K isotherm (Figure 6.2b).

6.2.4 Model Series

This study contains a successive series of model runs that first benchmarks the lithosphere as a thermal boundary layer and subsequently tests the implications of temperature steps and rheological heterogeneities. Table 6.3 lists the various models that this study contain, including the thermal boundary layer models (labelled as BoundLyr_), the model series with a hot upper mantle to the left of the domain referred to as hot backarc (TempModel_) and the models with variant rheological compositions (RheoMdl_).
All models contain a specified temperature field at their onset. Figure 6.2a gives a 1-D representation of the initial temperature structure. For the conductive branch, the temperature input model is prescribed by the $q_{\text{surf}}$ and the corresponding base of the thermal lithosphere (BTL). The temperature structure of the asthenosphere is approximated by the mantle adiabat (equation 6.4) with a gradient of $\sim 0.5$ K/km (i.e. for $\alpha$ of $3.0 \cdot 10^{-5}$ K$^{-1}$ and $c_p$ of 1212 J kg$^{-1}$ K$^{-1}$). For a depth of the conductive lithosphere base at 160 km (i.e. 1600 K isotherm), the mantle adiabat gives a temperature of $\sim 1800$ K at the base of 660 km deep model domain (Figure 6.2a).

The temperature structure of the initial input model may be varied. For the conductive upper part, the depth of the 1600 K isotherm and the corresponding surface heat flow are changed between different models. Also the adiabatic part of the input temperature structure may be varied. Lowering the $\alpha/c_p$ ratio means reducing the gradient what leads to lower temperatures at the bottom of the model box and an overall cooler convecting asthenosphere. Some previous studies benchmarked mantle convection with an iso-thermal or iso-viscous asthenosphere [among others Shapiro et al., 1999]. However, the herein discussed models all have a specific initial temperature gradient with a thermal expansion coefficient ($\alpha$) of $3.0 \cdot 10^{-5}$ K$^{-1}$, as given in Figure 6.2a.

Differences in initial temperature structure input will be characterized by the depth to BTL. The modelling considers a lateral temperature discontinuity outlined by a step in the depth to the base of the BTL (Figure 6.2c). The lithosphere to the left of the model domain contains a surface heat flow above 80 mW/m$^2$ and a corresponding depth of the BTL less than 80 km (i.e. typical for a continental backarc, see Currie and Hyndman [2006]). The lithosphere to the right of the model domain contains a surface heat flow of $\sim 50$ mW/m$^2$ and a corresponding depth to the BTL of $\sim 160$ km and considered representative of cratonic lithosphere [e.g. Artemieva and Mooney, 2001]. The consequences of these initial temperature structures (i.e. especially from large temperature transitions in the upper mantle) on dynamics of the mantle flow and lithosphere form the focus of this study. Another factor that is taken into account is the compositional structure of the backarc and cratonic mantle lithosphere.

The models predict temperature and displacement fields. They further give the predicted surface heat flow and deflection response. These free surface deflections, commonly referred to as dynamic topography, result from the pushing up and down of the Earth’s surface in response to viscous stresses driven by internal buoyancy forces within the mantle. Previous studies show how the predicted topography in such models is a result of both the surface isostatic response to the deforming crust and the surface response to flow dynamics in the underlying mantle [e.g. Pysklywec and Shahnas, 2003].

### 6.3 Modelling mantle convection and lithosphere temperature structure

#### 6.3.1 Convection below a lithosphere as boundary layer

The first series of benchmark models test the evolution of the lithosphere as a thermal boundary layer in exchange with upper mantle convection (BoundLyr model series; see Table 6.3). This test was done to parameterize a restrained (semi-stationary) model in regard to the lithosphere thermal structure and to serve as template for the follow-up models. The asthenosphere to lithosphere boundary (i.e. the base of the thermal lithosphere – BTL) is governed by the interplay between the conductive heat transfer and consequent heat loss to surface and convective heating at its base.
The study does not address the precise nature of mantle lithosphere instabilities that may arise over time and for which various scenarios have been proposed, including mantle delamination [Bird, 1979], convective removal [Morency and Doin, 2004] or gravitational instabilities [e.g. England and Houseman, 1989]. The initial temperature structure is laterally uniform with a conductive geotherm for the lithosphere set by the 1600 K isotherm (see Figure 6.2 and Tables 6.1 and 6.2 for model set-up and parameter definition).

For the first model (BoundLyr1; Table 6.3) the depth of the BTL is defined at 160 km and Figure 6.4 shows the consequent evolution in temperature and displacement fields for a series of time-slices. The first convection cell is activated with a shear at the lower left, after which convection cells in the asthenosphere start forming gradually eastward in the box. With a thickness of the convective mantle and an aspect ratio of ~1, the first-order convection cells have a width and depth of 500 km. With convection, hot mantle

Figure 6.3

Benchmark model testing the stability of the lithosphere for different initial thicknesses of the thermal boundary layer. (a) Evolution for a thermal boundary layer of initially 160 km thickness is presented by four time-slices at 10, 30, 50 and 100 Ma and showing the temperature and displacement fields.
material ascends and replaces the colder sub-lithospheric mantle. Heat from the zones of mantle upwelling subsequently propagates by conduction into the overlying lithosphere. Convection is established throughout the box by 20-30 Myr. At first, the lithosphere conductively cools and thickens, because convection has not developed yet throughout the model domain. Once convection is established, mantle heat is transferred to the lithosphere base where small-scale instabilities develop and the depth to the BTL declines and stabilizes around 130-170 km.

This is different for a model that has no initial shear applied to the lower left corner (BoundLyr2), and for which convection is consequently initiated much later. For instance, for a model with an initial 160 km-thick thermal lithosphere, convection starts only after 180 Myr. Convection is initiated from instabilities at the lithosphere base after it has conductively cooled and thickened up to 250 km. This difference in evolution for models BoundLyr1 and BoundLyr2 is also shown when plotting the change in thickness of the thermal lithosphere through time (Figure 6.4). The average depth to the BTL (1600 K) are shown, both with an initial thermal base at 160 km. The difference in timing for convection to start, in 10 Ma for the first and up to 180 Ma for the latter, is reflected in the curves. After convection establishes, both models exhibit comparable thickness fluctuations between 130-170 km. The third model (BoundLyr3) gives the thickness evolution of a lithosphere with an initial 80 km thickness. This model contains a hot upper mantle that is prone to convection. Over time however, the lithosphere loses more heat from conductive transfer than the amount of heat that enters at its base from convective hot mantle upwising, and consequently thickens to 140-170 km.

Many numerical studies have addressed the thickness evolution and stability of continental lithosphere as a result of compositional buoyancy and thermal boundary layer instabilities (i.e. tectosphere) [e.g. Jordan, 1978; Shapiro et al., 1999]. Numerical modelling indicates that conductive cooling and thickening and gravitational instabilities and thinning together establish a semi-stable continental lithosphere [e.g. Houseman and McKenzie, 1981; Houseman and Molnar, 1997; Conrad and Molnar, 1999; Lenardic et al., 2003]. Studies that make global comparison of the thickness of long-lived cratons in regard to the conductive heat evolution give comparable ranges of ~175-185 km [Artemi-

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**Figure 6.4**

![Time-series curves of the evolution in lithospheric thickness distribution over the box dimension of 3000 km for the different models including different definitions of thermal boundary layer’s initial stage (see Table 6.3).](image-url)
In the thermal boundary benchmark models (Figure 6.3) similar thicknesses are reached after the lithosphere first conductively cools and thickens and subsequently thins as result of gravitational instabilities and convection. Lithosphere thicknesses of 150-200 km display sufficient stability over the time-scales of the model runs and are taken as representative for the cratonic lithosphere.

### 6.3.2 Edge-drive convection around thermal discontinuity at lithosphere base

The principle that a convection cell establishes around a lateral temperature discontinuity in the upper mantle without alternative trigger was recognized in numerical experiments of continent-ocean transitions [e.g. King and Anderson, 1998]. However, the effect of such a lateral temperature transition for the evolution of continental intraplate lithosphere has only recently attracted more attention [e.g. van Wijk et al., 2008].

<table>
<thead>
<tr>
<th>Table 6.3</th>
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<tbody>
<tr>
<td><strong>Model</strong></td>
</tr>
<tr>
<td>BoundLyr1</td>
</tr>
<tr>
<td>BoundLyr2</td>
</tr>
<tr>
<td>BoundLyr3</td>
</tr>
<tr>
<td>TempModel1 (reference model)</td>
</tr>
<tr>
<td>TempModel2</td>
</tr>
<tr>
<td>TempModel3</td>
</tr>
<tr>
<td>TempModel4</td>
</tr>
<tr>
<td>TempModel5</td>
</tr>
<tr>
<td>RheoModel1 (reference model)</td>
</tr>
<tr>
<td>RheoModel2</td>
</tr>
<tr>
<td>RheoModel3</td>
</tr>
</tbody>
</table>

List of numerical experiments describing the change in model definition relative to reference models. For the detailed rheological parameter description is referred to Table 6.1.

*eva and Mooney, 2001*.
The first series of models (i.e. the TempModels; Table 6.3) have an explicit temperature structure at onset that define a lateral temperature discontinuity between a hot and thin upper mantle lithosphere to the left and a cool and much thicker mantle lithosphere to the right of the model domain. The TempModel1 has an initial temperature structure with a depth of 100 km to the base of the thermal lithosphere to the left with a transition zone (250 km width at x = 500 km) to the thicker platform lithosphere to the right (thickness of 160 km thickness; Figure 6.2c).

TempModel1 (Figure 6.5) shows how an initial convection cell forms around the thermal discontinuity. A strong descend of mantle material occurs from underneath the thin lithosphere downward along the thermal discontinuity to the deeper situated base of the continental platform lithosphere. At the edge of the cold platform lithosphere, material sinks deeper in mantle and is being replaced from both sides. The edge drive convection initiates particularly strong upward flow underneath the hot and thin lithosphere to the left of the discontinuity (Figure 6.5). Afterward, the mantle flow spreads through the entire model domain with wide zones of hot mantle upwelling and narrow zones of cold descending mantle. These convection cells in the asthenosphere, with an aspect ratio of \( \sim 1 \), have a width and depth in the order of 500 km.

To further test the dynamics of the transition zone, the contrast in thickness between the hot and thin versus cold and thick and lithosphere was varied as depicted in the model comparison chart (Figure 6.6). The temperature discontinuity at the lithosphere base was amplified by reducing the thickness of the thin lithosphere to 60 km and applying a 200 km thickness for the platform lithosphere (Figure 6.6). This TempModel2 shows again convection in the box starting around the temperature discontinuity. The convection then propagates under the continental lithosphere that heats up and thins slightly to 160 km thickness. Over time the amplitude of the thermal step in the BTL remains significant and stable.

The TempModel2 model develops a much stronger flow along the discontinuity when comparing the ascent of hot mantle and descent at the craton edge with the results from TempModel1. The thermal discontinuity shifts 500 km eastward after 30 Myr due to the strong uprising hot mantle of TempModel2 that laterally replaces the cold mantle lithosphere. After 100 Myr, the discontinuity has moved 1000 km toward the craton lithosphere. While shifted, the discontinuity maintains a strong amplitude with 150 km for the craton lithosphere in strong contrast with the 80 km-thick extended domain of thin and hot lithosphere.

The model comparison charts also presents panels with lateral surface-heat flow profiles for each of the time-slices of the different models (Figure 6.6). The discontinuity at the BTL is reflected as primary feature in the surface heat flow distribution. The evolution of the discontinuity is well reflected in the surface heat flow distribution with a difference of 30 mW/m² across the discontinuity after the mantle flow establishes. Furthermore, the lateral change in surface heat flow of TempModel2 follows well the eastward shift of the discontinuity at the BTL.

### 6.3.3 Evolution of the thermal discontinuity

The asthenosphere to lithosphere boundary (i.e. the base of the thermal lithosphere – BTL; Jaupart and Mareschal, [1999]) is governed by the interplay between the conductive heat transfer and consequent heat loss to surface and convective heating at its base. After the first convection cell develops spontaneously after a few Myr at the discontinu-
The evolution of TempMd1 (reference model) shown in five time-slices at 1, 10, 30, 50 and 100 Ma. The model shows a lateral transition in thermal structure of the lithosphere marked by change in depth to the base of the thermal lithosphere (BTL). The initial contrast is set from a thin (100 km) and a hot lithosphere at the left to a continental platform lithosphere to the right with a thick (160 km) and cool thermal lithosphere. The temperature and displacement fields are shown for each time-slice and temperature profiles are plotted to the right and derived from vertical slices at (I) x = 250 km, (II) x = 1250 km and (III) x = 2250 km. The BTL is shown by a temperature contour and marked with arrows in the temperature profile where it intersects with the mantle adiabat.
ity, adjustments to the BTL occur from the interplay between conductive heat transfer in lithosphere and mantle dynamics (i.e. convective flow and instabilities) at its base. The trace of the BTL (i.e. 1600 K isoline) outlines well the evolution of the thermal discontinuity over time. The BTL also depict drips that develop especially at the base of the thicker and colder cratonic lithosphere. This study does not address the precise nature of such instabilities that may arise over time and for which various scenarios have been proposed, including gravitational instabilities [e.g. England and Houseman, 1989] that, under certain circumstances may lead to convective removal [Morency and Doin, 2004] or delamination [Bird, 1979] of the mantle. This study instead focuses mainly on how the temperature discontinuity behaves in response to the mantle dynamics.

TempModel1 and TempModel2 in fact show that the thermal discontinuity imposed at onset is not readily eliminated by such instabilities, but instead remain a significant feature. Widening of the thermal discontinuity occurs for TempModel1 (Figure 6.5). From 10 to 50 Myrs, the temperature contrast between the upper mantle of the backarc and stable platform has flattened and migrated eastward. However, over a 100 Myr time-span, the thermal transition does not entirely dissipate (Figure 6.5). Heat conduction diffuses lateral discontinuities in the lithosphere if not otherwise maintained, whereas mantle dynamics has both the capability to destroy or support thermal discontinuities [Schmeling and Marquart, 1993]. This effect of strong edge-driven convection in support of the lateral temperature discontinuity is well illustrated with TempModel2. This second model exhibits enhanced edge-driven flow around a higher amplitude discontinuity that better preserves the thin lithosphere at one side by stronger upwelling and thick lithosphere at the other side due to enhanced downwelling.

Also several studies that addressed the fundamental issue of continental lithosphere longevity included in their models lateral temperature discontinuities at defined continental edges [e.g. Jordan, 1978; Shapiro et al., 1999]. The main focus of those studies were how conductive cooling and thickening followed by gravitational instabilities and thinning ultimately establish a ‘semi-stable’ continental lithosphere [e.g. Houseman and McKenzie, 1981; Houseman and Molnar, 1997; Conrad and Molnar, 1999; Lenardic et al., 2003]. The studies that contain edges to their ‘semi-stable’ platform lithosphere [e.g. Jordan, 1978; Shapiro et al, 1999; Lenardic et al., 2003] describe how mantle drips develop at the edge of cold platform lithosphere that consequently widens and flattens. This shows that together with the stability of continental platforms itself, also the accompanying temperature transitions reach long-term stability alongside the cratons [e.g. Jordan, 1978; Shapiro et al, 1999; Lenardic et al., 2003].

In conclusion, a lateral discontinuity at the base of the thermal lithosphere establishes edge-driven mantle convection. This way, the thermal structure of the lithosphere directs the development of upper-mantle flow. In turn, strong upper-mantle dynamics that establish at the discontinuity may counteract the tendency of the conductive lithosphere to diffuse strong temperature discontinuities within the thermal boundary layer. This interplay can be understood in terms of top-down and bottom-up tectonics between the lithosphere and upper mantle convection [e.g. Anderson, 2001].

6.3.4 Edge-driven versus subduction driven corner flow

The above models established convection cells by instabilities developing at the BTL in response to a strong downward flow along the temperature discontinuity. These mantle dynamics occur at a lateral temperature discontinuity that is already in place, whereas
Comparison chart of three TempModels at 1, 10, 30, 50 and 100 Myr. Model TempModel1 is the reference model with a step in the BTL of 100 to 160 km from the backarc to the continental platform, TempModel2 yields a step in the BTL from 60 to 200 km, and TempModel3 is the same as the reference model except of a shear of 3.2 cm at the lower left.
valuable alternative studies have addressed the formation of such a discontinuity in case of backarc thinning of continental lithosphere juxtaposed to a thick continental craton. Drag along a subducting plate causes a downward current and upwelling away from the slab underneath the overriding plate. Hyndman et al. [2005] and Currie et al. [2004, 2008] quantify the upper mantle dynamics underneath a continental backarc as result of subduction driven corner flow. Such subduction driven corner flow might lead to thinning of the mantle lithosphere in the overriding backarc position of the subduction system [Arcay et al, 2006; Currie et al., 2008].

While we consider the role of edge-driven flow to any thermal discontinuity of the lithosphere base away from active plate margins, continental backarcs against craton lithosphere form a good example with a very pronounced thermal upper mantle discontinuity. Edge-driven convection can only start operating after subduction driven corner flow has etched out the discontinuity. However, interplay may occur between the subduction driven corner flow and edge-driven convection once the discontinuity is established. In this study, the subducting plate is being omitted, but the related corner flow can be simulated using a shear imposed at the lower left base of the model (TempModel3) that triggers a counter-clockwise convection cell (Figure 6.6; at 1 Myr).

Depending on the width of the imposed hot backarc (i.e. 500 km in these models) and the width of the corner flow and edge-driven convection cells, these two mechanisms both contribute to upwelling underneath the backarc. The zone of strong mantle upwelling is similar to the upwelling in the TempModel1, but the initial shear in TempModel3 introduces an alternative trigger. After 20 Myr, once convection is established throughout the domain, there is no difference in evolution between TempModel1 and TempModel3. Therefore, in the rest of this study the influence of subduction-driven corner flow is disregarded, although it might enhance the upwelling at the backarc-side of a discontinuity when a higher basal shear is imposed (i.e. simulating faster subduction rates).

Studies that successfully modelled subduction driven corner-flow [i.e. Currie et al., 2004; 2008] indicated the difficulty to produce simultaneously high temperatures beneath the volcanic arc and the backarc above the subducting slab. Models that only included the corner-flow effect showed insufficient to support a hot and thin continental lithosphere of the overriding plate away from the subducting slab [Currie et al., 2004]. Traction-driven corner flow above the subducting slab is limited by the rate of subduction, and convection driven by upper mantle temperature variations (also edge-driven convection) is more rapid. More recent models addressed shear-induced gravitational instabilities in the continental backarc above the subduction driven corner-flow [Currie et al., 2008]. These models highlight the important interplay between the initiation of flow by subduction-driven traction and thinning of the overriding continental lithosphere under influence of gravitational instabilities of weaker hydrated upper mantle.

Our study imposes an initial temperature discontinuity and addresses not the thinning of mantle lithosphere by upper mantle flow and gravitational instabilities at the lithosphere base, for which the above referred studies provide viable subduction related mechanisms. However, once a temperature discontinuity at the base of the thermal lithosphere has formed, the traction that results from the edge-driven flow along the discontinuity appears stronger than flow derived from subduction-driven traction.
Temperature and heat flow evolution from base conductive lithosphere to surface for model TempMdl1 at 30, 50 and 100 Myr. The central panels give the distribution in vertical heat flow for the upper 260 km of the model domain. The upper panels show the vertical heat flow distribution for three horizontal slices at 10, 30 and 60 km depth through the model domain.

6.3.5 Mantle versus surface heat flow patterns

First-order surface heat flow trends are commonly taken as proxy of the larger lithosphere temperature structure and with mantle derived heat being a critical ingredient [Jaupart and Mareschal, 1999; Mareschal and Jaupart, 2004]. Such first-order heat flow contrasts typically coincide with variations in depth to the base of the thermal lithosphere (i.e. the BTL). Temperature variations in the upper mantle define the shape of the BTL and determine the mantle derived heat flow into the lithosphere. Mantle convection models examine the dynamics that underly the thermal variations in the upper mantle [e.g. Schmeling and Marquart, 1993]. However, linking mantle flow to lithosphere and surface heat flow patterns in numerical studies is not straightforward [e.g. Lenardic and...
Figure 6.7 depicts the vertical temperature and heat flow distribution for the upper 260 km of the model domain together with lateral crustal and surface heat flow profiles. After 30 Myr, convection cells are well developed throughout the model domain and hot upper mantle has risen to the base of the lithosphere (Figure 6.5). For instance, TempModel1 produces about 55 mW/m² of vertical heat flow at the base of the thermal lithosphere that is from a heat-advection dominated asthenosphere into a heat-conduction dominated lithosphere. Up to 60 Myr after the onset of the model does the surface heat flow signal not fully reflect the incoming mantle heat. Especially the 30 Myr time-slice (Figure 6.7; TempModel1) shows clearly the strong difference in vertical heat flow between the mantle and crustal depth-slices and at surface. In the ideal case of steady thermal lithosphere conditions, the observed surface heat flow equals the sum of mantle flow and crustal heat production. In principle, under steady-geothermal conditions and for TempModel1 that contains no crustal heat production, surface heat flow would equal the heat flow from the mantle. That this is clearly not the case is due to the strong transient heat effect of the conductive lithosphere.

In fact, heat that advected to the base of the lithosphere shortly after the initiation of mantle convection takes 60 Myr to conduct through the overlying thermal lithosphere. Shortly after convection cells establish heat accumulates at the base of the BTL and causes a steep temperature gradient in the lower mantle lithosphere (i.e. in the thermal boundary layer zone; Figure 6.2b). It produces a strong rise in vertical heat flow through the mantle lithosphere with 50 mW/m². At the same time, heat has not yet transferred upward in the lithosphere, resulting in a still smaller average temperature gradient for the lithosphere (i.e. for the craton only 8 K/km at 30 Myr; Figure 6.7) and a lower heat flow of 35 mW/m² through the overlying crust. Over time, the surface heat flow increases also when the average base of the thermal lithosphere remains approximately stable. The strong discrepancy in vertical heat flow between the mantle lithosphere and crust at 30 Myr dissipates over time. The craton to the right records only a minor transient heat effect with 15 mW/m² surface heat flow after onset increasing to 25 mW/m² after 100 Myr.

As said the crustal heat production is not included in these above described models. The effect of crustal heat production, however, is shown in TempModel5 (i.e. Figure 6.7; panels to the right) with higher crustal heat flow in the upper part of the crust. It takes much more time for the conductive lithosphere to transfer the heat from the mantle to surface than for heat generated within the crust. The surface heat flow after 30 Myr is 40 mW/m² higher, relative to the model without crustal heat production (Figure 6.7), when the extra heat from the mantle still has to enter the crust.

In conclusion, first-order thermal structure, i.e. the thermal step at the BTL, is reflected in the surface heat flow distribution. Heat transferred through the conductive lithosphere exhibits a strong transient heat effect so that first-order thermal discontinuities, introduced by mantle upwelling, take time to be reflected in the surface heat flow pattern. Smaller amplitude and wavelength variations in mantle heat flow and BTL are cancelled in the thermal boundary layer [Schmeling and Marquart, 1993; Mareschal and Jaupart, 2004].
6.4 Lithosphere response at thermal and rheological discontinuities

6.4.1 Lithosphere deformation and surface deflection

The modelling also returns a free-surface deflection signal of the upper bounding surface that may be considered as the topographic response. The predicted topography in our models is the response to changes in the density structure and to stresses induced by flow dynamics [e.g. Boullier and Keen, 1982; Pysklywec and Shahnas, 2003]. The contribution from mantle flow, commonly referred to as dynamic topography, results from the

Profiles at selected times for model TempMd11, showing the evolution over the horizontal box domain $x = 0 -> 3000$ km, with for (a) surface deflection (b) depth to base thermal lithosphere (i.e. 1600 K isotherm (c) crustal thickness (d) horizontal surface strain rate.

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Profiles at selected times for model TempMd11, showing the evolution over the horizontal box domain $x = 0 -> 3000$ km, with for (a) surface deflection (b) depth to base thermal lithosphere (i.e. 1600 K isotherm (c) crustal thickness (d) horizontal surface strain rate.

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pushing up and pulling down of the Earth’s surface in response to viscous stresses driven by internal buoyancy forces within the mantle.

Dynamic topography is sometimes also defined as the component of topography that cannot be explained by the isostatic response to crustal density or layer thickness variations. The free-surface deflections rendered by the numerical models contain density variations in the vertical columns across the model domain. Beside changes to the crust, the isostatic response in the models is also a function of temperature dependant density variations in the mantle. Furthermore, the models contain stress-induced non-isostatic contribution to the free-surface deflection that either comes from mantle flow dynamics or from stresses generated in the lithosphere itself. In this study, however, no tectonic horizontal boundary forces are considered for the lithosphere. Any response in stresses and strain in models presented in this study ultimately sources in the temperature discontinuity initially imposed at the base of the thermal lithosphere.

For the above described BoundLyr and TempModel series, the different zones where hot mantle rises versus cold mantle descends are reflected in the deflection profile by zones of positive versus negative topography, respectively. For instance in model BoundLyr1, where convection establishes by a basal shear under a 140-160 km-thick mantle lithosphere, the mantle convection cells produce free surface deflections with maximum amplitudes of 100 m and wavelengths corresponding to the width of the convection cell (~500 km). For models with a shallower or deeper initial BTL (i.e. BoundLyr4 with BTL at 80 km-depth or BoundLyr5 at 240 km depth) the free surface deflection patterns are similar.

The hot and less dense upper mantle of the backarc in model TempModel1 gives a stronger free surface response with 1000 m of positive elevation immediately after onset of the TempModel1 (Figure 6.6; 1Ma). This strong positive elevation also sets a topography of -300 m all across the platform, that has no geodynamic meaning other than that it reflects the net conservation of volume in the model domain. The integration of the deflection profile over the model domain remains essentially equal to zero because of

**Figure 6.9**

History curves of the surface deflection response for various TempModels. Three history curves are given for each model containing averaged values over the backarc domain with \(x = 0 \rightarrow 500\) km, the proximal platform \(x = 1250 \rightarrow 1750\) km and distal platform domain \(x = 2250 \rightarrow 2750\) km. (a) for TempModel1 (b) for TempModel2.
the conservation of mass and incompressibility of the material. Consequently, absolute values of the free-surface topographic prediction are disregarded, only the relative topographic spatial and temporal variations are meaningful.

However, the free-surface deflection patterns of interest in the TempModels are still significant with amplitudes of 300-600 m and result from cold mantle descent due to convection cells or gravitational instabilities at the BTL (Figure 6.6; 10 Myr). Over the duration of the model run, different gravitational instabilities develop at the BTL that produced negative surface deflection (e.g. Figure 6.5; TempModel1 at 50 Myr).

The evolution of the surface deflection for TempModel1 is further shown with profiles for selected times across the horizontal box domain (Figure 6.8). The topography profiles are plotted along with profiles of the evolution of the BTL, crustal thickness and horizontal surface strain. The first two contribute to the isostatic response, whereas the latter reflects the internal response of the lithosphere related to the dynamics of the imposed thermal discontinuity. The most prominent features in the profiles of the BTL at different times are the strong disturbances. Here, the BTL, as 1600 K isotherm, is affected by gravitational instabilities or dragged deep into the mantle. The difference in thickness of the thin lithosphere to the left versus the thick lithosphere to the right, that is noticeable in the BTL-profile at 1 Myr, is much less apparent for the subsequent time-slices. Amid of the disturbances, the BTL gives for 30-100 Myr a ~90 km depth for the initially thin lithosphere transitioning between x = 750-1250 km to a depth of 120-130 km.

The crustal thickness curves (Figure 6.8c) exhibits strong thinning for the domain with hot and thin lithosphere and distributed thickening occurs all along the cratonic crust to the right. The changes in crustal thickness are isostatically compensated and reflected in the free-surface deflection response (Figure 6.8b). The thinning of the crust to the left is related to tensional stresses, which propagate upward through the lithosphere, also reflected in the horizontal strain rates distribution at surface (Figure 6.8d). The extension occurs above the strong upward mantle flow central underneath the thin lithosphere with horizontal flow in opposite directions at the interface with the thermal lithosphere base. At surface the horizontal strain exhibits a rate of $1.0\cdot2.0\cdot10^{-16} \text{ s}^{-1}$. Over a time-span of 30-50 Myr, such strain rates thin the crust with a factor 0.2-0.3, that is about 10 km for an initial thickness of 35 km. The strong mantle flow around the step in the BTL exerts high stresses into the upper mantle and crust that push and re-distribute crustal material in the craton.

6.4.2 Mantle and lithosphere rheology structure

Flow in the asthenosphere initiates traction at the mechanical boundary that causes stresses to propagate into the rigid lithosphere. With no tectonic forces acting on the lithosphere at the side-boundaries, the crustal thinning in the above models result purely from mantle dynamics due to edge-driven flow. The nature of the asthenosphere-lithosphere interface is important for understanding these deformation patterns in effect of the edge-driven flow. For instance, the mantle flow in the models is repeatedly deranged by instabilities that develop at the BTL and that are significant in magnitude (see evolution of the BTL in Figure 6.8b). However, they appear not to control the lithosphere deformation pattern. Instead, the lithosphere deformation is especially controlled by mantle flow around the temperature discontinuity and by the hot upper mantle left of the discontinuity.

The character of the upper mantle temperature discontinuity and interaction be-
Geodynamics at thermal and rheological discontinuities in the mantle lithosphere

The mechanical strength of mantle lithosphere is defined by the temperature dependent viscous rheology. The rheology of both the asthenospheric and lithospheric mantle is determined by olivine creep parameters. Differences may exist due to reduced water content by dehydration and chemical depletion of older mantle lithosphere [Jordan, 1978]. A dry olivine rheology for the mantle lithosphere increases significantly the effective viscosity contrast with the underlying convecting mantle [Currie et al., 2004; Lenardic et al., 2003]. The question is whether the viscosity structure that sets the mechanical asthenosphere-lithosphere interface is the consequence only of a thermal boundary layer or also marked by a compositional change in water content or depletion of the olivine. Models with a compositional distinct lithosphere, for instance by considering a depleted mantle lithosphere, decouple the boundary layer from convective instabilities in the upper mantle [e.g. Jordan, 1978; Shapiro et al, 1999; Lenardic et al., 2003].

Figure 6.10 shows the viscosities and rigidities are calculated for the different crustal and mantle compositions (see also Table 6.1) and are calculated following the creep law (Equation 6.2) for a linear temperature range of 600-1800 K (Figures 6.10a and 6.10b). Quartzite and diorite describe crustal rheologies with low resistance to flow at crustal temperatures above 700 K. The olivine mantle rheologies [Hirth and Kohlstedt, 1996] are mechanically stronger at these temperatures and start flowing with a rate of $10^{-15}$ s$^{-1}$ for temperatures above 1100 K (i.e. with respectively lithosphere stresses and viscosities in range of $\sigma = 1$-100 MPa and $\eta = 10^{20}$-$10^{22}$ Pa·s). In addition to curves for dry and wet olivine creep parameters are also curves for extra-wet and moderate-wet olivine shown, representing alternative rheologies relative to the normal wet olivine [Hirth and Kohlstedt, 1996] with the first holding a weaker rheology with $A_{\text{extra-wet}} = 1/10 A_{\text{normal-wet}}$ and the moderate-wet holding a rheology between wet and dry-olivine ($A_{\text{moderate-wet}} = 10 A_{\text{normal-wet}}$).

The viscosity structure of the mantle lithosphere is not only important for understanding the coupling with the underlying convective asthenosphere, its rigidity also plays an important role in the discussion of the bulk stresses that the mantle lithosphere can support [Ranalli and Murphy, 1987; Jackson, 2002; Handy and Brun, 2004; Burov and
Watts, 2006]. Estimates of the mantle lithosphere rigidity vary in the range of 10-1000 MPa. Such upper limits to the mantle lithosphere rigidity corresponds to viscosities up to $10^{23}$ to $10^{25}$ Pa·s in case of ductile flow under average lithosphere strain rates of $10^{-16}$ s⁻¹ or a cohesion coefficient of 10-1000 MPa in case of brittle failure with low internal friction and under high confining pressures. The viscosity structure, computed in the models as a function of temperature, contain an upper limit to assure numerical stability by keeping the calculated viscosities within a reasonable range and prevent local extremes. The viscosity range may confine the mantle lithosphere below its actual rigidity when the upper limit is set to $10^{23}$ Pa·s, whereas an upper limit of $10^{25}$ Pa·s includes a wider range for the high viscid upper mantle lithosphere.

Lateral changes in the upper mantle rheological structure may occur as result of variations in the temperature or compositional structure. The above discussed TempModels already contained a lateral rheological contrast due the temperature discontinuity between the hot and thin lithosphere to the left, juxtaposed to cold and thick (craton) lithosphere to the right. As result, the rigidity of the upper mantle was much lower for the first in comparison to the latter as result of higher temperatures so that viscous flow occured at shallower depths. Thus, the already present rheological transition in the TempModels will be enhanced when adding a distinct dry olivine rheology for the craton mantle lithosphere to the right.

A dry olivine mantle lithosphere for the craton and the upper limit to the bulk rigidity of the mantle lithosphere might be critical parameters that we examine with a follow-up series of models.
6.4.3 Rigidity of the upper mantle lithosphere and response to mantle flow

Figure 6.11 compares the compositional and rheological (i.e., viscosity) structure at 30 Myr for the earlier discussed TempModel1 with anew RheoModel1 (listed in Table 6.3). A dry olivine mantle lithosphere is introduced for the craton domain to the right, whereas its temperature structure is similar to TempModel1. Along with a stronger dry olivine mantle lithosphere of craton is also the upper limit of the viscosity range increased to $10^{25}$ Pa·s and its brittle failure defined by a cohesion coefficient of 120 MPa.

Comparison of TempModel1 and RheoModel1 (Figure 6.11) shows clearly the difference in viscosity structure due to a dry olivine upper mantle and increased upper limit of the viscosity range. The model renders viscosities up to $10^{25}$ Pa·s, whereas TempModel1 is restrained to $10^{23}$ Pa·s. The RheoModel1 thus supports higher deviatoric stresses (well above 10 MPa), except of the lower crust in the backarc domain that still has a lower viscous rigidity. TempModel1 contains a lithosphere that deforms significantly as result of edge-driven mantle flow. Particularly the crust thins along the backarc Cordillera domain and thickens in a distributed manner across the craton (Figure 6.11). For the RheoModel1 instead, the lithosphere exhibits a minor displacement field and the crust remains at constant thickness of 35 km (Figure 6.11). The comparison shows that mantle flow at discontinuities at the base of the lithosphere produces critical stresses that may or may not initiate yielding depending on sensitive rheological parameters of mantle lithosphere. Increasing the upper viscosity limit to $10^{25}$ Pa·s implies that higher viscosities are allowed for the mechanical strong mantle lithosphere of the craton that, for given model produced stresses, is much less prone to flow.

Thus, the models show the critical role of the upper mantle lithosphere rigidity on the occurrence or absence of lithosphere deformation from edge-driven mantle flow acting at its base. Furthermore, the plots of the viscosity structure also outline how the base of the lithosphere forms not only a thermal boundary layer but also marks a mechanical transition [e.g., Schmeling and Marquart, 1993]. The low viscous asthenosphere reaches high flow rates in range of $10^{-12}$-$10^{-15}$ s$^{-1}$ and in the more rigid lithosphere deformation occurs under strain rates in order of $10^{-15}$-$10^{-17}$ s$^{-1}$.

6.4.4 Geodynamic implications from upper mantle rheological structure

Figure 6.12 shows the evolution of the RheoModel1 in five subsequent time-slices along with temperature gradients and viscosity profiles for three vertical slices across the model box (at 250, 1750 and 2725 km). The viscosity structure, being a function of temperature and strain rate, is affected by the evolution in temperature and deformation fields. The viscosity profile for the platform section shows a rigid and thick mantle lithosphere whereas strength in the mantle lithosphere of the hot backarc is almost absent.

The temperature and rheological vertical profiles are plotted for each time-slice alongside the model evolution for backarc and proximal and distal platform positions (Figure 6.12). For the backarc, over time, the conductive geothermal gradient heats up, and alongside the viscous rigidity of the mantle lithosphere declines. The model contains well established mantle convection cells, similar to those of TempModel1 (Figure 6.5) that start with edge-driven flow around the thermal step from the backarc to platform (located at x=500-750 km). The model exhibits again wide zones of hot mantle ascent and localized drips at the lithosphere boundary layer.

However, comparison of TempModel1 and RheoModel1 (Figures 6.5 and 6.12, having
The evolution of model RheoModel1 depicted in five time-slices at 1, 10, 30, 50 and 100 Ma. The model corresponds to model HtBkArc1 (Figure 6.5) except of a dry olivine rheology for the platform mantle lithosphere.

The same initial temperature structure, suggests that the RheoModel1 better preserves the shape of the BTL. The lateral thermal discontinuities appear more stable when supported by an accompanying compositional contrast that supports a more rigid platform.
Geodynamics at thermal and rheological discontinuities in the mantle lithosphere. Whereas the temperature discontinuity migrates eastward for TempModel1, due to the vigorous mantle flow at the discontinuity (Figure 6.5), this discontinuity stays in place with the added rigidity from a dry olivine composition for the platform lithosphere in RheoModel1 (Figure 6.12).

The evolution of RheoModel1 is plotted together with the other RheoModels (Figure 6.13) and can be compared with the TempModels (Figure 6.6). The RheoModels exhibit roughly similar mantle flow as for the different TempModels for the same initial temperature structures. Thus RheoModel2 contains a similar initial temperature structure as that of TempModel2, both having a stronger the edge driven convection and overall mantle flow as result of the stronger temperature discontinuity between the backarc and craton.

RheoModel2 further illustrates the effect of the compositional upper mantle lithosphere to the stability of the discontinuity. Both models exhibit the strong edge-driven convection. This lead to a strong craton-ward migration of the discontinuity in TempModel2, whereas it stays in place over 100 Myr for RheoModel2 due dry olivine upper mantle for the craton lithosphere.

These differences in response between the RheoModels and TempModels also effect the response of the overlying lithosphere, i.e. the responses in the surface heat flow and deflection is different. For the reference model TempModel1, the lateral contrast in surface heat flow between the Cordillera and craton diminished over time. Differently for RheoModel1, for which the the sharp transition from high surface heat flows in the backarc domain and low values for the craton stays in place (Figure 6.13). Also the other RheoModels preserve the step in the surface heat flow profile forever a time-span of 100 Myr. As the step in the BTL (Figure 6.13) stays in place between x= 500-750 km, so does the heat flow signals in response. For the surface heat flow evolution over the course of 100 Myr, again the heat flow increases slowly due to the gradual adjustment of a conductive lithosphere to advected heat in the convecting mantle.

Also the surface deflection response is different (Figure 6.13). The free-surface deflection response of the lithosphere are much more stable (Figures 6.12a and 6.12d) compared to the TempModels (Figure 6.6). In absence to deformation, the crustal thickness remains constant at 35 km and uniform along the model domain (Figure 6.13). Consequently, the only factor to the surface deflection is the dynamic mantle topography (i.e. mantle density variations due to temperature and up- and downward mantle flow). The rheological transition from wet to dry olivine enhances the corner flow effect, supports the hot backarc anomaly and maintains its positive elevation (Figure 6.13) at 100 Myr. As a result, the more rigid crustal mantle lithosphere of RheoModel1 preserves the thermal and rheological discontinuity and maintains the corner flow and lithosphere response at the initial position of the discontinuity.

The dry olivine mantle lithosphere for the craton together with higher viscosities (i.e. up to $10^{25}$ Pa·s) represent better the lithosphere above a vigorously convecting asthenosphere. The hot mantle upwelling along with the edge driven flow at the thermal transition still tend the crust in the overlying backarc to thin, which occurs in RheoModel2 as result of stronger mantle flow. Here, the undulations (i.e. ~1000 m in amplitude) which develop onto the positive topography of the backarc are related to redistribution of crust within the backarc domain.
Comparison chart of the RheoModels, with RheoModel2 (similar to RheoModel1) (reference model) plotted together with RheoModel3 and RheoModel4. The evolution of the models is shown in five time-slices at 1, 10, 30, 50 and 100 Myr.
6.5 Application to south-western Canada

6.5.1 The hot upper mantle underneath the Canadian Cordillera

The geodynamics of a lateral temperature transition in the upper mantle finds application in the southern Canadian Cordillera (Figure 6.14). The crustal and upper mantle architecture of the Canadian Cordillera to North American Craton have been extensively mapped [e.g. Clowes et al., 1995; Cook, 1995; Zelt et al., 1996; Burianyk et al., 1997; Bouzidi et al., 2002]. It is broadly agreed that the Cordillera marks a zone of major lithospheric thinning [e.g. Gough, 1986; Jones and Gough, 1995; Hyndman et al., 2005] that stands in contrast to the much thicker lithosphere of the North American craton to the east. The thin and hot Cordillera lithosphere is outlined by high heat flow and an uppermost mantle beneath the Omineca and Intermontane belts (Figure 14a) of the Cordillera and by high electrical conductivity and a thick low seismic velocity zone at shallow depth [e.g. Gough, 1986; Jones and Gough, 1995].

The thin Cordilleran lithosphere, by Hyndman et al. [2005] referred to as a hot orogen and continental backarc, is considered being related the upwelling of hot asthenosphere (Figure 6.14d). Numerical modelling studies have considered a mantle corner flow driven by traction of a subducting plate as cause of mantle-flow driven thinning of the continental backarc [Currie et al., 2004, 2008]. The North American continental lithosphere forms the overriding plate to the subduction of the Juan de Fuca plate. The corner-flow that establishes in the mantle between the two plates thins the overriding continental backarc (Figure 14d). This thinning must have occurred since Eocene time when the subduction of the Juan de Fuca established after major plate re-configuration [e.g. Engebretson et al., 1985].

While the mechanisms leading to the thinning of Cordilleran lithosphere have been well addressed, is the geodynamic implication of the hot Cordilleran upper mantle standing against the cold lithosphere of the Canadian craton another interesting topic for numerical modeling. The results in this study show how the evolution in upper mantle convection, initiated by edge-driven convection at the backarc to craton transition, is reflected in the surface deflection and heat flow patterns of the overlying lithosphere. This edge-driven flow only came into effect after thinning of the Cordilleran mantle lithosphere presumably occurred in Eocene time.

Direct comparison of these responses to field observations is difficult. As discussed, the free-surface deflection response of the models is affected by conservation of volume of a limited box-size so that no absolute amounts but relative variations count. Furthermore, heat flow spatial and temporal trends of the model focus mainly on the contribution from the mantle. Direct comparison with heat flow observations can not be made, firstly because crustal heat production was omitted in several of the models. The other reason is that the surface heat flow shows a box-wide gradual increase as artefact of mantle heat transfer that only starts after mantle convection establishes at onset of the model and a conductive lithosphere that only fully reflects this initiation after ~60 Myr. Instead, in reality mantle derived heat may vary over time in response to changes in mantle dynamics, but evolves from an existing mantle flow pattern and an already established thermal boundary layer.

Nonetheless the effects of dynamics around an upper mantle temperature discontinuity on the heat flow patterns in the overlying lithosphere are still manifest and significant. The surface heat flow distribution produces a distinct transition with 30 mW/m²
contrast between the hot backarc and cold craton that follows the discontinuity in case it migrates craton-ward. The hot lithosphere of the Canadian Cordillera produces elevated surface heat flow of ~80 mW/m² in strong contrast with the 50 mW/m² of the craton to the east. The transition is located at the Cordilleran Foreland Belt which basement heat flow evolution is relevant for burial-temperature and petroleum history reconstruction. Understanding the geodynamics of the strong temperature discontinuity in its subsurface is critical to understand basement heat flow fluctuations.

6.5.2 Cordillera to craton transition: a thermal and rheological discontinuity

The lateral temperature discontinuity also coincides with a rheological transition. The Cordilleran lithosphere is not only hot and thin but also weak and whereas the cold and thick craton lithosphere is mechanically much stronger [e.g. Lewis et al., 1992; Lowe and Ranalli, 1993; Hyndman and Lewis, 1999]. The contrasting rheological stratification between the Cordillera and craton not only results from the temperature dependant viscous rigidity that is evidently lower for the hot Cordilleran mantle lithosphere than for the cold craton mantle lithosphere. Also, the upper mantle composition may be different with higher water content for the Cordilleran upper mantle due to subduction related water enrichment above the continental backarc [Arcay et al., 2005, 2006] versus a mantle lithosphere of the craton that is chemically depleted [Jordan, 1978].

The Cordillera to craton transition thus exhibits a strong rheological discontinuity for which this study earlier estimated an approximate integrated strength difference of 0.5-2.0·10⁴ GPa·m versus 10.0-50.0·10⁴ GPa·m for the Cordillera and platform lithosphere, respectively (see Chapter 5). Previous estimates of the rheological transition along a profile from the Cordillera to craton outline a comparable contrast in rheological structure [Wu, 1991; Lowe and Ranalli, 1993; Hyndman and Lewis, 1999] and is further confirmed by coherence analysis of Bouguer gravity and topography [Flück et al., 2003]. The discontinuity has even been addressed in recent high resolution tomography studies [van der Lee and Frederiksen, 2005; Mercier et al., 2009; Sigloch et al., 2008].

The models in this study show the effect of a rheological transition on the evolution of an upper mantle temperature discontinuity. They outline that a rheological transition from wet to dry olivine will support a temperature discontinuity in maintaining the contrast and keeping it from migrating by edge-driven traction. The model results show that the temperature discontinuity can migrate craton-ward by the traction of the edge-driven flow until it meets a rheological transition that keeps the discontinuity in place.

This means for the Canadian Cordilleran to craton temperature discontinuity, which probably established by lithosphere thinning due to backarc mantle upwelling, afterwards might have migrated eastward and established at the rheological transition. When mantle upwelling and lithosphere thinning occurred in Eocene time due to the activation of the Juan de Fuca subduction, the temperature discontinuity probably established shortly after at its current position at the western limit of the Canadian craton underneath the Cordilleran eastern foreland belt.

6.5.3 Hot Cordillera lithosphere and post-orogenic deformation

Until the Eocene, the subduction and accretion history coincided with major contraction of the Cordillera. Since Eocene time, the plate configuration along with the tectonics of the Cordillera changed with shortening localized at the pro-wedge close to the active
margin. Instead, the Cordillera to the east became subject to the formation of several core complexes and half graben systems, expressing both tensional forces and viscous creep activation of the crust [e.g. Parrish et al., 1988; Vanderhaeghe et al., 2003; Gordon et al., 2008] (see also Figure 6.14a and 6.15c).

It has been argued that the Cordillera underwent tectonic inversion from compression to extension from Late Paleocene to Eocene times [e.g. Constenius, 1996; Lowe and Ranalli, 1993]. Core complexes may have formed under the load of a thickened crust and lithosphere that failed after large-field compressional forces were removed during plate-reconfiguration [e.g., Constenius, 1996; Liu, 2001]. However, clear examples exist of crustal thinning of continental backarcs under a far-field compressional stress regime [e.g. England and Houseman, 1989; Jolivet and Faccenna, 2000]. The scenarios postulated for the thinning of the mantle lithosphere underneath the Canadian Cordillera require no tensional tectonic forces when attributed to hot mantle upwelling [Gough, 1986; Jones and Gough, 1995] or gravitational instabilities at the base of the lithosphere [Ranalli et al., 1989; Lowe and Ranalli, 1993;]. Whether the occurrence of a hot lithosphere base due to upwelling and convective instabilities also initiates thinning in the overlying crust is addressed in several modelling studies [e.g., England and Houseman, 1989; Liu, 2001].

The deformation response in the here presented models indicate extension and thinning of the crust above the hot upper mantle as result of traction at the lithosphere base from edge-drive mantle flow at the discontinuity. The transfer of edge-driven traction into the overlying lithosphere varies strongly with the viscous rigidity of the mantle lithosphere. However, the models do show that when the strain from the edge-driven mantle flow propagates into the overlying mantle lithosphere of the discontinuity, the overlying crust tends to thin. This thinning occurs without horizontal tectonic forcing operating on the side-boundaries of the lithosphere. Thus the presence of a hot upper mantle and the temperature discontinuity trigger thinning of the crust and consequently no far-field extensional forces would be required. This implies the establishing of the Juan de Fuca plate subduction and asthenospheric upwelling beneath the overriding plate may induce the thinning of lithosphere and crust, without the need from change of far-field compression to extension.

### 6.5.4 Edge-driven flow and Cordilleran surface deflections

The surface deflection patterns in response to edge-driven convection need to be combined with other evident factors in account of the topography around the Canadian Cordillera to craton transition. For instance, the edge-driven flow around the temperature discontinuity is located underneath the foreland belt that exhibits a strong flexural component because of the build-up (and removal) of an orogenic surface load (Figure 6.14).

The models produce high topography above the hot and thin upper mantle lithosphere because of an isostatic response and additional dynamic support from the upward flow in the mantle. The isostatic response occurs by the hot and hence low density mantle asthenosphere. Dynamic support is attributed to upper mantle convection cells (Figure 6.14d) that have a width of 400-600 km and cause surface deflections of similar wavelength as flexural response of the lithosphere. Such wavelength deflections, induced by general mantle flow patterns, are inhibited in case of a strong mantle lithosphere.

Studies on the exhumation of the Cordillera and craton indicate an uplift and exhu-
Figure 6.14

(a) The tectonic make-up of the southern Canadian Cordillera. (b) Surface response in terms of exhumation for the Omineca Belt, Foreland Belt and Foreland Basin. (c) Lithosphere structure across the Cordillera derived from geophysical probing [after Clowes et al., 1995]. (d) Postulated upper mantle flow scenarios in explanation of mantle lithosphere thinning and thermal discontinuity: (I) edge-driven mantle flow, this study; (II) Subduction-related corner flow [see Currie et al., 2004, 2008]; (III) mantle upwelling central underneath the Cordillera [e.g., Jones and Gough, 1995] possibly related to presence of a relic subducted slab which sinking through the 660-km discontinuity (IV) may cause large-wavelength uplift [e.g. Mitrovica et al., 1989; Pysklywec and Mitrovica, 2000]. This occurs against a possible backdrop of ‘normal’ mantle convection (V). (e) Outline of P-wave velocity anomalies (dVp / Vp %) under the Cordillera along a transect following the US-Canada border [adopted after Sigloch et al., 2008]. Positive P-wave anomalies point to bodies with higher velocities and typically low temperatures whereas negative anomalies to lower velocity and typically hotter bodies. The Juan de Fuca plate is imaged with a +3 dVp/Vp% in the upper 200 km and connected with a large +1-2 dVp / Vp % at 400 km depth underneath the Cordillera. Also the cold craton to the east is depicted with a positive 2-3 dVp/Vp% velocity anomaly. In strong contrast stands the upper mantle underneath the Cordillera with a negative 3-4 dVp/Vp% P-wave anomaly.
formation in post-orogenic Eocene time occurring over an extremely large wavelength that is labelled as epeirogenic and attributed to deeper-rooted dynamic topography [Mitrovica et al., 1989; Gurnis, 1992; Pyskylywe, and Mitrovica, 2000]. Mantle flow that roots deeper than upper mantle edge-driven flow is required to explain both large wavelength subsidence of the foredeep during the orogenic phase as well as large-wavelength uplift in the post-orogenic phase [e.g. Mitrovica et al., 1989; Russell and Gurnis, 1994; Pyskylywe and Mitrovica, 2000]. Particularly models of sinking slabs of earlier subducted lithosphere through the 660-km mantle transition can produce vertical motions at even a continent-scale [e.g. Mitrovica et al., 1989; Russell and Gurnis, 1994; Burgess et al., 1997; Pyskylywe and Mitrovica, 2000].

The surface expression of deeper mantle flow dynamics may exhibit larger wavelengths than both the flexural signal from the lithosphere and effects from upper mantle flow dynamics discussed in this study. Already Beaumont [1981] pointed out that the Cretaceous foreland subsidence in southern Canada was too wide in extent to be attributed only to continental flexure under the Cordilleran orogenic load. Yet, the upper mantle flow dynamics operate at a much smaller wavelength and its impact is much more dependent on the rheological structure of the upper mantle (lithosphere).

The upper mantle edge-driven flow developing underneath the foreland belt was accompanied by deeper mantle flow dynamics in support of post-orogenic epeirogenic motions (see Figure 6.14d). Both the large wavelength dynamic topography and the several hundred meters of negative and positive dynamic topography due to the edge-drive convection must be added to the post-orogenic flexural rebound signal [Beaumont, 1981]. The dynamics from the deep mantle helps explaining vertical motions beyond flexural wavelengths (Figure 6.14b) whereas the upper mantle flow adds both extra subsidence and uplift within the flexural wavelength for which otherwise a reduction in flexural rigidity or stronger surface load will be required.

6.6 Conclusions

This study addresses how much response in the lithosphere can be expected for a configuration with no other geodynamic trigger than an upper mantle lateral temperature discontinuity (i.e. in absence of any plate boundary or other forces). The numerical experiments in this study show that the effect from edge-driven convection at an upper mantle temperature discontinuity is large enough to produce also a noticeable lithosphere response. The edge-driven flow also affects crustal deformation and surface heat flow and deflection pattern, although with a strong dependency on the rheological structure of the mantle-lithosphere.

The amplitude of the discontinuity influences the vigour in edge-drive flow and the amount of traction along the base of the lithosphere at the discontinuity. A stronger amplitude increases the traction, that accentuates the contrasts and increases tendency of the discontinuity to migrate craton-ward. However, the temperature discontinuity remains well preserved and also stays in place when supported by an enhanced rheological contrast. A rheological contrast marked by a normal wet olivine mantle lithosphere (similar to the rest of the mantle) against a dry olivine rheology for the colder and thicker lithosphere maintains the temperature discontinuity and edge driven-flow. This shows that the rheology of the mantle lithosphere exerts a top-down influence on mantle flow.

The difference in geodynamic response to variant temperature and rheological discontinuities is also reflected in the crustal deformation and surface deflection patterns.
The deflection pattern exhibits a positive isostatic response above the thin lithosphere, due to higher temperatures and therefore lesser density of the upper mantle. The surface deflection also records a small direct response to upward or downward flow in the mantle. Depending on the overall rigidity of the mantle lithosphere, the edge-driven convection induces a positive topography signal at the backarc-side of the discontinuity and a negative signal on the craton-side with topography contrast up to 300-400 m. Also the surface heat flow response, while entailing a significant transient component, over time reflects the thermal discontinuity well. The surface heat flow distribution produces distinct transition with 30 mW/m² contrast between the hot backarc and cold craton that follows the discontinuity in case it migrates craton-ward.

A particular example of an upper mantle temperature discontinuity is the hot Canadian Cordilleran lithosphere juxtaposed to a much colder and thicker North American craton. The transition is located at the Cordilleran Foreland Belt which basement heat flow evolution is relevant for burial-temperature and petroleum history reconstruction. The geodynamics around a strong temperature discontinuity in its subsurface is critical for understanding basement heat flow fluctuations. The temperature discontinuity underneath the Cordilleran Foreland Belt is positioned at the transition between a hydrated backarc and a depleted craton mantle lithosphere. The temperature discontinuity may have established under the Foreland Belt due to this rheological transition and shortly after hot mantle upwelling started in Eocene time under Cordillera backarc. Due to a transient effect and possible east-ward shift of the temperature discontinuity must the basement heat flow have increased since then to present-day values.